THE SEDIMENTOLOGY AND STRATIGRAPHY OF RED BEDS IN THE WESTPHALIAN A TO C OF CENTRAL ENGLAND

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Abstract

Red beds in the Westphalian A to C of Central England belong to the Etruria Formation. Examination of exposures and cores has allowed subdivision of this Formation into three Facies Associations.

Facies Association I forms the gradational base of the red bed sequence. The sediments resemble the underlying Productive Coal Measures, consisting dominantly of grey organic-rich mudstones. Much of this Association is formed by seat-earth palaeosols, which occasionally contain red pigment.

Facies Association II forms most of the Etruria Formation, and consists largely of dark red mudstone. Throughout there are mottled horizons which are interpreted as lateritic palaeosols. Occasional upward-fining sand bodies containing lateral accretion surfaces are present.

Facies Associations I and II are interpreted as fluvial deposits. The rivers were dominated by suspended load and deposited thick overbank sequences. In Association I the overbank areas were swampy, while in Association II good drainage prevailed.

Facies Association III comprises red mudstones and palaeosols, and a range of conglomerate and sandstone sheets and channel fills. This Association occurs near inferred fault bounded margins of the depositional basin. It is interpreted as the deposits of muddy alluvial fans deposited in a humid tropical climate.

The petrography of the Etruria Formation is described. Derivation was from an area of mixed composition which was undergoing tropical weathering. The diagenetic alteration of the sandy sediment occurred soon after deposition in fresh water conditions. Red pigment in the mudstones was mainly inherited from a lateritic mantle in the source area, but was also partly generated by oxidation during water table fluctuation soon after deposition.

Compilation of stratigraphic data shows a complex pattern of diachronism between the Etruria Formation and the Productive Coal Measures. Red bed deposition began during Westphalian A in the slowly subsiding southern margins of the depositional basin, and spread progressively northwards from Westphalian B to late Westphalian C in response to uplift of the Wales-Brabant ridge.

He had been eight years upon a project for extracting sunbeams out of cucumbers, which were to be put into vials hermetically sealed, and let out to warm the air in raw inclement summers.

Swift.

Le suprême charme qu' on trouve à lire une page de de Selby est qu' elle vous conduit inexorablement à l'heureuse certitude que des sots vous n' êtes pas le plus grand.

du Garbandier

Preface

This study of the Etruria Formation was started in 1976 at a time when economic pressures were causing the closure of many brick- and tileworks in the Central England coalfields. The consequent loss of clay pit exposures to the pressure of waste disposal was threatening to leave large areas without representative exposures of the Formation. The scale of this decline can be seen from the following figures. In 1900 there were 90 brickworks listed in Kelly's Directory for Staffordshire which were probably working the Etruria Formation (39 in North Staffordshire; 8 in the Cannock area; 7 in the Aldridge area; 36 in the Black Country). In 1976 there were only 16 brickworks working Etruria Formation clays in Staffordshire (8 in North Staffordshire; 2 in the Cannock area; 2 in the Aldridge area; 4 in the Black Country). Of 40 exposures examined in the present study, 19 have since been partially or completely infilled, are threatened with infilling, or have become degraded.

It was felt that, as a matter of urgency, a record should be made of the geology of this Formation. To this end this thesis includes considerably more illustrative material than would otherwise have been necessary. It is hoped that the available exposures provided a representative sample of the Formation.

At the same time large amounts of new data were becoming available as a result of the National Coal Board's 'Plan for Coal' and 'Plan 2000' exploration programmes. The National Coal Board has very kindly allowed access to these data, subject to the following minimal conditions:

i) that, to protect the confidentiality of some data, no depths from surface are given, and in most cases the thicknesses of coal seams not specified;

ii) that the mentioning of individual coal seams relates only to their stratigraphic significance, and has no implication as to any present or future economic potential.

In some of the area under consideration, the re-arrangement of local authority boundaries has resulted in changes in county names. As all literature pertaining to the Westphalian employs the county names used prior to 1971, this practice has been continued in this thesis.

Acknowledgements

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Data has been obtained from a large number of industrial undertakings. Without the help and advice of staff in these enterprises the project could not have been completed.

Access to borehole cores and records has been allowed by the Western and South Midland Areas of the National Coal Board (Deep Mines), and by the Opencast Executive of the National Coal Board. Special thanks are due to Bob Hoare (Western area) and Alan Jones (South Midland Area) for granting this permission. My task has been made easier by the assistance of Lloyd Boardman, Dave Mellor, Paul Norman, Keith Whitworth, Phil Richardson and Ian Fulton, who have guided me through the NCB filing system and made arrangements to view cores of the Allotment, Playground and Kibblestone boreholes. Ros Todhunter of the Opencast Executive made arrangements for me to examine the Street's Lane borehole cores.

Permission to visit exposures has been granted by all companies approached. Mr. S. Davenport and Mr. K. Pate of G.H. Downing Ltd. (now Steetly Brick Ltd.) allowed unlimited access to their quarries, and enabled me to examine the Rosemary Hill borehole cores. Thanks are also due to: D. Platt & Sons Ltd.; J. Caddick & Son Ltd.; Mr. P.R. Powell; Haunchwood Lewis Ltd.; Hawkins Tiles Ltd.; Hinton, Perry & Davenhill Ltd.; Ibstock Brick (Aldridge) Ltd.; Ibstock Brick (Himley) Ltd.; Stourbridge Brick Co. Ltd.; Baggeridge Brick Co. Ltd.; Barnett & Beddows Ltd.; Wilnecote Brick Co. Ltd.; Butterley Building Materials Ltd.; Stoneware Ltd.; Blockleys Ltd.; Stanley Bros. Ltd.; A. Adams & Sons Ltd.; The Lilleshall Co. Ltd.; Michelin Tyre Co. Ltd.; London Brick Buildings (Garden Division); Land Reclamation Co. Ltd.; and Polymeric Treatments Ltd. Advice as to the ownership and location of exposures was given by the minerals officers of the following County Councils: Staffordshire (Mrs. L. Donovan); West Midlands (Mr. J. Boldon); Shropshire (Mr. T. Butler); and Warwickshire (Mr. T. Long).

The study has benefited from discussion with members of the Institute of Geological Sciences who have worked on the Westphalian. I would particularly like to thank Richard Hamblin, Michael Calver, Wyndham Evans and Albert Wilson. Information from the Sidway Mill borehole is reproduced by kind permission of the IGS.

I would like to thank the technical staff in the Geology Department at Keele for their ready and cheerful assistance: in particular David Kelsall (photography); David Emley (geochemistry); and Peter Greatbach (thin section preparation). The thesis has been efficiently typed by Jackie Dray and Julie Robinson (both of Shell UK) and Joan Cliff. Vic Haddon (Shell UK) helped with the reproduction of some of the line diagrams, and with the printing.

Throughout the study I have been greatly helped by the accumulated experience of Bob Hoare, Lloyd Boardman, David Thompson and Peter Turner, who have spent many hours discussing coal measures and red beds. Mavis Butterworth, Stephen Daly, Nick Kusznir, Graham Lees, Peter McCabe, Nico Rossel and Dave Stewart have all helped by providing advice, criticism and various forms of practical assistance. I would like to thank all of these for having shown an interest in Carboniferous red beds at a time when it was neither popular nor profitable to do so.

Finally, I would like to express my deepest gratitude to my future wife, Jane Whieldon, without whose encouragement this thesis would not have been written.

CONTENTS

Chapter 3. Facies Association I: predominantly organic rich sediments deposited in poorly drained alluvial swamps and lakes

Chapter 4. Facies Association II: well drained alluvium deposited by meandering rivers.

CHAPTER 1

General introduction; historical and regional review

1.1 Extent and aim of study

Red beds in the Westphalian A to C rocks of the English Midlands are mainly referred to the Etruria Formation. This is a distinctive group of predominantly red mudstones with subsidiary sandstones and conglomerates, which overlies various horizons of the productive Westphalian B and C coal measures. Rocks of this type are also locally interbedded with, or underlie, productive coal measures; these have no formal stratigraphic name. In most published accounts, the Formation is referred to as the 'Etruria Marl' or 'Etruria Marl Formation', the former being the original name, proposed by Gibson (1901) after a type locality in Stoke-on-Trent.

The use of the term 'Marl' is incorrect as the majority of the clays contain less than 10% of calcium carbonate. It is not used in this thesis.

No detailed study of the sedimentology or stratigraphy of the Etruria red beds has previously been undertaken, and the most recent account of the origin of the red pigment (Hoare 1959) predates the work of Van Houten (1968, 1972) and Walker (1967, 1974) which has fundamentally altered concepts of the formation and significance of red beds.

In this study the depositional environments of the Etruria Formation are interpreted on the basis of facies analysis of measured sections of exposure and borehole core. The petrography of the sediments is described. The occurrence of the red pigment is compared with that in other red beds and discussed in the light of Walker's (1974) model for red-bed diagenesis in humid environments. The regional stratigraphy of the Westphalian A to C is described, and the palaeogeographic evolution of the Midland area during this period is discussed.

1.2 Geographical occurrence; nature of exposure

The study area is largely occupied by a Permo-Triassic basin. Westphalian rocks outcrop around the margins of the basin, and in three major horst structures within it (Fig. 1). The nomenclature of coalfield areas used in the text is shown in Fig. 2. There are extensive outcrops of the Etruria Formation in the North Staffordshire, Mid-Staffordshire and South Staffordshire areas; locations of exposures in these areas are shown in Figs. 3, 4 and 5. Other exposures are illustrated on Fig. 1. Apart from descriptions of individual features in the text, all exposures are briefly described in Appendix 1.

The Westphalian A to C throughout the area of study lacks major sandstone bodies. The outcrops of the Etruria Formation and the productive coal-measures are largely indistinguishable, generally giving rise to a flat, monotonous topography which is extensively covered by Quaternary sediments, made ground, and buildings. Locally the base of the Westphalian D Halesowen Formation forms a steep escarpment, the slope of which is formed by the top 50 m of the Etruria Formation. There are no significant natural exposures; all the exposures that have been examined are in working or recently abandoned brick - clay pits.

1.3 History of previous research:

1.3.1 Stratigraphy

The first subdivision of the Westphalian to include the Etruria and

Halesowen Formations was that of Murchison (1839), who recognised the Carboniferous age of the Coalport Beds (in part equivalents to the Halesowen Formation) in North Shropshire, and referred them to the Upper Coal Measures. In South Staffordshire he regarded the Etruria Formation as the Lower New Red Sandstone, conformably overlying 'Upper Coal Measures'. In this area Jukes (1859) correctly placed both Etruria and Halesowen Formations in the Carboniferous, describing an Upper Coal Measures sequence of 'Red Coal Measure Clays' overlain by 'Halesowen Sandstones' succeeding the Productive Coal Measures.

In the late Nineteenth Century a tripartite division of the Westphalian into Lower, Middle, and Upper Coal Measures was generally adopted. The divisions were arbitrary, that between the Lower and Middle Coal Measures being based largely on the number and quality of coal seams present (e.g. Hull 1881), while the Upper Coal Measures included all overlying Carboniferous rocks which were devoid of workable coals. These generally consisted of the Etruria Formation and sometimes the Halesowen/Newcastle Formation, although in the latter case the descriptions are ambiguous (e.g. Hull op cit). The base of the Permian was generally placed at the base of the Keele Formation (Hull 1869) although occasionally as low as the base of the Etruria Formation (Davies 1877).

Gibson (1901), describing the upper part of the North Staffordshire Westphalian, recognised the conformable relationship of the Keele Formation to the underlying coal measures, and its Carboniferous age. He named four formations after type localities in the Stoke-on-Trent district and used them to define the Upper Coal Measures as follows:-

Keele Series	: red and purple sandstones and marls.
Newcastle Series	: grey sandstones and shales with thin coal seams.
Etruria Marl Series	: mottled red and purple marls and clays; thin bands of green grit very characteristic.
Blackband Series	: grey sandstones marls and clays; numerous thin coal seams and Blackband ironstones; base at Bassey Mine coal.

Gibson noted the similarity of sequences in North and South Staffordshire, Nottinghamshire, and North Wales, and suggested that the North Staffordshire sequence should be regarded as a standard for the Upper Coal Measures.

The first biostratigraphic subdivision of the Westphalian was introduced by Kidston (1894, 1905), who erected a fourfold classification on palaeobotanic grounds (Fig. 6). Arber (1916) summarized flora obtained from the Etruria Formation in South Staffordshire and from the Blackband Formation in North Staffordshire, and noted that there was an anomaly in the species present, those from the Etruria Formation in the south being apparently older than those from the Blackband Formation in the north. The possibility of a diachronous relationship between the Etruria Formation and the productive coal measures was suggested by Eastwood et al (1923) and, more forcefully, by Whitehead and Eastwood (1927). The introduction of a more sophisticated biostratigraphic classification based on non-marine bivalves (Trueman 1933; Dix and Trueman 1937; Trueman and Weir 1946) (Fig. 6) confirmed this suggestion (Wills 1935; Trueman

1947). Surprisingly, Wills, in a later publication, played down the importance of this diachronous relationship, suggesting that the variable position of the base of the Etruria Formation resulted from local unconformity (1956 pp.62, 68).

The diachronism has since been described in more detail by Earp (1961) and Boardman (1978) in North Staffordshire, and by Poole (1965, 1970) in South Staffordshire and the Wyre forest.

The stratigraphic nomenclature of the rocks making up the Westphalian A to C Etruria Formation, together with the overlying Westphalian D formations, is derived from a series of Geological Survey maps, together with their associated memoirs. (Gibson 1905; Barrow et al. 1919; Eastwood et al. 1923; Whitehead and Eastwood 1927; Whitehead et al. 1928; Mitchell 1942, 1945; Whitehead and Pocock 1947; Mitchell et.al. 1961; Geological survey of Great Britain 1978). The cumulative effect of these is to divide the Westphalian in each coalfield into a number of units, each of which have formation status within that coalfield, and which may be broadly correlated over the whole area (Fig. 7). Inevitably anomalies arise between adjacent sheets, which are discussed in Chapter 9. In addition, the Westphalian is usally subdivided into Lower, Middle and Upper Coal Measures, the boundaries between these being taken at the Gastrioceras subcrenatum, Anthracoceras vanderbecki, and A. cambriense Marine Bands (Stubblefield and Trotter 1957) (Fig. 6). Although this equates the traditional tripartite classification of the English Westphalian with the standard classification adopted on the European mainland, it suffers from the grave disadvantage that the A. cambriense marine band is not found in large areas of the Midlands, the onset of red bed formation having occurred before this marine incursion (Trueman 1947, pp. lxxxi - ii; Poole 1970; this thesis). The continued popular application of the term 'Upper Coal Measures' to the Etruria Formation in areas such as South Staffordshire and Warwickshire where the base of the Formation predates the A. cambriense marine band (e.g. Bennison and Wright 1969) renders it ambiguous and confusing. For this reason the tripartite classification of the Westphalian of Stubblefield and Trotter (1957) is not employed in this thesis.

For convenience, the term Etruria Formation is used in the present study to name all the rocks of the red bed facies hereinafter described, irrespective of their stratigraphic position (i.e. including those which underlie productive coal measures, in the extreme south of the area under study). This stratigraphic scheme is illustrated in Fig. 8. The formal lithostratigraphic nomenclature of the Formation is discussed in Chapter 9. The standard European chronostratigraphic system (Fig. 6; Ramsbottom et al. 1978) is employed throughout.

1.3.2 Other previous research

The Etruria Formation sandstones and conglomerates in South Staffordshire were described by Murchison as volcanic grits formed "from the detritus of submarine volcanoes, which were in activity towards the close of the accumulation of the coal measures." (1839 p. 468). He also recognised the difference between these and the intrusive igneous rocks in South Staffordshire. The presence of igneous material and abundant clay matrix in these sandstones caused them to be regarded as tuffs by Jukes (1859) and as " true ashes, or ... derived from the comminution of lavas " by Barrow (in Gibson 1905).

Gibson (1905) noted the presence of quartzite clasts, and described the coarser sandstones in North Staffordshire as " not unlike the coarser bands of the Millstone Grit". Whitehead and Eastwood (1927) recognised that the conglomerates in South Staffordshire are lenticular, and consist mainly of fragments of Cambrian quartzite, with subsidiary amounts of Silurian sandstone and igneous material.

The distinctive ceramic properties of the Etruria Formation mudstones were described by Plot (1686) and, in some detail, by Dobson (1850) and Boulton (1917). Their petrography was first described by Robertson (1931). By comparing their chemistry with that of fresh and weathered basalts the latter concluded that the characteristic red clays of high iron content were derived from the decomposition and denudation of contemporaneously erupted basalts, a view strongly criticized by Boulton (in discussion of Robertson op cit).

A detailed study of the petrography of the sandstones was made by Williamson (1946). He listed a variety of grain types and diagenetic features, and concluded that the rocks were derived from a Pre-Cambrian and Lower Palaeozoic metamorphic, igneous, and sedimentary terrain in Central England. He was not able to verify Robertson's conclusions.

Depositional environments for the Etruria Formation have been suggested by several authors, most of whom have also commented on the palaeoclimate. Trueman (1947) stated that the red bed facies represents a marginal facies of the productive coal measures. Wills (1935) suggested that there was no marked change in climate connected with the transition from coal-bearing to red-bed sedimentation. Subsequently (1950) he suggested that the climate during Etruria

Formation deposition had been hot, with seasonal rainfall. Later (1956), in his detailed study of the Midlands Upper Carboniferous, Wills concluded that the climate had been semi-arid. He recognised that the change from productive coal-measure to red-bed deposition was accompanied by the influx of poorly sorted, locally derived sediment, which resulted from contemporaneous tectonic block movement. The facies represented "deposition in marshy tracts that were at times flooded, and at times retained wind blown lateritic dust on their damp surfaces". The sandstones and conglomerates were deposited as alluvial fans and screes. Red pigment was derived from lateritic soils in the source area.

In a series of papers Trotter (1953a, 1953b, 1954) described red beds from the Lancashire coalfield which were demonstrably formed by deep oxidation of grey coal-measure type sediments beneath the Permo-Triassic land surface. Hoare (1959), working in the Coalbrookdale and Cannock (Mid-Staffordshire) areas, was able to distinguish between red beds formed in the this way below the sub-Permo-Triassic and sub-Halesowen Formation unconformities, and those of Etruria Formation type which he regarded as primary red beds, formed in a manner similar to that suggested by Wills (1956). Studies of the clay mineralogy of the Etruria Formation by Keeling and Holdridge (1963) and Keeling (unpublished abstract, 7th IAS Conference 1967), using the non-quantitative IL/MA method (Keeling 1961), concluded that deposition had taken place in a lacustrine environment which was alternatively fresh water and hypersaline. In an earlier paper Holdridge (1959) had examined the chemistry of some fifty Etruria 'Marl' samples in detail. He noted a rhythmic alternation in iron content in two carefully sampled localities and suggested (p. 320) that successive periods of lateritization had occurred during deposition.

Malkin (1961) described material obtained from the full thickness of the Etruria Formation during the sinking of the No. 3 shaft at Wolstanton Colliery, North Staffordshire. He recognised the presence of rooted horizons and plant material throughout the sequence, and was able to correlate carbonaceous horizons in the succession with other sections recorded in the northern part of the coalfield. He concluded that the depositional environment was not dissimilar to that of the Productive Coal Measures. This view was echoed by Barnsley (1965) in his study of the Cannock Chase area of Mid Staffordshire.

The junction between the Etruria and Newcastle Formations at Metallic Tileries, Chesterton, North Staffordshire was described in detail by Pollard and Wiseman (1971). The occurrence of algae of marine affinities in limestone nodules containing a fauna of Spirobis and non-marine bivalves was regarded as indicative of a transition at this junction between a hypersaline environment, in which the Etruria Formation had been deposited, and brackish conditions of deposition of the lowest Newcastle Formation.

1.4 Palaeogeographic setting

1.4.1 Setting within the context of northern Europe

Westphalian deposition in northern Europe occurred within two distinct tectonic and environmental zones (Fig. 9). To the north of the Variscan front an interlinked series of 'paralic' basins extended from Ireland to Central Poland, bounded to the north by Caledonian and Pre-Cambrian upland areas, and to the south by the contemporaneously uplifting Variscan mountains (Hedemann and Teichmüller 1971). Within the Variscan fold belt deposition also took place within isolated fault-bounded 'limnic' basins.

Deposition of the British Westphalian took place within the belt of 'paralic' basins. (These basins are so called from the occasional marine incursions which occurred during the deposition of their fill). The 'paralic' belt consists of a number of continuous depocentres (frequently described in the literature as basins) in which more or less continuous deposition occurred. These are separated by structural highs, variously described as 'shoals', 'highs', 'positive areas', and 'islands', which were zones of little or no subsidence during the Namurian and early Westphalian, but which were increasingly rejuvenated during Westphalian C time, acting as local sediment sources (Bless et al. 1977).

In the British mainland area four sedimentation provinces can be recognised (Calver 1970): the Scottish, Pennine, South West and Kent provinces (Fig. 10). The Scottish and Pennine provinces were separated by the Southern Upland 'high', which was mainly onlapped during the Namurian, and which accumulated a thin Westphalian cover (Ramsbottom et al. 1978). The Pennine province was separated from the remaining two provinces by the Wales - Brabant high. Some marginal areas of this high were onlapped during the early Westphalian, but it is likely that connection between the Pennine and South Western/Kent provinces during the Westphalian A to C was only intermittent, if it existed at all. (Wills (1956) proposed a connection between South Wales, and the West Midlands, but the evidence for this is tenuous). The Wales - Brabant high was breached during the Westphalian D, when the coal measures of Oxfordshire were deposited unconformably on pre-Carboniferous rocks.

In this thesis, the area under consideration forms the south western corner of the Pennine province (Fig. 10).

Before Westphalian C the dominant sediment source for the Scottish and Pennine provinces lay to the north. During Westphalian C time uplift of blocks within the 'paralic' belt, and of the Variscan mountains, led to a dominance of southerly and some locally derived sediment in Belgium and Germany (Bless et al. 1977). This change of sediment derivation pattern has not been documented as such in Britain , although the extensive occurrence of locally derived debris in the Etruria Formation (described in Chapter 7) suggests that rejuvenation of local structures was widespread in the later Westphalian C in the southern part of the Pennine province. This topic is elaborated in Chapter 10.

1.4.2 Global setting

The geometrical reconstruction of continental crustal areas in the Upper Palaeozoic published by Bullard et al (1965) predicted a near equatorial position for northern Europe. This position has been confirmed by Turner and Tarling (1973). Recent palaeomagnetic determinations made on haematite rich mudstones from the Etruria Formation (Besly and Turner, in press) have given an equatorial latitude (inclination of - 1 deg.) for Central England during the Westphalian B to C, with a pole position of 35 deg. N, 191 deg. E. In the reconstructions of continental landmasses given by Turner and Tarling (op. cit.) and by Scotese et al. (1979) the north European area was situated on the eastern margin of a large land area, and separated from the Upper Palaeozoic Tethyan Ocean by the Variscan mountain chain, and a small shelf sea area to its south east (Fig. 11).

This reconstruction of the continental land masses has considerable implications for the interpretation of the palaeoclimatological conditions which prevailed during the Westphalian. In the idealized continental precipitation map produced by Robinson (1973), the European area, near the eastern margin of a continent straddling the equator, would have been situated in a broad belt of heavy precipitation. The effect of the creation of the Variscan mountain chain, to the east of the North European area, would have been to narrow this belt of high precipitation considerably (Ziegler et al. 1979); (Fig. 12). The equatorial humid belt would also have been intermittently squeezed still further by the global migration of all climatic belts towards the equator, as a result of the Gondwanaland glaciation (Glennie 1981) (Fig. 13). The tendency of these two effects to reduce the latitudinal extent of the equatorial humid belt may be expected to have become more influential through the late Carboniferous, owing to the increased intensity of Variscan mountain building, and the growth of the Gondwana glaciation. The former of these events reached an acme in the Westphalian C/D, while the latter reached its maximum extent in the early Permian.

The progressive narrowing of the equatorial belt of high precipitation in northern Europe during the later Carboniferous may be anticipated to have had two effects, which may locally be recorded in the sedimentary record. Firstly, the formation and preservation of coals would have become much more dependant on latitude. Secondly, the narrowing of the equatorial belt of high precipitation would have made the climatic conditions progressively more unstable through the late Carboniferous. These effects would jointly have tended to reduce the altitudinal extent of dense topical vegetation with a consequent effect on run off, and to produce an increasingly marked seasonality in the precipitation pattern. These may be expected to be reflected in the sedimentary sequence by a progressive decrease in the preservation of coals, and possibly by the appearance of sedimentary features suggesting intermittent run off. Features of the latter type have been recorded by Broadhurst _et al._ (1980) from early Westphalian sediments in Lancashire.

1.5 Notes on sedimentary facies descriptions and interpretations

The use and implication of the term 'facies' have been discussed by Reading (1978), who lists four different and popularly employed implications of the term. These are objective classifications of sedimentary rocks in terms of:

i) their lithology and internal structure;

ii) the inferred process of their deposition;

iii) the inferred environment or goup of environments in which they were deposited;

iv) the broad tectonic environment in which they were deposited.

In previous descriptions of the Westphalian in Central England the term 'facies' has been used very loosely. Various authors (e.g. Poole 1966) have used the "Etruria Marl facies" or the "Productive Coal Measure facies" in a broad sense, with lithostratigraphic, environmental, and tectonic connotations.

In this thesis the term facies is employed in two distinct ways. In the descriptive chapters (Chapters 3 to 6) twenty depositional Facies are described and interpreted. These describe rocks which differ recognisably from one another in composition, colour, grain size, nature and scale of sedimentary structures, etc. in ways which are recognisable in the field. Where borehole core material has been used, many of these criteria still apply, but where distinctive criteria take the form of large scale sedimentary structures, these cannot be recognised, and the identification of Facies is less certain and relies heavily on context.

Facies described in this way correspond to the first of the implications described by Reading (op. cit.). These Facies are grouped into three Facies Associations, which are groupings of Facies which usually occur together, and which are inferred to have been deposited within a related set of depositional environments. The Facies and Facies Associations described in this thesis are summarized in Table 1.

In chapters 7 to 10, which are concerned with more general aspects of the Etruria Formation, the term facies is used in a more general way to differentiate between the rocks of the Etruria Formation (the Etruria facies or red bed facies) and those of the Productive Coal Measures.

While this use of the term is founded upon the purely descriptive use in the ealier chapters, it does have an environmental implication. Although this may seem slightly confusing, this slight change of meaning is quite apparent when seen in context, and obviates the repeated usage of cumbersome phraseology. The descriptive use in distinguished in Chapters 3 to 6 by the word Facies being spelt with a capital letter.

1.6 Organization of text

The thesis is organized as follows:

Chapter 2: introduction to the description and nomenclature of palaeosols.

Chapters 3 to 5: description and interpretation of individual Facies.

Chapter 6: description of relationship between three major Facies Associations.

Chapters 7 and 8: description of sedimentary petrography of the Formation, and description and interpretation of the diagenesis of the Formation.

Chapter 9: summary of the stratigraphy of the Formation.

Chapter 10: synthesis of data from preceding chapters to reconstruct palaeogeographic changes during deposition of the Formation; concluding discussion.

Detailed descriptions of individual exposures are not given in the main text: the typical Facies and Associations are illustrated by representative extracts from complete sections. The details of individual exposures named in the text, and of other exposures, are to be found in Appendix 1. Co-ordinate locations given for exposures and boreholes refer to the British Ordnance Survey national grid. The approximate geographical locations of all boreholes and exposures which have been studied sedimentologically are given on Figs. 1 and 3-5.

All of the vertical sections of exposures, and the detailed vertical sections of palaeosols, have been drawn to a standard key. This is summarized in Fig. 14.

CHAPTER 2

Palaeosol description, nomenclature and classification

2.1 Prefatory remarks

Throughout the Etruria Formation the red mudstones of Facies 8 and 9 (described in 4.2.1, and 4.2.2) contain horizons of variegated red, grey, ochre, and purple colouration. These horizons have in the past been referred to as 'mottled' (e.g. Gibson 1901), 'variegated' (Hoare 1959), or 'brecciated' (Mitchell et al 1954). The only published interpretation of these features is by Hoare (in discussion of Richardson and Francis 1971), who suggested that such features might have formed soon after deposition, by desiccation or by brecciation induced by mass-flow.

Because of their frequent association with root traces, thin coals, and their characteristic leached profile, these horizons are interpreted in this thesis as palaeosols.

A number of other palaeosol types, resembling the seat-earths typical of coal measure sequences (Huddle and Patterson 1961), are also present. As the study of palaeosols, other than calcretes (e.g. Allen 1974 a; b; Steel 1974), is not well represented in the British and North American literature, the following section of the thesis briefly discusses the methods of study and nomenclature that will be used.

2.2 Definition of the term palaeosol

The confused and sporadic record of the study of pedogenic features in ancient sedimentary rocks has largely resulted from the interdisciplinary nature of pedology. The term palaeosol itself is used in three distinct and different ways (Gerasimov 1971), to describe:

i) relict or residual properties in present day soils which indicate an earlier phase of pedogenic evolution, possibly under different conditions of climate, drainage, or vegetation;

ii) pedoliths - sediments produced by the erosion and re-deposition of modern or ancient soils, preserving to some degree properties acquired during soil formation;

iii) autocthonous buried soils, preserved *in situ* in their entirety, or with the upper parts of their profiles eroded.

The palaeosols encountered in the British Westphalian fall into the last of these categories.

As palaeosols, as defined here, indicate a substantial period of non deposition and pedogenic evolution, they cannot include sedimentary deposits which were colonized by plants during a rapid phase of deposition. These contain root traces, but no other feature indicative of soil formation. In recent environments such deposits might be classified as alluvial soils (e.g. Paijmans *et al* 1971).

2.3 The soil profile : processes, horizon nomenclature, and profile development

At the time of its formation, a soil is a zone of interaction between the parent material, the groundwater, the atmosphere, and the associated biota. This interaction produces a profile of superjacent soil horizons, which are divided into three groups. In presently forming soils, many of these horizons are diagnostic of distinctive soil formation processes, and are defined and named. Two systems of horizon nomenclature are commonly used (Birkeland 1974). In one, horizons are divided into three groups: A, B and C horizons. The uppermost of these, the A horizons, correspond to the zone of maximum weathering, and are enriched in organic material. Material is removed from the A horizons by solution leaching or eluviation, the downward movement of fine grained solid particles. Material thus mobilised accumulates in the underlying B horizons, by precipitation and concretionary growth and/or by accretion of the eluviated material. Both A and B horizons are usually intensely bioturbated by both root and animal action, and retain no trace of the depositional sedimentary structure of the parent material. The transition between the soil and the underlying unaltered parent material is formed by the C horizon, in which both the structure of the parent material and pedogenic structures are visible.

In an alternative nomenclature system (Soil Survey Staff 1960) A horizons forming the surface of the soil are called surface horizons and all other horizons are subsurface horizons. All horizon types are rigorously defined.

The types of diagnostic horizons occurring within the one soil profile are, obviously, not mutually exclusive. The classification of entire soil profile types is usually achieved on the basis of the co-occurrence of distinctive horizon types, and on the thickness and maturity of the horizons present. Approaches to palaeosol profile classification are discussed in 2.7.1.

In this thesis the formal identification of soil horizon types in palaeosols is not attempted. The reasons for this are elaborated in 2.7. Other authors have, however, made quite specific horizon identifications in ancient palaeosols, using both the A, B, C horizon terminology (Buurman and Jongmans 1975) and the Soil Survey Staff (1960) system (Meyer 1976; Retallack 1977a). The latter system is also used by Yaalon (1971) in a discussion of the preservation potential after burial of various soil horizon types and pedogenic features. It is therefore necessary for the present discussion to define some of the more frequently encountered horizon types, and the approximate relationship between the two nomenclature systems. These definitions are to be found in Appendix 2, section 1.

Soil profiles are classified, on the basis of their combinations of diagnostic horizons, into Sub-orders and Orders (Birkeland 1974). This style of classification is common to the many systems of soil classification used in different countries. As the Soil Orders thus defined are broadly process-related, Orders defined in different classification systems can usually be equated, despite differences in terminology. The palaeosols described in this thesis are classified, broadly, at Soil Order level (see 2.7.2). The physical and chemical processes which lead to soil profile development of some of the major soil orders are summarized in Appendix 2, section 2.

2.4 Preservation potential of pedogenic features and soil horizons

Yaalon (1971) has divided soil horizons and some other soil features into categories of varying potential for preservation after burial. The categorisation is based on the type and the speed of the processes which give rise to their distinctive features.

i) Horizons and features which have a high preservation potential are those which involve the irreversible concentration of material into concretions composed of a chemical phase which is immobile in the soil. This may occur by the influx and precipitation of material in solution, or by the removal of all but the most insoluble material by extensive leaching. Immobile concretion forming compounds include iron oxides and calcium carbonate. Peat and soluble salts may also be immobile, if the subsequent conditions of early diagenesis are favourable to their preservation. Horizons from which large amounts of iron or clay have been eluviated may also have a high potential for preservation, although, in a sedimentary sequence, these may be difficult to distinguish from the textural variation in the original sediment.

A variety of horizons and features in this class have been recognised in pre-Quaternary palaeosols. The most frequently described are horizons of pedogenic calcium carbonate enrichment in fossil caliche soils (Allen 1974a, 1974b; Gile and Hawley 1966; Steel 1974). Pedogenic iron oxide concretions and, by inference, oxic horizons have been described by Abbot et al (1976) and by Kulbicki and Vetter (1955). Most coals formed from autochthonous peat can be regarded as buried organic soils, although this is not an approach which has commonly been adopted.

ii) Horizons and features of moderate and low preservation potential are formed, within the soil profile, by reversible processes. Once a soil is buried, these will tend to be destroyed.

Yaalon _(op cit)_ argues that features which have formed rapidly tend to degenerate rapidly. Horizons with the lowest preservation potential are those which are characterised solely by features which are unlikely to survive compaction, and those which are characterised by small concentrations of soluble salts or organic matter. In addition, several horizons (ochric, cambic, etc...) do not possess any distinctive characteristics which are likely to survive burial.

Features of moderate preservation potential include horizons which are enriched in clay and/or iron oxides and/or humus. Buurman and Jongmans (1975) have described an Oligocene podzol in which an eluviated A horizon and a spodic B horizon, enriched in clay and iron oxides, are present. These features are carefully defined on both field and thin section criteria. This sediment has apparently not undergone oxidizing diagenesis. The mobility of iron during red bed development (Ixer _et al_ 1979; Slânska 1976; Turner 1974) should be borne in mind when considering the preservation potential of palaeosol horizons containing fine grained accumulations of iron compounds.

2.5 Effects of intermittment deposition on horizon and profile development

Apart from the question of preservability of individual diagnostic horizons and of soil profiles, a further complication is introduced into the study of palaeosols by the effects of contemporaneous deposition or erosion.

Three effects may be anticipated (Fig 15): polyphase soil development on a site undergoing progressive change in drainage conditions during subsidence; intermittment deposition during soil formation giving rise to a modified soil profile; and the erosion of horizons of palaeosol profiles, which may give rise to incomplete profiles.

Polyphase soil development occurs when, through an externally controlled change of environmental conditions, the type of soil being formed on a particular site changes. This results in a soil which contains relict characters of an early, different soil type, unless the profile development of the later soil compltely obliterates features of the earlier soil. The most common cause is a change in the drainage conditions. In recent soils this most often reflects comparatively short term climatic fluctuation. In palaeosols preserved in alluvial sequences, polyphase soil formation probably reflects the effect of subsidence on the position of the water table.

Intermittent deposition of sediment during soil formation may be anticipated to have two effects (Fig 15). If the sediment increments are small, a profile may be developed in which individual horizons, especially the B horizon may be anomalously thick. Larger sediment increments will give rise to stacked palaeosol profiles in which overprinting of horizon characters may occur, the A horizon of one soil becoming the B or C horizon of the soil developed on the same site after the addition of a sediment increment. Sequences of the latter type have been described by Freytet (1971), who found that in many cases it was not possible to identify individual profiles or even individual horizons, so great was the interference produced between properties of successive soils.

2.6 Palaeosol features other than horizon and profile development

Apart from the recognition of ancient soil horizons and profiles it has been claimed that fossil soils may also be recognised by their macroscropic field appearance, and by various small and microscopic pedological features.

On outcrop scale, irrespective of whether individual horizons are identifiable, palaeosols are frequently pigmented by colours which are unusual in sedimentary rocks, for instance, ochre, pink, brown, "wine dregs" (Freytet 1971), or a characteristic violet or purple (McBride 1974: Ortlam 1971). A palaeosol may also be marked by a loss of bedding, and a gradational base and sharp, possibly erosive, top (Ortlam op cit).

Pedological features and soil fabrics have been exhaustively classified by Brewer (1964). Many authors have recognised some of his textural features, and used them as evidence for soil formation in ancient sediments.

Brewer's system for the classification of soil fabric takes the form of a hierarchical system of peds, pedological features, and S-matrix (pp 134-148). The definitions of these features, together with claimed ancient fossil examples and comments on preservability, are listed below:

i) Peds are individual natural soil aggregates, larger than the grains of the soil material, consisting of clusters of soil particles which are separated from adjoining peds by planes of weakness, formed by voids or by planes of preferentially oriented illuviated clay (cutans).

Peds generally do not exceed 5 cm in size. Because they are defined by voids or clay features it is very unlikely that such features will survive compaction. They have, however, been inferred in pre-Quaternary palaeosols by Retallack (1976) and Watts (1976).

ii) Pedological features are formed by reorganisation and segregation of soil materials during pedogenesis. Brewer (op cit) recognises three main groups: features formed by chemical mobilisation and segregation (e.g. concretions, mottles); features formed by mechanical mobilisation and re-arrangement (e.g. cutans, slickensides); and fossil organic features, resulting from the infill of roots and burrows.

Concretions are divided by Brewer into crystellaria (void filling) and glaebules (displacive and/or enclosing soil material). This distinction has been observed in descriptions of ancient palaeosols by Allen (1974a, 1974b), but is usually difficult to apply. Concretions of a variety of minerals have been recorded in ancient palaeosols. Other than descriptions of ancient calcretes, mineral species include iron-manganese and iron oxides (Freytet 1971; Ortlam 1967, 1971; Retallack 1976), siderite (Huddle and Patterson 1961; Retallack 1976; Roeschmann 1971), gypsum (Buurman and Jongmans 1975; Ortlam 1967, 1971) and chalcedony (Ortlam 1967, 1971). In all of these cases it should be borne in mind that such concretions can be formed or modified after burial of the soil (Roeschmann 1971).

Another group of features formed by chemical segregation are ferruginous mottles, which are formed by concentration of iron oxides into diffuse patches in hydromorphic soils (see Appendix 2). Mottles have been described from ancient palaeosols by Buurman (1975, 1980), Buurman and Jongmans (1975), and Freytet (1971).

The main group of features formed by physical segregation are cutans - laminated accumulations of clay and/or sesquioxides and/or organic material. These line voids and form coatings on peds and inherited grains of the parent material. They are formed by the accumulation of eluviated material in B horizons. Criteria for their recognition in thin section are given by Brewer (1964). Clay cutans have been recognised by Meyer (1976), Retallack (1976) and Watts (1976), bounding fossil peds. Cutans have also been recognised in thin section by Buurman and Jongmans (1975) and Terruggi and Andreis (1971).

In view of the potential for forming oriented clay aggregates by soft sediment deformation and compactional and tectonic shearing, it seems unlikely that cutanic features can be unambiguously recognised in ancient sediments. The same observation applies to slickensides, although these have been described as palaeosol features (Huddle and Patterson 1961). Even if such features are preferentially concentrated in identifiable palaeosols, it seems likely that their presence may reflect different patterns of early compaction in root churned and non rooted sediment.

Fossil root and burrow traces (termed pedotubules by Brewer, op cit) have been widely described from fossil soils (e.g. Huddle and Patterson 1961; Freytet 1971; Buurman 1975; Retallack 1976).

iii) S-matrix. This is an concept introduced by Brewer (1964 pp 302 ff) to describe the 'matrix' surrounding the pedological features in a soil. The s-matrix comprises skeleton grains, which are chemically stable grains, usually of quartz inherited from the parent material, and plasma, consisting of clay minerals, organic matter, and fine grained or colloidal sesquioxides.

Apart from the skeleton grains, most of this material is too fine grained to be observed directly with the optical microscope. The majority of it appears isotropic, either owing to its grain size, or to the masking effects of iron and manganese oxides. However, in most soils, aggregates of clay grains occur in which there is a degree of preferred orientation, and which are thus birefringent. Although some of the textures thus observed are analagous to, and may be inherited from, textures observed in sediments which have not been affected by pedogenesis, Brewer regards the majority of the textures as being of uniquely pedogenic origin. He has termed them plasmic fabrics, and erected a complex descriptive classification (op cit pp 308-318) based on the type of the aggregates, their orientation, extent, degree of mutual orientation, and their relationship to voids and skeleton grains.

Several authors have described such textures from ancient sediments (Allen 1974b; Freytet 1971; Retallack 1976, 1977b). In particular, Terrugi and Andreis (1971), recognising such textures in a study of the Cretaceous Chubutian Group of Patagonia, have adhered closely to Brewer's concept of the uniqueness of some of these textures to soils.

While not questioning Brewer's assertion of the uniqueness of these textures, the uncritical recognition of them as indicators of pedogenic modification in ancient sediments must be questioned. Apart from the possibility of oriented clay aggregates being produced by compactional or tectonic shearing, such pedogenically produced textures must be strongly affected by compaction and diagenesis. Schiller (1980) has shown that the degree of orientation of clay aggregates is markedly affected by the thickness of overburden loading. While Tertiary clay sediments buried by up to 500m have a 40° to 60° dispersal of orientations, Carboniferous shales buried at 1000m have a dispersal of only 30°, and this is reduced to less than 20° at a burial depth of 1800m. Schiller also notes that the presence of fine grained organic matter (as might be expected in palaeosols) markedly increases clay mineral orientation on compaction. Such burial compaction is bound to affect plasmic fabrics. There is also the possibility that clay mineral textures could be obscured by diagenetic recrystallization with increasing burial.

Although a number of potential plasmic fabrics have been observed during the present study, they have been formed of rather larger clay mineral grains that those illustrated by

Brewer, and in no case has it proved possible to differentiate them unambiguously from textures induced by compaction or deformation. They are therefore not regarded as diagnostic palaeosol features.

2.7 Nomenclature and classification of palaeosols

2.7.1 Approach adopted by other workers

At the 1970 Amsterdam Symposium a series of recommendations for the techniques of study, nomenclature, and classification of palaeosols was adopted by the Working Group on the Origin and Nature of Palaeosols (1971). The recommendations include: that palaeosols should be studied using the same methods as those used for modern soils; that modern soil nomenclature should be used in describing palaeosol features and horizons; and that palaeosols should be classified using a commonly employed internationally recognised soil classification system.

These recommendations were derived by a committee of soil scientists and workers on Quaternary palaeosols. They have consciously or unconsciously, been adopted by several workers on pre-Quaternary palaeosols, notably Freytet (1971) and Retallack (1976, 1977a, b). While pedological methods may be readily applicable in Quaternary soils which have been comparatively little affected by diagenesis, these workers do not seem to have considered the practicability of applying them to more modified, older material.

Soil fabric. The terminology of Brewer (1964) has been extensively adopted in palaeosol descriptions (references in 2.6). Where it is certain that features are of pedogenic origin it seems reasonable to employ it. However the potential for production of pseudo - pedogenic features during diagenesis, non-pedogenic diagenetic modification of soil features, and mimicry of soil microfabrics during compaction, remains unknown. The terminology of Brewer should therefore be employed with more caution than it has been to date.

Horizon nomenclature. Only two authors have specifically named palaeosol horizons, other than in fossil caliche soils. Buurman and Jongmans (1975) recognise A, B and C horizons, and where possible, add subscripts to record specific properties of the horizon such as the presence of illuvial humus concentrations. This system conforms to the horizon nomenclature used in the FAO/UNESCO soil map of the world, which is summarized in a modified form by Birkeland (1974 pp. 5-7). Retallack (1977a, b) attempts to use both this system and the USDA diagnostic horizon nomenclature (Soil Survey Staff 1960). While this approach might have been justified if all of the diagnostic horizons in question were of high preservation potential, his recognition of ochric, cambic, and argillic horizons must be regarded with suspicion. The USDA system is very precise, horizons often being defined on the basis of laboratory measurement of properties such as organic carbon content and exchangeable cation content, which are bound to change during diagenesis. The majority of ancient palaeosol horizons have probably been too much altered to comply with the criteria used in this system.

The approach of Buurman and Jongmans (op. cit.) is entirely justified, as it does not require a horizon lacking diagnostic features to the identified in any detail. Use of the USDA terminology introduces into the geological literature a very complex and specific nomenclatural system, which, if applied, will largely be used inappropriately to name horizons which do not process the requisite diagnostic properties.

Palaeosol classification. While more or less standarized systems of soil fabric and horizon nomenclature do exist, there is no universally accepted sytem of classification of soil types. The basis of several of the more commonly used systems is discussed with particular reference to tropical soils by Young (1976).

When one comes to attempt a comparative classification of ancient palaeosols, the problems are multiplied. It is difficult to identify features and horizons with sufficient certainty to deduce the soil forming process. Also, many buried soils in ancient rocks were formed in alluvial areas, whose soils are very poorly described in recent environments. In the majority of soil classifications, soils in areas of active deposition are grouped together, and called 'alluvial and other immature soils'. Profile descriptions are few and inadequate, compared with the volume of literature on mature soils on lithified bed rock material.

Two approaches to ancient palaeosol classification are possible. Freytet (1971) and Retallack (1977a) both attempt to classify palaeosols in terms of existing classification of modern soils, the latter scrupulously adhering to the spirit of the Working Group (1971) recommendations. Freytet related the palaeosols in the Eocene alluvium of south western France to the soil classification of the ORSTOM (French overseas scientific research organisation; see Young

1976). This attempt is fairly successful, in that the classifatory system, being based on natural macroscopic features, allows for a degree of subjectivity. Retallack classified the Triassic palaeosols in the Sydney area of Australia in terms of the USDA 7th Approximation system (Soil survey staff 1960) and in terms of two local Australian systems. Both the U.S. and the Australian systems are 'artificial' systems, in that soil types are arbitrarily defined on the basis of precise limiting values of the thicknesses and properties of the soil horizons present and the nature of the boundaries between them (see discussion in Young 1976 pp 249-254). As with horizon nomenclature, the USDA system of classification, and other similarly based on the use of arbitrary criteria, are unsatisfactory for ancient palaeosols. They can seldom be applied accurately and correctly; and they introduce into the geological literature a large, complex, and unnecessary new vocabulary.

In contrast, Buurman and Jongmans do not name the palaeosol that they describe in terms of any classifactory system. Instead they are content to identify the processes which operated during formation of the soil, and the order in which they occurred. This approach is much more satisfactory: it does not involve the application, with dubious accuracy, of unfamiliar terminology; it avoids the problems caused by the lack of a standard classification system; and it is much more useful to the sedimentologist in reconstructing the processes which occurred at the time of deposition and modification of the rock in question.

2.7.2 Approach adopted in this thesis

From the foregoing it is evident that rigid and uncritical adherence to the principles set out by the Working Party (1971) report is neither necessary nor desirable in the study of pre-Quaternary palaeosols.

In the present study, geological terminology is used throughout, with occasional reference to Brewer's (1964) fabric nomenclature. Horizons are only tentatively divided into A, B, and C, when there is evidence that leaching or pedogenic accretion of material has taken place. The palaeosols are only loosely referred to a classificatory system; instead they are directly compared with modern soils which are forming in sedimentologically analogous areas. This comparison between ancient and recent serves to suggest soil forming processes which may have occurred during the Upper Carboniferous.

CHAPTER 3

Facies Association I: predominantly organic rich sediments deposited in poorly drained alluvial swamps and lakes.

3.1 Introduction

Although sediments belonging to this Facies Association are volumetrically insignificant within the Etruria Formation, they form the stratigraphic and environmental link between the Etruria Formation and the paralic coal measure sequence which forms the bulk of the Westphalian in the Pennine basin. A brief sedimentological description of these rocks follows. It is drawn from the single exposure of the base of the Etruria Formation, at Ibstock Himley quarry (SO896902; Fig. 5), and from cored sequences through the base of the Formation in Kibblestone (SJ911361; Fig. 3), Playground No. 8 (SJ972124; Fig. 4), Street's Lane No. 1083 (SJ972061; Fig. 4), and Rosemary Hill Nos. 1-5, 9, 11, and 12 (SJ830460; Fig. 3) boreholes. Information is also included from the Blackband Formation in North Staffordshire, which underlies and passes laterally into the Etruria Formation. These data come from Boardman (1978; 1981; and personal communication), although some of the interpretations are by the present author.

3.2 Facies Descriptions

3.2.1 Facies 1: grey mudstones containing detrital plant material

This Facies occurs in all of the sections examined. The mudstones vary from dark to medium grey in colour (N3 to N5) and are occasionally silty. Horizontal lamination is occasionally present, and comminuted plant debris is commonly concentrated in individual laminae. Where large concentrations of plant debris are present, listric surfaces (closely spaced undulose polished fracture surfaces) are developed parallel to the bedding.

In most cases no depositional lamination is present, the sediment having been disturbed by rooting. Stigmarian roots are almost ubiquitous, and there is usually a gradation upwards into a palaeosol.

Spheroidal concretionary siderite grains (sphaerosiderite), usually not greater than 2 mm in diameter, occur commonly, either disseminated throughout the mudstone, or in aggregates of up to 5 cm.

This Facies overlies palaeosols of types 1 and 2, or sediments of Facies 2 and 3 (thin siltstones and sheet sands) or of Facies 5 and 6 (varved mudstones and ironstones). In one instance, in Ibstock Himley section 1 at ca. 9.50m, mudstone overlies a thin coal seam which forms the top horizon of a palaeosol. The mudstone contains abundant rolled coalified logs, up to 20 cm in diameter and some exceeding 4 m in length.

Interpretation

The fine grained nature of this Facies implies deposition in very quiet conditions. The preservation of organic matter and the grey colouration suggest a permanently waterlogged environment. Although the presence of occasional horizontal lamination demonstrates that the deposition of the muds took place from suspension in quiet water, the universal presence of roots shows that water depths were seldom too great for colonization by vegetation. Using modern tropical vegetation as an analogue (Tricart 1977), water depths probably did not exceed 2 m.

A closely comparable environment, described from the Recent and Holocene, is to be found in the poorly drained swamp environment of the Atchafalaya valley (Coleman 1966). In his descriptions, sediments accumulating in ineffectively drained floodplain areas closely resemble the sediments of Facies 1, and a similar interpretation is suggested. The major differences are that the Atchafalaya sediments contain abundant pyrite and calcite. The lack of the latter in the Carboniferous sediments is explicable by Coleman's observation that calcite is rapidly replaced by siderite after burial.

The sediments in this facies are closely comparable to those in Facies Association 3B in Scott's (1978) description of the Westphalian B coal measures in Yorkshire. He interpreted this Facies Association as floodbasin or floodplain deposits.

3.2.2 Facies 2: silty mudstone and siltstone containing plant material

Grey silty mudstone and siltstone occur in beds up to 50 cm thick in most of the grey beds encountered at the base of the Etruria Formation in the Rosemary Hill boreholes. No internal structure has been seen. Plant debris and/or roots are occasionally present. The Facies is usually associated with thin sands of Facies 3, and its interpretation is discussed with that Facies.

3.2.3 Facies 3: grey, sheet sandstones

Grey sheet sandstones containing plant material have been observed in Rosemary Hill boreholes Nos. 1, 2 and 11 (one correlatable unit), Nos 5, 9, 12 and 14 (one unit in B/H 9 at the same horizon as channel deposits in B/H 5 and ? in B/H's 12 and 14), and in section 1 at Ibstock Himley quarry.

The sandstones occur in sheets between 0.20 and 1 m thick. They are usually composed of fine grained argillaceous sand, although medium and coarse sands are found in this Facies at Ibstock Himley quarry, where the sand sheets are proximal to a channel sand body. Individual sheets are sometimes made up of amalgamated, thinner units. The sand sheets are often interbedded with siltstone to form 2 to 4 m sequences of coarser sediment within grey, Facies 1, mudstones.

Some of the sheet sandstones show well developed ripple cross lamination, marked in the lower sheet at Ibstock Himley (section 1 at 11.95 m) by concentrations of mica along foresets, and by concentrations of plant debris in the sand at 26 m in Rosemary Hill B/H No. 9. The lower sand sheet at Ibstock Himley contains Stigmaria roots, and detrital plant fragments including delicate leaf fronds of Sphenophyllum. In Rosemary Hill No. 9 B/H, the sand at 25.25 m contains a basal lag of coal pebbles.

In the Ibstock Himley quarry, sand sheets of this Facies surround upright and tilted Sigillaria tree trunks, standing up to 2 m high (Fig. 16). Upright trees are also recorded from the Blackband Formation in North Staffordshire (Gibson 1905 p. 333), but it is not clear in which Facies they occur.

At Himley the sand sheets pass laterally into channel fill deposits (Facies 4) (Fig. 17). The nature of this relationship is discussed under Facies 4. The same relationship can be inferred between sand units in Rosemary Hill boreholes, Nos. 5 and 9 (Fig. 18).

Interpretation

From their unique association with sediments deposited in swamps in an alluvial basin, these sheet sands were evidently deposited by the rapid influx of faster moving, sand laden water into this quiet environment. The lateral passage into channel fills shows that, in two cases, these flows were channel sourced. The internal structure of the sand sheets is consistent with an origin either in bank levêes of a channel or as crevasse splay units. (See 4.2.3 for fuller discussion of these depositional mechanisms). This interpretation is also proposed for the siltstones of Facies 2 which are usually associated with Facies 3 sandstones.

As few examples of these Facies have been observed, and those mostly in cores, it is difficult to distinguish between the deposits of the two suggested mechanisms. It is, however, felt that the coarser, thicker sheet sands associated with channels (Rosemary Hill No. 9 B/H, Ibstock Himley) are most likely to be levêe sands. Allen (1965) regarded the occurrence of upright trees in interbedded sand/shale overbank sequences as a characteristic feature of levêe deposits.

Scott (1978) interpreted thin sheets sands in his Facies Association 3B (Facies 1, 2, 3 in this thesis) as crevasse splay deposits.

3.2.4 Facies 4 - channelised sandstone bodies

This Facies has been observed in one surface exposure, section 1 at Ibstock Himley quarry. It may also be present in Rosemary Hill borehole No. 5, and possibly in boreholes Nos 1, 2 and 3, but the quality of the core recovery in these sections does not contribute much to the description of the Facies.

A complex, channelised sand body is exposed (1980) at the same stratigraphical horizon on both sides of a deep trial pit in the Ibstock Himley quarry. The two faces of this pit are less than 50 m apart. On the east face the channelised form is readily visible (Fig. 19), cutting through mudstones of Facies 1. The fill consists of fine to (dominantly) medium grained lithic sandstone. There is a basal lenticular lag of coarse sand. Sedimentary structures include ? upper flow régime horizontal lamination, and large scale (60-80 cm thick) trough cross bedding. The sandstones contain abundant woody debris and impressions of Lycopod tree bark. The situation of this exposure (at the top of a quarry face) precluded measurement of a detailed section. Although no accurate palaeocurrent data was obtained, it is apparent from the form of the channel margin that the channel runs nearly parallel to the quarry face.

The laterally equivalent horizon to this channel is not exposed to the south. To the north (towards the main part of the quarry) the channel fill is thinner, the base of the channel cutting down less into underlying sediments. Some 50 m to the north, the horizon of the channel is represented by interbedded sand and siltstone, although the channel margin is this direction was not exposed.

The exposure on the west face of the trial pit is more complex. Although a channel form is visible (Fig. 17), it does not fall in line with the apparent trend of the channel body exposed in the east face. Furthermore, unlike the exposure in the east face, the fill of this 'channel' conists of three units, which appear to have been rapidly deposited. The upper two of these drape the 'channel' margin, and pass laterally to the south into thick, Facies 3, sheet sands. In the base of the 'channel' feature there are two upright trees, surrounded by the lower two units of the fill, and truncated by the third. These trees are filled with fine sand and siltstone (Figs. 20 and 21). The lower two units of the fill of the channel consist of cross bedded fine to medium grained sandstone. These are overlain at the margin at the 'channel' by a thin grey silty mudstone, which is eroded by the upper unit in the centre of the channel, and which also passes laterally into the interbedded sand sheets to the south. The upper unit of the fill consists of coarse, lithic sandstone.

Although, as a result of poor core recovery, little internal structure could be seen in the sandbody cored in Rosemary Hill boreholes nos. 1, 2, 3 and 5 it is apparent that this 6 m thick body shows a fining upward sequence, and has a basal intra - and extraformational conglomerate (Fig. 18).

Sandstones forming fairly thick (c.6m) sandbodies are known to occur in the Blackband Formation in North Staffordshire (Boardman, personal communication). These sand bodies are not laterally extensive, and may consist of channel fills similar to the one observed at Ibstock Himley quarry. The only sandstones observed in surface exposures and cores by Boardman (1981) were 0.30 m sheet sands (probably Facies 3) in overbank mudstone sequences.

Interpretation

Because of the small number of examples, the interpretation of this Facies is speculative. The sand body in the east face of the Himley trial pit is obviously a channel deposit. From the association with overbank facies claystone and palaeosols this channel was an alluvial channel. As the width of the body is not accurately known, it is impossible to estimate the depth/width ratio of the channel, and thus also not possible to determine whether the channel exhibited any meandering or braiding. Similarly, the sedimentary structures in this exposure are not diagnostic of any particular style of channel behaviour. From the occurrence of facies 3 sheet sands (? levêe or crevasse splay) at the same stratigraphic level as the channel, and on each side of it, it would however appear that the channel did not migrate appreciably, and that the sand body was deposited as the fill to a single period of channel cutting.

On the west face of the trial pit the sequence of events was more complex. Although a channel was initially cut through overbank deposits at the same stratigraphical horizon as those cut by the channel in the east face, the history of the fill of the channel in the west face was different. While the east face channel was filled with cross bedded sands, indicative of sustained current action, the west face chanel was filled by individual influxes of sand, some of which drape the channel margin and pass laterally into levêe or crevasse splay sheet sands (Fig. 17). Furthermore, after the channel was cut, it remained free of accumulating sediment, and lacked strong current conditions for long enough periods to allow the growth of two substantial Sigillaria lycopod trees in the base of the channel. The first influxes of sand, filling the lower part of the channel, surrounded the trunks of these trees. Subsequently the upper parts of the trunks were truncated.

It is evident that the fills of the two channels exposed on the opposite sides of this trial pit were not deposited under the same hydrological conditions. From the coincidence of their locations, stratigraphic horizon, and lateral relationship to the same set of facies 3 sand sheets, these channels must have been related. The sequence of their development was probably as follows:

i) the west face channel was cut, probably in one high flow régime episode, but no sediment was deposited in it, and it existed either as part of the backswamp or as a clear water channel which was colonized by trees. Tricart (1977) has recorded preferential colonization by trees of abandoned channel segment of the Japura and Solimões Rivers in Brazilian Amazonia. His explanation for this is that these trees benefit from the supply of dissolved minerals brought in by floods. This may have encouraged tree growth in comparable circumstances in the Carboniferous.

ii) subsequently the channel was recut, at a slightly different orientation. The fill of this later cut, seen in the east face, accumulated under sustained currrent conditions. Sand transported in the later channel spilled over the banks during floods to fill the earlier channel in discrete phases, and form overbank sand sheets, possibly in levées. The channel was filled and abandoned fairly rapidly, and was not able to develop a meandering or braided pattern.

It is also possible that the channels in the two faces were both open and active at the same time, forming part of an anostomosing channel system.

The restricted lateral extent of major sandbodies in the Blackband Formation suggests that similar rapid cutting and filling of fairly low sinuosity channels may have been fairly common in this Facies Association.

3.3 Palaeosols in Facies Association I

Two types of palaeosol are present in this Facies Association. Both have, in the past, been referred to as "seat-earths", and both are often overlain by thin coal seams. As the coals observed in this study always overlie palaeosols, and as, in modern environments, peats are regarded as soils, these coal seams are regarded as a polyphase soil developments of the two recognised palaeosol types, and are described in relation to them.

3.3.1 Palaeosol type 1 (seat earth)

Palaeosol profiles of this type have been observed rarely in the present study, although they are extremely common in the productive Westphalian coal measures in Europe and North America. They have been described by Huddle and Patterson (1961), Roeschmann (1971), and Wilson (1965).

The palaeosols consist of grey mudstone, which is occasionally silty and often carbonaceous. In the field they are occasionally distinguishable by their lack of bedding and often abundant slickensiding. There is a gradational downward transition into bedded sediment, and usually a sharp upper boundary, overlain by a coal seam or strongly carbonaceous mudstone, or by bedded sediment.

The most conspicuous pedogenic feature of these palaeosols is the abundant occurrence of carbonaceous roots (as opposed to their sporadic occurrrence in Facies 1 and 2). The roots are randomly oriented. All of the roots that have been identified have been appendages and major root systems of Stigmaria, the branching root system of lycopod trees.

Type 1 palaeosols usually contain siderite concretions. These occur in two forms: a) as sphaerosiderite; and b) as ellipsoidal and mammillated concretions (sensu Brewer 1964) between 5 and 10 cm in diameter. Sphaerosiderite occurs as spherulitic concretions, not exceeding 1 mm in diameter. The concretions are found either randomly distributed, or in aggregates associated with root structures. The remaining distinctive features of palaeosols of type 1 are the presence of listric surfaces and slickensides. Listric surfaces are closely spaced, sub horizontal, undulose and polished fracture surfaces. These are characteristically developed in carbonaceous mudstones, to which they impart a shaley appearance. Slickensides are randomly oriented to sub-vertical concavo-convex surfaces, often with striations indicating vertical or oblique displacement. Neither are likely to be original pedogenic features.

Type 1 palaeosols show very little, if any, horizon development. Individual palaeosols do not usually exceed 2 m in thickness, but thicker stacked sequences often occur as a result of intermittent sedimentation during soil formation. Because of the lack of pronounced horizon development, it is difficult to separate phases of soil development in such a sequence, and it is by no means certain that even thin palaeosols of this type really represent a single phase of pedogenesis on a non-aggrading surface. A typical profile is that found at 56.70 m in Playground No. 8 borehole (Fig. 22):

Thickness m.	
0.05	Strongly carbonaceous mudstone
0.30	Light grey mudstone, slickensided, with abundant stigmarian appendages.
0.20	Light grey mudstone, blocky fracture, containing 2 cm siderite nodules.
0.50	Bluish grey silty mudstone with Stigmaria; siderite nodules in basal 20 cm.

Parent material: Medium grey, horizontally laminated silty mudstone, with carbonaceous plant fragements.

Interpretation

Palaeosols of this type can be compared closely with alluvial soils at present forming in backswamp areas of aggrading fluvial systems, for instance the Fly River (Paijmans et al 1968) and the Sepik River (Haantjens 1979), both in Papua New Guinea. Such soils tend to be a at a very early stage of development, showing little evidence of leaching, or of downward movement of clay or humus, and thus little or no B horizon development. They consist of rooted alluvium overlain in some cases by peaty surface layers. In some cases such soils are heavily gleyed.

In the Type 1 palaeosols, thin carbonaceous upper horizons (less than 10 cm thick) may be interpreted as organic surface layers of alluvial soils. Where a coal seam is present it is likely that this represents a subsequent phase of soil development, in which the "seat-earth" (rooted mudstone) part of the palaeosol was not involved. The coal seam is the compacted residue of a peat accumulation, which was probably up to ten times the thickness of the coal. Peats form in condition of extreme waterlogging and acidity. There is no circulation of water below the surface, and as a result the body of the peat is strongly anaerobic (Fitzpatrick 1971). Thus plant and animal life is restricted to the surface, and any underlying alluvial material is not part of the soil. Peats are termed histosols or organic soils (Birkeland 1974). They are widespread in the backswamps of the alluvial systems in Papua New Guinea previously mentioned (Paijmans et al. op. cit.; Haantjens op. cit.). The presence of a coal seam overlying a seat earth thus indicates polyphase soil development on the site, with initial development of an alluvial soil, followed by development of an organic soil.

The ubiquitous occurrence of siderite in these palaeosols merits further consideration, as siderite is rarely described in recent soils. The conditions of siderite formation are discussed by Curtis and Spears (1968) and Berner (1971). The essential conditions are low Eh, resulting usually from anaerobic bacterial decomposition of organic matter, and zero sulphide activity. As brackish water contains dissolved sulphate, which is rapidly reduced to sulphide where there is anaerobic bacterial activity, conditions suitable for siderite formation are limited to freshwater, waterlogged environments.

Although siderite is rarely described in modern soils, this may reflect the paucity of detailed descriptions of soils forming in waterlooged alluvial settings. These soils are often regarded as being immature, exhibiting little pedognic evolution, and thus of little interest ot many soil scientists. Fitzpatrick (1971) records siderite as a mineral phase developed at the base of peats. The highly acidic water associated with peat contains large amounts of iron in solution, dissolved, presumably, in part as organic complexes. When this water leaves the peat, iron compounds are precipitated. At the surface of the peat, reaction with atmospheric oxygen gives rise to ferric hydroxide precipitation to form bog iron are (e.g. James 1966). The siderite described by Fitzpatrick may be envisaged to form when acidic solutions containing complexed iron migrate into the less acidic, but strongly reducing environment of a sediment underlying the peat. Sphaerosiderite in alluvial palaeosols may thus, by inference, have formed by the downward migration of complexed iron in acidic solutions derived from an organic rich soil surface horizon.

The syndepositional, pedogenic origin of sphaerosiderite in these palaeosols is illustrated by the association of oxidized sphaerosiderite with hydromorphic mottling in the gleyed alluvial palaeosols of Type 2 (see 3.3.2).

The formation of larger siderite concretions in palaeosols admits of no modern analogy. These concretions are almost certainly of early diagenetic origin (cf. Coleman 1966; Roeschmann 1971).

The non-pedogenic features of the palaeosols - listric surfaces and slickensides - seem to occur preferentially in palaeosol horizons. This is probably not fortuitous.

Listric surfaces occur most frequently in mudstones containing detrital organic debris. A complete gradation appears to exist between bedding surfaces with carbonaceous plant fragments, and listric surfaces. This gradation accompanies an increase in the amount and decrease in the size of the organic debris present. In vertical thin section, carbonaceous listric mudstones consist of flattened specks of organic matter in a strongly birefringent clay matrix (Fig. 23). This texture, the unistrial fabric of Brewer (1964), results largely from compaction. The high degree of preferential orientation of the clay probably results from it having been sandwiched between layers of organic material. The role of organic matter in causing preferential clay orientation during compaction has recently been illustrated by Schiller (1980).

Listric texture is thus the compaction product of a structureless mud containing fine grained organic matter. Although such a lithology does not necessarily form part of a soil, the common occurrence of listic textures with other palaeosol features suggests that such textures do form preferentially in palaeosols. (The presence of listric surfaces is often regarded as indicating the presence of a seat earth palaeosol by National Coal Board geologists).

Slickensides are common in modern soils, often found on cutanic surfaces separating peds (Brewer 1964). Apart from the low preservation potential of such features in ancient buried soils, there is no evidence in this group of palaeosols for clay illuviation having occurred, without which ped and cutan formation is not possible. Huddle and Patterson (1961) note that slickensides are a common feature of seat-earths, and suggest that they are formed by compaction and dewatering of clay sediments, associated with the decay and collapse of plant roots, in particular of the fleshy stigmarian root types. Where carbonaceous plant roots are present in the palaeosols under consideration they often have slickensides associated with them. This supports an origin for all the slickensides comparable to that suggested by Huddle and Patterson.

Sedimentological implication

Type 1 palaeosols were formed in permanently reducing conditions of high water table, in which the intermissions between phases of deposition were too short to allow the development of a more evolved soil. In cases where clastic supply was cut off for longer periods, colonization by abundant vegetation led to two phases of soil development, in which an alluvial soil was succeeded by an organic soil. Both of these soil types are typical of the soils which develop at present in backswamp areas of alluvial systems.

The extremely widespread extent of some of the coal seams in this group of palaeosols (e.g. in the Blackband Formation in North Staffordshire: see 6.1.2) suggests that the transition from alluvial soil formation to organic soil formation sometimes resulted from cessation in overbank sedimentation over wide areas, which, with continued subsidence led to increased waterlogging and peat formation.

3.3.2 Palaeosol type 2: seat earths with a contemporaneous oxidized horizon

These are palaeosols similar to those of Type 1, but containing brown pigmentation and/or red mottles. Only one example has been observed in the field, at Ibstock Himley Quarry. Here, apart from the pronounced brown colour, the macroscopic appearance of the palaeosol is similar to that of a Type 1 palaeosol.

The main pedogenic features in this type of palaeosol are:

i) Roots. As in palaeosols of type 1, lycopod roots are extensively developed. In addition non-stigmarian root systems are occasionally present, best seen at 11.45 m in section 1 at Ibstock Himley (Fig. 24). These roots are dominantly vertical, not greater than 2cm. wide, with a vertical extent of about 20 cm. They have infrequent downward branches. At Himley they are preserved as siderite concretions. In one instance, at 29.20 m in Kibblestone borehole, roots of this type are infilled by aggregates of 0.25 mm cystals of pyrite.

ii) Sphaerosiderite. Spherulitic siderite is extremely common in this group of palaeosols. As it is often oxidized in, or associated with, oxidized mottled patches which are interpreted as pedogenic mottles (see iii) below), it must be pedogenic in origin.

iii) Mottles. Diffuse red mottles up to 3 cm across are developed in otherwise unoxidized carbonaceous palaeosols of this group. The mottles are usually a dull greyish red colour, and are preferentially developed around roots and aggregates of sphaerosiderite. Examples have been observed in the Rosemary Hill and Playground No. 8 cores. In a more extensively oxidized type 2 palaeosol at 53.20 m in the Playground No. 8 core the whole of the lower horizon is oxidized to a dull grey and greyish red colour, and sporadic mottles of bright red haematite are developed.

The presence of these red mottles in an otherwise unoxidized claystone sequence is difficult to explain, unless they were formed soon after deposition, before the permeability of the sediment, and thus its access to oxygen, was lost by compaction. The association of the red mottles with root traces in palaeosols strongly suggests that such early oxidization was pedogenic.

Both listric surfaces and slickensides are common non-pedogenic features of this group of palaeosols.

Two typical profiles of brown pigmented palaeosols are:

1. Palaeosol at 22.70 m, Rosemary Hill No. 5 borehole.

m.	
0 - 0.01	Mudstone, strongly carbonaceous, dark grey, many listric surfaces.
0.40	Core fragmentary. Mudstone, medium grey, slickensided.
0.40	Mudstone, deep grey, slickensided, contains listric surfaces and some roots. 0.10 Mudstone, silty: as previous.
0.70	Mudstone, slightly silty: grey, with diffuse brown colouration and some sphaerosiderite.
0.50	Mudstone, slightly silty; deep brown.

Underlain by carbonaceous top horizon of subjacent palaeosol.

2. Palaeosol at 20.85 m, Rosemary Hill No. 2 borehole.

m	
0.45	Mudstone, dark grey; carbonaceous
0.30	Mudstone, deep brown, much sphaerosiderite.
0.80	Siltstone, deep grey with brown tinge; contains sphaerosiderite.

The upper 0.40 m of the first of these profiles may belong to a separate, superjacent palaeosol.

The red - mottled palaeosols are very similar, except that a zone of more complete oxidation causes the lower part of the profile to be red or red mottled.

Two typical profiles are:

1. Palaeosol at 14.50 m. Rosemary Hill No. 11 borehole.

m	
0.90	Mustone, dark grey; with abundant listric surfaces and slickensides.
0.80	Mudstone, dark grey; with sparse sphaerosiderite.
0.70	Mudstone, grey with diffuse red patches; abundant listric surfaces and slickensides; some pyrite.
Parent material:	Silty mudstone, dark grey, horizontally laminated, carbonaceous plant fragments.

2. Palaeosol at 53.20m, Playground No. 8 borehole (Fig. 25).

0-0.01	Mudstone, black; carbonaceous.
0.80	Mudstone, deep grey; listric, some carbonaceous roots; 4 cm. siderite concretions at base.
0.30	Mudstone, medium grey; many Stigmaria appendages. Red pigment concentrated on listric surfaces, and rarely in diffuse patches not greater than 1cm. across.
0.30	Mudstone, medium grey; some Stigmaria appendages. Much disseminated fine grained sphaerosiderite. Red

pigment occurs in diffuse 1cm patches, sometimes forming haloes around carbonaceous roots. 10cm thick horizon of siderite concretions near base, some formed of sphaerosiderite aggregates.

Interpretation:

Palaeosols of this type have clearly undergone a more complex evolution than those of type 1. In the early stages of their formation, the soil was probably similar to type 1, with the formation of siderite occuring at this time. (Larger siderite concretions, such as those illustrated in Fig. 24, may have formed during post burial diagenesis).

The occurrence of oxidized mottles in a soil dominated by reduced iron and organic matter is probably due to gleying, owing to water table fluctuation. The association of mottles with roots and other weak zones in the palaeosols is reminiscent of the description of recent hydromorphic mottling in grey soils by Bloomfield (1964). The presence of a carbonaceous surface horizon suggests, however, that the soils were dominantly waterlogged.

The poor profile development of these soils, and their association with coals, indicate that they were alluvial soils. Comparable recent soils, in which slightly improved drainage has allowed partial oxidation in an otherwise completely reduced sediment, have been described by Slager and van Schuylenborgh (1974). Two of the profiles that they describe, the Ma Retraite and Santo profiles, are of soils developed on the Young Coastal Plain in Surinam. Here coastal progradation following the establishment of the present seal level, ca. 6000 years ago, has led to improved drainage conditions in what were formerly waterlogged coastal swamps. The area is now one of fresh water swamps with intermittent seasonal desiccation (Brinkman and Pons 1968). The soil profiles are characterised by peat or organic-rich surface layers, in which iron is all in the reduced state, and heavily gleyed B horizons. These are dominantly grey, but contain hydrated ferric oxides concentrated along fractures, and forming concretions which fill root channels. The parent materials are dark grey, organic rich mudstones.

A closer analogy between these soils and the type 2 Etruria palaeosols is not possible. The parent material of the Surinam soils was deposited under brackish conditions, which has led to the presence of a large amount of pyrite in the soil profiles. This feature is largely lacking in the Etruria palaeosols.

Hydromorphic mottling has frequently been described in ancient palaeosols (Buurman 1975, 1980; Freytet 1971; Retallack 1976).

It is not known why the ferric pigment in some of these palaeosols is brown. Brown pigmented rocks are described in association with red beds by McBride (1974), but only in rocks of coarser grain size. Where the pigment is not the result of weathering, he regards the brown colouration as resulting from the small amount, or 'improper distribution' (sic) of haematite pigment.

Sedimentological implications:

Both brown and red pigmented seat-earths are described from coal measure sequences by Elliot (1968), who regarded their partial oxidation as the result of their having developed in slightly better drained conditions than completely reduced seat-earths. He demonstrated that, in the Westphalian A and B of Nottinghamshire, there was an inverse relationshipship between the number of brown seat-earths and the thickness of sediment present, implying that drainage conditions were better in areas of lower subsidence. Also, a belt of brown seat-earths is present immediately above a major sand body of limited lateral extent. Here better drainage conditions resulted from differential compaction.

In the Nottinghamshire Westphalian studied by Elliot (op. cit) red pigmented seat-earths are only present in very thin sequences found towards the inferred edge of the depositional basin. Elliot used this stratigraphic relationship to suggest that red pigmented seat earths have undergone more prolonged oxidation during pedogenesis than the brown pigmented seat-earths. This implies that the red pigmented palaeosols were deposited in a better drained environment.

A close analogy may be drawn between the oxidation which affected palaeosols and the well drained swamp environment described by Coleman (1966) in the recent overbank areas of the Atchafalaya River. Here oxidation of the backswamp sediments takes place as a result of very slight improvements in drainage. The nature of such drainage improvement in the Upper Carboniferous rocks under discussion is considered in 6.1.3 and 8.5.

The presence of coal seams overlying some palaeosols of type 2 implies that, during subsidence, some of these soils underwent polyphase evolution, presumably from intermittently drained alluvial gley to poorly drained alluvial soil to organic soil.

3.4 Facies 5, 6, and 7

These Facies have not been observed in the grey, coal bearing Facies Association I sediments observed in the present study. Facies 5 and 6 do, however, have exact or close counterparts in Facies Association IIA, and a brief description is therefore necessary.

The facies descriptions and interpretations included here are derived from descriptions of their occurrence in the Blackband Formation of North Staffordshire by Gibson (1905), and by Boardman (1978, 1981, and personal communication). Comments by the present author are indicated by square brackets.

3.4.1 Facies 5 - entomostracan 'limestones'

The 'limestones' of this Facies form two widespread marker horizons at the base of the Blackband Formation, which have been found in all areas where that Formation can be distinguished from the diachronous Etruria Formation in the North Staffordshire area (see 9.3).

The 'limestones' vary from highly calcareous mudstones to fairly pure limestones composed of varying proportions of lime mud and bioclastic debris. The faunal elements present are the same as those in the shelly facies of the Blackband ironstones (Facies 6), viz: non-marine bivalves, ostracods, *Spirorbis*, *Euestheria*, and fish remains. Gibson observed that ostracods are the dominant faunal element, and proposed the term 'Entomostracan limestone' used here to replace the more commonly used term '*Spirorbis* limestone' (e.g. Bennison and Wright 1969), the occurrence of *Spirorbis* being comparatively rare.

The descriptions referred to above all relate to limestones observed in North Staffordshire. No unequivocal examples of this facies have been observed by the author, and descriptions of limestones from other parts of the study area are not available. Limestones are recorded from the Etruria (or equivalent) Formation in North Wales, Lancashire, and Warwickshire (see 4.4.2), but do not seem to occur within grey, coal bearing facies outside North Staffordshire.

3.4.2 Facies 6 - Blackband ironstones and oil shales

This Facies, which occurs only in North Staffordshire, has not been observed during the present study, but is present in a grey, coal bearing intercalation in the Etruria Formation in the Chesterton area (Gibson 1901). The Facies is characteristically developed in the Blackband Formation, and at horizons down to the Great Row coal.

The Facies consists of siderite ironstone, containing abundant sapropel and minor ankerite and calcite. It occurs in massive beds, up to 1.50m thick. The whole thickness consists of repeated alternating layers of dark and light grey-brown colours. These bands correspond to laminae of sapropel rich and sapropel poor ironstone, and are generally between 1 and 6mm thick. The sapropel material includes the kerogenous alga *Botryococcus*. The horizontal lamination of the sediment is occasionally disturbed by rooting.

In thin section the siderite occasionally occurs in peloids of microcrystalline grain size. Miospores have been observed preserved by siderite, in a non-compacted shape.

This Facies always occurs immediately overlying the coal horizon of a palaeosol. The two lithologies are occasionally interbedded. The ironstones are laterally extremely persistent, occuring over large areas of the North Staffordshire coalfield. In the western part of that area the ironstones exhibit a lateral facies change into impure, siderite rich, limestones. This change takes place initially by the appearance of layers of shell material in the ironstones. The fauna consists of bivalves (Anthraconauta phillipsi), ostracods (mainly Carbonita spp.), annelids (Spirorbis), crustacea (Euestheria) and fish remains. In some cases the calcareous shell material is replaced by siderite.

In the upper part of the ironstone layers there is usually a gradation into a thin (not greater than 0.30m) oil shale. This consists of extremely sapropel rich mudstone, containing large concentrations of the Botrococcus kerogenous alga. In the lower part of the ironstones there is usually a downward gradational into pure sapropel material, which forms a cannel coal roof to the underlying coal seam.
Where both an ironstone and an oil shale are present, there is a tendency for their thicknesses to vary laterally inversely; i.e., where the oil shale is thick, the underlying ironstone is thin, and vice-versa.

The environmental interpretation of this Facies is discussed in 3.4.4.

3.4.3 Facies 7: 'varved' mudstone

This Facies has not been observed in the sediments which are intercalated with, and immediately underlie the Etruria Formation. It has thus not been observed in the course of this study, although it occurs at several horizons in the underlying Blackband Formation.

The Facies consists of grey claystone, showing a rhythmic alternation of light and dark grey laminae, which resemble varves. The colour difference is caused by variation in the content of fine grained carbonaceous material between the laminae. Bedding planes are sometimes crowded with a fauna of non-marine bivalves (Anthraconauta phillipsi), ostracods, annelids (Spirorbis) and crustacea (Estheria), and with plant fragments.

The 'varved' mudstones are always intimately associated with Blackband ironstones (Facies 6) and/or limestones (Facies 5), and are discussed together with these two Facies.

3.4.4 Interpretation: Facies 5,6, and 7

These Facies are all regarded by Boardman as being lacustrine deposits. He divides them into two associations.

a) Entomostracan limestones. These are interpreted as the deposits of lakes which were well oxygenated, and occasionally widespread. Water depths were probably very shallow. The calcareous deposits imply that the water was hard, and the presence of Euestheria suggests that slightly brackish conditions occasionally prevailed. In view of the lack of any palaeontological or sedimentological indications of proximity to marine waters, this probably resulted from strong evaporation (cf Pollard and Wiseman 1971; see 4.4.2).

b) Blackband ironstones, oil shales and varved mudstones. These facies occur together overlying coal seams, in a vertical sequence; coal - cannel coal - blackband ironstone - oil shale - varved mudstone - any Facies Association I mudstone or siltstone lithology (Fig. 26). They are therefore interpreted as a related succession of deposits, formed in progressively deepening lakes which developed during subsidence of the peat horizons (which now form coal seams) following the cessation of plant growth and prior to the resumption of clastic sedimentation.

The earliest phase of lake development occurred as plant growth in the underlying peat ceased and accumulation took place of the finely macerated organic debris which now forms the cannel coal roof of the coal seam. With the establishment of an open water body, the oxygenated conditions at the surface of the lake caused the oxidation of highly acid, iron-rich waters draining from surrounding unflooded areas of peat. This oxidation gave rise to a precipitate of 'ochre' (ferric hydroxide), which underwent rapid early diagenetic reduction to siderite in the lake bottom sediment pile. At the same time regular, possibly seasonal, fluctuations of the amount of organic matter precipitated led to the deposition of a siderite rich sediment with repeated thin organic laminae. [The process of iron concentration and precipitation suggested is similar to that involved in the formation of modern bog iron ores (e.g. Blatt _et al._ 1972 p.575)]

The lack of fauna in the ironstones in most of the North Staffordshire area suggests that for much of the time the lakes were unsuitable for colonization by any fauna, as a result of the large algal production. Only in the western part of the area, perhaps near a tectonically controlled margin of the lake(s) (see 9.3), was any faunal colonization present. [This probably reflects deposition in shallower water, which may have had two effects:

a) to reduce algal production as a result of increased water turbulence;

b) to raise the lake bed above any seasonal or permanent layering that may have been developed in areas of greater subsidence.]

[The limnology and bulk chemistry of the lacustrine system responsible for the deposition of the Blackband ironstones are not discussed in detail by Boardman, and fall outside the scope of this study. In particular, the question of the supply of carbonate for siderite formation remains unanswered.]

Towards the end of the formation of the Blackband ironstone layers, the supply of iron dwindled. This probably resulted from deepening of the lakes, with a consequent reduction in the area of peat swamp from which local derivation of iron and bicarbonate rich waters occurred. As long as conditions of low clastic input lasted sedimentation of algal-rich mudstone continued, forming the oil shale which overlies the ironstone layer.

Clastic infill of the lake is represented by the "varved" mudstone facies. The lack of silt and regular alternation of organic rich and organic poor laminae demonstrate that sedimentation was entirely from suspension [unlike the superficially similar lacustrine coal-measure sediments described from a comparable setting by Haszeldine in Leeder et al. (1981)]. Periodic (?seasonal) fluctuation of organic production is still indicated, but this was insufficient to render the water hostile to the development of a typical fresh water fauna. Infilling of the water body is represented by an upward gradation from this Facies to any of Facies 1 to 3, marking a reversion to poorly drained swamp deposition. The influx of clastic sedimentation which ended each phase of ironstone/oil shale deposition was possibly due to the breakdown of vegetation which was acting as a clastic trap during the development of lacustrine facies.

3.5 Sequential relationships within Facies Association I

Depositional sequences within this Facies Association are characterised by the following measured sections: Rosemary Hill No. 11 borehole. 6m–20m (T.D.) (Fig. 27); Rosemary Hill No. 3 borehole, 14m–33.79m (T.D.) (Fig. 28); and Mitchell's Wood Opencast site (SJ832508; Fig. 3), between the Red Shagg and Red Mine horizons (Fig. 29: data from Boardman 1981).

The pattern displayed is similar in all these examples. The regular occurrence of palaeosols indicates repeated phases of non deposition and plant colonization, in some cases followed by peat formation. The distribution of the other floodplain lithofacies appears random within packets of sediments deposited between phases of pedogenesis. Siltstone and sandstone sheets occur near the top of such a packet at 26m–28m in borehole no. 3, and towards the base of such a packet at 11m–13m in borehole no. 11. There are insufficient data to determine whether these patterns are either typical or significant. The distribution of these thin sands in a vertical sequence is presumably controlled by the behaviour of nearby channels, which is not known.

These sequences are very similar to those described in a coal measure overbank association by Scott (1978), and are interpreted similarly. Subsidence in the swamp surface was balanced by phases of sediment deposition, between which palaeosols developed. The nature of individual packets of sediment between palaeosols was controlled by the proximity and behaviour of the channel systems which fed water and sediment into swamps. Because of a lack of examples (only one has been observed) the relationships at alluvial channels to alluvial swamps in this Facies Association is not known.

The sequence in the Blackband Formation is similar, except that the lacustrine facies association occurs at the base of a thicker sediment packet, which overlies a thick coal. This coal, unlike the thin coals in the sequence, is regionally extensive, occurring, with its ironstone, over the whole of North Staffordshire. The occurrence of a thicker than usual, regionally extensive coal, a regionally extensive lacustrine development and a thicker than usual sediment sequence between the coal and the next palaeosol, all combine to suggest that the subsidence mechanism giving rise to these phenomena was different to that controlling deposition of the swamp sequences which lack lacustrine facies. In the latter case the dominant subsidence control on facies was that induced by compaction set against a background of regional subsidence. In contrast regional subsidence was probably the main control on the occurrence of extensive coals and lake deposits.

This may explain why lacustrine facies and extensive coals have not been observed in Facies Association I sediments outside North Staffordshire. The pattern and origin of this subsidence is discussed in Chapter 10.

The occurrence of red pigmented, Type 2 palaeosols in borehole No. 3 (Fig. 27) is typical of sequences in this Facies Association. Its significance is discussed in 6.1, and 8.3.

CHAPTER 4

Facies Association II : well drained alluvium deposited by meandering rivers

4.1 Introduction

The rocks making up this Facies Association form the majority of the Etruria Formation. They consist of thick sequences of red and variegated mudstones with minor sandstones, and of occasional sandstone bodies. These two sediment types are interpreted, respectively, as the floodplain and channel deposits of meandering fluvial systems. To render their description and discussion more digestible, the two types are subdivided into Facies Association IIA (floodplain sediments) and Facies Association IIB (channel deposits).

4.2 Association IIA : predominantly well drained floodplain sediments

4.2.1 Facies 8 : red mudstones

Red mudstones occur in units up to 3m thick (Fig. 30). Together with red silty mudstones (Facies 9) they are extensively exposed in working brick-clay quarries, and make up the majority of the Etruria Formation. They are usually massive, and characteristically weather to a spheroidal blocky texture. The dominant colour is 10R 3/4, dark reddish brown. Internal structure is rarely visible, although a faint horizontal lamination is sometimes visible on freshly broken surfaces. It consists of an alternation of non-silty and slightly silty bands 1 to 2mm thick.

In those mudstones which show no depositional structure, individual root traces are occasionally visible, picked out by pale grey pigment or by an ochre coloured, sometimes concretionary ?goethite fill to the supposed root tube. More commonly a sporadic mottling of 1mm patches of grey (N7), ochre (2.5Y 7/6), and purple (5P4/2) pigment suggests that the sediment was extensively disturbed, probably by root action. Where this texture occurs there is usually an upward gradation into a palaeosol (see 4.6.1).

Apparently structureless red mudstones which also lack sporadic colour mottling may still, however, be rooted; although root traces have not been observed in this lithology in surface exposures, they are seen in fresh core material as oxidized, listric traces (Fig. 31) resembling the *Stigmaria* appendages observed in the alluvial palaeosols in Facies Association I.

Some horizons in the mudstones are calcareous and contain calcareous concretions. These take two forms: i) 5 to 10cm flattened spheroidal concretions occuring at one consistent horizon; and ii) spherical concretions ranging in size between 0.2 and 4mm. The first type are uncommon, having been observed only at Manor and Rosemary Hill Quarries (SJ885445 and SJ830460; Fig. 3). In the latter case, not observed *in situ,* the concretion surrounded the oxidized trace of a *Neuropteris* (pteridosperm) leaf bearing branch. The small spherulitic concretions, known locally as 'shot' or 'shotty', are ubiquitous in North Staffordshire, where they occur randomly distributed through mudstone and silty mudstone units up to 6m thick. They are also found as a reworked constituent of sheet sands (Facies 10) and minor channel fills (Facies 11) in the overbank association, in some instances forming the majority of the sand sized particles in these bodies. From these occurrences it is evident that such small concretions are of very early diagenetic origin.

In rare cases small concretions of spherulitic siderite, often partially oxidized to haematite, occur within red mudstones. Examples have been found at Wilnecote Quarry (SK220000; Fig. 1) and in Playground No. 8 borehole (SJ72124; Fig. 4) cores.

Desiccation cracks have been observed in Facies 8 at three localities. In Section 1B of Manor Quarry, Fenton (SJ885445; Fig. 1) and at Himley Wood (SO89911; Fig. 5) they immediately underlie thin sheet sands (Facies 10) and have sand fills. The maximum observed depth is 1.50m and widths vary from 0.5 to 5cm. In both cases the desiccation crack fills are distorted owing to differential compaction between them and the surrounding mudstone. In Wilnecote Quarry (SK220000; Fig. 1) desiccation polygons were exposed on the quarry floor following excavation in the summer of 1977. The polygons, 15 to 35cm across, were in a red mudstone unit. The desiccation cracks, up to 4cm wide, were preserved as a pale grey mudstone (Fig 32). When seen in vertical section, the desiccation cracks were easily mistaken for mottles produced by root activity; it is thus likely that such features have been misidentified in other vertical sections.

Interpretation

The fine grained nature of the sediment and the presence of horizontal lamination suggest deposition out of suspension in quiet water. The ubiquitous presence of rooting and the occurrence of desiccation features indicate that water depths were very shallow. As in Facies 1, the analogue with modern tropical vegetation suggests that water depths did not exceed 2m. Plant growth may have taken place in shallow water. Alternatively water cover and sedimentation were ephemeral, and plant growth took place on an exposed floodplain, on which the rate of sediment accumulation was too high for the formation of a soil profile.

Deposition of fines from suspension in a water cover which was either intermittent or shallow suggests an extensive, near subaerial environment. From the regular association with meandering channel deposits (Facies Association IIB), and with fresh water organisms (in Facies Association I) the depositional environment is interpreted as an alluvial plain and/or floodbasin. In the former, the lack of ponding caused by depositional topography would have led to a well drained, intermittently inundated area, while the latter would have been permanently waterlogged, if not actually submerged.

Although the deposits of these two environments can easily be subdivided on the basis of the palaeosols which occur in association with Facies 8 (see 4.3), it is not easy to differentiate the degree of waterlogging which prevailed at the time of deposition of sediment which lacks well developed palaeosol profiles. The usual characters of swamp sediment (iron in Fe II state, siderite, abundant plant debris and organic material) have usually been destroyed by syn- or post-depositional oxidation. The only criterion that can confidently be applied is the nature of contained concretions. Coleman (1966) in his description of the overbank sediments in the Atchafalaya valley distinguishes clearly between the early diagenetic history of poorly- and well-drained areas. In poorly drained areas iron is in FeII state, and the dominant concretion types are pyrite and siderite. By contrast, in well drained areas iron is in FeIII state and goethite and calcite concretions form. Where siderite, or haematite pseudomorphing spherulitic siderite, is present in Facies 8 red mudstones, one may be reasonably confident that deposition took place in a waterlogged area. The much more common occurrence of extremely early calcite concretions suggests that deposition on a well drained floodplain was more frequent.

Where such independent criteria are lacking, it is reasonable to assume that dominantly well drained conditions prevailed. The very high iron content of the majority of the mudstones (see 7.1) suggests that the iron has never been in the potentially more mobile FeII state since deposition. This is in accord with Tricart's (1972 p.65) observation that, on non-waterlogged floodplains of humid tropical rivers in Africa, the overbank alluvium is brown in colour (i.e. contains FeIII), and suggests deposition in well drained conditions.

The diagenetic history of iron in these sediments is discussed further in 8.3 and 8.4.

4.2.2 Facies 9 : red silty mudstone and siltstone

This Facies consists of mudrocks which are recognisably silty to the touch in the field. In some cases the content of silt and fine sand is such that they are better described as muddy siltstones. Two main types may be distinguished:

i) Laminated silty mudstones. These consist of horizontally interlaminated silt or silt-rich laminae and silt-poor laminae, 1 to 3mm thick (Fig 33). The silt-rich lithology is usually picked out by patchy grey colouration, while the silt-poor material is deep red brown (10R 3/4). Small scale trough cross lamination is sometimes visible (Fig 34). The horizontal lamination is often only crudely developed, perhaps being better described as crude horizontal bedding, recognisable horizontal partings being up to 20cm apart.

Laminated siltstone and silty mudstone units are usually sharply based, and most often pass gradationally upwards into:

ii) Apparently structureless silty mudstones. In surface exposures these have a similar appearance, other than in their colour, to Facies 8 mudstones, showing no obvious internal structure, and weathering to a blocky texture. Rooting is ubiquitous, root traces being visible as bleached zones or filled by concretions of ?goethite. More commonly a random colour mottling is present as mm to cm size patches of ochre (2.5Y 7/6 to 2.5Y 7/4), pale red (10R 6/2), purple (5P 4/2), or light grey (N7). These apparently structureless silty mudstones pass gradationally upwards into Facies 8 mudstones or into a palaeosol.

Silty mudstone units are usually not greater than 3m thick, and are commonly 1-2m. The laminated lithology is much less common than the rooted lithology, and always passes upwards into it.

Locally, both types are calcareous, and contain 'shot' concretions. In places these are reworked into grain supported sheets, which are described under Facies 10.

In the Springfield south quarry (SJ863442; Fig. 3) the bedding surfaces of the laminated facies are crowded with completely oxidized impressions of Calamites stems. Although not seen elsewhere in exposures, oxidized plant fragments have been observed in this facies in the Kibblestone (SJ911361; Fig. 3), Street's Lane (SJ972061; Fig. 4), and Allotment No. 1 (SJ947268; Fig. 1) boreholes.

Interpretation.

The depositional environment of this Facies was broadly similar to that of Facies 8. The higher proportion of silt probably indicates a more proximal position relative to the alluvial channel system. The presence of lamination, especially in the lower and siltier parts of the units (type i) silty mudstones) suggests that the deposition of these units was initially rapid, outstripping the capacity of vegetation to disturb the sediment by rooting; alternatively this part of the sediment was not vegetated. The presence of cross lamination indicates some reworking by traction currents during deposition. These were, however, exceptional, the poor sorting and mud content of the sediment precluding any sustained current action.

The random colour mottling present in the majority of this Facies is, as in Facies 8, probably the result of root disturbance and incipient soil formation. This also accounts for the lack of internal structure. Both indicate that, during deposition of the majority of Facies 9 sediments, emergent vegetated conditions predominated.

The internal structure of fine grained overbank deposits formed adjacent to a fluvial channel has been described in some detail by Klimek (1974). Describing recent alluvium in the dissected terraces of the Wisloka river in Poland, he recognised that proximal deposits form units of between 20 and 40cm thickness, each of which consists of three layers; a lower, sharp based, structureless loam, with abundant compressed plant material at its base, a layer of cross laminated sand, and an upper layer of structureless loam. The basal structureless layer is formed by rapid sedimentation from suspension during the first phase of the flood, owing to the cover of grass and low shrubs slowing the heavily laden current. The cross laminated layer is formed when more rapid current conditions occur after vegetation has been buried, and the upper layer by renewed sedimentation out of suspension during the waning flood stage. The cross laminated sand unit is not always present. The most important factor in proximal overbank deposition is the entrapment and filtering of sediment by vegetation. This eliminates the coarser fraction, so that in the more distal parts of the floodplain the overbank deposits consist mainly of clay.

The sediments described by Klimek may not be analagous to those in the Etruria Formation, in that the Carboniferous vegetation may not have included low growing plants. However, Scott (1979) describes a levée/ river bank palaeofloral assemblage from the Westphalian Coal Measures of northern Britain, in which the _in situ_ flora consisted of Pteridosperms and _Calamites._ The former were comparatively low growing, and the latter, being easily flattened, would have had the capacity to entrap sediment in the initial stages of the flood, and subsequently to have been easily buried after falling.

A number of similarities exist between the laminated rocks in Facies 9 and the sediments described by Klimek. The crude 20cm horizontal bedding may represent the deposits of individual flood events. The sporadic occurrence of horizontal and cross lamination is due to intermittent and rapid suspension and traction sedimentation respectively, and the plant debris, particularly the concretations of _Calamites_ at Springfield South Quarry, may represent plant rich units as described by Klimek. Individual sequences such as those documented by Klimek have not been seen. These might be revealed by more detailed study, but it is possible that the combination of an oxidizing phase of diagenesis and compaction may have destroyed characteristic features such as plant material and sedimentary structures defined by concentrations of organic material.

In terms of the larger scale alluvial morphology of the depositional system, the sediments of Facies 9 were deposited as proximal floodplain sediments. These could have formed part of a levêe, or have been deposited on a topographically undifferentiated floodplain.

The sediments described by Klimek were deposited on a levêe. Similar sedimentary structures have been described by Singh (1972) from dissected levêes of the Gomti River in India. While both of these examples contain thin sand units, levêes consisting almost entirely of silt and mud are reported from the Angabunga River (Speight 1965), and the Fly River (Löffler 1977 p.90). Tricart (1972 p.67) notes that levêes of humid tropical rivers may be only slightly more sandy than the deposits of their associated backswamps. In tropical fluvial systems, silt levêes are only well developed in the lower reaches of the floodplain, where, by virtue of the cohesive banks that they form, they confine flow and suppress the development of meanders (Holmes and Western 1969; Löffler 1972 p.90; Speight 1965). Deposits of levêes of this type would have a high preservation potential.

In situations where vigorous active meander development is taking place, levêes are discontinuous and poorly developed (Blake and Ollier 1971; Löffler op cit). Owing to their repeated erosion on the cut bank side of the channel, they have a low preservation potential.

In the Etruria Formation, meandering channel deposits (Facies Association IIB) are found at all stratigraphic horizons at which

Facies 9 silty mudstones occur, except where these occur in Facies Association III (see 5.2.6 ff). It is therefore very probable that proximal levêe deposits will have been preserved infrequently. Distal levêe deposits stand a higher chance of preservation. In the event of the meandering channels having had poorly developed and discontinuous levêes, it may be expected that the coarser parts of the suspended overbank load will still have been deposited near the active channel system. These would be genuine, topographically undifferentiated floodplain deposits, and might be expected to fine outwards from the channel. Apart from their higher preservation potential, sequences deposited in this way might be indistinguishable from channel bank levêe deposits.

The lack of documentation of structures and sequences in recent overbank alluvium makes it impossible to differentiate between levêe and floodplain sediments in this instance.

The same observations regarding the red pigmentation apply as in the case of Facies 8. Further discussion of this will be found in Chapter 8.

4.2.3 Facies 10 : Sheet sandstones

The sheet sandstones in this Facies are almost invariably associated with silty mudstones and siltstones of Facies 9, and occur preferentially at and near the base of these silty mudstone units.

The majority of the sheet sandstones are between 5 and 40cm thick. They are composed of fine to medium grained sand, with some admixed silt. The sheets commonly fine upward, and in occurrences in the Dudley area (Smithy Lane (SO908898; Fig. 5); Himley Wood; Ibstock

Himley) coarse sand and granule material occurs in the basal 5-10cm of the sheets. In several cases (Atlas south (SK044021; Fig. 4) section at 6.20m; Redhurst Wood (SJ969052; Fig. 4) west section 1 at 0.05m; Ibstock Himley section 2 at 4.65m; and Himley Wood at 19.40m) the entire sheet is composed of coarse sand or granule material. In Goldendale (SJ850519; Fig. 3) and Springfield North (SJ861445; Fig. 3) quarries, the dominant grain supported components in thin sand sheets are intraformationally derived mm size spherulitic calcareous concretions ('shot'). At section 3 in Manor Quarry, a sequence of seven thin fining upward grain supported 'shot' conglomerate layers occur in the basal 0.70m of the section.

The thin sandstones are generally green in colour, except where they contain abundant silt and clay size matrix, in which case they are red.

The bases of the units are always sharp; the tops are generally sharp, although in a few cases there is a rapid gradation, over 1 or 2cm, into silty mudstone. In a few cases some cross-lamination is visible.

Generally, exposure is inadequate to follow these sand bodies laterally. In Knutton quarry (SJ828468; Fig. 3) a group of thin sandstones occurs in a 2.20m thick unit of Facies 9 silty mudstone (Fig 35); the upper sand sheets thicken and pass laterally into a small channel fill (Figs 36 and 37) 8m wide and 2m deep, which cuts some of the lower sand sheets in the group. All of these thin sands, both those cut by and those associated with the channel, thin laterally and die out within 50m in each direction of the channel. This packet of sheet sands also shows a tendency for the sheets to become thicker upwards. Laterally impersistent sand sheets of this Facies, dying out horizontally within 50-100m, have also been observed at Chesterton and Keele quarries (SJ827493 and SJ788441; Fig. 3).

Elsewhere, for instance at Himley Wood and Ibstock Himley, groups of thin sandstones are laterally persistent over 100m, although individual thin sandstones are mutually erosive.

Interpretation

From their sharp boundaries and the presence of cross lamination it is evident that these thin sheet sands are the deposits of short-lived eposides of flow competent enough to transport sand bed load, or of high enough energy to transport sand in suspension, into a low energy depositional environment otherwise dominated by sedimentation of fines from suspension. These conditions can be met by two processes in an alluvial setting. Crevasse generation and outflow of sand laden sediment from an adjacent alluvial channel during flood periods (Allen 1965) produces a short lived period of high flow regime, affecting a comparatively small area of the floodplain. Alternatively, unconstricted flow over larger areas of the floodplain may occur during sheet floods. These may be caused by much larger scale bank collapse, or by a massive, short lived overloading of the entire alluvial system following excessively heavy rain.

In the case of examples in which the sand sheets are laterally impersistent, a crevasse splay origin is most probable. Leeder (1974) regarded limited lateral extent as a common feature of crevasse splay deposits. The thickening upward packet of sands, with associated channel, seen at Knutton probably represent a small crevasse lobe prograding onto the floodplain, with a crevasse channel cutting through earlier parts of the lobe, and its fill spilling over to form thicker, later sheets in the lobe (cf crevasse delta processes described in Elliott 1974). Thickening and coarsening upward crevasse lobe sequences have also been described by Friend et al (1981) from the Escusaguás fluvial facies in the Oligocene Campodarbe Group of northern Spain.

Examples of sand sheets with limited lateral extent are restricted to North Staffordshire, where only meandering channel coarse members are present. In the Dudley area, the thin sandstone packets are more laterally extensive, a feature regarded by Heward (1978) as characteristic of sheetflood deposits in distal alluvial fan settings. As meandering channel deposits are fairly uncommon in this area it is possible that here the thin sheet sand Facies is a distal equivalent of the graded sand and conglomerate sheet Facies found in these proximal Facies Association (Facies Association III; Facies 18,19). There is, however, insufficient evidence to justify separating the laterally extensive sheet sands from the remainder of Facies 10.

4.2.4 Facies 11 : Isolated small channels in overbank association sequences

Small channels, ranging from 1.00m to 5m in width and 0.20m to 2m in depth, occur sporadically as isolated features in overbank mudstones and silty mudstones of Facies 8 and 9. In two cases, at Metallic Tileries (SJ840498; Fig. 3) and Knutton quarries, these channels are closely associated with sheet sands of Facies 10. The channel in the latter exposure has been mentioned in 4.2.3.

The channels are all concave based, with sharp tops to their fills, where these can be distinguished from the surrounding sediment (Fig 38).

Three types of fill have been observed:

i) massive coarse sand. The fill of the channel consists entirely of coarse sand, with a varying amount of granule material, apparently deposited as one unit. At Spoutfield (SJ863464; Fig. 3) and in the lower small channel at Springfield South no internal organization is visible, and, in the latter, the sand is poorly sorted. In the upper channel at Springfield South the fill is cross bedded and fines upwards.

In the channel in Knutton Quarry, the fill is composed of seven units, all but one of which are less than 0.70 thick, and are draped over the topography of the channel sides; the remaining unit is lenticular, thickening to 1.10m, and forming the major part of the fill at the centre of the channel (Figs 36, 37). On the south side of the channel each unit passes laterally into a thinner sand sheet, interbedded with silty mudstone.

ii) interbedded sandstone and mudstone. In these instances the channel fill consists mainly of deep red silty mudstone, which is hard to distinguish from the silty mudstone through which the channel is cut. Four examples have been recorded. At Chesterton, the channel is c.1.50m deep, and is recognisable by a group of 2 to 5cm thick, horizontally and ripple cross laminated siltstones lying parallel to the channel side, at a strong angle to the regional dip (Fig 39). At Springfield North the fill of a channel 1.20m deep and 5m wide consists of red silty mudstone, the form of the channel being picked out by thin (less than 25cm thick) coarse sand lenses, and by a slightly siltier fill in the basal 0.50m.

At Spoutfield quarry, two small channels were observed (Fig 40), cut into Facies 9 silty mudstones immediately overlying a sandbody of Facies 14 (see 4.5.3). The lower channel (A in Fig 40), cutting down into the upper part of the sandbody, was probably about 8m wide and up to 2m deep; the upper channel (B in Fig 40) was about 5m wide and 1m deep. Both have symmetrical silt/mudstone fills.

iii) conglomerate. Two types of conglomeratic channel fill occur: orthoconglomerates of extra-basinal origin, in which granule and larger clasts are admixed with poorly sorted sand; and paraconglomerates of extra- and/or intra-formational clasts in a silty mud matrix.

Three nearly parallel channels with extraformational fills were exposed in plan on the floor of Wilnecote Quarry during the summer of 1977. Over a length of some 10m they were nearly straight, averaging 1.30m in width and 0.40m in depth, some 20m apart. The fill consisted of a polymictic granule conglomerate, containing clasts of indurated shale less than 1cm in diameter. No internal structure or organisation were visible. A channelized body 0.40 thick and of unknown width infilled with coarse sandstone and granule conglomerate cuts a Facies 10 siltstone sheet at 18.90m in Redhurst East quarry (SJ972052; Fig. 4).

Two types of intraformational clasts are found in small channel fills: reworked spheroidal calcium carbonate concretions ('shot'); and mud intraclasts. The former type are found as a paraconglomerate in a silty mudstone matrix in an irregularly based channel 0.50m deep and

5.00m wide in Redhurst West, section 3. Here the intraclasts presented a polished appearance in hand specimen, and might have been mistaken for very well rounded quartz grains. A similar situation obtains in a body of unknown size c. 1.90m beneath the coal seam in Wilnecote Quarry. Examination of this specimen in thin section revealed the presence of mudstone intraclasts, 2 to 5mm in diameter, recognisable by a higher silt and haematite content than the enclosing mudstone matrix. The body also contained reworked calcareous 'shot' concretions. The cemented intraclasts were not visible in hand specimen. Other occurrences of this rock type may have been overlooked.

Interpretation

The lack of internal organization and poor sorting in the fills of the massive sand and conglomerate filled channels suggests that they were very short lived, being perhaps cut and filled in one flood event. Where internal organization is present, in the Knutton example and in the channels with an interbedded sand/silt and mud fill, it is probable that the channel was rapidly cut through silt and mud sediments which are usually relatively resistant to erosion, and subsequently infilled over a period of perhaps a few years, the sand units in the fill representing subsequent phases of high flow.

There are four possible mechanisms for the formation of small, rapidly filled channels in the overbank areas of an alluvial system.

1) Secondary drainage systems in the interfluve areas of larger river systems. In extensive alluvial plain areas terracing, resulting from internal or external controls, is supposed to give rise to temporarily stabilized geomorphic surfaces above the base level of the active

alluvial system (Allen 1974). These surfaces may undergo an extended period of soil development, and may have smaller drainage systems developed on them, feeding the local precipitation into the main rivers. No modern described example is known to the present author.

Sequences containing channel deposits of this supposed type have been described from the Lower Old Red Sandstone of South Wales by Allen and Williams (1979). The deposits of the ephemeral influve drainage system (Type B system) consist solely of intraformational calcrete and mudstone dêbris, and are completely lacking in the quartzose sandstone and exotic conglomerate clasts found in the deposits of the major (type A) river systems. Allen and Williams regard this as evidence that the sedimentary fill was derived from within the bounds of the depositional system; and thus the minor drainage systems were independent of the major channels, which carried an extraformational bed load. While an intraformational sediment fill is a characteristic feature of some of the minor channels in the Etruria facies rocks, most of them also contain polymictic, extrabasinal material. In view of the comparative rarity of sand bodies in the Etruria Formation sequence as a whole (see Figs 105-114 and appendix 1) it is unlikely that this could have resulted from intraformational reworking. Furthermore, the type B deposits of Allen and Williams (op. cit) were developed on a stable surface undergoing pedogenesis, and generally overlie palaeosols. The minor channels in the Etruria Formation are always associated with Facies 9 silty mudstone units which mark a reversion to active depositional conditions following a period of pedogenesis, rather than deposits of a system developed during a period of non deposition and soil formation.

ii) Spillway channels. These are a distinctive overbank channel form, described from the Lower Indus river by Holmes and Western (1969). In this river most overbank flow during flood periods occurs by overtopping, creating continuous and extensive bank levêes. Locally, where well developed levêes have already been formed, some of the flood escapes down small overflow channels, termed spillways, which act as feeders, carrying such coarse material as is in suspension in the flood down the levêe slopes and into the flood basin, where it is deposited either in the fill of these channels, or as small levêes associated with them. These flow events are distinct from crevasse events, in that the channels originate at the crest of the levêe, rather than a breach being formed. Coarse material from the bedload of the main channel should thus be excluded from the deposits of such channels and their associated sheet sands.

iii) Crevasse channels. Prolonged re-use of the same crevasse breach in a levêe can lead to the formation of a channel which feeds flood discharge out into the floodplain, and which cuts through associated levêe and crevasse splay sediments. The development of crevasse channels, with associated minor deltaic lobes, is best modelled by Elliot (1974) for interdistributary bay environments in deltas. A similar arrangement can be postulated for alluvial environments. This interpretation has already been advanced for the association of a small channel with a packet of thin sheet sands at Knutton (see 3.2.3).

iv) Tie channels and overflow channels. These terms were introduced by Blake and Ollier (1971) to describe minor channels in the overbank areas of the Fly River alluvial plain, Papua New Guinea. Overflow channels link lakes to the main channel system, the flow being unidirectional; tie channels link lakes and floodplain areas to the main channel, the flow being in either direction, depending on the relative levels of water. Overflow and tie channels are found in association with active meander belts which have poorly developed levées. They are generally less than 5m wide, and are periodically choked by weed, which might encourage filling by fine grained sediment.

The minor channel Facies is of such limited occurrence that it is not possible to interpret it unequivocally. Where it is obviously associated with crevasse splay deposits, it is reasonable to interpret it as a crevasse channel facies. Where small channels are present without associated sheet sands, two interpretations are possible. They may represent tie channels (*sensu* Blake and Ollier i.e. linking a lake to a river channel). Alternatively they may represent more distal crevasse/flood channels, which fed flood waters into the overbank area, but which were too distant from the channel for bedload material to reach them in significant quantities.

The two alternative interpretations suggested are thought unlikely. Interpretation as the channels of a secondary drainage is improbable as the channels occur in the wrong position in the succession, and contain too much extraformational material. Interpretation as spillway channels is rendered improbable by the frequent occurrence of coarse and conglomeratic material.

4.3 Palaeosols in Facies Association IIA

4.3.1 Introduction

Two distinct types of palaeosol are present in this Facies Association, together with common occurrences of intensely colour mottled mudstone, which are probably of pedogenic origin, but within which identifiable soil profile types are not recognisable.

4.3.2 Palaeosol type 3 : post depositionally oxidized alluvial soils

These palaeosols consist of deep red, purple and grey mudstone, with an extremely strong development of the listric texture characteristic of Type 1 palaeosols (alluvial soils). The grey colouration is usually found at the top of the profile, which is infrequently thicker than 50cm (Fig 41). In some cases the grey upper horizon contains an appreciable quantity of disseminated carbonaceous material, and in a few cases (eq. at 9.00m in the section at Wilnecote Quarry) a coal seam of up to 10cm thickness is preserved at the top of the palaeosol. In cases where a significant amount of carbonaceous material is present a symmetrical zone of 10-20cm above and below the carbonaceous horizon is unoxidized or only partially oxidized, and may contain carbonaceous roots and plant debris.

Other features of this Palaeosol Type are the presence of haematite filled concretions of Stigmaria and other root systems, and of partially or completely oxidized siderite concretions.

The root concretions occur as distinctive, friable ochre pigmented concretions, which sometimes contain enough haematite to be identifiable by XRD, and at other times consist of soft clayey material which is probably goethite. Stigmarian systems up to 2m long have been excavated at Manor Quarry, near the base of section 1B, and, when unweathered, such concretions may faithfully preserve the morphology of the root system (Fig 42).

Siderite concretions of two types have been observed. Ellipsoidal, mamillated concretions (shape terms sensu Brewer 1964) between 5 and 10cm in size occur in a red palaeosol at the base of Wilnecote Quarry. They consist of coarsely crystalline siderite, identified by XRD. Their outer skins are oxidized to haematite, and they are surrounded by red mudstone. Sphaerosiderite occurs in red horizons of palaeosols in Wilnecote Quarry and in the Playground No. 8 core. Spherulitic haematite, almost certainly formed by the oxidation of sphaerosiderite, is also present in the Playground core.

Little profile development is present in this type of palaeosol. A typical section is that at 1.95m in section 1B at Manor Quarry (Fig 43):

m	
0-0.05	Bedded haematite (Facies 12: see 4.4.1)
0.03	Dark grey, slightly carbonaceous mudstone
0.35	Mudstone, mainly deep red, with abundant listric surfaces and some 2cm sphaeroidal, soft nodules of ochre pigmented clay and ?haematite.

Interpretation

These palaeosols are interpreted as alluvial soils (Palaeosol Type 1) which have subsequently been oxidized during early burial. This conclusion is indicated by four main lines of evidence:

i) The abundance of listric surfaces is reminiscent of palaeosols of Type 1, and suggests that the sediment was organic rich during early compaction.

ii) The presence of Stigmaria implies that the soils formed in permanently waterlogged conditions (Scott 1979).

iii) The presence of siderite or oxidized relicts of siderite suggest concretion formation in an organic rich, reducing chemical environment, either during soil formation or during early burial. Where post depositionally oxidized ironstones (Facies 12) overlie the palaeosols, these also indicate an initial reducing, organic-rich depositional environment (see 3.4.2 and 3.4.4)

iv) The pigments present, especially the purple and grey colours, are reminiscent of those seen in coal bearing sequences affected by post depositional oxidation below the Permian land surface, described by Mykura (1960). In progressively more intense stages of alteration, shales and mudstones change from grey to deep red through an intermediate purple stage. Ironstones become brown and then red, and carbonaceous mudstones and coals retain their dark colour, while losing most or all of their organic content.

By this analogy, the tendency for dark coloured listric mudstones to occur at the top of Type 3 palaeosols presumably reflects the presence, before oxidation, of a carbonaceous surface horizon, or of a coal. This is also suggested by Malkin (1961 p.56) who, writing of the Etruria Formation, suggests:

"it is possible that the thin coals now seen actually represent much thicker seams which have not been completely destroyed by reddening, and have protected the surrounding strata from its effects".

Such a concept of postdepositional destruction of carbonaceous horizons is supported by the symmetrical, non oxidized zones observed in the present study above and below carbonaceous horizons in red claystone palaeosols.

Sedimentological implications:

The occurrence of post depositionally oxidized alluvial soils and, by inference, coals (organic soils) implies that the overbank surface must at times have been poorly drained, and that at these times low clastic input allowed the formation of alluvial and organic soils in floodplain swamps. The sequence of sediments thus formed were affected by penetrative oxidation soon after burial, before permeabilities in the mudstones were reduced by compaction to prevent circulation of oxidizing groundwater.

The sedimentological relationships are discussed further in 6.1, and the post-depositional oxidation in chapter 8.

4.3.3 Palaeosol type 4 : Highly evolved palaeosols

The most prominent feature of this type of palaeosol is the pronounced colour banding that it presents in exposures (Fig 44). In cases where a single phase of soil development has apparently occurred, and there has been no erosion of the upper part of the palaeosol profile, three distinct colour horizons are visible (Fig 45);

i) an upper horizon of pale grey mudstone;

ii) an intermediate horizon, generally of varicoloured mudstone with abundant concretions;

iii) a lower horizon, forming a gradational transition into the parent material.

Apart from the upper boundary of the palaeosol, which is usually the slightly erosive base of the overlying unit of Facies 8 or 9, all of the boundaries between these horizons are very diffuse and gradational. Recognition of these horizons becomes difficult when soil profiles formed by several phases of soil development separated by deposition are superimposed on one another, or when the upper part of the palaeosol profile has been eroded prior to the resumption of deposition.

The upper horizon consists of massive, structureless pale grey or pinkish grey mudstone (Fig 46). The colour is not usually uniform, there being abundant diffusely bounded patches with a very pale red or purple pigment, up to a few cm in size. Where the underlying parent material is silty, this upper palaeosol horizon is lacking in silt. In many cases this horizon is slickensided, and in combination with the depletion in silt content, this causes the horizon to weather more rapidly than the surrounding mudstones.

The most abundant and distinctive features of this palaeosol type, other than the exposure-scale colour banding, occur in the intermediate horizon. There is a considerable diversity of features, which are better or less well developed in different examples. The following description is thus, to a degree, a generalization.

The intermediate horizon is most obviously characterised by an abundance of colour mottling (Fig 47). The mottling patterns are usually without any obvious organisation, but locally a network of pale grey 'pipes' or 'channels', 1 to 5cm wide, surrounds areas of purple, ochre, and red pigment. The mudstone forming these channels is, from its colour, presumably depleted in FeIII. The purple pigmented areas (not seen in Fig 47) represent zones in which FeIII oxides are relatively depleted, the purple colour resulting from the very fine grain size and disseminated nature of the remaining haematite. This pigment is also commonly present in the interconnected 'pipes' or 'channels', and in some cases occupies vertical 'channel' or fissure features up to 50cm deep, which taper downwards (Fig 48). (These features, in common with all of the mottling features in the intermediate horizon, are distinguished only by the type of pigment present, the mudstone texture remaining the same throughout). In the yellow pigmented areas the FeIII oxides are mainly hydrated (goethite). The red patches are zones in which the properties of the parent material have remained relatively unaltered, and in which net accumulation of haematite has occurred, the latter indicated by a deeper and/or brighter red colour than the typical dark reddish brown of the mudstone facies. The areas of net haematite enrichment form distinct specks, representing incipient diffuse concretions (Fig 49).

Root traces are abundant in this horizon. They are usually filled with a yellow goethite pigment, and occasionally have a concretionary fill (Fig 50). The roots are always sub-vertical, mainly straight, with occasional branching. Such roots have a diameter of ca 1cm, and a vertical extent of up to 50cm. Stigmarian types have not been found, except where polyphase soil development has occurred (see 4.3.5).

Apart from the diffuse haematite segregations already described, nodular haematite concretions of ellipsoidal and mamillated shape (sensu Brewer 1964) are occasionally present (Fig 51). They consist of pure, extremely well crystallized haematite (determined from XRD), and often contain segregations of white, pure kaolinite.

The lower horizon forms a gradation into the parent material. It consists mainly of red mudstone, heavily veined with purple pigment and traversed by 1 to 2mm wide 'pipes' and veins of white material. Both white and purple pigmented features die out downards. Root traces, if recognisable, consist of bleached root scars without any concretionary fill (Fig 52).

In examples of this palaeosol type which are apparently more mature, a number of modifications are apparent in the intermediate horizon, which in such cases can be subdivided into two or more horizons. Commonly, the upper part of the horizon contains the features described above. The lower part of the horizon is characterised by the development of diffuse haematite concretions with irregular outer boundaries (Figs 53,54). These consist of 1 to 10cm aggregates of smaller haematite concretions, typically 1 to 2mm in size, which, in the field, weather slightly proud of the surrounding mudstone. In thin section, the concretionary haematite occasionally shows a crude pisolitic structure. The small concretions are usually sharp edged, but sometimes show a diffuse boundary, passing laterally into yellowish red or purple mudstone, containing disseminated very small (less than 10) specks of haematite pigment. Segregations of haematite also occur in a fibrous habit, which probably represent haematised plant material. In all cases the concretionary haematite encloses quartz grains, indicating growth by replacement rather than displacement.

These concretions are generally intricately veined with pale grey clay, and surrounded by a halo of pale grey material (Fig 55). The aggregates are separated one from another by sub-vertical channels 1 to 2cm wide, which, from their size and, in horizontal section, their cylindrical shape were probably root channels (Figs 53,56). Occasional goethite filled root channels are present. In the upper part of horizons containing these concretionary aggregates, the concretionary material also occasionally consists of goethite, producing a lithology of extremely unusual lithology and texture (Fig 57).

In most occurrences of this concretion type, the concretions are disseminated, forming 10% to 30% of the rock. In two instances, however, at Chesterton and Manor Quarries there is, within horizons containing such concretions, a gradational upward increase in the proportion of concretionary masses present, over a thickness of ca. 1.50m. In the Manor examples the concretionary patches also increased in size upwards, from ca. 4 x 4cm. at the base of the horizon to ca 10 x 20cm x 0.80m above the base. In both examples, the concretionary masses coalesced at the top of the horizon to form a semi-continous layer, with its top occurring at a clearly defined level in the palaeosol profile.

The thickness of horizons and profiles varies considerably. The upper horizon ranges from 0.20m to a maximum of 1.00m. Much of this variation depends on the degree to which the horizon has been eroded. It is occasionally absent owing to erosion. The interediate horizon varies from ca 0.50m to ?ca 2.00m in thickness. The maximum thickness of thicker intermediate horizons is difficult to judge, as these horizons, when thickly developed, often show signs of having formed by the stacking of several successive soil profiles. Typical sections through single and stacked palaeosol profiles are shown in Figs 58-60.

Stacked soil profiles are easily recognised by the repetition of one distinctive horizon, usually that containing the distinctive diffuse aggregates of haematite concretions (Fig 59). In very thick sequences of superimposed maturely developed palaeosols there is often a horizon of massive ochre coloured mudstone, without concretions, occuring in the centre of the profile, between the horizons of disseminated concretions belonging to the successive palaeosol profiles (Fig 60).

Geochemistry

Geochemical data on these palaeosol profiles are sparse. Partial analyses have been carried out on one profile at Rosemary Quarry, Cannock (G. Lees personal communication), one profile at Utopia Quarry (P. Turner personal communication), and on a number of profiles at

Bradwell (= Bentley quarry: SJ848505; Fig. 3) and Goldendale Quarries (Holdridge 1959). (The latter author did not specifically recognise his analysed profiles as containing palaeosols, but the descriptions of 'grey' and 'mingled' horizons seem unambiguous). In addition, a profile from Manor Quarry has been analysed in detail by Besly and Turner (in press).

Two distinct patterns are evident, the main variation being in the relationship of iron and silica content (Fig 61). In one group of profiles (Fig 61A) there is a marked upward decrease in iron and silica, and a concomitant increase in alumina. Profiles of this type are those at Rosemary Quarry (Fig 62), Utopia Quarry, the lower part of the stacked profile at Manor Quarry (Fig 63) and the profiles in Bradwell Quarry. The Manor quarry profile contains concretionary haematite, which gives a high anomalous iron content to the intermediate horizon. In the second group of profiles (Fig 61B) there is an upward decrease in iron, accompanied by an upward enrichment in silica. This pattern is shown by the profiles in Goldendale Quarry, and by the upper part of the stacked profile in Manor Quarry (Fig 63). In the latter case, the upper of the stacked profiles shows less advanced pedogenetic features than the lower one, containing only slight ferruginous mottling and no haematite concretions in its intermediate horizon.

Interpretation

The interpretation and reconstruction of this type of palaeosol presents great problems, in that a wide variety of soil types can give rise to features similar to those described above. The presence of abundant segregations and concretions of haematite is, however, characteristic of tropical and subtropical soils, and modern analogues will be restricted to soil types developed in these areas.

There are immediate similarities apparent between the palaeosols of this type in the Etruria Formation, and palaeosols described from the Reading Beds by Buurman (1980), and from the Upper Cretaceous to Lower Eocene of Languedoc by Freytet (1971). The most notable points of similarity are the dominance of red colouration, the presence of ferruginous mottles, and, in the case of the Languedoc palaeosols, of disseminated haematite concretions (present in Freytet's "Palaeosol type 1 - non calcimorphe", p.264). Both authors interpret the palaeosols that they describe as hydromorphic alluvial soils, showing colour mottling due to gley or pseudogley development.

Although a similar interpretation is possible for some of the Etruria palaeosols (see 3.3.2), it is felt to be unlikely in the case of the well developed palaeosol profiles for two reasons. Firstly, the palaeosols described by Buurman and Freytet seem to lack the pronounced development of colour and chemical profiles present in the Etruria examples. Secondly, and more significantly, the palaeosols described by both authors occur in thick stacked sequences, making up almost the entire sequence of sediments under discussion. The red colouration in the Reading Beds is produced by pseudogleying, and, as this is produced at, or just above the ground water table, the underlying parent sediment below the water table is dark grey in colour and organic rich. A thick red bed sequence is only produced by pervasive pseudogley formation during sediment accumulation, resulting in thickly stacked palaeosol profiles. In the Etruria Formation, however, single palaeosol profiles, or profiles containing two or a maximum of three stacked palaeosols occur at intervals in the sequence, separated by mudstone and other Facies which show no sign of having been deposited as organic rich sediments (see 4.2.1, 4.2.2, 7.1, and 8.3).

The characteristic features at the Type 4 Palaeosols in the Etruria Formation are:

- the presence of a strongly leached profile, showing consistent depletion of iron and _either_ depletion _or_ concentration of silica in the upper horizon (which is presumably an A horizon), and concentration of iron in the intermediate (?B) horizon.
- the presence of mottling, and patchy segregations of haematite in the intermediate horizon.
- the presence of nodular and diffuse haematite concretions in the intermediate horizon.

The two main tropical soil types that contain such features are laterites _sensu lato_ (e.g. Young 1976) and some tropical podzols. The latter, known as Ultisols in the U.S. 7th Approximation soil classification (Soil Survey Staff 1960), are not specifically recognised in other classifications, and are often described under the general term 'ground water laterites', the difference between lateritic and podzolic features not being specified. Ultisols are best described by Mohr _et al_ (1972).

There are two main 'lateritic' features in the Etruria Type 4 palaeosols. The discrete nodules, and especially, diffuse concretions of haematite occuring in the intermediate horizon strongly resemble the nodular variety of lateritic ironstone defined by Pullan (1967 and personal communication). The examples that he cites, from Northern

Nigeria, are formed by progressive accretion of ferric oxides in areas which are initially mottled. A similar relationship exists between mottled and concretionary patches in Etruria palaeosols. The concretions described by Pullan also frequently enclose quartz grains and other particles. The formation, in some cases of more or less continuous concretion layers is reminiscent of indurated lateric ironstone layers described by Mohr et al (1972).

Secondly, the upward depletion of both iron and silica in the upper part of some of the palaeosol profiles suggests that lateritic weathering processes were involved in the formation of the soils (cf. analyses in Mohr et al op cit). Lateritic processes were suggested by Holdridge (1959) to account for the chemical variation he observed at Bradwell Quarry, although he did not recognise the presence of palaeosols.

'Podzolic' features of these highly evolved palaeosols are as follows. The analytical results of the upper part of the Manor profile (Fig 63), and those from Goldendale Quarry are in a trend similar to that described from an Ultisol (Fig 64) by Slager and Van Schuylenborgh (1970). There is also a striking resemblance between the appearance of this soil, the Kennedy Highway profile (illustrated in colour by Mohr et al, 1972 p.271) and some Etruria Formation palaeosols. The Kennedy Highway profile, developed on Pleistocene clays in the coastal plain of Surinam, is characterised by a prominent bleached upper (A) horizon, and by the development of mottles and nodules in the intermediate (B) horizon. The superficial resemblance is strengthened by the association of this soil with a series of gleyed alluvial soils in the less well drained coastal plain areas, which may be broadly comparable with palaeosols of Type 2 (see 3.3.2).

From the above description it is apparent that palaeosols of Type 4 showing 'podzolic' and 'lateritic' features can only be distinguished by chemical profile analysis. The 'podzolic' type may show less obvious development of haematite concretions, and a slightly less thick and mature profile, but, by and large, the field appearance of the two types is not dissimilar. It is felt here that, in an ancient sedimentary sequence, it is not practical, or useful, to separate these types. For this reason, this highly evolved palaeosol type is interpreted to be a 'latosol', a vague term favoured by Young (1976) to describe all soils with lateritic tendencies. Descriptions of latosols developed on recent or Pleistocene alluvium are, in any case, rare, and so close analogies with modern soils are not possible. However it appears from sketchy descriptions by Bleeker (in Paijmans et al 1971) that there is very little difference between lateritic and podzolic tropical soils in their early stages of development on such materials. The difference between the two, in terms of process, is that in the podzolic soil silica is immobile in the upper horizon, while in the lateritic soil it is leached together with the iron. The relationship between the two is controlled by the intensity and duration of weathering. In the soils described by Bleeker (op. cit). it may be inferred that there is an evolutionary trend from podzolic to lateritic soils. Such a relationship is also suggested by Mohr et al (1972) in the formation of lateritic duricrusts, and may have existed between 'podzolic' and 'lateritic' palaeosols in the Etruria Formation.

Further consideration of this 'latosol' palaeosol type is hampered by the lack of modern analogues. Profile descriptions of recent laterites (e.g. Hallsworth and Costin 1953; Mohr et al 1972) concentrate on mature upland profiles forming on bedrock parent materials. Very few profiles developed on fine grained sedimentary parent materials have been described. Descriptions of upland profiles are, in any case, unlikely to be pertinent to the interpretation of soils formed on depositional alluvial landforms. Upland laterite profiles form in well drained, geomorphologically stable areas. The behaviour of the groundwater table in such an area is largely controlled by progressive erosional downcutting of the drainage system during laterite formation, causing a progressive drop in groundwater level (Mohr et al 1972). Many characteristic features of mature laterite soils, (the indurated horizon, pallid zone, etc....) arise directly from such a progressive water table drop (Mohr et al : op cit). In contrast, water table behaviour in a depositional setting, while possibly affected in the short term by climatically controlled fluctuations and drainage system incision, is dominantly controlled by depositional topography and tectonic subsidence. As subsidence proceeds there will always be a tendency for drainage conditions to undergo a secular change towards waterlogging, prior to the resumption of sedimentation. This is the opposite of the process inferred by Mohr et al for the production of mature, indurated laterites, and it is thus unlikely that such soil profiles will be produced in an active depositional setting.

Amount of time needed for formation

In the absence of a good recent analogue, it is difficult to judge the period that was necessary for the formation of a palaeosol of this type. Some of the less evolved profiles may be compared with the podzol described by Retallack (1977 : Long Reef Palaeosol Series), for which he estimated a period of formation of ca. 2000 years. The more mature profiles, containing well developed horizons containing haematite concretions, probably took considerably longer to form, perhaps in the region of 5000-6000 years.

Sedimentological implication:

The presence of palaeosols of type 4 in this Facies Association implies that parts of the alluvial surface were well drained, for considerable periods and at regular intervals, during the deposition of the Etruria Formation alluvium. The relationship of these areas to areas in which alluvial soils of type 3 formed is considered in 6.1.4. The behaviour of the water table, and the controls on drainage, during deposition of the Etruria Formation are discussed in Chapter 8.

4.3.4 Palaeosol type 5 : Palaeosols containing calcareous concretions

In two exposures, at Springfield South Quarry and at Bradwell Quarry, palaeosol profiles are present which, superficially, resemble rather poorly developed type 4 palaeosols, except that they contain concretions of calcite in their intermediate horizons. These concretions are spheroidal or ellipsoidal, and range from 2cm to 10cm in maximum diameter. The palaeosol at Bradwell forms the lower part of a ?polyphase soil profile, being overlain by a strongly carbonaceous mudstone.

Interpretation:

Although calcium carbonate concretions form in a wide variety of soil types, they are not common feature of the alluvial soils, gleys, and podzolic/latosol soils which are the predominant palaeosol types in the Etruria Formation. The palaeosol type may be compared to the "Paléosol type 2 - peu calcimorphe" of Freytet (1971), who regarded this type as a moderately evolved alluvial soil, formed under conditions of deep water table with little or no gleying. In the Campanian of Saint-Chinian area, palaeosols of this type occur in association with alluvial palaeosols containing no carbonates. Freytet gives no indication of any possible climatic or parent material control on the development of these two palaeosol types. It is possible, in the Etruria Formation examples, that they reflect periods of slight aridity, in which evaporative loss was sufficient to concentrate carbonate minerals in the soil profiles by the 'per ascensum' mechanism of Goudie (1973). Alternatively, during a period of low rainfall, calcium carbonate introduced into the soil by leaf fall or from aeolian dust may not have been removed by the strong leaching which characterised the formation of soils in well drained areas during most of the deposition of the Etruria Formation.

4.3.5 Polyphase soils:

In some instances, palaeosols of type 4 lack a typical leached upper horizon, and in its place have a horizon of claystone containing abundant listric surfaces, which is either dark red or grey to black in colour. In the latter case, the horizon may be carbonaceous. A typical profile is that from Knutton Quarry (Fig 65). In this, and several other examples, the listric mudstone horizon is overlain by a postdepositionally oxidized bedded haematite ironstone (Facies 12 - see 4.4.1).

Interpretation:

Such a listric mudstone occurring at the top of an evolved latosol type palaeosol represents a partly or completely postdepositionally oxidized alluvial soil (palaeosol type 3) developed on, and thus modifying the upper horizon of, a pre-existing more evolved palaeosol profile.

Such a composite soil profile would have formed during the following sequence of events, summarized in Fig 66. During a period of consistently good drainage a latosol soil profile formed. At this time the site of soil formation was probably remote from active alluvial channels and areas of deposition, and may have been topographically elevated, either as a result of pre-existing alluvial topography, or owing to incision due to tectonic activity or to some other external base level change (see 8.5). As the site of soil formation subsided, so it became increasingly waterlogged, and some deposition may have occurred. On this surface, previously the surface of the latosol, a poorly drained alluvial soil formed. Following renewed deposition, this upper, alluvial soil profile may or not have been post depositionally oxidized.

This development of two contrasting soil types on the same site, the latter overprinting the features of the former, is typical of polyphase soil formation. As the latosol palaeosol type tends to occur in thicker profiles than the alluvial palaeosol type, the lower parts of the latosol tend to survive unmodified.

4.3.6 Palaeosols of non-identifiable types

In addition to palaeosols whose mode of formation may be inferred, there are a number of palaeosols which do not suggest an obvious interpretation.

The first group consists of thick (up to 15m) sequences in which all of the sediment is veined, slickensided, sometimes intensely mottled, and contains root traces and concretions. An example, from Playground No. 8 borehole, is illustrated in Fig 67. While such sequences contain features similar to those found in palaeosols with a recognisable profile, a complex superimposition of profiles has given rise, in this case, to a succession in which no single profile can be recognised. A similar problem was encountered by Buurman (1980) and Freytet (1971). The latter referred to such succession as "complex stacked palaeosols".

The second group of non-identifiable palaeosols are very thin and, presumably, poorly developed. They occur within completely red mudstone sequences throughout the Etruria Formation. A typical profile is at 7.90m in the section at Springfield North Quarry (Fig 68).

m	
0.25	Light grey, non carbonaceous, mudstone
0.30	Mudstone, mainly khaki colour, with a few grey root traces

Parent material : Red mudstone

The distinctive palaeosol features are the extremely sharp top and gradational base, and the irregular boundary between the grey and brown horizons. Pipes of grey material extend irregularly downwards into the underlying brown horizon by up to 10cm. These may represent root traces. They are not seen in all exaples of this palaeosol type.

The original soil type represented by these palaeosols is not known, although the development of a thin grey upper horizon with a leached appearance suggests that such palaeosols may be immature precursors of the leached 'latosol' type palaeosols (palaeosol type 4).

4.3.7 Summary of sedimentological and hydrological implications of palaeosols in Facies Association IIA

In terms of the sequence of depositional events which contributed to the accumulation of a sequence of alluvial sediments, a palaeosol, as opposed to a rooted mudstone (e.g. in Facies 8 and 9), represents a period of non deposition and subaerial exposure of sufficient duration for a soil profile to develop, with some recognisable soil horizon development.

Depending on the type of soil, the time involved may have been between a few hundred years or less (for the formation of gleyed podzolic and alluvial soils : cf Retallack 1977), a few thousand years (for the formation of mature podzolic soils : cf. Retallack <u>op. cit.),</u> to ten to twenty thousand years (for the formation of mature caliche soils : Allen 1974).

The pedogenetic reconstruction of the palaeosols also allows reconstruction of the groundwater behaviour in the alluvium both during and after deposition. This in turn casts light on the climate, and on the floodplain topography.

The two palaeosol types in the Etruria Formation reflect distinct patterns of groundwater behaviour in the floodplain sediments during periods of non deposition. The evolved palaeosols (type 4) formed during periods, probably of between two and ten thousand years, during which no deposition took place and well drained conditions prevailed. (The figures used here are arbitrary : two thousand years is the time assumed for the evolution of a podzolic evolved soil, for instance the upper part of the profile analysed at Manor Quarry (cf. Retallack 1977); ten thousand years is an arbitrary figure taken for the evolution of a latosol type evolved soil, as no descriptions of modern analogues are available). If, as is suggested in chapter 8, the post depositional oxidation of ironstones and organic rich horizons took place during periods of lateritic soil development on the alluvial surface, the water table must have been at a considerable depth during such phases of pedogenesis, possibly approaching the depths of 18m recorded by Rawitscher (1948) below lateritic soils in the Campos Cerrados savanna in southern Brazil.

Two interpretations of such a drop in the groundwater level are possible. Firstly, evaporative water loss may have caused a water table drop, recharge being less than loss as a result of avulsion of the main fluvial channel system to a distant part of the alluvial plain, or as a result of a strongly seasonal climate (Rzóska 1974). Secondly, base level change may have caused temporary incision of the alluvial system, either as a result of tectonic uplift and/or folding (e.g. Paijmans et. al. 1971), or owing to processes inherent in the behaviour of the river system itself. These possibilities are further discussed in 8.5.

The alluvial palaeosols (type 3), by contrast, formed during periods of non-deposition during which waterlogged conditions prevailed, at least for part of the time. This suggests that, as soon as sediment supply ceased, continued subsidence created swampy zones in which had previously been a well drained area. The period during which pedogenesis could take place on such a site was limited, as its low lying nature would make it a preferential site for renewed deposition.

Where there is an alluvial palaeosol phase overlying an evolved palaeosol, this indicates a period of alluvial soil development in swamps which formed as a result of a rise in water table in the previously well drained area. This could have resulted from subsidence, from infilling of valleys incised into the alluvium, or from changes in surface topography on the alluvial plain as a result of continuing deposition.

The possible evolutionary trends of palaeosol development are summarised in Fig 69.

4.4 Minor Facies in Facies Association IIA

4.4.1 Facies 12 : Bedded haematite rocks.

Examples of this facies have only been found in North Staffordshire, at Manor, Chesterton, Knutton, and Rosemary Hill quarries, and in the Silverdale New Mine (Fig. 3) access drifts.

The rock consists of massive, bedded haematite in isolated individual layers, usually not exceeding 6cm in thickness (Fig 70). No internal structure has been seen, although in one sample from Manor quarry some faint laminar structures are visible, which may be traces of bivalve shells. In some instances the layers of the haematite rock are only 1-2cm thick, and are discontinuous when traced laterally.

This Facies always overlies a palaeosol of type 3 (post depositionally oxidized seat-earth : see 4.3.2) or type 4 (latosol : see 4.3.3, 4.3.5). In almost all cases it rests on the thin, heavily listricated, usually black and carbonaceous mudstone which represents the former peat surface horizon of these soils. At Knutton lenses of haematite occur within this carbonaceous layer.

One sample, taken from a horizon ca. 50m above the base of the Formation in the Silverdale New Mine access drift, presented a mm-scale speckled appearance in hand specimen. This sample proved, in thin section and XRD, to be composed of ca. 50% siderite and 50% haematite. No residual depositional structure was visible. This horizon probably correlates with the Blackband ironstone occuring in a thin grey bed intercalation in the Etruria Formation in the Chesterton area, recorded by Gibson (1901).

Interpretation:

In view of the rarity of haematite forming the primary mineral phase in Phanerozoic bedded sediments, this Facies must have been formed as the result of a post-depositional process, either altering a pre-existing iron rich phase, or concentrating iron into a haematite rich horizon.

The occurrence of the bedded haematite rock overlying a palaeosol, or the remains of the carbonaceous or coal surface horizon of a post depositionally oxidized, type 3 palaeosol, invites comparison with the usual occurrence of the Blackband ironstone facies (Association I, Facies 6) overlying a coal seam. The comparison is strengthened when the vertical sequences in which both iron rich facies are found are compared. Both overlie palaeosols which had similar soil characteristics during the last phase of pedogenetic evolution before deposition was resumed (swamp/alluvial soil in the case of the Blackband facies; post depositionally oxidized swamp/alluvial soil or swamp phase of polyphase highly evolved soil in the case of the bedded haematite facies). Both are overlain by overbank sediments, in which, usually, rooting becomes more intense upwards. The major difference between the two is that, in the red bed association, the entire sequence has been affected by early post depositional oxidation.

The alternative mechanism by which such a haematite-rich rock might have been formed would have involved post-depositional remobilisation and concentration of ferric iron. In an alluvial environment this could be achieved by a process of lateritization. However, the lack of palaeosol features associated with this facies (other than the palaeosol which often underlies it), and the lack of resemblance between the facies and the haematite rich horizons observed in unquestionable palaeosols (see Figs 53-57), make this interpretation unlikely.

The bedded haematite facies thus represents post depositionally altered bedded siderite ironstone of Facies 6. The depositional environment is discussed in 3.4.4.

4.4.2 Entomostracan ("Spirorbis") limestones (Facies 5)

As mentioned in 3.4.1 limestones comparable to those found in Facies Association I occur throughout the red beds of Facies Association II. Although no unequivocal examples have been observed during the present study, such limestones have been recorded from the Etruria Formation by Gibson (1905: - North Staffordshire), by Wills (1950; 1956), and by

Wedd et al (1928:- equivalent facies in Ruabon Formation, North Wales). As recently as the mid 1960's a limestone of this type was exposed in a now overgrown quarry in the Etruria Formation, to the north of the Metallic Tileries quarry (A.A. Wilson, personal communication).

There appears to be no significant difference between the limestones in the Etruria Formation from those seen in the underlying Blackband Formation (Gibson 1905 p.334), and so the observations and inferences made in 3.4 apply to limestones in Facies Association II as well.

An interesting occurrence of lenticular limestones has been recorded at the boundary between the Etruria and Newcastle Formations in the Metallic Tileries Quarry (Pollard and Wiseman 1971). Although containing a lithology and fauna typical of other "Spirorbis" limestones, these nodules are characterised by the occurrence of codiacean and porostromatate algae, both as a detrital component and in life position. Such algae occur in the Lower Carboniferous of southern Scotland and northern England in environments ranging from shallow marine to non marine. It is thus difficult to assess their palaeoenvironmental significance in this instance. Pollard and Wiseman regarded the limestone as having formed in a restricted "pool" environment "transitional from the semi-arid terrestrial conditions of hypersaline lakes of the Etruria Marl to the brackish water lagoon of the lowest Newcastle Beds". As the present author regards both the Etruria and Newcastle Formations as predominantly freshwater alluvial deposits, the interpretation of Pollard and Wiseman requires some modification. The suggestion of Neale (in discussion of Pollard and Wiseman op cit) that the presence of algae may reflect alkalinity or the availability of calcium carbonate in the water was not accepted by

Pollard and Wiseman, who considered that high salinity was the major environmental factor responsible for their presence. However, conditions suggested by Neale could well have occurred in a hard water lake in an alluvial setting, especially in the case of high evaporative water loss. If this were the case, this limestone probably formed in conditions similar to the other "*Spirorbis*" limestones.

4.5 Facies Association IIB Sandstones deposited in meandering alluvial channels

4.5.1 Introduction

Within this Facies Association there are three major lithofacies which together make up composite sand bodies: massive sandstone; interbedded sand/siltstone and mudstone showing inclined accretionary bedding; and red mudstone and siltstone.

As there are only a limited number of exposures the inter-relationship between the three Facies is not fully understood, although they are demonstrably present as lateral equivalents within the same composite sand body. In the following sections the characteristics of the Facies are described, after which their lateral relationships and their interpretation are discussed.

4.5.2 Facies 13: Massive sandstone

The rocks forming this Facies consist of thickly bedded sandstone, in units which do not usually exceed 2m in thickness. The sandstone is medium to coarse grained. Only one laterally extensive exposure of this facies has been observed, at Metallic Tileries quarry (Fig 71). Here, a sand body composed entirely of this Facies fills a shallow channel, which is exposed oblique to the apparent palaeocurrent direction.

Where the bases of units of this Facies are visible they are always erosive, either into floodplain sediments of Facies Association IIA, or into other sediments of Facies Association IIB. In Chesterton Quarry the basal erosion surfaces of massive units in sand bodies 1 and 3 show, respectively, linear groove casts and prod casts.

In fresh exposures little internal structure is visible. In weathered exposures it can be seen that the dominant internal structure is trough cross bedding, in sets of between 10cm and 90cm (Fig 72), with occasional intercalations of ripple cross lamination (Fig 73). The paucity of weathered exposures precludes generalization of the internal sequence of structures and grain size within units of the Facies. However, upward fining seems to be usual, both within entire units and within individual sets of cross bedding. The bases of units usually contain a lag conglomerate of granules and occasional pebbles of extraformational material, and of pebble size intraformational mudclasts. Sporadic wood dêbris is also present.

In sandbody 1 at Chesterton there is an upward decrease in the size of cross bedded sets concomitant with the upward decrease in grain size. In this sand body 10 to 30cm thick tabular cross bedded sets are present towards the top of the massive sandstone unit. At Manor Quarry the major features of the organisation of sand body 2 seem to be series of poorly defined dipping planes extending through the entire thickness of the sand body (Fig 74). These features invite comparison with features of Facies 14 (described in 4.5.3), and suggest that there may be a complete overlap of sand body types between Facies 13 and 14.

4.5.3 Facies 14: Interbedded sand-, silt-, and mudstone showing inclined accretionary surfaces

This Facies consists of interbedded sheets of sandstone and of siltstone and/or silty mudstone. When viewed in the correct orientation, the individual sheets can be seen to possess depositional dips of up to 10°, and have a generally sigmoidal form (Figs 75-77).

It has not been possible to determine an unequivocal palaeocurrent direction for most of the sand bodies of this Facies that have been observed. The presence of trough cross bedding as the most easily visible sedimentary structure in the vertical face exposures of the sigmoidal sandstone sheets makes it extremely difficult to obtain accurate palaeocurrent measurements. In most cases, the sandstones have not been sufficiently weathered for sedimentary structures to be visible. In a few examples palaeocurrent direction has been judged by estimating the orientation of trough axes. In two cases direct palaeocurrent measurement has been possible. Four measurements at Rosemary Hill Quarry and two measurements at Metallic Tileries (Fig 81) have shown palaeocurrents at a high angle to the depositional dip of the sigmoidal units. These results are in accord with estimates derived from trough axis orientation.

These scanty palaeocurrent data, together with the sigmoidal shape of the interbedded sand/mudstone sheets and their resemblance to other described ancient examples (see 4.5.5), have led to the interpretation of sand bodies of this Facies as lateral accretion deposits. Hereafter individual sand/silt sheets are called lateral accretion sheets, and the sand body produced by lateral stacking of these sheets is referred to as a laterally accreted sand body. This may occur as an isolated body, surrounded by fines of Facies Association IIA, but more usually forms part of a composite sand body, which may include units of Facies 13 and 15. This descriptive nomenclature is summarized in Fig. 78.

The least complex exposure of this Facies is in the northern corner of Knutton Quarry (Figs 75-77). The majority of the following description of the Facies is drawn from features seen in this exposure.

i) The laterally accreted sand bodies have erosive bases, cutting either into floodplain sediments of Facies Association IIA, or into other Facies in the alluvial channel association (Figs 75,79,80). In several examples, notably at Metallic Tileries (sand body 2: Figs 79-81) and at Lightwood (SJ921409; Fig 3) (Fig 82) the lateral accretion sheets in the sand body amalgamate near its base to form a thick basal sheet, which resembles the massive sandstones of Facies 13.

ii) The lateral accretion bedded sand body is composed of sigmoidal lateral accretion sheets of sandstone, separated by thinner interbeds of mudstone and siltstone. Near the base of the best exposed lateral accretion bedded sand body at Knutton the sandstone sheets are up to 40cm thick (Fig 83).

iii) The sandstone lateral accretion sheets are composed of granule sized to fine grained sand. They are sharp based and are sometimes observed to fine upwards. In the basal 2m of the Knutton sandbody this gradation is from granule to medium sand; in the upper parts of the lateral accretion sheets the grain sizes involved become finer.

iv) There is a rapid decrease in thickness in each lateral accretion sheet in the most steeply dipping part of the accretionary topography. At Knutton this occurs about 2m above the base of the sandbody (Fig 84). The thinning is accompanied by a decrease in grain size, from medium to fine sand. Towards the top of the lateral accretion bedded sandbody individual sheets consist mainly of mudstone and siltstone, and such sand sheets as are present are very thin. Thus, each lateral accretion sheet contains a lateral sequence of grain size and thickness changes which corresponds to the thinning and fining upward sequence observed in a vertical section through successive lateral accretion sheets in the sand body.

In addition to the general upward thinning and fining observed in individual lateral accretion sheets, groups of sheets sometimes show an anomalous thickening some way above the base of the sandbody, which gives the upper surface of the sets a cusped appearance (Figs 77,84). The overlying sheets thicken and thin to drape the irregularity in the depositional surface thus created.

v) Near the base of the sand body individual sand lateral accretion sheets are usually trough cross bedded in sets of the thickness of the lateral accretion sheet. In higher parts of the lateral accretion sheets ripple cross lamination is present. (Internal structure has only been observed in a few clean and well weathered exposures: cross bedding at Rosemary Hill, Spoutfield (Figs 85,86), Knutton; cross lamination at Rosemary Hill, Lightwood). The orientation of these structures has already been discussed.

vi) Within the lateral accretion bedded sand bodies there are angular non-sequences between successive packets of lateral accretion sheets. These may take two forms: a) slight erosive truncations nearly paralleling the dip of the lateral accretion surfaces, affecting most of the thickness of the lateral accretion bedded sand body (Type A, Fig 87; Lightwood Quarry, Fig 88; Knutton Quarry Fig 75,77); b) subhorizontal erosive features, which truncate several lateral accretion sheets at a particular level in the sand body, and which are draped by subsequent lateral accretion sheets (Type B, Fig 87; Rosemary Hill Quarry, Figs 89,90).

vii) When viewed in exposures parallel to the inferred direction of palaeoflow (Fig 91; Knutton Quarry) the sandstone lateral accretion sheets appear horizontally bedded, and might be mistaken for a thick development of Facies 10 sand sheets. Closer inspection shows, however, that, along the current direction, the lateral accretion sheets show a complex pattern of lenticularity, mutual erosion, and splitting and rejoining. (In the example illustrated the sand sheets can be followed into an exposure normal to the palaeoflow: Fig 75).

vii) Soft sediment deformation involving lateral accretion sheets has been observed at two localities. In sand body 1 at Chesterton Quarry load casts were present beneath one lateral accretion sheet, penetrating 10cm into the underlying siltstone. At Knutton Quarry, one sandstone lateral accretion sheet is slump folded into recumbent folds, loaded into the underlying siltstone, and thickened at the toe end of the folds (Figs 84,92). The direction of slumping appears to be in the direction of maximum depositional dip.

viii) The proportion of sand to fines varies greatly between individual lateral accretion bedded sand bodies. At one extreme the majority of the lateral accretion sheets are composed of sand, and the sand body resembles the massive sandstone facies. At the other extreme virtually no sand may be present, being limited to a thin lag near the base of the laterally accreted sand body. In such circumstances lateral accretion surfaces are often difficult to see.

Much of the Springfield North quarry composite sand body consists of siltstone, in which lateral accretion surfaces are picked out only by slight colour changes (Figs 84,85), and by the presence of some sandstone at the base of the sheets.

This variation in dominant grain size extends to the separate laterally accreted units within the same composite sandbody, adjacent units often showing a marked difference in sand content (e.g. at Knutton Quarry, Fig 76).

4.5.4 Facies 15: Siltstone and mudstone associated with Facies 13 and 14

Apart from the silt- and mudstone interbedded with the sandstone lateral accretion sheets of Facies 14, siltstone and mudstone also occur as distinct units within sand bodies composed predominantly of Facies 13 and 14.

i) Red mudstone and silty mudstone, similar to that in Facies 8 and 9, occurs as the fill of discrete channel forms within the composite sand body at Springfield North Quarry (Figs 95,96), and in the isolated sand body 2 at Metallic Tileries Quarry (Figs 81,97). The mudstone fill in these channels appears to form a plug, in both cases draping lateral accretion sheets, which form the visible margin of the channel form.

ii) In two exposures (Chesterton Quarry sandbody 1, Fig 98; Manor Quarry sandbody 1, Fig 99) the upper parts of composite sand bodies, both massive and laterally accreted, are cut by small channels, filled with red mudstone (in Manor Quarry) or interbedded red mudstone and green sandstone (at Chesterton). Unlike the mud filled channel forms in i) above, which extend to the base of the sand body, the bases of these channels do not cut down further than 0.30m above the erosional base of the composite sand body. In the Chesterton example the channel fill is symmetrical.

4.5.5 Interpretation of Facies 13 and 14

This group of related Facies is interpreted as the deposits of point bars developed in meandering channels. From their ubiquitous association with continental flora and fauna, these channels are regarded as fluvial channels.

i) Massive Facies (Facies 13)

The interpretation of sand bodies composed of this Facies is difficult, owing to the infrequent and poor exposures available. In the sand body 1 at Chesterton Quarry, the massive unit occurs in a composite sand body which, elsewhere in the quarry, contains laterally accreted deposits. This, together with the internal fining upward and upward decrease in the size of sedimentary structures, suggests that the entire body is the deposit of one point bar.

Within this point bar, the trough cross bedded sets were probably formed by the migration of subaqueous dunes on the lower part of the bar surface, with smaller scale structures forming higher on the bar surface (cf. Allen 1970). The lenticular conglomerates and intraformational conglomerates at the bases of the units represent channel lag deposits. The intraclasts would have formed by the erosion of bank material, or possibly by the erosion of clay drapes deposited at low flow stage. Smith (1972) has shown in flume experiments that such mudclasts can only survive limited transport, and must therefore be of local origin.

Why the massive unit at Chesterton, alone of the several point bars represented in this composite sand body, should lack lateral accretion surfaces is not clear. Comparison of this exposure with sand body 2 at Manor Quarry (Fig 74) and with the sand body at Himley Wood Quarry (Fig 93) suggests that there is a continuous spectrum between sand bodies of Facies 13 which lack large scale sigmoidal structures, and sand bodies of Facies 14 in which they are present. Lateral accretion surfaces are only preserved in heterolithic sand bodies.

From this it would appear that the 'massive' sand bodies were deposited by channels in which either : the stage fluctuation responsible for the preservation of lateral accretion surfaces did not occur (see ii) below); or there was a higher than usual content of bed load sediment. The presence of 'massive' units in composite sand bodies consisting otherwise of laterally accreted units suggests that the former alternative was not the case.

It should also be stressed that the observations made of these sand bodies do not give any indication of their relationships in an up- or down-current direction: it is possible that, in either of these directions the massive units may pass laterally into more heterolithic units containing lateral accretion surfaces. If this is the case, the presence of a 'massive' unit may simply reflect the location of the exposure with respect to its radial position in the point bar.

The massive unit (sand body 1, Fig 71) in Metallic Tileries Quarry is clearly of channelised origin, although the nature of channel behaviour is not known.

ii) Interbedded Facies (Facies 14) showing lateral accretion

The preservation of the traces of the instantaneous topography of accretionary banks of meandering channels has long been recognised in the deposits of tidal channels (Van Straaten 1951; Reineck and Singh 1973). Allen (1963) termed this feature "epsilon cross bedding", and subsequently (1965) recognised it as a feature of inferred fluvial point bar deposits. Recently it has come to be regarded as one of the few relatively unequivocal diagnostic features of meandering fluvial coarse member deposits (Collinson 1978; Jackson 1978, and references listed therein), and has been extensively recognised in ancient fluvial sediments, notably in those of the early Tertiary in northern Spain (Puigdefabregas 1973; Nami 1976; Puigdefabregas and Van Vliet 1978; Nijman and Puigdefabregas 1978; Stewart 1981; Friend et al 1981). The rocks of Facies 14 in the Etruria Formation very strongly resemble those illustrated and described by Nami (op. cit) and Puigdefabregas and Van Vliet (op. cit).

The well developed lateral accretionary bedding in the Etruria Formation sand bodies suggests that deposition on the bar surface was sporadic, and that, for most of the time, bed load transport did not take place, deposition if any being only from suspension. Intermittent sand transport and deposition is borne out by the internal structure of some of the lateral accretion sheets observed. The erosive bases, upward fining, and the upward decrease in bedform scale indicate that each sheet is the product of one waning flow event. Bedload transport along the channel probably only occurred at flood stage, and deposition on the bar surface probably took place after these events.

The thinning and reduction in grain size in the upper part of the lateral accretion sheets reflects the decreasing capacity of the flood stage flow to transport coarse sediment in the shallower water flowing over the bar surface. Water levels in the channel must, however, have remained near bankfull for some time after the cessation of bedload transport, in order to deposit the suspension load silt and clay drape over the upper part of the point bar. This, together with diagenetic considerations discussed in 8.2 and 8.3, suggests that there was not a pronounced low stage.

Deposition of individual lateral accretion sheets in one waning-flow episode is comparable to the process inferred by Mossop (1978) for thicker (up to 1m) sheets in the epsilon cross-bedded fluvial sandstones in the Cretaceous McMurray Formation in Alberta. A comparable mechanism is implied by Puigdefabregas and Van Vliet (1978) for epsilon cross-bedding in Eocene fluvial sandstones in Northern Spain. No recent analogue has been satisfactorily described.

The only detailed facies description of recent point bar sediments composed of interbedded mud and sand is by Taylor and Woodyer (1978). In point bars of the Barwon River, Australia, sand lateral accretion sheets have gradational bases and often exhibit reverse grading in their lower part. There are no erosional contacts. Taylor and Woodyer regard these features as reflecting the river hydrograph. A rather slow rise in discharge up to flood peak causes the maximum sediment transport to precede the discharge peak by 14 days. Sand deposition therefore takes place under increasing energy conditions during this period.

Applying the reverse of this analogy, it may be inferred that the discharge in the Etruria rivers fluctuated much more sharply, with a very rapid rise to flood discharge. During this period bed scouring took place, giving rise to the erosive contacts and angular non-sequences at the base of the lateral accretion sheets. Bed load deposition took place during the waning stage. The marked lenticularity of the lateral accretion sheets in the direction parallel to flow (Fig 91) suggests that deposition took place rapidly. Subsequently fall out of fines from suspension draped the newly deposited sheet of bedload material.

Deposition of bedload material under similar circumstances, albeit on a smaller scale, has been described in tidal flat channels by Bridges and Leeder (1976). Their observations provide an insight into the origin of the angular non-sequences within the epsilon cross-bedded units in the Etruria Formation. In the tidal channels exceptional discharge builds up very rapidly during tidal ebb flow and following heavy rain in the hinterland area. Point bars are deposited which consist of epsilon cross-bedded interlaminated mud and silt. Scour of the point bar surface takes place in one of two ways. The lower point bar surface may be scoured by spiral eddies downstream from the flow separation point at unusually high discharge (Fig 100). These scours may be covered during waning flow, and further lateral accretion sheets may develop, giving rise to a sequence (Fig 101) similar to that observed in the example at Lightwood Quarry (Fig 88). The analogy is rendered more convincing by the truncation being limited to the lower part of the cross bedded unit.

Alternatively, during high discharges, the point bar may be cut off by a chute, which subsequently generates a new series of lateral accretion sheets by outward migration (Fig 102). In such a high discharge, the entire convex bank of the channel might be removed, and similarly reconstructed by outward growth of lateral accretion sheets. In both cases these lateral accretion sets may rest on the truncated remains of those formed by the earlier point-bar. Such a process could account for the extensive, near-horizontal scour seen at Rosemary Hill (Figs 89,90), although this sequence is not exactly comparable with that predicted in Bridges' and Leeder's model, the scoured surface occurring some distance above the base of the meander belt deposit.

The model presented by Bridges and Leeder (op. cit.) demonstrates that point bar scour of the two types described above occurs at exceptional discharge in progressively tighter meander bends. In meanders with a high ratio of radius of curvature to flow width simple sigmoidal epsilon cross bedding is formed. As this ratio decreases, scour of the surface is succeeded by chute development with formation of more extensive, near horizontal non-sequences. As meander loops evolve and their radius of curvature decreases, an orderly arrangement of these features in the laterally accreting point bar sediments might be expected. While it is likely that erosive features within epsilon cross bedded sandstones in the Etruria Formation were controlled by meander curvature, the exposures are not large enough for an orderly arrangement of scour types to be detectable.

The model of Bridges and Leeder also requires the generation of a point bar platform in meander loops where active scour of the bar surface is taking place. No evidence for the formation of such a feature has been found in the Etruria sand bodies, with the exception of the exposure at Rosemary Hill, where the subhorizontal sand sheets overlying an erosive surface in the epsilon cross-bedded sand (B-B in Fig 90) may represent such a feature. The cuspate forms observed in some lateral accretion sets at Knutton (Figs 77,84) may also have been generated by very short lived bar platforms, but they do not have scours associated with them.

The nature of the exposures is such that it is difficult to assess the larger scale erosive relationships within individual laterally accreted sand bodies, or within composite sand bodies composed of a number of laterally accreted units. One observation is possible. In the sand body at Knutton Quarry, and in the upper laterally accreted unit at Rosemary Hill Quarry, there are erosional disconformities which separate bundles of lateral accretion sheets which dip in the same direction. (Figs 76 and 90). In both cases the basal fill of the channels which deposited the lateral accretion sheets underlying the erosional disconformity is preserved (marked C on both Figures). This arrangement is very similar to that described by Friend et al (1981; Fig 4.39) from the distal facies of the Eocene Campodarbe Group of Northern Spain. They suggested that such an arrangement might result from channel migration occurring in discrete jumps, leaving the channel shape partially preserved. This behaviour may have reflected deposition in a zone of transition between braided and meandering channel types. It certainly indicates an unstable and rapidly changing channel morphology.

A similar, although not necessarily identical pattern of channel migration is consistent with features of the interpreted fluvial point bar deposits in the Etruria Formation. The poor sorting and sometimes coarse grain size of the sands suggest a fairly high gradient, and the comparative rarity of small erosive features of the Bridge and Leeder (op. cit) type suggests that the meandering reaches did not often attain a very sinuous pattern.

Three interpretations are possible for epsilon cross bedded units which pass downwards into a massive sand body at their base.

i) A similar explanation may be suggested to that of Puigdefabregas and Van Vliet (1978) for the "thicker" (6-10m) sandstone bodies in the Puente de Montanana area. They suggest that point bar deposits that show epsilon cross-bedding down to the base of the sand body were deposited in channels which were dry between phases of discharge. Those which only show epsilon cross-bedding (i.e. discontinuous deposition) in the upper part of the sand bodies were deposited in channels which continued to carry water at a low water level during periods of minimum discharge. Puigdefabegas and Van Vliet specify that deposits of this type form thicker sand bodies in the sequence and were deposited by the larger channels in the alluvial system. There is no comparable relationship between sand body thickness and type of sand fill in the Etruria Formation examples (see 4.5.7). Furthermore it seems unlikely that, at minimum discharge, bed load of coarse sand and granule grade could be transported, or that cross bedding of the size observed could have been formed. No fines or sedimentary structures formed at low stage have been observed in these basal sands, although they are of low preservation potential. Also no indications of emergence (roots, red colouration) have been seen in upper parts of such sandbodies. The preservation of organic matter in the point bar deposits suggests that the point bar sediments were permanently waterlogged, and this suggests that the channels were permanently fairly full of water.

ii) An analogy may be drawn with the Barwon River depositional model (Taylor and Woodyer 1978). The amalgamated cross bedded sands may represent deposits of a permanently active channel forming a 'low' point bar platform, while the overlying units showing inclined accretionary surfaces may have formed on an upper point bar platform where deposition was only active in floods (cf. Taylor and Woodyer Fig. 13). Apart from differences already mentioned between the Barwon sediments and those in the Etruria Formation, this model is not favoured. Other than the possible bar platform sediments observed at Rosemary Hill (Fig 90), no evidence for the existence of bar platforms has been seen, particularly in places where clay channel abandonment plugs are present, which might drape such features.

The same criticism applies to this model as to that of Puigdefabregas and Van Vliet; namely, that the formation of a massive body of sometimes coarse and granule grade sand containing dune scale cross bedding is incompatible with processes acting at low stage flow. (The sand grade sediment in the Barwon River is of very fine to medium grain size only).

iii) The amalgamation of accretionary sigmoidal beds at the base of the sand bodies may simply reflect the fact that, during periods of high discharge, large bedforms were present near the base of the channel. The scouring effect of these would have removed any mud/silt drape deposited since the prevous flood event, and scoured into the underlying sand sheet. In this way sand deposited would become amalgamated with sand sheets deposited during earlier flood events.

4.5.6 Facies 15: siltstone and mudstone associated with meandering alluvial channel sandstone Facies

i) Mud and siltstone filling channel forms which extend to the basal erosion surface of epsilon cross bedded sand bodies (e.g. Figs 95, 96, 97) are interpreted as the fills of meander loops which, following cut off, became ox-bow lakes. Similar clay plugs filling cut off meander loops are described in recent sediments by Fisk (1947), and in ancient sediments by Stewart (1981) and Nijman and Puigdefabregas (1978). It is notable that, in the two latter descriptions, the abandonment fill contains interbedded thin sand layers. These have not been observed in all of the Etruria examples, the interpretation being based entirely on shape and relationship to other point bar lithofacies. It is possible that the fills were homogenised by bioturbation, such a texture now being obscured by the deep red colouration developed during subsequent diagenesis.

In both examples of such clay plugs the plug apparently rests directly on a normal, fully developed point bar surface. This suggests that abandonment was sudden rather than progressive. In the Springfield North example, which forms part of a composite sand body, this implies abandonment by neck cut-off (cf Walker and Cant 1979), or by avulsion of the entire channel system. In the Metallic Tileries example the clay plug is associated with a simple sandbody consisting of a single laterally acreted unit. In this case avulsion must have occurred.

ii) Mud and siltstone filled channels which do not extend down to the basal erosion surface of the fluvial channel sand bodies (Figs 98,99) cannot be interpreted as channel abandonment deposits. Instead they must have been created by a process which gave rise to a channelised topography on the point bar surface. This could have been formed by swales associated with scroll bars, or by chute channels cut across the bar surface. In view of the comparative rarity of these features, the latter interpretation is preferred. Swale deposits illustrated in descriptions of ancient point bar deposits (e.g. Puigdefabregas and Van Vliet 1978) have a comparatively small depth compared with the thickness of the sandbody in which they occur, and have a regular lateral distribution in the top of the sandbody. They also do not erode into the point bar deposits. An isolated clay filled channel cut into epsilon cross bedded point bar deposits is illustrated in Walker and Cant (1979; Fig 6). This feature, similar to such channels seen in the Etruria Formation, is also interpreted as a clay filled chute channel (P.J. McCabe; personal communication).

4.5.7 Channel parameters

The presence of exposures in which complete epsilon cross bedded sets are visible and oriented approximately at 90° to the inferred palaeocurrent allows direct measurement to be made of the size of the channels which deposited the point bar sediments.

The width of the channel may be calculated by the method of Allen (1965), who states that point bars extend for two-thirds of the distance across the channel. Bank full width(W) is thus calculated by determining the average horizontal width of the epsilon cross bedded units in one point bar deposit, and multiplying by 1.5. Some investigators define the bank full depth (D) of a meandering channel as being equal to the thickness of the epsilon cross bedded sandbody that it deposits. However, Etheridge and Schumm (1978) suggest that the depth thus derived is too great, and have suggested, on the basis of experimental studies, that this figure should be multiplied by a constant of 0.65 to obtain a more realistic bank full depth. This correction has been applied here.

Five exposures of point bar sediments in the Etruria Formation are large enough for palaeochannel measurement to be made, demonstrably exposing the full width and thickness of lateral accretion sets at approximately 90° to the palaeocurrent direction. These are at: Knutton Quarry (Fig 75); Metallic Tileries (sand body 2, Fig 81); Rosemary Hill Quarry (north face above access road, lower unit of stacked sand body, Fig 90); Springfield North Quarry (Fig 95); and Himley Wood Quarry (Fig 93).

As the measurements made in the field were not highly accurate, the results can only be taken as order of magnitude estimates. In all five cases similar channel widths (20-30m) and depths (2-4m) are obtained. The results are the same for a sand body consisting largely of sand (Himley Wood) as for sand bodies consisting of a massive basal unit and interbedded upper part (Metallic Tileries sand body II) and for sandbodies consisting wholly of interbedded sand and silt/clay lateral accretion sets. The difference between these types of point bar deposit is therefore not a function of channel size.

Width/depth ratios for the five measured examples vary between 6 and 10. This places these channels within the category of suspended load channels in Schumm's (1968) classification of alluvial channels. This is consistent with the mud dominated nature of the Etruria Formation alluvium as a whole, sands forming only about 15% of the sequence.

Further arbitrary estimates of channel characteristics can be calculated from the width and depth, using the equations of "Method II" of Etheridge and Schumm (1978). Taking an average figure of W = 25m and D = 3m, a sinuosity of 1.99 is obtained, with meanders having a wavelength of ca. 500m. The radius of curvature of the meanders would have been ca.150m. By comparison with other palaeochannels whose dimensions are tabulated by Etheridge and Schumm (1978), the Etruria channels were rather small, the most closely comparable in size being those described from the Jurassic of Yorkshire by Nami (1976). The small channel size is consistent with the comparatively small catchment areas and palaeodrainage lengths inferred from palaeogeographic maps (see chapter 10).

4.5.8 Summary of features of palaeochannels involved in deposition of Facies Association IIB

The meandering channels which deposited the sand bodies in Facies Association IIB were comparatively small ('average' width 25m; 'average' depth 3m), and apparently did not develop a very high degree of sinuosity. As a result point bars were deposited in which a comparatively simple pattern of lateral accretion is observed, an in which localized scouring of the point bar surface and the cutting of chutes only occurred rarely.

The sediment load transported in the channels consisted mainly of fines, carried in suspension. The bed load fraction, however, was texturally and compositionally immature, owing to the comparatively short sediment transport paths which existed in the depositional basin. The compositional immaturity of this sediment is discussed in Chapter 7.

The preservation of well developed lateral accretion surfaces indicates that the discharge of the rivers fluctuated strongly. Transport and deposition of bed load material took place only in periods of high discharge; at other times deposition of fines from suspension blanketed the layers of sand deposited during flood stages. The lack of desiccation features or red colouration in the upper parts of the point bar sediments suggests, however, that the channels always contained water, and did not have a pronounced, or very long, low stage.

4.6 Relationships between Facies in Facies Association II

4.6.1 Vertical sequences

Several types of vertical sequential arrangement can be recognised in this Facies Association. For convenience these are subdivided into four groups.

1) Mudstone sequences containing no siltstone or sandstone, and composed largely of palaeosols

Sequences of this type fall into two: those which over an appreciable thickness, of say 5-10m, contain only palaeosols of type 3 (post depositionally oxidized alluvial soils); and those which over comparable thicknesses contain only palaeosols of type 4 (evolved and polyphase soils).

The former type, typified by a section from Kibblestone borehole (Fig 103), contains abundant slickensided and listric mudstone containing roots and plant fragments. The large proportion of alluvial palaeosols in the sequence is reminiscent of sections in Facies Association I (e.g. Figs 27-29). In the Kibblestone section illustrated, red colouration is poorly developed, the rocks being mainly a dull purplish red colour. This probably reflects the low initial iron content of sediments deposited, initially, in a reduced state. There is a complete gradation in sequences of this type between completely reddened sediments, as seen in sections at Manor Quarry (where Facies 12 oxidized ironstones are present), through sediments in which some organic matter has survived (as in the section illustrated in Fig 41), to sequences which have hardly been affected by post burial oxidation, and which are thus probably best ascribed to Facies Association I. Examples of the latter type occur below 52m in Playground No. 8 borehole (see Fig 153), and between 27m and 34m in Kibblestone borehole.

A characteristic sequence of mudstones containing abundant evolved palaeosols is found in Playground No. 8 borehole (Fig 104). As described in 4.3.6, the development of evolved palaeosols in this section has been such that interference and overprinting between successive soil profiles has rendered individual profiles unidentifiable.

Such sequences indicate the persistence of zones of low clastic input on the alluvial swamp or alluvial plain surface. Such a low input reflects, in the case of sequences characterised by alluvial soils, persistent remoteness from active alluvial channel systems. While this may have been induced by tectonic subsidence, such an explanation seems unconvincing, as an area of persistently higher subsidence would tend preferentially to be occupied by a channel system following avulsion. Instead, such areas of persistent swamp sedimentation are regarded as having formed during periods when the active channel systems' avulsion pattern was such as to reduce the clastic supply to some areas for a long period.

Sequences dominated by alluvial palaeosols are common in the lower part of the Etruria Formation in all areas, and reflect a continuation of the transition from alluvial systems characterised by swampy overbank areas (Facies Association I) to those characterised by well drained floodplains (Facies Association II). The transition between these two Facies Associations is further discussed in Chapter 6.

Sequences dominated by evolved palaeosols, as that in Playground No. 8 borehole, are uncommon; in other mudstone dominated sequences, for instance that at Metallic Tileries Quarry (Fig 30) the palaeosols occur at discrete intervals, do not form complex superimposed profiles, and are generally poorly developed.

Circumstances in which a sequence, such as that in Playground No. 8 borehole, might form require a long period of low clastic supply, while the good drainage conditions necessary for the formation of evolved palaeosols persisted. Such a combination suggests a depositional setting remote from active channel areas, in which good drainage conditions and low clastic conditions were maintained by the presence of a topographic high on the depositional surface. This may have been a fairly short lived feature, formed by pre-existing alluvial topography, or a more permanent feature generated by basement controlled tectonic activity.

<u>ii) Mudstone (Facies 8) sequences with intermittent palaeosols, and often containing silty mudstone and siltstone (Facies 9) and sheet sandstones (Facies 10).</u>

Sequences of this type form the majority of the Etruria Formation. At their simplest they are exemplified by the lower part of the section exposed at Redhurst Wood East Quarry (Fig 105). Here an 18m thick sequence of mudstones can be subdivided into six units, 2.5m to 5m thick, each of which consists of red mudstone passing gradationally upwards into an evolved (type 4) palaeosol. In similar sequences elsewhere, for instance at Bentley Quarry, polyphase palaeosols are present.

More commonly the 'packets' of sediment between palaeosols contain silty mudstone, and sometimes siltstone. In most cases in which this occurs, the sediment between palaeosols forms a fining upward sequence, in which siltstone and/or silty mudstone (Facies 9) pass gradationally upwards into red mudstone (Facies 8). This in turn shows an increasing amount of colour mottling upwards, and passes gradationally into a palaeosol. Typical examples of such sequences are illustrated in Figs 106 and 107, from which it is apparent that there is a considerably variation between fining upwards sequences of this type, both in their thickness and in the proportion of the sequence occupied by siltstone or, more often, silty mudstone. The thickest individual fining upward units that have been observed are ca. 7m thick (e.g. at Lightwood and Springfield South Quarries - Fig 108). Both of these examples consist mainly of silty mudstone. It is possible that much of the apparent upward fining in such examples is due to the loss of quartz (as silt sized grains) through leaching during soil formation.

The boundary between successive fining upward units is usually sharp. In some instances, the upper leached and varicoloured horizons of the palaeosols are not preserved beneath the overlying base at the succeeding fining upward unit. This suggests that the bases of these units sometimes are slightly erosional. An erosive relationship can be demonstrated at ca 8m in the section at Springfield North Quarry, where the silty mudstone overlying an eroded palaeosol contains a basal concentration of calcareous 'shot' concretions reworked from the underlying mudstone sequence.

Very rarely a coarsening upward sequence is observed in the interval between palaeosols (Fig 107).

Where sheet sandstone bodies (Facies 3) and small isolated channels (Facies 4) occur, they are always found in the silty mudstone or siltstone, in the basal part of fining upward sequences such as those just described (e.g. Fig 108).

Palaeosol types occurring in the upper part of these fining upward sequences include alluvial soils, usually oxidized after burial, evolved soils and polyphase evolved soils.

Alluvial overbank sequences which consist of alternations of unrooted lithology and palaeosols were obviously deposited during alternating phases of fairly rapid deposition, during which little rooting of the sediment could occur, and non deposition, during which a soil formed on the deposition site. Sequences consisting entirely of mudstone, such as that at Redhurst Wood East (Fig 105) were presumably deposited at a site remote from active channels, and thus from the loci at which silty and sandy sediment was introduced onto the floodplain surface. The presence of fining upward sequences in the overbank alluvium indicates a depositional site more proximal to the active channel systems, and within reach of coarser material introduced by crevassing or bank overflow of the channels.

Three explanations are possible for the presence of fining upward sequences in the overbank alluvium

i) A fining upward sequence may have been produced during the migration of an active fluvial channel away from a particular depositional site, with a concomitant decrease in the influence of proximal overbank depositional processes.

ii) A fining upward sequence may have formed in response to the evolution of the channel complex during its occupancy of a particular site, and, in particular, in response to the progressive formation and growth of mud rich levée banks.

iii) A fining upward sequence may have been produced during the abandonment of a channel complex, during which less bedload material was transported, and a lower proportion of the flood discharge channelled onto the floodplain by the channel in question.

The corollary of the first explanation is that coarsening upward sequences should have been formed elsewhere in the floodplain, during the approach of a migrating channel, or by progradation of crevasse lobes sourced by a stable channel. Sequences of the latter type have been described by Friend et al (1982) from the Oligocene Campodarbe Group of Northern Spain.

The rarity of such coarsening upward sequences suggests that sequential organization of the floodplain sequences did not, by and large, occur in response to movement of the adjacent channels.

The second and third explanations are more attractive. However, in this case, it is difficult to envisage the exact relationship between channel processes and overbank processes. It is evident from the number of lateral accretion sets present in the channel sandbodies of Facies Association IIB that repeated major flooding occurred during deposition of these bodies. Yet, in almost all cases, the thin sheet sands of crevasse splay origin (Facies 10), occurring at the bases of the floodplain fining upward sequences, are very few, often only a single sheet sand, or at the most, three or four sheets. It seems unlikely that so few of the repeated floods in the channel complexes penetrated onto the adjoining floodplains. Either very few of such floods carried sand onto the floodplain, or floods were largely constricted; in either case the genesis of the floodplain fining upward sequence would have occurred during the phase of maximum activity in the adjoining channel, rather than after abandonment of the channel.

The most likely explanation is that active crevasse formation, and the associated deposition of sand on the floodplain surface, only occurred during the early stages of the evolution of a particular channel complex, soon after avulsion. Because of the suspension dominated load large mud levée banks would have been formed by the active channels, which would have been more resistant to breaching than banks composed of sand or heterolithic lithologies. Thus, from a fairly early stage in the development of an individual channel complex, the proximal overbank areas would have been starved of bedload transported material. The only sediment reaching the overbank areas in such a case would be suspended load, transported by floodwaters which overtopped the levées. Further bank construction and/or ultimate abandonment of the channel system would cut off this supply in time. This combination would result in the formation of a fining upward sequence.

If this explanation is correct, the silty mudstones of Facies 9 may indeed represent levée deposits (see discusion in 4.2.3).

iii) Mudstone (Facies 8) and siltstone/sandstone (Facies 9 and 10) sequences, containing channel deposits of Facies Association IIB

Problems of exposure have made it difficult to observe the sequential relationship of the channel deposited sandbodies of Facies Association IIB to the overbank sequences of Facies Association IIA. The sequences illustrated here are thus taken from borehole core sections. While these are undoubtedly more or less complete, their interpretation is clouded by the difficulty of distinguishing between in-channel and overbank deposits in the siltstones and interbedded sandstones occurring in the upper parts of the sandstone bodies.

Three typical occurrences of channel sandbodies in vertical sequences are shown in Figs 109-111. These three sequences have in common a simple fining upward pattern, strongly resembling that described as diagnostic of fluviatile sediments by Allen (1964, 1970) and others (summarized in Collinson 1978). Owing to the constraints of core sample material, it is not possible to identify lateral accretion sets in these sandstone bodies. However, the sandbody in the Rosemary Hill section (Fig 109) correlates with a body exposed at the surface (Figs 89,90), and the predominance of lateral accretion structures in other exposed sandbodies makes it likely that the sandbodies in the Allotment and Kibblestone sections were also deposited by meandering channels.

As in the interpretation of Allen and others (op cit), the components of these fining upward sequences may be interpreted as follows:

a) the fining upward sequence within the basal sandbody, resting on an erosive surface, represents the deposit formed by an actively migrating alluvial channel (as discussed in 4.5).

b) the siltstone dominated sequence overlying the basal sandstone unit represents either the upper bar deposits of the channel, or the deposits of the levêe or proximal floodplain. Silt dominated deposits showing pronounced lateral accretion, and thus of channel origin have been observed in several places, notably at Springfield North Quarry (Fig 95).

c) the overlying mudstone sequence represents more distal floodplain (or floodbasin) deposition following migration or abandonment of the active channel.

d) the capping palaeosol formed after the cessation, for some time, of deposition.

These fining upwards sequences strongly resemble those described previously from the floodplain association, except in having a channel deposit at their base.

iv) Sequences dominated by sandstone (Facies 13)

Only one example of such a sequence has been observed, at Bayton Works (SO702753; Fig. 1) in the Wyre Forest (Fig 112). This is the only significant section exposed in the Wyre Forest. However, for reasons which are outlined in 9.8.3, it is believed to be typical of much of the Etruria Formation in this area.

The 15m thick section consists of three fining upward sequences composed mainly of medium to fine grained sandstone. The upper two sequences have erosive bases. Although weathering has obscured internal structures, enough is visible to confirm that these sandbodies are similar to Facies 13 bodies, as seen in the main outcrop areas (see 4.5.2). These bodies fine upwards into overbank mudstones containing pedogenic textures. The mudstones are, however, very thin, being erosively overlain by the base of the succeeding sand body.

The fining upward sequences at Bayton do not themselves differ much from those contained in channel deposits which have been described previously. The sequence as a whole differs in the proportion of mudstone to sandstone present in each fining upward unit, and in the apparent presence of a channel deposited sandbody in every fining upward unit. The latter contrasts with all Facies Association II sections measured elsewhere, in which channel deposited sandbodies are present in the minority of fining upward units.

These differences may be accounted for by considering the position of the Wyre Forest in the context of the basin as a whole. Its marginal nature implies proximity to the source area, while the comparatively thin sequence preserved and dominance of red beds (see 9.8 and 10.2) testify to a low subsidence rate. Both would favour a more sand rich alluvium, and the preservation of thin sequences of overbank fines.

4.6.2 Lateral relationships between Facies in Facies Association II

The nature of the exposures in the Etruria Formation precludes field observation of the lateral relationship between the various Facies making up Facies Association II. In the only group of closely spaced boreholes, at Rosemary Hill quarry, the correlation between observed sections of this Association is not sufficiently good to establish lateral relationships. Such relationships can therefore only be inferred.

The lateral relationship between the alluvial channels and the associated overbank areas can be inferred by considering the Etruria Formation sequences as accumulations of sediment packets, successive packets being separated by palaeosols which mark successive depositional hiatuses. Excluding the sediments observed at Bayton in the Wyre Forest, which are atypical of Facies Association II in the majority of the area under consideration, three types of sediment packet are distinguishable, corresponding to those with features typical of the first three types of vertical sequence described in 4.6.1. These sediment packet types are: i) packets of mudstone, containing abundant and repeated palaeosols; ii) packets of silty mudstone and mudstone, containing relatively fewer palaeosols, and sometimes having crevasse splay and minor channel and/or levée sediments at their bases; iii) packets containing major fluvial channel deposits at their bases.

The increasing proportion of silt and sand size sediment, and the increasing thickness of sediment separating successive palaeosol horizons from sequences of type i) to those of type iii) suggests that these distinctive types of sediment packet have a distal to proximal relationship, relative to the fluvial channels in the alluvial system (Fig 113). The most proximal sequences contain deposits of the major channel itself.

This inferred lateral relationship between sequences containing channel and overbank deposits allows a reconstruction of the relationship between the active channels and the associated overbank areas. This is summarised in Fig 114, which takes the form of a hypothetical section drawn normal to the channel system. This figure is largely self explanatory: one aspect only calls for further comment.

The comparatively rare occurrence of channel deposits in vertical sequences - on average one depositional packet in every ten - suggests that individual channels did not migrate laterally to a great degree. This bears out the conclusion drawn on the basis of the internal organisation of the channel sand bodies (in 4.5.5), that the channels were rather unstable, and did not construct very extensive point bars. Channels of this type shifted mainly by avulsion, resulting in a high preservation potential for overbank sequences lacking channel deposits.

The sequence of events leading to the accumulation of a typical Etruria Formation succession, and summarizing aspects discussed in 4.6, is illustrated in Fig. 115.

CHAPTER 5

Facies Association III - basin margin alluvial fan and fluvial sediments

5.1 Introduction

In vertical sections through this Facies Association, 50% or more of the sequence is usually composed of red mudstones, silty mudstones, and palaeosols, identical to Facies 8 and 9 and the palaeosols of Facies Association IIA. The distinctive feature of Facies Association III is the occurrence of paraconglomerates and orthoconglomerates, forming bodies of a variety of shapes, which are commonly of apparently non-channelized origin.

5.2 Facies descriptions

5.2.1 Facies 16: paraconglomerates, showing little or no internal ordering

The rocks in this Facies consist of extremely poorly sorted matrix supported conglomerates, containing clasts of intra- and extra-formational origin which range in size from granules to cobbles and boulders, up to 1.50 m in maximum length. In one example (Fig. 116), in section 13 at Redhurst Wood west quarry (SJ969052; Fig. 4), the extraformational clasts range from 5 to 30 cm in diameter, and consist of acid lavas and tuffs. These boulders are very well rounded, and are supported in a silty mudstone matrix. The more frequent examples in the Telford area at Blockley's quarry (SJ683118; Fig. 1) (Figs. 117 and 118), and at Donnington Wood quarry (SJ710113; Fig. 1) (Fig. 119), contain extremely angular, virtually non-abraded, cobbles of vein quartz and a variety of metamorphic and intrusive igneous rock types, supported in a matrix of very argillaceous granules and coarse sand.

These conglomerates all rest on an erosive base, which may take the form of an irregular, slightly scoured nearly horizontal surface (Fig. 119), or form part of a deeply incised channel (Fig. 120).

Little internal organization is visible within the conglomerates. In the example of Redhurst West Quarry (Fig. 121) the basal 0.90m. of the conglomerate contains only intraformational clasts of mudstone, similar to that immediately below the erosive base. The main part of the deposit contains both intra- and extraformational material. In the example at the top of the main face at Blockley's Quarry, lenses of clast supported conglomerate are present both in the middle (Fig. 117) and at the base (Fig. 118) of the deposit.

Interpretation

The matrix supported nature of these conglomerates, together with their lack of internal organization suggests that they were deposited by debris flows. Similar conglomerates are thus interpreted by Rust (1978), and Larsen and Steel (1978).

There are insufficient exposures to determine whether any clast orientation is present, but the impression derived from the exposures illustrated in Figs. 116-118 suggests that, in these cases, there is not. The thin clast supported lenticles in the example from Blockley's quarry (Figs 117, 118) may represent brief periods in which the top (in the case of Fig. 117) or the entire thickness of a debris flow deposit was reworked by flowing water after deposition, and lost some or all of its matrix by winnowing. Alternatively, these thin lenticles could be the bases of normally graded sequences deposited by rather more fluid, less viscous, flows (cf. Reineck and Singh 1975).

The intraformational clast horizon at Redhurst West quarry is a unique occurrence. From the similarity in lithology between the intraclasts and the underlying mudstone, these particles seem to have been transported for only a very short distance. This mud flow is considerably thicker than any of the others observed, and it is likely that the main body of the flow was 'frozen' (cf. Johnson 1970 - cited in Middleton and Hampton 1976), and moved as a rigid plug, riding on a thin, constantly deforming sole zone. Within this zone, intraclasts from the underlying mudstone could easily be incorporated into the flow, while the extraformational clasts would not be able to enter this zone because of the rigid state of the majority of the flow. This feature is not seen in the flows (e.g at Blockley's quarry) with a sandy matrix, as 'freezing' of these flows, by dewatering of the matrix, would have taken place very rapidly, and resulted in an abrupt cessation of flow.

The pronounced erosive base to some of the examples of this Facies is an anomalous feature, as debris flows are sometimes claimed not to have the capacity to cut signficantly into their substrate. It is likely, especially in the case of the Redhurst Wood example, that these flows moved along channels which were already in existence.

5.2.2 Facies 17: Clast supported conglomerate occurring in a laterally restricted channel or valley fill

Only one exposure of this Facies has been observed, filling the upper part of a large scour or small valley at the eastern end of the north face of Redhurst West quarry (Fig. 120). This 'valley' is between 7 and 10 m deep, at the deepest area observed, and at least 70 m wide. The full width may be inferred from the shape of the exposed portion to have been at least 150 m. The lowest part of the valley fill consists of the mud flow described under Facies 16. (The facies relationships of this valley are described in section 5.3.)

The majority of the conglomerate is massive in appearance, showing an overall arrangement into near horizontal units 0.60 m to 1.00 m thick (Figs. 122, 123). Within these units there is usually no visible depositional structure, although occasionally poorly defined horizontal bedding (Fig. 122, 123) or cross bedding is visible. The clast size decreases upwards within individual units. Maximum clast sizes range from 3-5 cm. Within these conglomerate units there are occasional small, flat topped sandstone lenticles (Figs. 123, 124). In some cases, the entire thickness of a conglomerate unit passes laterally into sandstone, the conglomerate in this case having the appearance of a bar. At one point there is a lenticular, horizontally laminated siltstone body which drapes the top of a conglomerate sheet (Fig. 124).

Towards the top of this conglomerate body the conglomerates become finer grained and pass upwards into coarse sands containing abundant granule material. These in turn are overlain by interbedded sandstones and siltstones of Facies Association IIA, which locally fill a small channel cut into the top of the conglomerate body.

Interpretation

It is difficult to make a definitive interpretation of this Facies on the basis of a single poor exposure showing no three dimensional control.

The most abundant facies, volumetrically, are closely comparable with the Facies Gm (massive or crudely bedded gravel) and Gt (trough cross bedded gravel) described by Miall (1977) in the context of braided river deposits. Although gravels can equally effectively be transported and deposited by meandering rivers (Jackson 1978), a braided steam interpretation for this Facies in the Etruria Formation is tentatively suggested in view of the following considerations.

i) The conglomerate body fills a laterally constricted erosional feature.

ii) The gradient of this feature was sufficient to allow the movement of the underlying debris flow.

iii) The distribution of silt drapes and sand lenticles suggests that the ca. 1.00 m thick depositional units within the conglomerate body represent the amplitude of the active bedforms in the channel at any time. As these units appear to be laterally extensive, this implies that the channel system had a high width to depth ratio, which, with a pebbly bed load, would tend to favour a braided pattern.

Within this interpretation the horizontally stratified and cross bedded conglomerates probably represent, respectively, longditudinal bars and interbar channel fills (Miall 1977). The flat topped sand lenticles may represent lower energy interbar channel deposits, formed at low stage or, more probably, following temporary diversion of the main flow. The silt drape represents deposition from slow moving water, either during falling stage draping a gravel bar, or during low water stage draping any part of the channel sediments. The scarcity of the latter two types of sediment reflects the low preservation potential of fines in the braided stream system.

5.2.3 Facies 18: Orthoconglomerates and conglomeratic sandstones

The conglomerates and sandstones in this Facies occur in a variety of sand body forms and sizes, and almost certainly overlap with Facies 16 to 20 in terms of depositional process. The occurrences have in common a tendency to occur in parallel sided sheets of great lateral extent, made up of one or several fining upward units. Occasionally channel fill features, of varying scale, are present.

The most typical examples of this Facies are to be found in the group of quarries to the north of Aldridge (Fig. 4). The typical macroscopic appearance is illustrated in Fig. 125, which shows a rain-washed fallen block from one of the units of this lithology at the base of the Highfield south quarry section. The block shows a graded unit, some 50 cm thick, grading upwards from a framework supported cobble grade conglomerate at the base, through finer grained framework supported conglomerate and pebbly sandstones to sandstone with disseminated small pebbles. In this example the graded unit forming the main part of the block is underlain by another thin unit with a conglomeratic base and sandy top, and overlain by the conglomeratic base of another unit.

Internal structure is rarely seen in thinner units (ca. 0.30 m to 0.50 m), but thicker units (thicker than 0.50 m) often show trough cross bedding. The sandy tops of several graded units in Utopia Quarry show ripple cross lamination, and ripple forms are occasionally preserved on the top surfaces.

At two places in Utopia Quarry (SK046026; Fig. 4)a graded conglomerate unit (probably the same unit in both cases) is overlain by 0.60 m of interbedded medium grained sandstone and siltstone, which forms units which have depositional dips of up to 10° (Fig. 126). Ripple forms preserved on one of these surfaces indicate flow at approximately 90° to the depositional dip.

The bases of conglomerate sheets formed of this Facies are usually slightly erosive, and often show infilled scour structures (e.g. Fig. 126), occasionally show groove casts (Fig. 127), and sometines show a complex pattern of shallow depressions up to ca. 50 cm wide, which sometimes amalgamate to give a polygonal pattern. The latter features are regarded as some form of load structure.

In several examples a more complex internal structure is present.

i) In a number of cases thicker conglomerate sheets are present, which are composed of several graded units. These may lack apparent internal structure (as in Fig. 125), or may contain cosets of trough cross bedding (Fig. 128). In the latter case, pebbles and cobbles are limited to the bases of individual cross bed sets. There is a continuum from sheets containing cosets of cross bedded, largely matrix supported conglomerates, through the lithology illustrated in Fig. 128, to sheets of pebble-free sand, which form Facies 19.

ii) Within sheets containing several interbedded conglomerate and sandstone units, lateral variation in the proportion of conglomerate to sandstone is observable, the two lithologies interdigitating (Fig. 129). In this exposure, the conglomerate lenses are clearly forming "bars" of some kind within the sand/conglomerate sheet.

This conglomerate Facies occasionally occurs in channels, which may either be isolated or associated with sheets of the same lithology.

Three forms have been seen:

i) Channels of greater than 1m depth with conglomerate fills. Typical examples of this type are seen in Utopia quarry (Fig. 130) and at the western end of the main face at Blockley's quarry (Fig. 131). Such channels have fills of conglomerate and pebbly sandstone, which may exhibit some crude horizontal bedding. These channels occur in isolation, or passing laterally into, or eroding into laterally extensive sheets of conglomerate and sandstone. In two examples, at BLockley's quarry (Fig. 132) and at Redhurst west quarry (Fig. 133) channels of this type have steep banks, with bank overhangs, or injection of bank material into the conglomerate fill. The fill of the latter example seems to belong to the same depositional event as the material forming an immediately overlying sand sheet (Fig. 134).

ii) Channel with sheet sand and mudstone fill. One example only has been observed. (Although it contains no conglomerate, its close association with sheet conglomerates justifies its description being included in the conglomerate lithofacies). The channel is poorly exposed in Redhurst West Quarry (Figs 134, 135). The eastern margin of this channel (right hand side in Figs.) consists of an erosion surface cutting down through mudstones and an earlier sheet sandstone, draped by a single bed of medium grained sand, 0.40 m to 0.60 m thick. Above this the channel is filled with silty mudstone. The westward continuation of the sand and mudstone fill is covered, but exposures between 20 m and 30 m to the west probably represent the other side of the channel. The margin at this point consists of a sloping surface, cut through a massive sandstone. The fill (Fig. 136) consists of interbedded units up to 0.50 m thick of medium to fine grained sandstone and siltstone, dipping at up to 30°. These are truncated at an angle by the overlying conglomerate sheet, which is demonstrably showing the regional dip. The dipping sand beds are apparently banked up against the channel margin.

iii) Small conglomerate filled scours. Small scours 0.40 m - 1 m deep and up to 6 m wide have been observed at Utopia quarry and Wilnecote quarry (SK220000; Fig. 1), occurring as isolated bodies or eroded remnants below conglomerate sheets. Their fill consists of pebbly sandstone or poorly sorted clast supported conglomerate.

In the Warwickshire area, conglomerates of this Facies, occurring in sheets and channels, are virtually monogenetic, consisting nearly entirely of disc shaped clasts of indurated shale (see 10.1.1 for discussion of provenance). No cross bedding has been seen in this lithology, depositional structure being limited to crude imbrication, which itself is very rare.

Interpretation

This Facies represents a 'mixed bag' of sediments, deposited under broadly similar conditions, but forming a spectrum, with end members similar to sediments of Facies 16 and Facies 19 to 21.

At its simplest, when this Facies occurs as sheets consisting of one upward fining unit, each sheet represents the deposits of a single high energy waning flow event, carrying pebbles and sand across an alluvial surface. In cases where the sheet consists of two or more graded units this process was repeated a number of times, but apparently without the development of any kind of organized braided or meandering stream system. The initial erosive effect of these flows must have been considerable, to judge by the presence of the deep and narrow channel with edges showing undercutting or soft sediment injection which occurs beneath the sheet conglomerates at Redhurst Wood west quarry, and at Wilnecote Quarry. Deposition of sediment in the lower parts of the sheets was rapid, preserving undercut features and grooves, and producing local load structures. In the later stages of flow, some bedload transport of sand took place, giving rise to the cross lamination observed, for instance, at Utopia Quarry.

Sediments very similar to these 'simple' examples of Facies 18 have been described from the Oligocene of the Ebro Basin, Northern Spain, by Nagtegaal (1966). Points of similarity between the Etruria Formation examples and these much better exposed Tertiary rocks are as follows: great lateral extent, in the order of kilometres; thicknesses of less than 3 m, with very constant thickness over hundreds of metres; loaded and erosive basal contacts; and the occurrence of large scale 'scour and fill' structures at the bases of some sheets, and as isolated channels. These 'scour and fill' structures are of similar size and shape to the channels observed in this Facies in the Etruria Formation. Nagtegaal interpreted the conglomerate sheets as sheetflood deposits. He stressed the role of soft sediment deformation in modifying the margins of channel structures, and in leading to apparent channel margin overhangs (cf. Figs 132, 133).

A similar interpretation is adopted for simple conglomerate sheets in the Etruria Formation. (Further discussion of the total depositional environment is reserved until sections 5.4 to 5.7). This interpretation may be elaborated in the light of several other features of the Facies.

i) Occurrence in the upper part of a conglomerate sheet of interbedded sandstones and siltstones showing depositional dip (Utopia Quarry). These dipping units resemble, on a small scale, the lateral accretion units developed in the meandering channel deposits of Facies Association IIB. This resemblance is reinforced by the current directions obtained from the interbedded sands. This feature may be interpreted as resulting from the development of very small scale meandering streams, probably not more than 1 m deep and 2 m-3 m wide, on top of the conglomerate sheet, for a short period after deposition.

ii) Sand and mud filled channel (Redhurst West Quarry). The nature of the contact between the dipping sand units at the western end of this feature and the sandstone through which the channel is eroded (Fig. 136) does not support the view that these sands have accumulated by any form of lateral accretion. Rather, it would appear that this angular relationship results from rotational collapse of the channel bank, and the consequent preservation of inclined layers of the sediment which formed the upper part of the bank, and which is not preserved <u>in situ</u> owing to subsequent erosion associated with the deposition of the overlying sheet conglomerate.

At the eastern end of this channel, the channel form is preserved by being draped with one medium grained sand layer. Otherwise the channel is mud filled. This channel was thus not cut and filled immediately. It must instead have been cut by a very high energy event, and subsequently filled with sands and muds deposited under much lower energy conditions.

iii) Presence of ? bars (e.g. Utopia Quarry) and cross bedded pebbly sands (e.g. Empire Quarry). Both of these features suggest that, in some cases, sheet conglomerates of this Facies were deposited by more perenially active alluvial systems, which may have developed some form of organization, probably in the form of braiding. Where present these features indicate a complete transition between this Facies, and Facies 17 (entrenched braided stream system - see 5.2.2).

These relationships are further discussed in 5.4, 5.5 and 5.7.

5.2.4 Facies 19: sheet sandstone bodies

Within sequences of laterally extensive units of interbedded conglomerate sheets and mudstones, occasional sheet like bodies occur which are composed exclusively of sand. In the best exposed example, in the western section of the Empire quarry (SK043023; Fig. 4) (see Fig. 141) a sheet-like body 1.40 m thick consists entirely of medium grained sand, forming one coset of 0.10 m to 0.25 m planar tabular cross bedding (Fig. 137), showing a unimodal current direction.

Interpretation

From one good exposure it is difficult to generalise about this Facies. From its thickness and internal structure it was obviously deposited in a manner different to that which deposited the sheet sands of Facies 10.

This sandstone sheet is closely associated with sheet conglomerates of Facies 18, some of which may have been deposited by a braided stream system, but the majority of which were probably deposited by short lived flood events. By inference, similar events were probably responsible for the deposition of sheet bodies composed exclusively of sand.

Bedforms comparable to those observed in this Facies have been described by Williams (1971) from recent sandy sheet flood deposits in Australia.

On the basis of the small amount of data available, if cannot be excluded that the cosets of planar crossbeds in this Facies are the product of the migration of transverse bars or sandwaves in a more perenially active braided stream (cf. Facies Sp of Miall 1977). Such a system would, however, have been short lived, to produce so small a sand body.

5.2.5 Facies 20 : isolated channel sandstone bodies

Two isolated channels, filled with medium to coarse grained sandstone, occur in a mudstone dominated sequence in Ketley quarry (SO898890; Fig. 5). It cannot be ascertained with certainty to which Facies Association this sequence should be referred, as no distinctive conglomeratic facies are exposed. However on the basis of the lack of well developed palaeosols (see 5.2.6), and the predominance of conglomeratic deposits in the surrounding area (see 9.7; Whitehead and Eastwood 1927), it is assumed that the sequence in this quarry is probably best referred to Facies Association III.

The channels (Figs 138,139) occur at approximately the same horizon, and both have the same directional trend. Both are ca. 3m deep at their deepest point, and were probably about 20m wide. (The nature of the exposure makes estimation of their width difficult). The channels are filled with crudely bedded medium to coarse grained sandstone. No internal structural was seen.

Interpretation

As these channels are significantly larger than channels (Facies 11) observed in the overbank sediments of Facies Association IIA, they must be interpreted as some form of major alluvial channel. From their cross sectional shape, it would appear that these channelised sandbodies conform to the 'simple ribbon' type of Friend et al (1979). These authors suggest that such a form of channel and channel fill results from a combination of the following factors: low or high flow strength; strong banks; flash flooding régime; and differential vertical movement of the alluvial area with respect to the base level of erosion. From the occurrence of graded conglomerate sheets and 'scour and fill structures' in other Facies within this Facies Association, it would seem that at least the first and third of these criteria were satisfied at the time of deposition of these channel fill sandstones.

The occurrence of these two similarly oriented channels immediately adjacent to one another and at about the same stratigraphic horizon may be fortuitous. It is possible, however, that the two channels mark two positions, not necessarily successive, of a single alluvial channel system which, as a result of strong stage fluctuation and rapid infilling of channels underwent repeated and frequent avulsion. Such behaviour resembles slightly that inferred by Friend et al (1981; p. 4.42 and Figs 4.38, 4.39) in low sinuosity ribbon sand bodies in alluvial fan sediments in the Campodarbe Group of the Palaeogene of Northern Spain. In this case ribbon sand bodies occur in vertically stacked sequences, produced by the upbuilding of a low sinuosity channel through repeated filling and re-cutting. Such a sequence of events might lead to the formation of a cluster of channel fills, if accumulation of fines between channel cutting events was sufficient to mask the original channel topography. Whether the two channels in the Ketley exposure under consideration were formed in this way by the same alluvial channel system must, however, remain speculative in the absence of more extensive exposure.

5.2.6 Other sediments and palaeosols in Facies Association III

Fine grained sediments: Within vertical sections through this Facies Association, a large proportion of the total sequence is composed of red mudstone and silty mudstone identical to Facies 8 and 9.

Palaeosols: Palaeosols are less abundant and less well developed in this Association than in Association II. Where present, the palaeosols are of Palaeosol Type 4 (highly evolved) or of indeterminate type. The best developed palaeosol profiles, showing pronounced leaching, have been observed at Blockley's Quarry. Carbonaceous palaeosols of Type 1 and their oxidized equivalents have not been observed in this Facies Association, although siderite occurs in a red palaeosol at Wilnecote Quarry. Coaly horizons are recorded within the Etruria Formation in areas where it contains large amounts of Facies Association III sediments (Whitehead and Eastwood 1927; see 9.7), but they cannot definitely be ascribed to this Facies Association in the absence of any sedimentological description.

5.2.7 Sedimentary dykes

Sedimentary dykes occur frequently in association with Facies 16 and Facies 18 (debris flow deposits and sheet conglomerates). Two distinct types are recognisable:

i) mudstone dykes intruded upwards through debris flow deposits. These take the form of anastomosing, 5 to 15cm wide near vertical dykes, sourced by the mudstone immediately below, and cutting the entire thickness of, matrix supported debris flow deposits. Examples have been observed at Redhurst Wood West Quarry (cutting the mudflow deposit illustrated in Fig. 116) and at Donnington Wood Quarry. In neither case does the intrusion extend into the sequence overlying the debris flow.

In one example at about 17.50m in the northern (downfaulted) section at Blockley's Quarry, several dykes of this kind are intruded upwards through a thin debris flow deposit and into a sequence of palaeosol and red mudstone. Granule material from the debris flow is incorporated into the intrusive material in the palaeosol horizon.

ii) sandstone/conglomerate dyke intruded upwards through a sequence of mudstones and conglomerates. The single exposure of this type of intrusion is in the eastern part of the main face at Blockley's quarry (Fig 140). The dyke is 2 to 10cm wide, and composed of coarse sand and granule to small pebble sized conglomeratic material, with an argillaceous matrix. It is apparently sourced from a laterally extensive conglomerate sheet, and intrudes upwards through at least 8m of overlying mudstone and conglomerate, including three palaeosols.

Both types of dyke show some degree of ptygmatic folding, resulting from differential compaction. This is particularly pronounced in the case of the sandstone dyke.

Interpretation

These features present a problem of interpretation. Intrusions of soft sediment are usually early features, formed by rapid, often load induced, dewatering of sediment soon after deposition. In such circumstances a soft sediment intrusion may be expected to penetrate the thickness of the body of loading sediment, but no further.

In the cases of the intrusions at Donnington Wood and Redhurst Wood West, such a mechanism may have occurred to form dykes cutting debris flows. It is, however, likely that the sediment immediately underlying these bodies was already in a dewatered state at the time of debris flow emplacement, in view of the complete lack of soft sediment deformation observed elsewhere in the mudstone facies. The liquefied sediment must therefore have formed in the sole zone of the debris flow itself. In both cases the debris flow deposits contain a high proportion of argillaceous matrix, which would have helped trap any pore waters and thus encouraged rapid fluid escape and sediment intrusion.

The intrusions in Blockley's Quarry cannot be explained by early post depositional water expulsion. Both of the examples cited cut through mudstones and palaeosol horizons, which implies a substantial time elapse between deposition of the containing rocks and emplacement of the intrusion. Furthermore, it is unlikely that after such a period sediment loading could have led to soft sediment intrusion as a result of rapid dewatering of a sand and conglomerate body.

The mechanism of emplacement of this latter type of intrusion may only be speculated upon. It may be significant that the quarry in which the intrusions are exposed is within 1km of the contemporaneous fault bounded margin of the basin. Any tectonically induced shocks caused by movement on this faulted margin may have caused slight fracturing in the sediment fill of the basin. If the coarse grained aquifers in the sediment fill were overpressured, this pressure might have been released through such fractures, giving rise to soft sediment intrusions. It seems unlikely that such pressures, in sandy sediments fairly near the surface, could have been due to burial overpressuring. It is possible that an artesian head of water may have been involved.

5.3 Relationships between Facies in Facies Association III

5.3.1 Spatial relationships between sand bodies:

The simplest arrangement of Facies is found in the exposures at the Highfield and Empire Quarries (Fig 141), and in the Wilnecote and Averill Quarries (Fig 142). In both of these the sequence consists solely of sheet conglomerates (Facies 18) and sheet sandstones (Facies 19), interbedded with red mudstones and poorly developed palaeosols. The sheet conglomerates consist of both simple, thin graded units, and of more complex sheets containing several graded units, and consisting in part of cross bedded pebbly sand. Small channels are present, occurring in isolation, and truncated by, or eroding into, the sheet conglomerates. In the Empire Quarry exposure, sheet sandstones (Facies 19) are present in the lower part of the sequence only.

The relationship between typical thin laterally extensive conglomerate sheets of Facies 18, and the thicker sheets of the same Facies, which contain several graded units and some bar structures, is seen in the north face of Utopia Quarry (Fig 143). Here thin conglomerate sheets and sandstone sheets occur regularly in the vertical section, apparently thickening and dying out randomly across the quarry face. Towards the western end of the quarry, however, two sheets thicken rapidly into an interbedded sequence of conglomerates and sandstones, containing both large and small conglomerate filled channels (Fig 130).

A better exposure of the relationship between conglomerate filled channels and laterally extensive conglomerate sheets is found at Blockley's Quarry (Figs 144-147). Here a variety of sandbody forms is present (Fig 145). Channelised bodies vary from small channels with single phases of fill (R in Fig 146) to larger bodies with two or more phases of fill (C in Fig 146). Channelised bodies occur in isolation, and both erode into and pass laterally into sheet like bodies (E and L in Figs 146 and 147). Some of the laterally extensive conglomerate sheets occur in shallow channels (Ch in Fig 147), while others consist of packets of conglomerate and sandstone units, with apparently great lateral extent and no erosional topography (S in Fig 146). (The sheets illustrated as Ch and S in Figs 146 and 147 all extend for considerable distances beyond the limits of the Figures).

Facies 17, conglomerates occurring in a laterally constricted 'valley' fill, have only been seen in one exposure, at Redhurst Wood West Quarry (Fig 120). The erosive 'valley' feature cuts through a packet of Facies 19 pebbly sandstone sheets interbedded with red mudstone (Fig 148).

5.3.2 Sequential relationships

A typical vertical sequence through the Facies Association is provided by the northern section of Blockley's Quarry (Fig 149). No overall sequential organization is apparent, beyond the presence of fining upward units formed by individual sand/conglomerate bodies. A similar lack of vertical sequential pattern is found in other exposures, where the sequences are more monotonous and lack well developed palaeosols and debris flow deposits.

5.4 Towards a model of depositional environment

From the occurrence within them of palaeosols, and from their interdigitation with sediments of Facies Association II, the sediments of Facies Association III are clearly alluvial in origin. The extremely angular and poorly sorted nature of the conglomeratic material, and the closeness of some of the exposures to documented fault bounded margins of the depositional basin (see Chapter 9) combine to suggest that the alluvium was of extremely proximal nature. However, the sequences and sand body types differ so much from those in existing facies models for ancient proximal gravelly alluvium deposited in alluvial fans (e.g. Collinson 1978) or in braided fluvial complexes (e.g. Miall 1978) that some discussion of the depositional processes is necessary before a conceptual facies model can be considered.

The principal differences between the conglomeratic sediments in the Etruria Formation and those described from other ancient gravelly alluvium are as follows:

i) Vertical sequences through conglomerate bearing sediments in the Etruria Formation usually contain at least 50% of mudstone, silty mudstone, and palaeosols. This is the case even in the most proximal settings containing extremely immature conglomeratic material (e.g. Blockley's Quarry - Figs 146,147,149). This compares with the negligible preservation of fines predicted by conventional facies models for arid zone alluvial fans (e.g. Collinson 1978) and for pebbly braided alluvium (e.g. Miall 1978).

ii) The scale of the conglomerate units is much smaller, both in thickness and in lateral extent, than that of other ancient alluvial conglomerates. Vertical sequences are much thinner than in the ancient alluvial fan sequences described by Steel and Aasheim (1978) and Heward (1978a,b) from arid and humid zone alluvial fan sediments respectively. The conglomerate units, other than the 'valley' fill at Redhurst Wood, are thinner than, and lack the internal organisation and lateral interconnectedness of, conglomerate units in ancient pebbly braided stream deposits, as described, for instance, by Rust (1978) in the Devonian Malbaie Formation in Quebec, or by Miall (1977).

iii) There seems to be a general lack of sequential organisation within vertical sequences through the Facies Association. This is in contrast to the organised sequences produced by lobe progradation in some ancient alluvial fan sequences (e.g. Larsen and Steel 1978).

5.5 Depositional processes:

Channelised sand and conglomerate bodies: The predominance of channels showing, apparently, one phase of fill is notable. Only one channelised sand body, in the eastern section of the main face at Blockley's Quarry, shows several phases of fill. These fills may be compared with the fill arrangements of the simple and complex ribbon type sand bodies described by Friend et al (1979). These authors suggest that such fill patterns in channelised sand bodies reflect the filling of the channel with sediment in a single phase, or a series of short phases of scour followed rapidly by the transport and deposition of sediment.

Sheet conglomerate bodies and sandstones: Although it has not proved practical to subdivide these Facies on the basis of the few available exposures, there is, at least in the conglomerate sheets, enough variation in internal structure to suggest two distinct modes of deposition for such sheets.

i) Where the conglomerate sheets are thin they often consist of one or several normally graded conglomerate and sand units. These represent the deposits of intermittent waning flow events in an environment otherwise dominated by the deposition of mud.

ii) The thicker conglomerate sheets contain cross bedding and bar structures, and were deposited by a more frequently active, probably braided, stream system.

If the lateral relationship to be inferred form Fig 143 is correct, the thin conglomerate sheets pass laterally into the thicker more organised sheets. They thus form 'wings' to the more organised conglomerate bodies, (in the sense of Bersier 1958 and Friend et al 1979), and probably represent the deposits of short lived phases of overbank flow generated by flooding in the more organised and more frequently active channel areas.

Debris flow deposits: The presence of debris flow deposits, in one case (Redhurst Wood) at a considerable distance from the margin of the basin suggests, but does not demonstrate, that this Facies Association was deposited on an appreciably sloping depositional surface. This slope cannot be quantified on the basis of the occurrence of debris flows, although Sharpe and Nobles (1953 - cited by Collinson 1978) observed debris flows moving on slopes as low as 1.4%.

Laterally constricted conglomerate body: The 'valley fill' conglomerate body at Redhurst Wood West Quarry (Facies 17) conforms much more closely to the existing depositional models for pebbly braided stream deposits (Miall 1978) than do any of the other conglomerate bodies. If this single occurrence is typical, it suggests that well developed, typical patterns of braided stream development were only developed in places where the channel system was incised and laterally constricted.

Mudstone, silty mudstone, and palaeosols: The presence of these Facies implies that, in areas away from the main fluvial channel (or sheet flow) activity, overbank depositional processes comparable to those in Facies Association II obtained. This suggests deposition on a fairly low palaeoslope.

5.6 Possible modern analogues

5.6.1 Distinctive features of tropical alluvial processes

It is proposed that the sediments of Facies Association III represent the deposits of alluvial fans, which formed aprons of conglomeratic sediment along the edges of the basins in which the Etruria Formation was deposited in areas where these were fault bounded. The distinctive features of the Facies Association, which differentiate it from other alluvial fan deposits, are attributed to it having been deposited in the vicinity of small, intermittently uplifted, block faulted source areas, in which patterns of weathering, erosion, and sediment transport were controlled to a large degree by the prevailing tropical climate.

Recent alluvial fans of this type are very poorly described in the literature, discussion of so-called 'humid' alluvial fans being restricted to those developed in temperate regions, and to the sand dominated fan being constructed by the Kosi River in northern India (Gole and Chitale 1966). The latter example may only be representative of one type of tropical alluvial fan, in that it is forming at the foot of a large and rapidly uplifting mountain area (the Himalayas), whose morphogenesis is taking place under non tropical conditions, with a consequent effect on the nature of the alluvium and on the river flow characteristics in the piedmont alluvial fan system. Under such circumstances, alluvial processes typical of the intertropical zone are modified, and the resultant alluvium resembles more that of the temperate zone (Tricart 1972 p.65).

In alluvial systems originating in less rapidly uplifted areas, the following geomorphological processes, characteristic of tropical zones, may be of importance in determining the type and facies of any alluvial fan deposits.

i) In areas of slow uplift, the rapidity and intensity of tropical weathering processes will result in the development of a thick mantle of regolith in the potential sediment source area. In most instances this weathered mantle consists dominantly of kaolinitic clays and iron oxides, with a proportion of sand depending on the parent rock (Tricart 1972 p.65). The bulk of the alluvium derived from such an area thus consists of fines. It is difficult in such circumstances to generate appreciable quantities of coarse detritus. This may, however, be achieved by several processes. If the weathered mantle is stabilised by vegetation, gullying may be encouraged. As the transition between regolith and unweathered bedrock is usually sharp, this results in the creation of a sediment which consists of an admixture of heavily weathered fine grained material and immature fresh boulders and cobbles of bedrock (Krynine 1936, in discussion of Sapper 1935). A similar bimodal mixture of coarse rock fragments and fines is produced by landslides and debris flows stripping the regolith and transporting underlying bedrock on steep valley slopes (Löffler 1977 pp 137-142; Tricart op. cit. p.61). When such material is reworked and transported by fluvial action, it gives rise to extremely coarse grained bedloads (Löffler op. cit. pp 133-135).

ii) The rate of weathering in tropical areas is such that the bedload component of streams may be degraded by chemical weathering more rapidly than by abrasion in transit (Tricart op. cit. pp 62-65). This means that tropical streams may tend to become depleted in gravel bedload very rapidly downstream. The rapidity with which this occurs depends on the rock types in the detritus, basic igneous and argillaceous sedimentary components breaking down rapidly, while quartzitic and other siliceous pebbles persist for a longer time, and thus a longer distance from the source. There may thus be a proportional downstream increase in the proportion of siliceous pebbles.

Although Tricart suggests that this process is universal, its importance is questioned by Löffler (op. cit p. 136) on the basis of his observations in New Guinea, where gravel bedload transport is common.

iii) The general high rainfall and clay dominated nature of tropical alluvium may encourage the occurrence of debris flows. Alluvial fans composed largely of debris flows are recorded from New Guinea by Löffler (op cit pp. 99, 142).

iv) Deep weathering may modify the alluvium after deposition by altering some of the coarse sediment fraction to secondary clay matrix.

5.6.2 Tropical alluvial fans in Papua New Guinea

The only alluvial fans that have been, and are being deposited under these conditions, and which are described in the literature, are in Papua New Guinea. These descriptions are all written from a geomorphological viewpoint, and it is thus only by inference that they can be used as recent analogues for the sediments of the Etruria Formation.

The best described fans are those of the tributaries of the Markham River in eastern New Guinea, especially those of the Leron and Maniang Rivers (Löffler op. cit pp. 87-89) and of the Erap River (Knight 1975). These rivers all drain southwards from the Finisterre/Saruwaged Ranges (part of the Palaeogene New Guinea and New Britain volcanic arc) into the Markham - Ramu graben systems. This area has been a zone of net subsidence since the late Tertiary, and the fans forming at present are depositional landforms.

The best documented fan, from a sedimentological viewpoint, is that of the Erap River. This is a comparatively small fan (Fig 152), having a catchment area of 470 km, a maximum radius of ca 12 km and a gradient of 0.67%. The fan surface is made up of three depositional components: a single, perenially active braided fluvial channel; levée areas adjacent to the channel; and overbank areas, best developed in the lower part of the fan, where deposition of silty and sandy clays takes place.

The perenially active channel is ca. 500m wide, and contains both active and vegetated braid bars. The bedload sediment consists of gravel and sand. At peak flows the maximum discharge of 30m3/sec. Although no flood depths are recorded, the effect of this increased discharge is for bank overtopping to occur, giving rise to levée, crevasse splay, and crevasse channel deposits. Although the main channel is locally incised at the fan head, the presence of levées causes it to be raised by up to 1.70m above the main fan surface. Crevasse channels, once established, may become permanent distributaries, and eventually capture and divert the main channel flow. (This process is described in great detail by Knight op. cit). In the early stages of such distributary development flow in the channel may be lost to groundwater. However, because of the perenially active nature of the main channel, and the comparatively moist subsurface conditions, this loss accounts only for a small proportion of the total water budget of the fan, and as a result sieve deposits and the development of suprafan lobes, as described in a semiarid fan setting by Hooke (1967), are not present. Knight (op. cit) gives no details of depositional processes in the overbank areas occupied by non gravelly sediments.

Further patterns of channel behaviour are described by Löffler (op cit) from the Leron and Maniang Rivers. The main fan stream on the upper part of the Leron fan is entrenched by up to 40m, and within this shallow valley has a well developed braided pattern (Löffler, Plate 45, p. 88). By contrast the Maniang River is not incised, and its main channel consists of a number of braided channels, some of which die out through infiltration, and some of which amalgamate to form larger braided channels. The migration behaviour of these channels appears to be controlled by the topography created by gravel deposition on the fan surface. The channel positions remain stable for a period of years, but then shift abruptly. Channel shifting occurs during overland flood events, in which unconfined flows several kilometres wide deposit gravel sediments over extensive areas, in sheets up to 0.50m thick. These flows occur during a single wet season, and after they have declined, the main braided channel may return to its original course, or may be abandoned in favour of a new course. The Leron River behaves similarly on the lower part of the Leron fan.

The largest of the Markham tributary river fans that has been described is the Leron Fan, which has a radius of 20km (Löffler op cit p. 100). In northern New Guinea larger alluvial fans are present along the northern side of the Sepik Depression, which is the westward continuation of the Markham - Ramu Graben system. These fans have a radius of ca. 40km, and are amalgamated to form a 'bajada' along the southern flanks of the Toricelli and Prince Alexander Mountains. They are described by Reiner and Robbins (1964), Reiner and Mabbutt (1968), and by Löffler (op. cit pp. 102-104). The fans are relict late Pleistocene features. It is uncertain to what extent their present fan morphology is a depositional feature, rather than having resulted from tectonic tilting.

Descriptions of sediment types making up the Sepik Depression fans are largely lacking. However two features are of relevance to the present study. Firstly the fans are described as consisting of "predominantly very fine grain, largely clay and silt [material]" with local interbedded sand and gravel (Löffler op. cit). Secondly, the rivers at present draining the fan surfaces may be considerably incised into the surfaces, by up to 20m, especially in the lower part of the fan surfaces (Reiner and Robbins, and Reiner and Mabbutt op. cit). This incision may be related to late Pleistocene eustatic events, but may equally well have been caused by degradation of the fan toes by the Sepik River.

In the context of the Etruria Formation sediments, the Papua New Guinea alluvial fans, although not described in any great detail, provide some basic information for the erection of a facies model. In particular they show that:

i) extensive alluvial fan formation can take place under humid tropical climatic conditions;

ii) that, provided source area denudation rates are not too high, such fans contain an association of facies in which appreciable quantities of muddy sediment can be deposited;

iii) that the fluvial channel processes on the fan surface have considerable variety, and can produce both sheet and channelised sand and conglomerate bodies of various sizes, within which a wide range of internal structures and organisation may be expected.

5.7 Discussion of Etruria Formation facies in the light of the Papua New Guinea alluvial sediments

5.7.1 Regional considerations

Apart from any similarities in facies which may be discovered between the Papua New Guinea alluvial fans and the Etruria Formation Facies Association III sediments, the Papua New Guinea fans are a suitable starting point for the creation of a facies model for these sediments for three main reasons:

i) palaeomagnetic studies (Turner, pers. comm; Besly and Turner, in press) indicate a tropical equatorial palaeolatitude for the Etruria Formation;

ii) the association of the Formation with coal bearing sediments with a non-seasonal flora implies that humid climatic conditions prevailed at the time of deposition;

iii) the Papua New Guinea fans are forming in a similar tectonic environment (see Chapter 10) and are depositional landforms.

5.7.2 Interpretation of Facies Association III facies

Facies 18 sheet conglomerate bodies

These may be interpreted as the deposits of non incised braided fan channels, such as those described from the Leron and Maniang Rivers. Thinner, less well organised sheets probably represent short lived phases of overbank flow, which deposited the single, waning flow graded units. Where the sheet conglomerate units are thicker, and contain bars and cross bedding, deposition was probably in a braided fan channel in a more stabilized phase. The lateral passage from thick, organized conglomerate sheets into thinner sheets, seen at Utopia Quarry, represents the deposits of a braided stream complex which was fairly stable, but which occasionally moved its course, or, during floods overflowed its banks, thus depositing thinner, less organized sand and conglomerate sheets as sheetflood or levêe deposits on the adjoining overbank area.

Facies 18 conglomerate filled channels

By analogy with the behaviour of the Erap River (Knight 1975), these channels represent various stages in the development of crevasse and avulsion channels from fan channels which had developed levêes. The smallest channels were cut and filled immediately, thus preventing their further development. In one case, at Blockley's Quarry, a multiple channel fill testifies to repeated cutting and filling. In examples where such channels are overlain by or pass into sheet sandstones, as at Redhurst Wood West Quarry, Blockley's Quarry and Wilnecote Quarry, such a crevasse channel probably preceded crevassing of the main fan channel.

Facies 19 sheet sandstones and Facies 20 channel sandstones. These Facies probably bear the same relationship to each as do the sheet and channelised conglomerate Facies, although probably at a more distal position on the fan surface.

<u>Facies 17 laterally constricted conglomerate body.</u> By analogy with the Sepik fans, the 'valley' that this conglomerate body fills was possibly incised as a result of degradation of the fan toes by a river system on an adjoining, downstream, floodplain area. An analogous situation is described by Heward (1978b). The probability of this interpretation is increased by the geographical position of this exposure (Redhurst Wood West Quarry), which, at ca. 20 km from the inferred bounding fault of the basin, marks the furthest progradation of conglomeratic facies into the basin so far encountered.

The apparently very well ordered nature of the braided alluvium probably reflects the development of a well ordered braiding pattern as a result of lateral constraint of the channel. This feature occurs in the incised part of the main Leron fan channel.

It might be argued that, in the event of valley incision in the lower part of a fan as a result of degradation of the fan toe, the base of the resulting valley would be filled with river deposited alluvium, rather than with debris flow deposits, as is the case at Redhurst West. As an alternative model, it may be suggested that the channel or 'valley' incision into the fan surface may have occurred in response to rapid tectonic uplift or tilting of the surface, possibly associated with movement on the bounding faults of the basin. In such a case, a fluvially deposited fill might not be expected in the base of the erosional feature. The spasm of tectonic activity which gave rise to the incision of the channel might also trigger the debris flows and soft sediment intrusions which have been observed in this Facies Association.

Facies 16 debris flow deposits. Debris flow deposits do not occur in the main parts of the well described Papua New Guinea fans. The two types in the Etruria Formation occur in distinct settings. The sandy debris flows only occur in quarries very near the bounding fault of the basin (at Blockley's and Donnington Wood quarries). These may represent materially derived very locally from the source area, and deposited on the fan surface, away from major fluvial channels, by debris flows started by landslides. The large muddy debris flow filling the bottom of the Redhurst Wood Quarry 'valley' must, on account of its distance from the source area, represent a major event in the hinterland area.

The inferred heavy vegetation in the overbank areas of the fans might be expected to have slowed or stopped debris flows within short distances. It is thus likely that flows which travelled long distances occupied scars formed by previous landslide or flood events.

Overbank deposits. Away from areas of fluvial channel activity, fine grained overbank deposits were able to accumulate in a manner similar to those in Facies Association IIA. It is unlikely that ponding of water occurred, and sedimentation may have taken place partly by plant filtering, and partly by the loss of water to the groundwater table. In view of the fairly high slope and probably good drainage, only evolved palaeosol types are likely to have formed. The comparative rarity of well developed palaeosols in the Facies Association probably reflects a propensity for erosion of the fan surface during periods of high discharge.

The occasional presence of siderite or of coaly horizons in Association III overbank sediments probably indicates deposition very near the level of the transition from fan into alluvial plain sediments.

5.7.3 Depositional model

The sediments of this Facies Association are regarded as having been deposited on mud dominated alluvial fans, sourced from fault bounded uplifted blocks, and passing laterally into alluvial plain sediments of Facies Association II. The elements of the model are reconstructed schematically in a block diagram in Fig. 151.

The size of the fans varied. In Warwickshire the fans probably never reached radii of more than 10-15km, while in South Staffordshire and Mid Staffordshire the fans had a maximum extent of ca 20km. This variation in size probably reflects the contrasting lithological composition of the conglomerates, those in Warwickshire being dominated by easily degraded shale clasts, while those in South Staffordshire were composed dominantly of quartzite (see 7.3 and 10.1). The size and extent of fans in the Mid Shropshire area and to the south west of Birmingham is not known, owing to insufficient data.

CHAPTER 6

Relationships between Facies Associations

6.1 Facies Associations I and II

6.1.1 Vertical sequences

The relationship in vertical sections between Facies Associations I and II has been observed at the base of the Etruria Formation. Two distinct sequential developments are present. The boundary between the Coal Measures and the Etruria Formation often consists of a transitional sequence in which grey beds and red beds are intercalated, the red beds becoming predominant upward. Alternatively there is sometimes a sharp boundary between the two Formations. The relationship between Facies Associations and lithostratigraphy and the precise lithostratigraphic nomenclature is discussed in Chapter 9.

The sedimentology of the topmost Coal Measures underlying the Etruria Formation has not been studied. It is evident, however, that in all areas, irrespective of the diachronous nature of the base of the Etruria Formation, the topmost Coal Measures are of alluvial origin. The facies are identical to those in Facies Association I, and are characterised by a predominance of rooted lithologies and palaeosols (unpublished NCB borehole data; E.L. Boardman, pers. comm.; see Chapter 9).

Sedimentary Facies in the grey - red bed transition

i) Intercalated sequences

Where the base of the Etruria Formation consists of grey and red intercalations, this zone is usually between 10 and 20 m thick, although in North Staffordshire intercalations of grey and red sediment occur over thicknesses of up to 100 m. Within the transitional sequence, red bed intercalations occur on two scales. On the smaller scale, intercalations of red pigmented mudstone have thicknesses of 0.50 to 2 m, separated by comparable thicknesses of organic rich mudstone. On the larger scale, groups of such thin intercalations may collectively form thicker red bed intercalations, up to 10 m thick, which again are separated by similar thicknesses of organic rich mudstone (Fig. 152).

Smaller scale grey/red intercalations are exemplified by the section at the base of the Playground No. 8 borehole (Fig. 153). The basal 3 m of this section consists of typical Facies Association I sediments:- laminated grey mudstone (Facies 1) and polyphase palaeosols of Palaeosol Type 1 (seat earths and thin coals). Unfortunately the borehole was terminated at the base of the section illustrated in Fig. 153; it may however be assumed, from the stratigraphic control provided by underground mine workings in the vicinity, that this 3 m section truly marks the top at the Coal Measure sequence.

The first appearance of red pigment in the Facies Association I sequence is found in the lower horizon of the seat-earth palaeosols. These are palaeosols of Palaeosol Type 2 (? gleyed alluvial soils: see 3.3.2); the red horizon indicates the development of a contemporaneously oxidized zone in the soil. Higher in the sequence the red intercalations become thicker, extending into the unrooted sediment below the palaeosols. Ultimately, at 10 to 12 m in the section, a thicker red intercalation includes a palaeosol, which, from its texture and its content of siderite and carbonaceous material, must originally have been of Type 1 or 2. Its pervasive red pigmentation shows, however, that it has been oxidized after burial to form a Type 3 palaeosol. Immediately overlying this palaeosol, at 12 to 14 m in the section is an evolved palaeosol of Type 4. From the predominant red colour, and content of Type 3 and Type 4 palaeosols, the section above 8 m evidently consists of sediments belonging to Facies Association II.

Similar sequences of red intercalations have been observed in the Rosemary Hill boreholes, although, in this case, the progressive increase in red pigment is not so apparent. A sequence containing Type 2 palaeosols with a red horizon is here sharply overlain by a completely red sequence, without such an obvious transition between the two.

Unfortunately no cored sequence through the larger scale (c. 10 m. thick) red bed intercalations has been observed during the present study. Core descriptions from NCB boreholes drilled between 1970 and 1976 (unpublished NCB records; E.L. Boardman personal communication) suggest that each of these intercalations contains a sequence similar to that in the Playground borehole section, consisting of Facies Association I mudstones containing Type 2 palaeosols with reddened horizons, passing upwards into more typical Facies Association II mudstones containing evolved palaeosols. From the NCB data it may be inferred that the thicker red intercalations have sharp, well defined tops, and gradational bases (see 9.3).

These inferences are to some extent borne out by the section exposed in the deep trial pit at Ibstock Himley quarry, (Fig. 154). Here a 7.5 m thick sequence of Facies Association II red beds is overlain by an intercalation of grey Facies Association I sediments. At the base of the red beds, between 0 and 1 m in the section, and at about 3 m, the red pigmentation is weak, and the sediment contains oxidized plant debris and sphaerosiderite. At these levels there are thin (0.30 m) intercalations of grey, slightly carbonaceous sediment. These features suggest that the base of the section is composed of post depositionally oxidized grey organic rich sediment. Higher in the basal red bed interval, a channel sandstone is overlain by red mudstone containing an evolved palaeosol profile. This is overlain successively by a slightly brown pigmented seat-earth (Type 1 palaeosol) and a thin coal seam, in turn overlain by grey organic rich sediments.

ii) Sequences with a sharp contact between red and grey beds

A clearly defined boundary between grey, organic rich, Coal Measures and Etruria Formation red beds has been observed in the cored sections of Kibblestone and Allotment No. 1 boreholes. Although in detail these sections differ slightly, they share the following features:
i) the basal part of the Etruria Formation contains red pigmented palaeosols of Types 2 and 3, and mudstones with oxidized plant remains;
ii) the basal part of the red bed sequence contains units of silty mudstone and siltstones, the lowest appearance of red pigment taking the form of a diffuse mottling of red in one of these coarser grained horizons.

6.1.2 Form and nature of occurrence of larger scale red/grey bed intercalations

In areas where the National Coal Board has drilled closely spaced boreholes (2 - 5 km apart) it is possible to identify the three dimensional geometry of the larger scale (2 - 10 m thickness) intercalations of red and coal bearing sediments. The best documented example is in the southern part of the North Staffordshire coalfield. There, in the space of ca. 15 km, the Blackband Formation passes laterally into the Etruria Formation by means of such 'larger scale' interdigitation of grey sediments, predominantly of Facies Association I (E.L. Boardman personal communication) and red beds which are predominantly of Facies Association II.

In this transition the red beds occur in tongues between the coal seam marker horizons (Boardman 1978). The tongues have tops which are concordant with the coals, and bases which are discordant. The red intercalations thicken in the direction in which the sequence as a whole thins, as judged by the spacing of the marker horizons. In this direction the red intercalations coalesce to form a continuous Etruria Formation sequence (Fig. 155).

In North Staffordshire there is sufficient borehole data for the red intercalations and the intervening areally extensive coal seam marker horizons to be mapped (Figs. 156-162). Although there are considerable uncertainties, especially in the stratigraphic interpretation of some of the older shaft and borehole data, three clear observations may be made.

i) The red intercalations formed prior to the deposition of the Bassey Mine coal seam occupied, at their greatest extent, an area in the southwest of the coalfield bounded by a more or less straight line running northwest to southeast from Bowsey Wood borehole to Groundslow borehole.

ii) The red intercalations formed after deposition of the Bassey Mine coal seam occupied progressively larger areas of the coalfield. In the red intercalations associated with the E.1, E.2 and E.3 groups of seams, formed after deposition of the Blackband coal seam and before the onset of the continuous red bed sequence, this progressive extension culminated in the whole of the coalfield area (up to the present outcrop of the Etruria Formation) being occupied by red beds.

The red intercalations in the Blackband Formation, between the Bassey Mine and Blackband coal seams, have a markedly lobate plan form. This is most pronounced in the red intercalations underlying the Hoo Cannel and Red Shagg seams (Figs. 158, 161). The area around Silverdale Colliery appears to have been preferentially occupied by coal bearing sediments during the formation of these red bed lobes.

iii) The coal seams in the Blackband Formation, occurring in the grey sediments separating the red bed wedges, all occupy a similar area, extending to a line running south eastward from Bowsey Wood borehole to Meafordhall No. 1 borehole, and thence to a less well defined line running north eastwards. There are two exceptions to this generalization. The Blackband coal seam has its maximum extent considerably to the north and east of the earlier seams; and the Hoo Cannel coal seam is absent in the Peacock's Lane and Hanchurch boreholes. This area, in which the seam is not developed, corresponds approximately to the centre of the red bed lobe developed in the sediments immediately underlying the coal seam.

In other parts of the area under study there is insufficient borehole data for the pattern of larger scale intercalations between red and grey beds to be traced in such detail. This lack of data results largely from the coring points in most boreholes outside North

Staffordshire having been well below the base of the Etruria Formation. It is, however, evident that a similar pattern of large scale intercalation is present elsewhere. Notable single vertical sequences of this type having been encountered in Mid Staffordshire in Devil's Dumble borehole (Fig. 188) and in Whittington Heath and Bowman's Bridge boreholes (Fig. 191). In both cases a diachronous relationship similar to that found in North Staffordshire may be inferred. Also, in the South Stafforshire and Wyre Forest areas the red beds found at the base of the Westphalian succession consist mainly of intercalations, up to 10 m thick, of red and grey, coal bearing sediments (Fig. 206). Core descriptions from the Alveley boreholes (NCB unpublished data) make it evident that these intercalated sediments consist of rooted claystones and palaeosols, referrable to Facies Association I and IIA.

Because of the very uneven data on the nature of the Coal Measure/ Etruria Formation transition, it is difficult to generalize as to the geographic distribution of transitional sequences characterised by the larger scale, repeated intercalation between grey and red beds. However, such data as are available (see Chapter 9) suggest that sequences of this type are commonest in areas where the diachronous relationship between grey and red beds is very pronounced. These areas coincide with marked changes in the thickness and succession in the underlying coal-measures, which in turn reflect the activity of basement structures. The sedimentological and tectonic interpretation of these larger scale intercalations is discussed in 6.1.5.

6.1.3 Inferred relationships of palaeosols within red bed intercalations

Unfortunately no observations have been made of the lateral variation in the types of palaeosol present within a single red bed intercalation. This relationship can, however, be inferred from the following observations, derived partly from unpublished NCB core descriptions.

i) Where red intercalations are thinly developed, the red pigment occurs as red mottles in type 2 (gleyed alluvial) palaeosols.

ii) E.L. Boardman (personal communication) has observed that, although there are no red beds in the Blackband Formation at Mitchell's Wood opencast site, North Staffordshire, the palaeosols immediately below the Red Mine coal seam contain exceptionally abundant sphaerosiderite. This occurrence correlates with a red bed intercalation which is extensively developed elsewhere in North Staffordshire. Sphaerosiderite in seat-earths is often considered to indicate slightly improved drainage conditions (Elliot 1968).

iii) The discordant bases of the red intercalations suggest that they might have been formed by facies progradation. The red intercalations seem to be characterised by an upward transition in the types of palaeosol present, from organic rich grey alluvial palaeosols (type 1) through red mottled gleyed alluvial palaeosols (type 2) to well drained, evolved, palaeosols (type 4).

If progradation was involved in the formation of these intercalations, it must have involved the progressive extension of well drained swamp and floodplain facies into the poorly drained swamp area in which the organic rich sediments were being deposited.

These observations, especially the last, suggest that each red intercalation was characterized at the time of its deposition by a lateral sequence of palaeosols which had characteristics of progressively better drainage in the direction of thickening of the intercalation (Fig. 163).

The discordant bases of the red intercalations are probably, in many cases, overprinted by penetrative post depositional oxidation, associated with the development of the well drained soils formed at the end of the phase of progradation. This may lead to the presence of post depositionally oxidized palaeosols in the vertical sequence, as found in the Playgound No. 8 section (Fig. 153).

6.1.4 Occurrence of post depositionally oxidized alluvial palaeosols (type 3) in Facies Association II

In addition to the occurrence of intercalated sequences of grey and red sediments of Facies Associations I and II, the presence in Facies Association II of post depositionally oxidized alluvial palaeosols (palaeosol type 3: described in 4.3.2) sheds further light on the relationship between the two Facies Associations. Before their organic content was destroyed during penetrative oxidation, these palaeosols, and possibly parts of the associated sedimentary sequence, resembled sediments of Facies Association I, and were presumably deposited under similar conditions.

The only two complete sections through the Etruria Formation that have been observed are those cored in the Playground No. 8 and Allotment

No. 1 boreholes. Unfortunately the Formation is truncated by the sub-Halesowen Formation unconformity in both of these boreholes and in the former it contains the anomalous sequence composed entirely of evolved palaeosols described in 4.6.1. In the Allotment No. 1 section, type 3 palaeosols only occur at the base of the sequence.

Elsewhere, type 3 palaeosols and polyphase palaeosols are concentrated near the base of the sequence. They are, for instance, only present in South Staffordshire at Ibstock Himley Quarry, which is in the basal 50 m of the Etruria Formation. In Warwickshire the coal seam and the siderite bearing palaeosols at Wilnecote Quarry occur within the basal 25 m. of the Formation and such features have not be observed in the higher parts of the Formation exposed in the nearby Stoneware and Kingsbury Quarries.

Post-depositionally oxidized organic rich sediments are more extensively developed in North Staffordshire. They are well represented in the lower part of the Formation, cored in Kibblestonne borehole. The lowest horizons at present exposed, at Rosemary Hill Quarry, contain a post-depositionally oxidized ironstone (Facies 12) and a polyphase palaeosol. In the middle of the Formation there are abundant polyphase and type 3 palaeosols, and oxidized ironstones, exposed in the Manor, Chesterton, and Bentley Quarries. In the middle of the Formation Malkin (1961) correlated a thin coal seam (the 'Chesterton Top Red Mine seam') between exposures at Chesterton and the shafts of Parkhouse and Wolstanton collieries, and several other organic rich grey horizons occurring in the same area. In the upper part of the Formation, exposed at Metallic Tileries and Spoutfield

Quarries, type 3 palaeosols are not prominent, although one example has been observed at Metallic Tileries.

6.1.5 Summary of relationships between Facies Associations I and II: sedimentological and tectonic interpretation

The relationship between these two Facies Associations is obviously complex. The most important features of the sequences where the two are intercalated are as follows.

i) Intercalations occur on two distinct scales, a 'smaller' scale of less than 2 m thickness, and a 'larger' scale of up to 10 m thickness. The 'larger' scale intercalations consist in part of groups of 'smaller' scale intercalations.

ii) The smaller scale intercalations are ubiquitous at the grey coal measure to red bed transition. The larger scale intercalations are probably only present near tectonically active hinge lines and basin margins.

iii) Organic rich intervals, in the form of alluvial palaeosols, coal seams and ironstones were deposited in sequences now composed of Facies Association II, especially in the lower part of what is now the continuous red bed sequence. These intervals were destroyed by post-burial oxidation.

iv) In the transition sequence between coal measures and the Etruria Formation, red intercalations of both scales are separated by extensive, correlatable organic rich horizons. Some organic rich horizons within the Etruria Formation are also extensive and correlatable.

v) Within the 'thicker' intercalations, and at the base of the continuous red bed sequence, there is a tendency for the types of palaeosol present to change upwards, from poorly drained to well drained types.

vi) Where the plan form of 'larger' red intercalations can be mapped, some of these intercalations have a nearly rectilinear boundary with laterally equivalent grey strata, while other have a lobate form.

The tendency for the palaesols to evolve from poorly drained to well drained types in the transitional sequence between the Coal measures and the Etruria Formations suggests that the change from organic rich to red bed sedimentation was caused by a change in drainage conditions in the overbank areas of the alluvial systems involved. The poorly drained backswamps suitable for the preservation of large amounts of organic matter were replaced by mainly well drained floodplains, on which evolved soil profiles developed.

Where the transition from Coal Measures into Etruria Formation consists of one thin transitional sequence (as in Playground No. 8 borehole - Fig. 153), this transition from poorly to well drained floodplain conditions probably resulted from a single phase of progradation of a topographically slightly elevated alluvial system over the pre-existing poorly drained backswamp area. This progradation may have resulted from an increase in clastic input, related to uplift in the source areas; or may have resulted from differential block movement affecting the subsidence rate in an entire depositional area (see 10.2), and causing a rapid change in drainage conditions over a wide area.

In the Playground No. 8 section (Fig. 153), the red mottled, type 2, palaeosols formed during the first phase of this progradation are overlain by a continuous red bed sequence. This contains, at 9-11 m, a type 3, post depositionally oxidized alluvial palaeosol, which is immediately overlain, at 11-14 m, by an evolved palaeosol. In this case the alluvial palaeosol was almost certainly oxidized during the period of improved drainage which accompanied the formation of the evolved soil profile. There is evidence higher in the Playground sequence for similar formation and post burial oxidation of alluvial palaeosols. Such soils probably formed in localised poorly drained hollows on the alluvial plain surface. These probably formed by a combination of differential subsidence and remoteness from the active alluvial channels which supplied sediment to the floodplain. Most alluvial and polyphase palaeosols found in the Facies Association II red bed succession probably formed in this way. Sediments with a high initial organic content deposited in such a setting must have stood a high chance of post depositional oxidation, in the event of a well drained evolved soil being formed on the subsequently deposited alluvium.

Where 'larger scale' intercalations of red and grey beds are present, it is evident that the controls on drainage responsible for the facies change were more complex. North Staffordshire provides the only sufficiently documented example (Figs 156 - 162). The virtually identical maximum extent of the coal seams, and thus minimum extent of red beds, in the Blackband Formation suggests that the maximum extent of peat formation at the time of deposition was tectonically controlled, by differential subsidence. This being the case, it would seem reasonable to suggest that the rectilinear limit to the red bed intercalations formed prior to the Bassey Mine coal seam was also caused by differential subsidence. This gave rise to intermittent periods of good drainage on the more slowly subsiding side of an inferred tectonic flexure.

The tendency for red bed intercalations to thicker and amalgamate in the directions in which the Coal Measure sequence as a whole thins demonstrates that the formation and preservation of red bed sequences must, to an extent, have been controlled by subsidence. It can however to demonstrated in the North Staffordshire area that the good drainage conditions needed to initiate red bed formation resulted, at least partly, from the creation of topography by sedimentary progradation. There are two lines of evidence leading to this conclusion.

i) The lobate form of the red bed intercalations in the Blackband Formation bears no relationship to the pattern of basin subsidence obtained from the maxiumum extent of coal seams in the Foramtion.

ii) The form of the lobate red bed intercalations is discordant with the regional isopach pattern (Figs, 164, 165), whereas, if differential subsidence were the sole control, some degree of concordance might be expected. This argument should not be overstressed. The only reasonably reliable isopach maps are of interval preceding the development of red intercalations, and of the thick interval which includes the whole of the Backband and Etruria Formations. The isopach pattern of these intervals may have been affected by subsidence effects post-dating the formation of the Backband Formation red bed intercalations. Both maps (Fig. 164, 165) are made on the basis of very scattered data.

The lobate form of the red intercalations suggest that red bed forming conditions prograded into the North Stafforshire depocentre in the form of fan shaped areas, sourced from a tectonically controlled hinge (Fig. 166). Slight evidence for these red bed lobes having had a slight topography is found in the lobe underlying the Hoo Cannel coal seam, which is anomalously absent in the Hanchurch and Peacock's Lane boreholes (Fig. 158). These are situated in the centre of the mapped area of the red bed lobe, fairly near the inferred tectonic hinge. The absence of the seam in this area suggests that peat formation did not occur over the topographically higher parts of this lobe, where slightly better drainage conditions still prevailed after the abandonment of the lobe and the reversion to poorly drained floodbasin swamp conditions.

In the absence of specific data regarding the distribution of Facies in these lobate red bed intercalations, their interpretation must remain speculative. It is suggested that they may represent a type of "terminal fan" (Mukerji 1976), formed within a zone of alluvial deposition by the progradation of a well drained floodplain into a tectonically controlled basinal area, in which the alluvium was dominantly poorly drained. Each such "fan" would probably have been sourced by the debouchment of a single fluvial channel complex into the swamp zone. Obviously, the facies present in such "fans" are very different to those described by Mukerji (op cit.).

Between phases of lobe progradation poorly drained conditions were re-established, and Facies Association I sediments deposited up to the 'tectonic hinge'. This may have resulted from migration of the clastic input beyond the area mapped. However the presence of regionally extensive coals in these grey bed intercalations suggest a large scale abandonment of clastic supply, which in turn suggests a regional subsidence control on the cessation of red bed lobe formation. With time sedimentary progradation became the dominant force, with successive red bed lobes occupying more of the North Staffordshire area. Extensive and correlatable coal seams in Facies Association I intercalations did, however, continue to form long after red bed deposition had become prevalent, giving rise to the thin groups of seams above the Blackband coal seam (E.1 to E.3, see 9.3) and, in the middle of the Etruria Formation, to the 'Chesterton Top Red' seam (see 9.3).

6.2 Facies Associations II and III

There is little information regarding the relationship between these Facies Associations. In the only exposure which undoubtedly contains sediments of both Associations, at Redhurst Wood West Quarry, sediments of Association III overlie those of Association II conformably in the west side of the quarry, and are channeled into them in the east side (Figs 120, 148). In the Kingswinford area a large scale interbedding of the two Facies Associations is apparent. The basal part of the Etruria Formation, exposed in the upper part of the Ibstock Himley Quarry (section 2) contains thin granule and pebble conglomerate sheets which may be distal representatives of the Facies 18 conglomerate sheets. Higher in the Formation in the same area Facies Association IIB point bar deposits are present at Himley Wood Quarry, while at the top of the Formation in Ketley Quarry Facies Association III sediments are present including thick sand sheets and channels of Facies 19 and 20.

This meagre evidence suggests that sediments of Facies Associations II and III are interdigitated on a fairly large scale. It is also possible that such an interdigitation may locally exist between Facies Associations I and III. An immature conglomeratic sandstone sheet, which may have been similar to those of Facies 18, was found in 1978, interbedded in Westphalian A grey Coal Measures at the Coalpit Lane open cast site in Telford (R. Todhunter, personal communication). It was unfortunately not observed by the present author.

CHAPTER 7

Petrography of Etruria Formation sediments

7.1 Fine grained sediments

The composition of the fine grained rocks in the Etruria Formation has been studied by Besly and Turner (in press), Holdridge (1959), and Keeling and Holdridge (1962). The following brief description is mainly a compilation of these sources, but ignores the clay mineral determinations of Keeling and Holdridge, as they were obtained by an inaccurate, non quantitative technique.

Besly and Turner have examined representative samples collected from all the stratigraphic levels in the Etruria Formation, in all areas of outcrop. The most notable feature of this study is the extreme monotony of the mineralogy of the Etruria Formation mudstones. X-ray diffraction shows that they consist of disordered kaolinite, quartz and haematite, with minor amounts of illite, muscovite, siderite and goethite.

The principal components of the mudstones are disordered kaolinite and quartz. The occurrence of these minerals is the same in both overbank mudstones (Facies 8 and 9) and in palaeosols. The interbedded grey, coal bearing mudstones at Ibstock Himley Quarry have a similar composition, as do other Westphalian coal bearing sequences, described by Ashley and Pearson (1978), and Pearson (1975). In all of these cases quartz occurs as disseminated silt and occasionally sand sized particles. The similarity of composition between both organic rich and red sediments suggests that both kaolinite and quartz are of detrital origin.

Muscovite occurs in all areas in trace quantities, but is abundant in the southern part of the South Staffordshire area. This strongly suggests a detrital origin. Illite is locally present throughout the area, but is characterised by an extremely ragged and poorly defined XRD peak, which possibly indicates that this clay mineral is only present in the form of degraded and strongly altered mica. Holdridge (1959) also infers the presence of trace quantities of detrital rutile, to account for the low but constant titania content (0.5 - 1.4%) determined by chemical analysis.

The remaining minerals determined in the mudstones are all of diagenetic origin. They are discussed in 8.2.

7.2 Sandstones and conglomerates

7.2.1 Introduction

In the present study it has proved impractical to undertake a systematic petrographic analysis of the sandstones in the Etruria Formation. The nature of these rocks renders the making of good quality thin sections difficult, owing to the abundance of clay matrix and the ease with which it is washed out during preparation. In addition, the poor sorting, with occasional large grain size and the great variety of lithic constituents would necessitate the preparation of a very large number of thin sections to allow a statistically representative study to be made.

In this section, therefore, only a brief qualitative account is given of the detrital petrography, sufficient for the purpose of discussing the lithostratigraphic and palaeogeographical aspects of the Etruria Formation.

The conglomerates and sandstones in the Westphalian of Central England fall into two distinctive compositional groups. One group is of local derivation, containing a varied suite of rock fragments and little feldspar. The second contains abundant feldspar, and a limited variety of rock fragments. The sandstones in the Etruria Formation belong to the former group, as, locally, do those in the grey Coal Measures. The majority of Coal Measure sandstones, however, belong to the second group, which is petrographically similar to the northerly derived Namurian sandstones of the Pennine area.

7.2.2 Previous petrographic descriptions of Etruria Formation sandstones

The petrography of the characteristic compositionally and texturally immature sandstones in the Etruria Formation has attracted the attention of workers from Murchison onwards (see 1.3.2). After it had been recognised that these rocks were not of volcanic origin, their distinctive immaturity was noted by several authors. Gibson (1901) recorded "pebbles of ..... Cambro-Silurian rocks with .... quartz, feldspar, greenstone, porphyry, together with fragments of Wenlock Shale and Carboniferous Limestone" in Denbighshire, and coarse breccias with large angular clasts of Cambrian Lickey Quartzite in South Staffordshire. Barrow (in Gibson 1901) regarded much of the material, especially the matrix, as being of igneous origin, and (in Gibson 1905 p.131) recognised fragments of rhyolitic lava. Robertson (1931) emphasized the content of igneous material to support his view that the Etruria 'Marl' was the product of denudation of contemporaneous basalts. Whitehead and Eastwood (1927) added vein quartz and Silurian sandstones to the list of detrital components occuring in South Staffordshire.

The most comprehensive petrographic study has been that of Williamson (1946), made on samples from North Staffordshire. He listed the following components: quartz, of sedimentary, metamorphic, and hydrothermal derivation; very rare feldspar; rock fragments, including granophyre, rhyolite, andesite, lithic tuff, and silicified vitric tuff.

7.2.3 Detrital composition of Etruria Formation Sandstones and conglomerates

i) Quartz. Quartz grains make up not more than 60% of the sandstones. Both monocrystalline and polycrystalline grains are present. Polycrystalline grains larger than -1Ø are regarded as rock fragments. Textural descriptions of polycrystalline grains are those proposed by Young (1976).

Typically, in a fairly well sorted sandstone of medium or smaller grain size, the majority of quartz grains are monocrystalline. Polycrystalline grains are more abundant in coarse sandstones. The monocrystalline grains show either non undulose or only poorly developed undulose extinction. Deformation lamellae are rare. The polycrystalline grains commonly show polygonised textures (Fig 167a) and elongation of the original quartz grains within the grain (Fig 167b). Less frequently, grains containing polyhedral textures and showing recrystallisation are present (Fig 167c).

The polycrystalline grain textures compare closely with those illustrated by Young (1976), and according to his criteria, indicate derivation from regionally metamorphosed sediments of low to moderate grade.

In one sample from North Staffordshire, quartz grains occur which have very well defined, non-undulatory extinction, are nearly euhedral, and lack inclusions. These are probably of volcanic origin (Blatt and Christie 1963), and strongly resemble the quartz phenocrysts found in rhyolite rock fragments observed in the same specimen (see iib) below).

In most other quartz grains abundant inclusions are present. Identifiable types include acicular inclusions of rutite and chlorite, which have been found in samples from North Staffordshire. Quartz containing such inclusions has been equated, respectively, with derivation from granites (Turner and Whitaker 1976) and from hydrothermal veins (Blatt _et al._ 1972).

ii) Rock fragments. The most remarkable aspect of the petrography of the Etruria Formation sandstones is the high content and great diversity of rock fragments present in them. The content of lithic material varies from ca. 10% in the well sorted, fine to medium grained sandstones present in North and Mid Staffordshire, to ca. 70% (i.e. all of the fragmental material) in the poorly sorted coarse sandstones and conglomerates occurring near the margins of the depositional basin in Warwickshire and Shropshire. Rock fragments of sedimentary, igneous, metamorphic, and hydrothermal origin have been observed.

a) Four distinct sedimentary rock fragment types have been recognised: quartzite, sandstone, siltstone and shale.

Quartzite is the dominant lithic component in many sandstones, especially in the southern part of the study area (see 10.1.1). The surfaces of the fragments are often deeply coloured by the surrounding matrix, either red by haematite or green/grey by chlorite. When broken the fresh quartzite is a creamy white colour, and shows a high reflectivity from crystal faces of quartz overgrowths developed on the quartz grains. In thin section, the rock consists of extremely well sorted and rounded monocrystalline, non undulatory quartz grains, heavily cemented by syntaxial overgrowths, and showing only slight suturing.

Other sandstone and siltstone clasts occur less frequently than the quartzite, and have only been observed in South Staffordshire. The sandstone is fine grained and well sorted. Both sandstone and siltstone have a patchy calcite cement, and contain fairly abundant clay matrix, which has usually been haematized.

Fragments of indurated shale are extremely common in some areas (see 10.1.1). They occur as chips, varying between 1cm. and 8cm. in diameter, and between 0.1cm. and 0.6cm. in thickness.

b) Five main types of igneous rock fragment have been recognised: porphyritic lavas or very fine grained intrusives of basic composition; similar lavas or intrusives of intermediate composition; welded acid tuffs; other indeterminate ?tuffs; and granite.

Fragments of porphyritic basic lava are fairly common in South Staffordshire. They are commonly rather small and extensively altered. In most cases their characteristic feature is a relict texture of lath shaped plagioclase phenocrysts. The texture is often preserved by the preferential replacement of either phenocrysts or groundmass by haematite (Fig 168). In these cases the remaining non-haematized components of the rock are replaced by clay and/or calcite. Because of the extensive alteration, these fragments are identified as having been of basic composition solely on the basis of the 'basaltic' textures exhibited by the replaced plagioclase phenocrysts.

Intermediate lava fragments contain euhedral orthoclase phenocrysts surrounded by a felted groundmass composed of fine grained plagioclase laths, of an opaque oxide phase, and of clay. Although the phenocrysts are sometimes comparatively unaltered, none have been observed which have been large enough to determine the feldspar composition. Williamson (1946) recorded the presence of andesite fragments in North Staffordshire, but his descriptions suggest that the grains were extensively altered. As with the basic lava fragments, intermediate lavas have only been observed in South Staffordshire in the present study.

Grains of welded tuff have been observed, in samples from North and South Staffordshire. They are very similar to rock fragments described by both Barrow (in Gibson 1905 p. 131) and Williamson (1946). The example from North Staffordshire consists of a microcrystalline quartz groundmass, in which flattened and distorted shards are clearly visible when viewed in plane polarised light (Fig 169a). The shards are pseudomorphed by areas of slightly coarser grained quartz. The ground mass also encloses a subhedral, embayed phenocryst of inclusion free quartz. In other examples (Fig 169b) glass shards, very similar in appearance to those illustrated by Scholle (1979 p. 34), are pseudomorphed by fine grained quartz and enclosed by a groundmass of microcrystalline quartz and haematite. Many other microcrystalline quartz grains have been observed, in which areas of slightly coarser grain size may be pseudomorphed features of acid lavas or tuffs. In one example, very well preserved perlitic texture has been observed.

Other types of rock fragments of probable tuffaceous origin are more abundant, but less distinctive. A commonly observed type consists of a nearly homogeneous mass of fine grained microcrystalline quartz, with small crystalline quartz grains, occasional patches of chlorite, and in some examples, small euhedral pseudomorphed plagioclase laths. Fragments of this type are tentatively identified as tuffs on the basis of their textural resemblance in thin section to some rock types in the Uriconian tuffs in the Wrekin area of Shropshire (see 7.3.2). The other common fragment type interpreted as a tuff consists of microcrystalline quartz in a groundmass of very fine grained dark material. This presumed tuff was probably more basic in composition than the more easily recognised acid tuffs.

Plutonic igneous rocks are not abundant among the lithic fragments, owing to their large crystal size, the ease with which they are disaggregated, and the ease with which feldspars and ferromagnesian minerals appear to have been altered during the diagenesis of the Etruria Formation (see 8.2). In one sample from North Staffordshire very fresh granules of 'granite' occur, consisting of quartz, plagioclase, and minor orthoclase. Many of the grains are disaggregated. The feldspars are extensively replaced by clay and carbonates. Wills (1950) recorded the presence of granite fragments in Warwickshire, and Williamson (1946) noted the occurrence of granophyre in North Staffordshire. A variety of locally derived plutonic rock fragments have been found in the Coalbrookdale area by Hamblin (personal communication).

c) Metamorphic rock fragments, like plutonic rock fragments, are not always easily recognisable in sand and granule sized grains, unless the grains possess a distinctive polycrystalline texture (Blatt 1967; Young 1976). Most metamorphic rock fragments containing pronounced foliation and/or ferromagnesian minerals disintegrate rapidly during transport (Cameron and Blatt 1971), and, as a result, only quartz rich types have a significant preservation potential.

Two metamorphic rock types have been recognised in the Etruria sandstones, other than those represented by the polycrystalline grain types already described. One type is a meta-quartzite, in which the quartz grains have intricately sutured boundaries and show a slight preferential flattening which might reflect a weak foliation (Fig 170). The second type consists of equigranular, unsutured quartz grains, traversed by veins of fine grained quartz. This type may be of hydrothermal origin.

d) Hydrothermal rock types are represented by large pebbles of vein quartz found in conglomerates in the Coalbrookdale area.

In addition to the identifiable detrital rock fragments listed above, there are two very distinctive detrital grain types of unknown origin (Fig. 171). The first of these consists of microcrystalline quartz grains (Fig. 171a, b). The second consists of grains which have been completely pseudomorphed by chlorite, or by aggregates of chlorite and of a carbonate mineral.

The microcrystalline quartz grains are occasionally partially recrystallized to give a mosaic of mega- and microquartz (terminology of Scholle 1979). They strongly resemble the chert grains illustrated by Scholle, but also resemble the fine grained recrystallized groundmass of grains which are of undoubled acid igneous origin. If characteristic relict textures are not preserved, it is impossible to identify the parentage of these grains.

The chlorite replaced rock fragments are often recognisable as framework grains (Fig 171a), but in many cases have undergone deformation during compaction, and now form part of the matrix (Fig 171c). These deformed clay aggregates fulfil all of the criteria of Dickinson (1970) for the recognition of modified detrital grains. Deformed lithic material extends in wisps between undeformed, rigid grains; the internal fabric of the fragments conforms to the margins of the enclosing grains; and large 'matrix' filled gaps occur in the framework (fig. 171c).

Williamson (1946) identified the chlorite mineral as chamosite. This identification has not been eliminated by the limited XRD work carried out during the present study. If correct this occurrence, of a mineral usually formed in shallow marine environments, is most unusual in the alluvial Etruria Formation sediments, and must reflect unusual conditions of early diagenesis.

Little can be said regarding the original composition of the rock fragments replaced by chlorite. Galloway (1974) regarded all chlorite-replaced fragments in the Tertiary volcanogenic sediments in the north east Pacific as being of basic or intermediate volcanic origin. This view is supported by the abundance of chlorite in Cretaceous and Tertiary volcanogenic alluvium in Alberta (Carrigy and Mellon 1964), and echoed by Dickinson (1970).

The only direct evidence of an igneous origin for the chlorite-replaced grains in the Etruria sandstones is found in one specimen (Fig 172), where chlorite appears to pseudomorph an aggregate of ferromagnesian grains. An igneous origin for the majority of these grains is, however, favoured, as it is difficult to envisage any other potential source for such a large volume of detrital material rich in iron and in silicates.

iii) Feldspars

Feldspars are uncommon in Etruria sandstones, having been found at only three localities. In two of these cases they occur in sandstones containing a high proportion of feldspathic lithic fragments, which suggests that the feldspar grains disaggregated soon before deposition.

In the Aldridge area a few rounded grains of indeterminate plagioclase occur in rocks containing abundant basic volcanic detritus. The grains are extensively replaced by siderite and clay. They may have originated as phenocrysts in the volcanic fragments. In the sample from Hem Heath, North Staffordshire both orthoclase and plagioclase occur, the latter of andesine composition. Both are fresh, although the plagioclase is locally partially replaced by carbonate. Both are presumably derived from the same source as the abundant fresh granite fragments which are present in the specimen.

Feldspars also occur rarely in rocks which do not have a noticeable content of feldspathic rock fragments. Such occurrences have only been observed in the area around Dudley, where a few rounded and altered microclines occur.

iv) Mica

Mica is found very rarely in Etruria sandstones, being limited to the Hem Heath sample, where it is associated with fresh granite fragments, and to the sandstones in the area south-west of Dudley. In the latter case, micas are locally fairly abundant, forming up to 10% of the sediment when concentrated in laminae of fine grained sand and silt.

The mica is in all cases white mica. Grains are often distorted, and show incipient or extensive diagenetic alteration, usually being split and dislocated along cleavage traces and partly altered to chlorite.

v) Heavy minerals

No heavy mineral extraction has been carried out in the present study. However, lists of heavy minerals in the Etruria Formation sandstones are given by Fleet (1925) for South Staffordshire, and Williamson (1946) and Malkin (1961) for North Staffordshire.

All three authors list the following species: opaque oxides (probably incorrectly identified as ilmenite and magnetite), zircon, tourmaline, rutile, garnet, and anatase. In addition Williamson records staurolite, green spinel, sphene and brookite, and Fleet lists apatite.

The main character of this heavy mineral distribution is its monotony, the species present being found in most heavy mineral residues. There is a noticeable lack of easily decomposed ferromagnesian species, which

Malkin (1961) regarded as indicating either deep weathering in the source area, a source area impoverished in such species, or very long transport times. While the first two of these suggestions may contribute to the lack of ferromagnesian minerals, it is likely that their absence may result as readily from post-depositional diagenetic alteration.

vi) Organic material

Detrital organic material is found quite frequently in the finer grained, green sandstones. Usually it occurs as carbonaceous impression fossils of plant material, most of which have not been extensively damaged, and have presumably not undergone lengthy transport. Less commonly lenticular masses of vitrain are preserved, which represent collapsed and flattened plant branches. Rounded 2mm. vitrain clasts have also been observed.

vii) Intraclasts

Intraclasts of mudstone commonly occur in the basal conglomerate of channel sandbodies (see 4.5). Less frequently they form the major detrital component in rapidly deposited units in lateral accretion dominated channel fills. In both cases they occasionally contain abundant sphaerosiderite. This may be a pre-depositional feature, but, by analogy to the development of sphaerosiderite in masses of chlorite which pseudomorph rock fragments (see 8.2.2), this may be a replacement feature.

viii) Detrital ferruginized material

In the red sandstones found in the Mid Staffordshire area near Cannock and Aldridge, two types of detrital ferruginized material are present.

The first consists of rounded clasts of pure haematite (Fig 173a). This is usually completely opaque in thin section, occasionally containing small inclusions of quartz or carbonate. In some cases inclusions of a deeply haematite impregnated clay mineral are present, which is probably kaolinite. This occasionally shows a crude vermicular texture, which may be related to void spaces in the enclosing haematite. These clasts occasionally show a pisolitic structure.

The second detrital ferruginised clast type consists of rock fragments with a prounced haematite rich rim near to, and paralleling their surface (Fig 173b, c). This rim is quite distinct from diagenetic grain coatings (Fig 175).

In both cases the preservation of the detrital clast shape contrasts strongly with the grain coating and pore filling habit of the authigenic haematite in these sandstones (see 8.2.3). This suggests that these grains are truly detrital.

7.3 Provenance of Etruria Formation sediments

7.3.1 Fine grained sediments

Two suggestions have been advanced for the provenance of the iron rich mudstones which form the characteristic lithology in the Etruria Formation. Robertson (1931) argued that these claystones were derived from contemporaneously erupted basaltic lavas. This argument was based on the high iron content of the mudstones, and on their resemblance to the weathered boles associated with contemporaneous basalts in South Staffordshire. The main drawback to this argument lies in the paucity of demonstrable extrusive basalts in the Westphalian, most of those in South Staffordshire having an intrusive origin.

An alternative origin for the iron-rich mudstones was proposed by Wills (1956), who suggested that they were derived from lateritic soils developed in the hinterland source areas. Such an origin is favoured by the results of the present study, although the evidence is somewhat circumstantial. The sediment has a distinctive residual composition, consisting mainly of disordered kaolinite, quartz, and haematite, and having a very low content of alkaline and, especially, alkaline earth elements. (It is suggested in 8.3.3 that the haematite was largely formed by dehydration of detrital ferric oxyhydroxides, and that it thus represents an original detrital component of the alluvium). It is highly likely that laterites were forming in the source area, in view of the equatorial palaeolatitude, the demonstrable humic tropical climate, and the presence of immature lateritic soils within the alluvium itself.

The high iron content of the mudstones does not indicate derivation from an igneous source, contrary to the views of Robertson (op. cit). Van Houten (1972) has demonstrated that the total iron content of the clay fraction of sediments derived from tropical red soils in Northern Columbia ranges from 3% to 9% (4.3% to 12.8% expressed as $Fe_2 O_3$), irespective of the parent material upon which these soils are forming. Most of this alluvium is derived from an upland area composed of schist, gneiss, granite, and diorite. The iron contents of two B horizons of soils developed on quartz diorite and amphibole schist are 15.6% and 13.4% respectively (22.3% and 19.1% expressed as Fe_2O_3).

This range of iron contents is very similar to that found in the Etruria Formation mudstones. This, together with the distinctive residual mineralogy, strongly supports the notion that the fine grained detritus was largely derived from a lateritic weathering mantle in the source area. Such an origin is supported by the presence of detrital ferruginised grains in some of the sandstones, which are regarded as being of probable lateritic origin (see 7.3.2).

Robertson's suggestion of derivation from contemporaneous volcanics cannot, however, be totally overlooked. The presence of basalt fragments in sandstones in South Staffordshire, especially around Aldridge, and the ubiquitous occurrence of detrital grains of inferred original basic composition, pseudomorphed by chlorite, both demonstrate that detrital basic igneous material was being supplied to the basin, and that, therefore, the degradation of basalts may have contributed to the high iron content of the mudstones.

Although pre-Westphalian extrabasinal sources for these basic fragments can be suggested (in 7.3.2), there is no evidence, at present, that these sources were extensive enough to provide a major proportion of the alluvium.

7.3.2 Sandstones and conglomerates

The composition and textural immaturity of sandstones, and the geographical distribution of conglomerates in the Etruria Formation indicate that the source area must have lain close to the south and west of the basins in which the Formation was deposited. This derivation pattern agrees with the limited palaeocurrent data available. The geographical distribution of Facies and the palaeocurrent data are described and discussed in Chapter 10.

As described in 7.2.2, it was realized by early workers on the petrography of the Etruria Formation sediments that many of the detrital components in the sandstones could be directly matched to pre-Westphalian rock types, known from outcrop in the areas bordering the coalfields. In many areas (see Chapter 9) these pre-Westphalian rocks are unconformably overlain by the Westphalian D Halesowen Formation or its equivalents. They may thus be considered to be representative of the outcrop geology of the Etruria Formation source areas, which were being denuded until shortly before the deposition of the Halesowen Formation.

Three main groups of detrital material are present:

i) Lower Palaeozoic sedimentary rocks. Types which have been recognised are:

- Lower Cambrian quartzite. The heavily cemented quartzite which forms a major detrital component, especially in South Staffordshire, is petrographically similar to the Lower Cambrian quartzites which have a wide distribution in the South Midlands, known variously as the Lickey Quartzite (North Worcestershire), the Hartshill Quartzite (Warwickshire), the Wrekin Quartzite (Shropshire), and the Malvern Quartzite (South Worcestershire). Recent exploration has shown this lithology to extend as far south as Minety, 14km. north west of

Swindon, Wiltshire (well Coole's Farm-1; Shell UK Ltd, unpublished data).

- Llandovery sandstones. The finer grained, occasionally argillaceous sandstone clasts, occurring in South Staffordshire, were compared by Whitehead and Eastwood (1926) to the Llandovery sandstones outcropping at Rubery. The correlatives of these sandstones have a wide distribution to the south west and west of the present Westphalian outcrop area (Earp and Hains 1971; Hains and Horton 1969).

- Cambrian shale. The indurated shale clasts present in the Etruria Formation in Warwickshire and South Staffordshire are petrographically similar to the shales present in the Middle Cambrian to Tremadocian in north Warwickshire. This similarity was first noticed by Eastwood et al (1923). These shales underlie the Westphalian in most of the Warwickshire coalfield, having been found in boreholes as far south as Kineton, 15km. south of Warwick. They probably have an extensive distribution in the southward continuation of the Worcester Graben, having been encountered in the Coole's Farm-1 well at Minety (unpublished Shell UK data) and in the IGS stratigraphic test well at Shewton, 16km. north west of Salisbury, Wiltshire. It is unlikely, however, that clasts of this shale could have survived prolonged transport. The distribution of shale clasts (see 10.1.1) suggests that they were locally derived from Cambrian shales exposed in the area between the western boundary fault of the Warwickshire coalfield, and the eastern boundary fault of the South Staffordshire coalfield. Cambrian shales directly overlain by Westphalian D Halesowen Formation have been encountered in this area in the NCB Trickley Lodge borehole (see 9.6).

ii) Fine grained igneous rocks. Two types of fine grained igneous rocks have been recognized:

- acid and intermediate tuffs and lavas. These are an abundant detrital component in all outcrop areas. A wide variety of acid tuffs is known from much of the inferred sediment source area, in the late Precambrian Charnian and Uriconian 'Series'. These are at present exposed in Warwickshire (the Caldecote Volcanic Formation at Nuneaton), Worcestershire (the Warren House Group at Malvern and the Barnt Green Group at Rubery), and in the Western and Eastern Uriconian in the Church Stretton area of Shropshire (Earp and Hains 1971; Hains and Horton 1969). Similar lithologies, although of unconfirmed age, have been encountered in the Netherton-1 well, 10km south east of Worcester (unpublished Ultramar data). The abundance of clasts derived from this source in the Autunian Clent Breccia led Wills (1956) to suggest that such rocks underlie a large part of the south Midlands.

The rock types present in the Charnian and Uriconian acid volcanics include welded vitric tuffs, crystal tuffs, and rhyolites, all of which show textures comparable with those found in detrital grains in the Etruria Formation. In particular, some of the Uriconian tuffs in the Wrekin area show a textural similarity in thin section to some of the non-diagnostic microcrystalline quartz grains found in the Etruria Formation. Dacites and andesites form a major component in the Uriconian, and also occur, as locally derived boulders, in the Caldecote Formation (Dunning 1975).

- basic volcanics. Although recognisable basalt fragments are only locally abundant (see 10.1.1), the ubiquitous occurrence of detrital grains replaced by chlorite suggests that fine grained basic igneous rocks may have formed an important part of the original sand fraction of the alluvium. The derivation of this basic material is a controversial question. As has been mentioned in 7.3.1, Robertson (1931) was of the opinion that basaltic material had been derived from contemporaneous and earlier Carboniferous lavas. This explanation has been criticized on the grounds of the rarity of preserved examples of such rocks, and the demonstration that most of those which do occur are intrusive. Although extensive vulcanism occurred in the Westphalian A and B in the East Midlands (E.L. Boardman, personal communication), there is no evidence either that this vulcanism persisted into the Westphalian C and early D, when the bulk of the Etruria Formation was deposited, or that these rocks were subject to erosion at that time. They would in any case have been too far away from the West Midland depositional basin to have contributed significantly to the Etruria alluvium in that area. The cessation of vulcanism by the end of Westphalian B is supported by the stratigraphic distribution of basic tonsteins (Spears and Kanaris - Sotirou 1979).

In the absence of proof of contemporaneous vulcanism, other potential sources of basic fragments must be considered. Basic volcanics form a considerable proportion of the Uriconian sequence in Shropshire (Earp and Hains 1971), and Ordovician basic intrusions are widespread in the Cambrian in North Warwickshire (Hains and Horton 1969). Basic material may have been derived from these sources.

It is also possible that, in the later stages of deposition of the Etruria Formation, some basic material may have been supplied by the erosion of contemporaneous basic intrusions, during the development of the sub-Halesowen Formation unconformity. This suggestion is elaborated in 7.3.3 and in Chapter 10.

iii) Plutonic and metamorphic rocks. Granite fragments may have been derived from intrusions in the Uriconian, Charnian, or Malvernian complexes (Dunning 1975). These rocks are little exposed and their extent is not known. The only known possible source for the metamorphic detritus, for which there is any evidence, is the Malvernian complex, and its correlatives in Shropshire (Dunning 1975). Again, the extent of these rocks below the post Carboniferous cover is not known.

iv) Ferruginised detritus. The presence of vermicular clay inclusions and pisolitic textures in the ferruginised clasts, together with their detrital appearance, suggests derivation from lateritic soil profiles developed in the sediment source areas. (cf. laterite ironstone textures illustrated in Mohr *et al* 1972).

7.3.3 Derivation by the erosion of earlier Westphalian material

During the deposition of Etruria facies rocks in the early part of the Westphalian, the sediment was probably all derived from outside the basin, from sources described above. In the later Westphalian C and earliest Westphalian D, however, a phase of gentle folding and block faulting occurred, which, while rejuvenating the extrabasinal source areas, also gave rise to an intra-Westphalian unconformity, whose effect was to denude Westphalian rocks in structures developed locally in all depositional areas except North Staffordshire. This phase of deformation was accompanied by the emplacement of the alkali basalt intrusions in the Wyre Forest and South Staffordshire. The stratigraphic evidence for this unconformity is described in Chapter 9 and its development charted in Chapter 10.

Wills (1956) and Hoare (1959) have both suggested that Westphalian and possibly other Upper Palaeozoic material which was eroded during the formation of this unconformity would have acted as a major source of sediment for Etruria facies rocks being deposited elsewhere in the southern part of the Pennine Basin. This would most noticeably have affected North Staffordshire, where there seems to be a continuous sequence in the Westphalian D, between the Etruria Formation and the Newcastle (= Halesowen) Formation.

While this seems a very likely proposition, the proportionate importance of this erosion as a source of Etruria Formation sediment cannot be appraised. If, however, such erosion included the erosion of basic intrusions, and possibly of any associated surface lavas, from the Coal Measures and Etruria Formation in South Staffordshire, this could account for some of the inferred basic detrital components in the Formation in North Staffordshire.

CHAPTER 8

Diagenetic History and Controls on Diagenesis of Etruria Formation Sediments

8.1 Introduction

As noted in Chapter 1, much of the previous work on the Etruria Formation has, often unconciously, been directly concerned with its diagenesis. Most authors have been concerned with the origin of the predominantly red colouration of the formation, and have usually been at pains to find a palaeoclimatic or tectonic control responsible for the formation of red beds during the Westphalian, usually to support a palaeogeographic reconstruction. The only author to have written about the petrography of the sandstones (Williamson 1946) devoted much attention to providing an explanation for the detrital origin of what is now regarded as authigenic matrix (see 8.2).

Within the Etruria Formation there are three groups of sediments with distinctive diagenetic histories: the green sandstones; the red sandstones; and the mudstones. The diagenetic history of each is described separately in the following chapter, after which the inferred relationship between the three is discussed, and a model erected for the controls on diagenesis.

8.2 Diagenesis of Sandstones

8.2.1 Green sandstones: description of diagenetic minerals

Most of the major sand and conglomerate bodies (i.e. alluvial channel and fan deposits) are green in colour, and have a broadly similar diagenetic pattern. The dominant feature is the development of a large amount of matrix, mainly by the chemical and physical disaggregation of unstable lithic grains. Three main diagenetic products are present: secondary matrix, quartz and calcite cements, and pyrite. Only the matrix and calcite cement contribute appreciably to the induration of the rocks.

i) Matrix. At first sight these sandstones appear to have a high content of a low birefringence matrix composed of very fine grained quartz and clays. Much of this apparent matrix is illusory, consisting of grains composed of microcrystalline quartz, whose original form is readily visible in plane polarised light (Fig. 171).

The true matrix material present is most commonly a green chloritic clay, which occurs as large discrete masses, and as pore infills. These large masses are totally replaced rock fragments, which have been distorted and squeezed between rigid framework grains.

The chlorite pore fills may in some cases also represent deformed and altered framework grains. In other cases the presence of aggregates of chlorite with well developed anomalous birefringence suggests that some of this clay may be authigenic. In places where such chlorite forms pore linings, the euhedral crystalline habit clearly shows an authigenic rather than replacive origin.

The hydrodynamic conditions of sedimentation of most of the sandstones make it unlikely that such matrix is of depositional origin (see Chapters 3 to 5).

ii) Cements. Two authigenic minerals, quartz and calcite, occur in the green sandstones, and locally act as cements.

Quartz occurs ubiquitously as syntaxial overgrowths on detrital grains. These are easily recognised in thin section, as many of the original grain outlines are marked by inclusions or dust rims. In some instances more than one phase of overgrowth is present. Euhedral grain terminations are often present.

Syntaxial quartz overgrowths usually predate the development of carbonate cements, which often surround and corrode them. The occasional presence of overgrowths with large embayments suggests that these formed as void filling cements prior to the alteration of unstable lithic fragments, which have subsequently lost their identity after alteration to clays.

Calcite occurs as a poikilotopic cement enclosing detrital grains in large (up to 5mm) masses which have apparently uniform extinction. Both quartz grains and rock fragments have extensively corroded boundaries and some grains appear to have been fractured by displacive carbonate cement.

In rarer cases, carbonates are present in a spherulitic form. such spherulites may take three forms:

a) Calcite occasionally occurs as fan-shaped aggregates of bladed crystals often with euhedral terminations. These may occur within original pore space, but are more frequent within chlorite masses representing altered rock fragments.

b) Calcite and siderite occur as small (less than 2mm) spherulites within chlorite masses representing altered rock fragments and within intraclasts.

c) Calcite occurs as highly porous spherulitic masses ca. 2mm in diameter, which grow from individual nuclei on the walls of cavities, and coalesce to form botryoidal cavity linings with a laminated appearance (Fig. 174). This type of spherulitic carbonate has only been seen in near-surface samples, and is regarded as being a feature of recent vadose zone weathering.

iii) Other authigenic minerals. The only other common authigenic mineral in the green sandstones is pyrite, which occurs as discrete aggregates up to 5mm in size. It usually occurs in the basal few centimetres of a sand body. Very rarely it is disseminated throughout a sand body, and locally acts as a cement.

8.2.2 Discussion of the diagenesis of green sandstones

The green sandstones in the Etruria Formation have a fairly simple diagenetic history. It is possible to recognise the effects of the following processes.

- precipitation of syntaxial quartz overgrowths.
- precipitation of early carbonate cement.
- breakdown of unstable rock fragments, these being replaced by chlorite and carbonates.

- mechanical deformation of replaced rock fragments to form matrix.
- authigenesis of spherulitic carbonates and pyrite.

These processes may have occurred approximately in the order listed, although some, notably the precipitation of carbonates and the alteration of unstable rock fragments, may have operated throughout the diagenetic history.

a) Precipitation of syntaxial quartz overgrowths

Overgrowths on quartz grains are formed by precipitation of silica from solution in the formation fluid. This silica can be derived from four sources: i) dissolution of detrital quartz grains, especially if their surfaces are pitted; ii) pressure solution between adjacent quartz grains; iii) corrosion of quartz grains by the development of carbonate cements; iv) dissolution and/or replacement of unstable detrital rock fragments and feldspars (i) - iii) Nagtegaal 1969; iv) Walker et al. 1978).

In most cases where the pre-overgrowth grain boundary is recognisable, there does not seem to have been any dissolution before overgrowth formation. Occasionally, however, the inclusion traces marking the original grain outline are irregular and pitted. In such cases dissolution of the grain preceded overgrowth formation, and, presumably, liberated silica for precipitation elsewhere in the rock. If such dissolution was extensive it is possible that some finer grained detrital quartz may have been removed completely (cf Nagtegaal 1969). This may have provided an important source of dissolved silica.

There is no evidence for extensive pressure solution having taken place between quartz grains. This probably results from the presence of less rigid rock fragment grains in the framework.

Dissolved silica may also have been derived from corrosion of quartz grains associated with the precipitation of carbonate cements. However, as most carbonate cements appear to postdate syntaxial overgrowths on quartz grains, this can only have contributed dissolved silica in the later stages of quartz overgrowth formation.

Silica may also have been derived extensively from the breakdown of minerals which were contained in the detrital grains which are now pseudomorphed by chlorite. Lack of knowledge of the original composition of these grains prevents a quantitative estimate of their importance as a source of silica, but it is probable that they formed the major source.

b) Precipitation of carbonate cements

The cations necessary for carbonate cement formation were probably derived from the breakdown of lithic grains, notably from feldspars and ferromagnesian minerals in the basic rock fragments. Carbonate may have been provided by solutions of atmospheric carbon dioxide in the pore water, or by the oxidation of organic matter.

The texture of carbonate cements in alluvial deposits is controlled by the position of the water table. Vadose zone cements have the form of microcrystalline grain coatings often with geotropic, dripstone textures; by contrast cements formed below the water table are usually sparry and pore filling (Stalder 1975). On this criterion, the sparry carbonate cements in the green Etruria sandstones formed below the water table.

The origin of the spherulitic carbonate cements is not clear. The occurrence of aggregates of bladed calcite, and of true spherulites, is limited to the interior of masses of clay of detrital or authigenic origin. Such spherulitic aggregates probably formed in the manner suggested by Spenser (1925), who regarded spherulitic carbonates as being analagous to devitrification spherulites in volcanic glasses, having grown by simultaneous nucleation and radial growth in a gelatinous medium. This implies that such spherulitic carbonates must have formed simultaneously with, or very soon after, the breakdown of the unstable rock fragments which gave rise to aggregates of chlorite, before the clay had been dewatered and compacted.

The origin of the highly porous, botryoidal calcite spherulites is also not understood. Williamson (1946) regarded these as ooliths, but recognised that they had grown in place. The oolitic morphology is spurious. Almost all the examples observed in the present study are obviously nucleated on the margin of the void which they are filling, and the rare complete spherulites observed are probably seen in sections cut parallel to the margin of such a void. Williamson had difficulty accounting for their growth, and concluded that they had grown from a gel like mass of carbonate and clay, which, in the case of cross bedded sandstones, must have infiltrated the sandy sediment soon after deposition.

The extremely fragile and porous nature of these spherulitic masses makes it unlikely that they could have survived compaction of the sediment. They are thus likely to be of post-compactional origin. In this case, it becomes necessary to account for the voids into which these spherulites have grown, as voids have a low preservation potential during compaction of argillaceous sands (Nagtegaal 1978).

It is suggested here that these spherulitic carbonates are weathering features formed in the present vadose zone. The texture has not been observed in core samples (Williamson's specimens were collected from surface exposures), and the microcrystalline nature of the calcite conforms to Stalder's criteria (op cit)for vadose zone cements. The origin of the void spaces is not known. They may have formed by ground water dissolution of pre-existing carbonate cements.

c) Breakdown of unstable rock fragments and formation of secondary matrix

Rock fragment breakdown and replacement has been described from many environments. It most commonly involves rock rich in ferromagnesian minerals, and occasionally feldspathic rocks. The products of alteration are usually carbonate or clay pseudomorphs of the original grain, and/or authigenic carbonates, clay minerals, and zeolites. If alteration takes place under oxidizing ground water conditions, the iron liberated in this process forms haematite precursors (Walker *et al* 1978). If the early diagenesis occurs in non-oxidizing conditions, iron is incorporated into iron-rich chlorites. This process, commonly associated with greywacke sediments (cf. Cummins 1962), has been recorded in shallow marine sandstones (Galloway 1974), and has obviously occurred to give rise to the green sandstones in ancient alluvial red bed sequences (e.g. Friend 1968).

The diagenetic formation of chlorite is usually regarded as a fairly late stage diagenetic feature, controlled by burial temperature (e.g. Galloway op. cit.). In the Etruria sandstones this does not seem to have been the case, as two independent lines of evidence suggest an early origin. Firstly, palaeomagnetic data (P. Turner personnal communication, Besly and Turner, in press; see 8.4.3) show that the diagenesis of the iron minerals carrying the chemical remanent magnetisation of these sandstones was completed during the Carboniferous, at which time the Etruria Formation sediments had probably only been buried to a maximum of 1000m, and probably, in most cases, to a shallower depth. Secondly there is no evidence of chlorite formation either in the underlying Coal Measures or in the overlying Halesowen/Newcastle Formation. Sediments in both of these Formations have undergone early diagenesis in reducing conditions, and have been buried to similar or greater depths. They contain iron bearing minerals - potential iron sources for chlorite authigenesis - in the form of micas, albeit in smaller quantities than the quantities of presumed iron rich lithic fragments in the Etruria Formation sandstones.

These observations suggest that the partial or complete replacement of detrital lithic fragments by chlorite (and possibly other authigenic components) occurred as an early diagenetic process, and was strongly compositionally controlled. This process probably occurred over most of the diagenetic history, and liberated material which subsequently recrystallized as authigenic quartz and calcite.

The compositional control is further emphasized by the observation that green, Etruria - like lithic sandstones occur sporadically in the Coal Measures, mainly in the marginal areas of Shropshire and South Staffordshire, and mainly at horizons just below the onset of the red bed Formation. These sandstones contain abundant authigenic chlorite.

d) Origin of pyrite

The diagenetic mechanism of pyrite formation is summarized by Berner (1971). Pyrite is formed by the reaction of iron minerals with hydrogen sulphide. The iron is derived from the finest grained, most reactive, iron bearing mineral phases in the sediment, usually amorphous iron oxides or chlorite. Sulphide is derived from bacterial reduction of dissolved sulphate, or from sulphur compounds in organic material. Sulphide derivation from sulphate in pore water is the commonest source, but occurs only when the pore water is of marine origin. The meteoric ground water in alluvial sediments is unlikely to contain appreciable sulphate. Beadle (1974) gives fresh water compositions from Africa showing that, in most cases, sulphate concentrations are negligable. Sulphide in the Etruria sandstones must therefore be organically derived, presumably from plant material. In this context, the concentration of sulphide at the base of channel deposits is not coincidental, since much plant debris may have been deposited as a channel lag. No direct association of plant material and pyrite has been observed, but this may only reflect diagenetic destruction of the organic matter. An identical occurrence of pyrite and plant material in the base of a fluvial channel deposit is described by Friend (1966).

8.2.3 Red Sandstones

Red sandstones occur in two distinct settings in the Etruria Formation. Firstly, the coarse sandstones and conglomerates in a limited area between Cannock and Aldridge, together with some examples of such lithologies in the Dudley area, are red. Secondly, many thin sandstones with a high matrix content, in Facies 8 and 9, are found to be red in all areas studied. Sandstones of the latter type are considered to have behaved similarly to the overbank sediments during diagenesis, and are discussed in 8.3.3 and 8.4.

Seven diagenetic products have been recognised in thin section in the red sandstones: secondary matrix; quartz overgrowths; carbonate cement; barite cement; haematite cement; authigenic kaolinite; and authigenic feldspar.

i) Matrix. As in the green sandstones, the apparent dominance of fine grained matrix in thin sections viewed under crossed polars is illusory, many of the detrital grains in these sandstones consisting of microcrystalline quartz. Some samples examined do, however, contain substantial proportions of a secondary matrix consisting of fine grained quartz, abundant disseminated haematite, and some clay, which is unidentifiable in thin section.

Unlike that in the green sandstones, the matrix in the red sandstones does not obviously occur as altered framework grains.

ii) Quartz overgrowths. As in the green sandstones, syntaxial quartz overgrowths are ubiquitous. Euhedral terminations are common. Locally the overgrowths act as a cement (Fig. 178). Syntaxial quartz overgrowths are an early diagenetic feature enclosed by, and thus preceding, the sparry carbonate and massive barite and haematite cements (q.v.).

iii) Carbonate cements. Carbonate cements occur as large poikilotopic pore filling masses, and in various textures as grain replacements.

Three distinct cementing textures have been recognised.

a) Undeformed poikilotopic calcite masses. These are masses of poikilotopic cement enclosing detrital grains in a manner similar to that observed in the green sandstones (Fig. 175).

b) Poikilotopic sparry calcite cement, showing curved cleavage traces (Fig. 176) and strained extinction patterns. This type of cement occurs distinctively in the very coarse and poorly sorted conglomerates. Masses of cement in crystallographic continuity form zones up to 6mm across. The cement is clearly sparry, but has strongly curved cleavage traces, which are reflected by the brush-type extinction shown under crossed polars. In some cases, large masses are subdivided into distinct crystallographic zones, each showing an extinction cross with a pseudo - uniaxial pattern. In a few cases cements of this type show an undulose extinction pattern, and in one case a crude lamellar twinning.

c) Poikilotopic rhombohedral carbonate cement showing partial replacement by haematite (Fig. 177). This cement occurs in poikilotopic masses around detrital grains, and shows partial replacement by haematite. Haematite occurs in rhombohedral masses and in lamellae parallel to cleavage or twin planes of the carbonate, both forms pseudomorphing features of the rhombohedral carbonate cement, and thus demonstrating that the carbonate was the earlier phase. Because of its apparent ease of alteration to haematite this carbonate mineral may be siderite. This has not, however, been confirmed by XRD or chemical means, and it is possible that a pseudomorhous replacement of calcite, with an external source of iron, may have been involved.

Cements of types b) and c) have only been observed in a few samples in each case. They appear to be mutually exclusive. Deformed cements of type b) and undeformed cements of type a) do occur together.

In addition to forming cements, carbonate minerals occur as grain replacements in two habits. In one of these, entire detrital grains are replaced by sparry carbonate, either in a well crystallised mass, or as an aggregate of small crystals. The replacive nature of such carbonates is recognisable by partial replacement leaving part of the original grain unaltered; the preservation of the original irregular grain outline; the preservation of impurities marking the original grain outline within the carbonate aggregate; and occasionally, the presence of abundant inclusions of clay and/or haematite within the replaced grain. In the second grain replacement texture isolated euhedral rhombs of a carbonate mineral occur within grains which have otherwise been altered to haematite and/or clays. The carbonate mineral in this case is probably dolomite.

iv) Barite. In some specimens from the Aldridge area, barite has been observed in thin section forming a poikilotopic cement. Quartz grains enclosed by the barite show rather poorly developed overgrowths, which suggests that this barite is a very early cement.

v) Haematite. Apart from the ubiquitous presence of disseminated haematite in the secondary matrix (q.v.), haematite occurs commonly as a massive pore filling cement and as a grain coating (Fig. 175), postdating syntaxial quartz overgrowths, but probably preceding the sparry carbonate cements. No texture can be observed in the haematite cement in most thin sections. In one section there is a hint that haematite forms an overgrowth on a detrital haematite grain. This possibility could be investigated using reflected light microscopy.

vi) Kaolinite. Kaolinite occurs infrequently as aggregates of euhedral vermicular books, completely filling pore space which has previously been partially occluded by the development of cementing quartz overgrowths (Fig. 178). The kaolinite appears, in thin section, to be fairly pure, being contaminated only by minor quantities of disseminated haematite.

vii) Authigenic feldspar. In one sample small tablet shaped euhedral crystals of authigenic albite have been observed, growing as a cluster into empty pore space (Figs. 175, 179). (This is the same sample as that which contained the most abundantly developed authigenic kaolinite).

8.2.4 Discussion of the diagenesis of red sandstones

The diagenetic changes that have occurred in the red Etruria Formation sandstone are as follows:

- precipitation of syntaxial quartz overgrowths,
- precipitation of barite cement,
- breakdown of unstable rock fragments and minerals to form matrix, with subsequent mechanical rearrangement. Alteration of grains.
- precipitation of haematite cement,
- precipitation of authigenic feldspar,
- precipitation of carbonate cement.

Textural evidence in thin sections suggests that the first two of these changes were early phenomena and occurred in the order given. The haematite usually lines pores which are filled by kaolinite or feldspar and the carbonate cement usually postdates haematite and feldspar (Fig 175).

a) Quartz overgrowths. The sources of silica for the generation of quartz overgrowths were probably similar to those in the diagenesis of the green sandstones (8.2.2). As alteration of detrital framework grains and mafic minerals probably liberated silica throughout the diagenetic process, it is likely that quartz overgrowths, although recognisably a very early diagenetic feature, continued to form throughout diagenesis.

b) Barite cement. Barite is a common early authigenic mineral in modern desert sands, forming the early 'desert rose' concretions. By analogy with such occurrences, the presence of barite rosettes in inferred arid zone alluvial deposits (such as the Trias of the English Midlands) has led to its being regarded as an early diagenetic mineral characteristic of arid environments. This view, which is probably spurious, is unlikely to be applicable to the Etruria Formation sediments.

In seeking an explanation for the occurrence of early barite cements in sediments deposited under humid conditions a major difficulty lies in identifying the source of the barium. In the case of the Etruria Formation there is no obvious source in the immediately adjacent sediment pile. Hawkins (1978) invoked expulsion of barium during dewatering of an underlying marine shale as the source for barite which replaces kaolinite in Westphalian A sandstones in Nottinghamshire. Such a source cannot be proposed here, as the nearest marine shale in the underlying Westphalian sequence is probably some 250m below the barite cemented sandstones. The source of both barium and sulphate remains unknown.

c) Alteration and replacement of lithic material

The processes of alteration of lithic material which may be inferred to have occurred in the red Etruria sandstones are similar to those described by Walker (1967a and 1976) and Walker et al (1978) from Cenozoic arid zone alluvium. Matrix has been produced by the hydrolysis and/or replacement of ferromagnesian minerals and feldspars under oxidizing conditions. Because the sediment is not of first cycle derivation, the minerals which have been altered were probably contained within lithic grains, mainly of basic composition, rather than occurring as discrete ferromagnesian grains, as in the example of Walker. Evidence for the presence of lithic grains of basic composition is present in the form of recognisable basalt grains, in which either or both phenocrysts and groundmass have been altered to haematite and/or clays and/or carbonates which pseudomorph the basaltic textures. The products of grain alteration, other than matrix, gave rise to the carbonates, haematite, kaolinite, and authigenic feldspar described below.

d) Formation of haematite cement

The formation of haematite cements in alluvial enviroments has been carefully documented by Walker (1967a, 1976) and Walker et al (1978). Haematite forms by the dehydration of so-called 'haematite precursors' - hydrated ferric oxides - which form as an alteration product of iron bearing detrital grains in an oxidizing meteoric ground water environment.

Although the recent to Tertiary sediments described by Walker and others have been deposited and undergone early diagenesis in an arid environment, there seems no reason why similar conditions should not occur in tropical zone alluvium, provided oxidizing conditions are maintained by there being good drainage in the alluvial pile immediately after burial. The controls on diagenetic haematite formation in sandstones are further discussed in 8.4.2.

e) Precipitation of kaolinite and of authigenic feldspar

Both of these minerals are commonly cited as the authigenic products of material derived from the breakdown of lithic material and, especially, feldspars (e.g. Walker et al, op cit).

f) Precipitation of carbonate cements

The conditions under which sparry cements formed were probably similar those in the case of the green sandstones (8.2.2), cations being derived from the alteration of lithic clasts and feldspars. The poikilotopic sparry texture suggests that these are not vadose zone cements, but the large pore volumes filled by calcite cement demonstrate that the cement is a fairly early diagenetic feature preceding significant compaction.

The origin of cements showing strained extinction and curved cleavage traces is not clear. Calcium rich dolomite having this form ("saddle dolomite") has been described by Radke and Mathis (1980). Within such dolomite lattice distortion results from compositional variation within individual growth laminae, crystal face intersections being enriched in calcium in relation to the remainder of the crystal, which is proportionately enriched in magnesium, iron, and manganese. Apart from the compositional difference (the Etruria cements are calcites), Radke and Mathis specify that "saddle dolomite" forms at temperatures between 60° and 150°C, which is a considerably higher range than that envisaged for the early diagenetic formation of cements in the Etruria Formation. In addition, the paragenesis of 'saddle dolomite' appears to be related to hydrocarbon migration and mineralization in fault and stratigraphic traps. If the Etruria cements are related to "saddle dolomite", they must have formed by replacement of a pre-existing cement phase during later diagenesis.

It is also possible that analogous lattices defects may be able to form at the temperatures of early diagenesis, especially, as here, in the presence of a readily available selection of cations. Further research on the composition and structure of these cements is needed.

The occasional occurrences of lamellar twinning implies that some of the deformation in these cements results from compaction.

Poikilotopic carbonate cements which have been partially replaced by haematite may have had one of two origins. They may have originated as early sparry calcite cements, and subsequently been partially replaced by haematite. Or they may have been deposited as early siderite cements, and have been partially oxidized to haematite. In the latter case, the sandstone would have undergone an early phase of diagenesis under reducing conditions similar to those which occurred in the green sandstones. Without chemical analysis to determine the composition of the carbonate phase present, there is no way of telling which of these mechanisms was responsible for the replacement of these carbonate cements by haematite.

8.3 Diagenesis in overbank lithofacies

8.3.1 Previous Work

From publications up to 1970 two consensus views have emerged as to the origin of red pigmentation in the fine grained sediments which make up the bulk of the Formation. These are exemplified by Hoare (1959) and Poole (1966).

Hoare regarded the red pigment as having been derived directly from red soils formed on upland areas. Such soils formed in response to improved drainage resulting from uplift during the 'First Malvernian Movements' (Wills 1956), an episode of gentle folding and block faulting during which areas marginal to the depositional basin were uplifted, and Westphalian A to C sediments within the basin were locally folded and denuded. The red Etruria facies sediments were thus in part recycled, oxidized Westphalian sediments, and in part derived from lateritized pre-Carboniferous sources.

The alternative view, summarized by Poole was that, while some of the sediment may have been red at the time of deposition, the content of oxidized plant remains, roots, and thin grey seat-earths made it likely that most of the sediment had originally been deposited in a reducing environment as grey, coal measure type deposits. Poole envisaged that oxidation at these sediments had taken place shortly after deposition, as a result of a drop in the water table resulting from slight regional tectonic uplift.

8.3.2 Current trends in the interpretation of red bed genesis

In recent years three models for the origin of red beds have been widely applied.

i) Krynine (1949, 1950) suggested that red alluvium could be produced as the result of the erosion and redeposition of red lateritic soils. Although Walker (1967 b) attempted to discredit this idea when he demonstrated that red detritus was almost absent in alluvium derived from red soils, the concept has survived and been extended in a modified form by Van Houten (1968, 1972). The latter documented recent muddy alluvium derived from areas of ferrallitic weathering, which contained similar amounts of ferric iron to those found in ancient red mudstones. The iron was in the form of hydrated ferric oxides, which gave the sediment a brown, rather than red, colour. Van Houten suggested that, with ageing, this hydrated ferric oxide would dehydrate to give haematite. His observations help to account for the comparative rarity of red recent sediments, and of brown sediments in ancient sequences. The thermodynamics of the dehydration process have been studied by Berner (1969), who produced experimental evidence to show that this reaction is favoured under natural conditions.

The origin for the red pigment in the Etruria Formation proposed by Hoare (1959) falls into this category.

ii) Friend (1966) suggested that red and grey colouration in fluvial sediments results from differences in Eh and pH, controlled by the position of the water table at the time of deposition. This was based on the observation that the drab and red colouration observed in fluvial fining-upward sequences commonly correspond to the channel and overbank deposits respectively. Friend postulated that this resulted from the water table dropping away from active channels, allowing oxidation of flood plain sediments, while reducing conditions were maintained in the channel deposits.

iii) Walker (1967a, 1976) and Walker et al (1978) have erected a sophisticated model for the origin of red beds in arid zone, first cycle sediments. By careful documentation of such sediments of Quaternary to Pliocene age in Mexico and the south western United States, they have conclusively shown that reddening increases progressively with age, as a result of the breakdown of iron bearing detrital minerals and rock fragments, the precipitation of a variety of authigenic minerals including hydrated ferric oxide precursors, and the subsequent dehydration of these precursors to form haematite. This process operates over a period of several million years.

Walker (1974) has extended this model to include the diagenesis of alluvium deposited under humid tropical climate conditions. In examples from Puerto Rico he was able to recognise progressive reddening in sediments of unspecified texture of Quaternary age. He ascribed this reddening partly to the breakdown of unstable detrital ferromagnesian grains, and partly to the ageing of detrital hydrated ferric oxides, in the manner described by Van Houten.

Apart from these syn, and early post-depositional mechanisms for red bed formation, Mykura (1960) has shown that red beds may be formed by much later penetrative oxidation of lithified sediments during periods of sub-aerial exposure and low water table conditions in arid climate zones.

8.3.3 Red overbank deposits in the Etruria Formation

It is evident from the sedimentological considerations discussed in 6.1.5 that red pigment has been generated in the overbank deposits in at least two ways.

In sequences of sediment containing palaeosols of type 3 (post depositionally oxidized alluvial soils) the sediment was, at the time of deposition, rich in organic matter. In the reducing, waterlogged conditions associated with such organic rich deposits, iron would have been in a reduced state, as is witnessed by the persistence of siderite in some of these now red palaesols. Haematite, or haematite precursors, were formed in the sediment by a phase of post-depositional penetrative oxidation, which partially or completely destroyed the organic matter present, and, in the absence of this organic matter, oxidized ferrous iron to ferric pigments or precursors. In contrast, sediments containing evolved palaeosols of type 4 show evidence of the formation of haematite, or haematite precursors, as concretions which formed during pedogenesis at the sediment surface during and shortly after deposition.

Inferences regarding the origins of haematite pigment in palaesol units are not necessarily applicable to those parts of the sequence which do not consist of palaesols. In the bulk of the red mudstones and sandstones of Facies 8 to 11 it is necessary to decide whether red pigment has been formed by post depositional oxidation, or syndepositionally, either by oxidation of reduced sediment, or by the deposition of alluvium containing detrital ferric iron.

The question of the initial oxidation state of iron in the sediment may be approached from two viewpoints, one that of bulk chemical composition, and the other a mineralogical viewpoint. These approaches have been discussed by Besly and Turner (in press).

As has already been indicated in 7.1, the mineralogical composition of the red Etruria facies mudstones is broadly similar to that of the grey, organic rich mudstones both intercalated with Etruria facies deposits and occurring elsewhere. The principal compositional difference between the two lies in the amount of iron present. Grey, organic rich sediments contain 2.7% - 4.7% total iron (expressed as Fe_2O_3), largely in the ferrous state, while the red mudstones contain 7.0% - 11% total iron, in the form of haematite and goethite (Besly and Turner op cit; Holdridge 1959).

The occurrence of immature lithic sandstones (described in 7.2) in association with both grey and red overbank sediments suggests that these fine grained sediments share a common provenance (7.3.1). This being so, it is necessary to explain the marked disparity in iron content between the two. Two explanations are possible: firstly, that both grey and red sediments were originally deposited as grey sediments with low iron contents, and that iron was introduced into the sediments that are now red during a subsequent phase of oxidative diagenesis; or, secondly, that the initial alluvium originally contained a larger amount of iron, in the form of fine grained hydrated oxides, which became fixed in the red mudstones during or shortly after deposition, but was reduced and partially mobilized in the grey organic rich mudstones.

In the first of these models, it is necessary to identify a source for the additional iron which became concentrated in the red mudstones, and a mechanism for its concentration. There is no evidence for the existence of unstable iron rich heavy minerals in the unoxidized sediment, which rules out the source of iron suggested, in a similar setting, by Walker (1974). Although there is an abundant potential source of iron in the basic rock fragments found in the sand bodies, it is difficult to envisage a mechanism whereby iron could have been removed during diagenesis from permeable sand bodies to relatively impermeable mudstones. This would reverse the trend of fluid flow (and thus of solute transport) usually encountered during the compaction of an interbedded sand/mudstone sequence.

A more likely explanation is provided by the second model. In this case the differentiation in iron content would have taken place at the time of deposition, the iron being rapidly reduced and mobilized in the waterlogged organic rich mudstones while remaining fixed in the red mudstones. The potential for the rapid reduction and organic complexing of iron in waterlogged conditions in the presence of abundant organic material is commented on by Bloomfield (1964).

The initial oxidation state of the iron in the sediment is also reflected in the nature of the diagenetic minerals present. Two of these are present: siderite and calcite. Siderite occurs mainly as small concretions in association with type 1 palaeosols (seat earth/ alluvial palaeosols) and with type 3 palaeosols (post depositionally oxidized alluvial palaeosols). It also occurs as a disseminated component and as small concretions in a number of red mudstones, which usually underlie alluvial palaeosols, which may or not have been post depositionally oxidized. (Examples are found in Wilnecote Quarry and Playground No. 8 and Kibblestone boreholes). In both palaeosols and mudstones siderite occurs most frequently as sphaerosiderite. Occassionally, in sequences containing type 3 (post depositionally oxidized) palaeosols, spherulitic haematite grains are found, which may represent post depositionally oxidized sphaerosiderite. The persistence of siderite in the red mudstones indicates that this mineral formed at an early pre-reddening stage of diagenesis, at which time the sediment probably contained a considerable quantity of organic matter.

In most of the red mudstone, siderite is absent, and the only concretionary species are haematite (contemporaneously formed in palaeosols) and calcite, which occurs mainly as small, spherulitic concretions. The latter is of very early diagenetic origin, as it is subject to sedimentary reworking (see 4.2.1). Such early concretionary calcite probably formed in oxidizing conditions (cf. Coleman 1966) to the exclusion of siderite.

The marked difference in iron content and the apparently mutually exclusive occurence of siderite and calcite in the red mudstones suggests that two distinct processes of reddening were involved in the early diagenesis of these rocks. Mudstones deposited under initially reducing conditions were characterised by ther presence of siderite and organic matter, and by low initial iron contents. During subsequent penetrative oxidation the organic matter was partially or completely destroyed, and, where all organic matter was destroyed, the sediment was reddened. Such red beds may be characterised by the anomalous presence of siderite in haematite bearing rocks, by type 3 palaeosols, by bedded haematite ironstones (Facies 12), and in some cases by the presence of coal seams. In contrast, red mudstones characterised by high iron contents and by early calcite concretions were deposited under initially oxidizing conditions. The iron in these parts of the sequence was deposited as a detrital hydrated ferric oxide precursor, which subsequently inverted to haematite.

Although two distinct origins for red pigment can be recognised in these mudstones, the occurence of non-palaeosol mudstones, containing residual siderite is very low, and without this criterion it is impossible, at present, to recognise mudstones which have been initially deposited in a reduced state. It may be expected that these will sometimes be recognisable by having a lower iron content than mudstones which have never been through a phase of reduction, but data are not yet available. It would, however, appear from the sections measured in the present study, and from results such as those of Holdridge (1959) that the bulk of the Formation is composed of mudstones with a rather high iron content, and containing early calcareous concretions. This would suggest that, apart from in palaeosols, most of the red pigment in the overbank Facies deposits is of detrital origin.

8.4 Relationships between diagenesis of sandstones and mudstones; timing of diagenesis.

8.4.1 Red overbank deposits with green sandbodies.

This association of diagenetic facies is very common in alluvial sediments (Friend 1968). Friend presented a simple explanation, whereby a drop in water table away from an active channel in a floodplain (rather than a flood basin) allowed oxidation of overbank deposits while the waterlogged conditions in the channel sediments prevented oxidation.

In the Etruria Formation such a control produced by water table fluctuations during deposition needs careful examination, as the presence of post depositionally oxidized coals, ironstones, and palaeosols clearly shows that penetrative oxidation took place in the sediment after burial. Two questions arise from this. If penetrative oxidation has occurred, why have the channel deposits not been affected by it?; and, if penetrative oxidation occurred as a result of groundwater movement (cf. Walker 1967a, 1976), why have the effects of this been concentrated in the less permeable mudstones instead of in the more permeable sandstones?

The presence of post-depositionally oxidized horizons in sequences dominated by mudstones implies that this phase of oxidation took place soon after deposition, before the free circulation of oxygenated groundwater was impeded by the loss in permeability which would have accompanied the compaction and cementation of the mudstones. Studies of recent muddy overbank alluvium indicate that lithification takes place comparatively rapidly, often within 10,000 years of deposition (Ho and Coleman 1968). This length of time is of the same order of magnitude as that suggested (in 4.3.7) for the upper end of the time span in which an evolved palaeosol may have formed. This coincidence of time scales suggests that in all cases the penetrative oxidation of organic rich horizons in the mudstone sequences occurred at the same time as the formation of evolved, well drained palaeosols on the alluvial surface. Organic material and siderite which was not oxidized at this time was then liable to be preserved in the sediment, partly owing to the drop in permeability in the mudstones, and partly owing to its increasing depth of burial with continued deposition. This would take it beyond reach of the oxygenated groundwaters introduced by the fluctuations in groundwater table which accompanied by the formation of well drained soils.

Examination of vertical sections measured through the Formation (Figs 103-111; Appendix 1) shows that any individual site in the depositional basin was only rarely occupied by a fluvial channel. This being the case, it is likely that the fluvial channels were incised into and/or overlay sediments which had been well away from channel influence, and affected by pedogenesis and its concomitant penetrative oxidation, for a considerable period of time. If a round figure of 10,000 years is taken as the maximum duration of formation of an evolved soil, a sequence of sediments such as that underlying the fluvial channel deposit in Fig 110 had probably been removed from channel influence for between 10,000 and 20,000 years at the time of channel avulsion. The sediments were thus probably thoroughly oxidized and fairly well lithified. The concentration of the effects of post-depositional oxidation in mudstones is probably entirely due to its having occurred soon after deposition in sediments which were persistently remote from active channels.

There is insufficient evidence as to the control of these long intervals separating phases of channel activity. Possible constraints are discussed in 8.5.

The lack of penetrative oxidation in channel deposits after their burial is more enigmatic. The most likely explanation again involves the very long recurrence intervals of the fluvial channels. The pore water within the sandbodies was non-oxidizing, owing to the presence of organic matter within the sand sediment. Fluid flow probably took place within the sand body, derived ultimately from infiltration of surface water near the margins of the depositional basin. Thus, at any time during the early diagenesis of the sandstones, a constant stream of non-oxidizing pore water flowed through them. This water was constricted within the sandbodies, as these were surrounded by very much less permeable mudstones, and, especially, underlain by fairly well lithified sediment. The hydraulics of the channel and overbank sediments were thus completely separate, so that even at periods of low water table and penetrative oxidation in the overbank sediments, the channel sandstones maintained a perched water table, which preserved reducing conditions in the sandbodies. Where sandy sediments were argillaceous, permeability was often reduced sufficiently to prevent circulation of reducing pore waters. In such cases, the sandy sediment behaved as the remainder of the overbank sediment, and was oxidized. This is visible in the upper parts of most channel sandbodies, and, most prominently, at Knutton Quarry (Figs 36, 37) where mainly red sheet crevasse splay sandstones pass laterally into cleaner, green sandstones in the fill of a crevasse channel.

8.4.2 Occurence of red sandbodies

The diagenesis of the overbank deposits associated with red mudstones was the same as has been described above. The red pigmented sandbodies may have formed in two ways.

1) These sandbodies may have undergone late, post burial oxidation during the Permian, in the manner described by Mykura (1960). Three principal lines of evidence support this suggestion.

Firstly, the red sandbodies occur only in a limited area of Mid- and South Staffordshire, within the present Cannock Chase - South Staffordshire Horst. In this area the Etruria Formation is separated from the basal Triassic unconformity (i.e. the Permian land surface) only by a thin interval of the Halesowen Formation, and is locally overlain directly by the Trias (unpublished NCB information). The Halesowen Formation is here in a red bed facies. Palaeomagnetic studies show that the haematite pigment in this Formation was formed during the Permo-Triassic (personal communication by P. Turner), by the oxidation of an initially siderite rich, probably coal bearing sequence. Further evidence for deep Permo-Triassic oxidation in this area is found at Lea Hall Colliery, some 15km. to the north east of Aldridge, where red pigment occurs in permeable sandstones in the coal measures up to 50m. below the basal Triassic unconformity. The magnetic properties of the red Etruria Formation sandstones are not as unequivocal as those from the Halesowen Formation, but suggest that at least some of the haematite pigment was formed after the Carboniferous (personal communication by P. Turner).

Secondly, if the carbonate cements which have subsequently been partially altered to haematite (8.2.3. iii) are, in fact, siderite this suggests strongly that the formation of haematite pigment in the sandstones significantly postdates the early phases of diagenesis.

The third line of evidence is circumstantial. The occurrence of red pigmented sandstones is not facies controlled, red sandstones occurring in sediments of Facies Associations IIB and III. This alone suggests that the diagenetic history was not controlled by syn- or early post-depositional factors.

ii) If the genesis of red pigment in these sandstones was a late diagenetic event, it may be expected that their early diagenetic history was similar to that of the green sandstones. Two features of the red sandstones are not consistent with such an early phase of "green" diagenesis. Firstly, the sandstones in the Aldridge area locally contain abundant basaltic and intermediate rock fragments, in which original igneous textures are preserved, although the mineralogy has been altered. Original textures have not been observed in the chloritic aggregates in the green sandstones, which are regarded as the alteration product of possibly similar basic rock fragments. Secondly, the occurence of early barite cement is limited to the red sandstones. Its occurence is anomalous if these sandstones have undergone the same early diagenesis as the green sandstones.

Both of these features suggest that, at least in some cases, the sandstones were affected by an oxidizing phase of early diagenesis, similar to that described by Walker et al (1978). Such conditions may have occurred if the sandbodies were able to dry out for prolonged periods, preventing the accumulation of organic matter which might subsequently act as a reducing agent (Nagtegaaal 1969). This might have occured in response to climatic fluctuation (cf. Nagtegaal op. cit.), but the existence of contemporaneous green sandbodies elsewhere in the basin suggests that this is unlikely. Prolonged drying out of the sandbodies may also have resulted from their having been deposited at a topographically higher level than the bulk of the Etruria facies, as probably occurred in the case of the Aldridge and in the Redhurst Wood area. However, there is not a simple link between colour and Facies, the Facies Association III sediments in the Telford area being mainly green, while fluvial channel deposits of Facies Association IIB at Cheslyn Hay and Aldridge are red.

In the light of the preceding discussion, it is not possible to determine which of the two mechanisms suggested was responsible for the genesis of red major sandbodies in the Etruria Formation, or, if both were involved, to what extent they contributed to the red pigment. This is an area in which further detailed palaeomagnetic and diagenetic study is needed. Until such work is undertaken, it is not possible to comment on the relationship of the red sandbodies to the diagenesis of the surrounding red mudstones.

8.4.3 Timing of diagenesis

Other than in the case of the red sandstones discussed above, the timing of the diagenesis of the Etruria Formation sediments has been established by palaeomagnetic studies (Besly and Turner: in press; P. Turner, personal communication). All of the rocks in the Etruria Formation posess strong chemical remanent magnetisation, a feature which reflects their high iron contents. In red mudstones, the principal remanence carrier is the fine particulate haematite, while in the green sandstones it is probably magnetite of detrital and/or authigenic origin (P. Turner, personal communication).

Both green and red rocks show similar magnetic properties, with a mean declination of 197°, and a mean inclination of -1°. This is consistent with other Upper Carboniferous results, and gives a pole position of 35°N, 191°E, which agrees closely with the Late Carboniferous pole for northern Eurasia based on the compilation of Irving (1979). The palaeolatitude of the Etruria Formation calculated from the mean direction of magnetisation is virtually on the equator.

The Late Carboniferous palaeolatitude determined from the Etruria Formation demonstrates that the diagenesis of the Formation occurred within a comparatively short period after their deposition. More precise timing of diagenesis may be deduced from consideration of the Upper Carboniferous magnetic reversal stratigraphy.

In all of the mudstones examined there is a single, reversed component of magnetisation. By contrast, directional changes during thermal demagnetisation of two green sandstone samples revealed a dominant reversed component of magnetisation, with a minor superimposed normal component.

The geomagnetic field was mostly reversed during the Carboniferous, but two normal events have been identified in the Westphalian A (Roy 1977) and the Westphalian C (Noltimier and Ellwood 1977), which provide palaeomagnetic marker horizons. The samples examined by Besly and Turner (op cit) probably range from early Westphalian C to early Westphalian D in age. The single reversed component identified in the mudstones suggests that they were magnetised either before or after the Westphalian C normal event, and implies that their diagenesis was completed comparatively soon after their deposition. The superimposed normal component identified in the sandstone samples suggests that their diagenesis was more prolonged, and spanned the Westphalian C normal event, or a later, but still Upper Carboniferous event.

These results are consistent with the idea, expressed in 8.4.1, that the diagenesis of the Formation was predominantly very early, and was controlled by groundwater behaviour. The longer duration of diagenesis in the sandbodies was probably due to their greater permeability, which allowed more prolonged exchange of pore waters and, thus, growth of authigenic minerals.

8.5 Sedimentological Controls on Diagenesis

Sedimentological observations indicate that the increasing importance of red pigment in the coal measure to Etruria Formation transition reflects a change in depositional style from an environment dominated by waterlogged swamp conditions to one of nearly permanent good drainage. For this sedimentological model to be convincing, it must be compatible with the inferred early diagenesis, particularly fluctuation of the water table during red bed formation.

Three sedimentological models are possible.

i) Improved drainage, and a drop in the water table in previously deposited alluvium, may have been caused by channel incision. Within the depositional alluvial system. This could have resulted from an external control such as tectonic movement, or from an inherent process in the alluvial system characterised by well drained floodplains.

Red bed formation as a result of tectonically controlled incision of alluvial channel systems has been described in Pleistocene sediments in south western Papua New Guinea (Paijmans et al . 1971). Here grey alluvial plain mudstones of the Lake Murray Beds, radiogenically dated to be less than 27,000 years old, have been locally dissected as a result of late Pleistocene folding and sea level changes (op. cit. pp. 62 - 63). The area straddles zones of monsoonal and humid tropical climate, which are characterised, respectively, by savannah and rain forest. In both climatic zones extensive reddening of the sediment has occurred during the formation of lateritic and podzolic soils. In the southern part of the area, on the Oriomo Plateau, gley, podzolic, and lateritic soils are present, all of which are characterised by the presence of iron concretions, and which are all accompanied by a greater or lesser degree of reddening (op cit. pp 85 - 87). Beneath one of the lateritic soils, excavations showed that red mottling was present in the underlying sediment to a depth of at least 40 ft. (12m).

This example confirms the potential for very rapid reddening of alluvium to occur in humid and monsoonal tropocal areas. It is difficult to assess the importance of such a reddening mechanism in the case of the Etruria Formation, as the scale of exposures is too small to demonstrate whether, or not, incision of the channels took place during deposition. Friend (1978) has expressed strong doubts as to the possibility of alluvial incision being a common occurrence in depositional landforms. However, in view of the demonstrable contemporaneous tectonism (see Chapter 10), it cannot be excluded that gentle folding and block faulting within the depositional basin may have led, locally, to reddening of the Etruria sediment in a manner similar to that described from Papua New Guinea by Paijmans et al.

If such deformation and incision occurred, their effects are more likely to have been concentrated in marginal areas of the depositional basin, where the Etruria Formation sequences are thinnest, the time span represented by the Formation is longest (see Chapter 9), and where widespread folding and denudation can locally be shown to have taken place during the late Westphalian C and early Westphalian D (see Chapter 10). By the same token, reddening by alluvial incision is unlikely to have occurred in the North Staffordshire depocentre, where the thickest sequence of the Formation is present, deposited in the shortest time interval (see 9.3), with no obvious associated unconformity. In any case, this model for reddening is inapplicable to intercalated red beds such as those found at the Coal Measure/Etruria Formation transition in North Staffordshire (see 6.1.2), where the occurrence of red pigment seems to be related to depositional landforms.

ii) The formation of red sediment may have been controlled solely by improved drainage arising from the progradation of topographically higher, and thus better drained, facies into areas hitherto occupied by waterlogged swamps.

This model is consistent with the progressive change in palaeosol types observed in the upward grey to red bed transition, and also with the conclusion that the bulk of the overbank material (Facies 8 and 9) in Facies Association IIA was deposited in an oxidized state. This control on the distribution of red and grey facies was thus probably the most important one in the transition from Coal Measure to Etruria Formation deposition. It does not, however, explain the penetrative, post depositional oxidation of coals and ironstones in the red bed sequence.

The progradation of better drained floodplain facies was probably tectonically controlled. The preferential occurrence of red beds in areas of lower subsidence rates (see e.g. 6.1.2 and Chapter 9) suggests that changes in subsidence rate may have been sufficient to control the deposition of red or grey facies. Phases of accelerated uplift in the sediment source areas may also have initiated progradation of well drained facies, by increasing the yield of sediment into the despositional basin.

iii) A mechanism whereby penetrative oxidation may have occurred through water table fluctuation without incision of alluvial channels may be speculated upon, in the light of an ecological study of the Upper Nile swamps of Sudan by Rzóska (1974). These swamps occur in a tectonic depression, around the junctions of the five principal rivers which combine to form the White Nile. The area is one of net deposition. Rainfall is strongly seasonal, with peak precipitations from June to November, and is negligable for the rest of the year. This distribution of rainfall is strongly reflected in the behaviour of the rivers in the swamp area, and of the ground water table. During the dry season, flow is restricted to the main alluvial channels, which have associated with them small areas of permanent swamp. Evaporation reduces the discharge of the rivers by at least 50% during their passage through the swamps (ca. 500 km). During the dry season, the small areas of permanent swamp are flanked by very extensive areas of low lying, topographically undifferentiated floodplain. By contrast, during the wet season the river channels are rapidly overtopped, and extensive inundation of the floodplain occurs, accompanied by the deposition of heavy clay.

The descriptions of soils and water table behaviour given by Rzoška (op cit) are, unfortunately, not detailed. It is apparent, however, that the soils of the permanent swamps are perenially water saturated, and resemble the alluvial soils regarded as modern analogues of Palaeosol Type 1. In the seasonally flooded floodplains, the dry season evaporation causes the water table to be permanently very deep ("largely inaccessible ... [in] .. boreholes"). The clay rich soils crack to a considerable depth during the dry season, but during the wet season these cracks close, and the soil becomes impermeable, preventing any recharge of the water table. (This very brief description of the soils suggests that they may be vertisols, characterised by the presence of swelling clays).

Although the facies in the Upper Nile Swamps cannot be regarded as a direct analogue of those in the Etruria Formation, it provides an interesting hint that water table fluctuation may occur in the alluvium of a depositional landform, without any externally or internally controlled incision. The fluctuations of the water table are climatically controlled, in this case by dry season evaporation. Although the soils developed on the Nile swamp alluvium may not be similar to the evolved palaeosols found in the Etruria Formation, lateritic soils are at present forming on an adjacent area of basement, which is only slightly topographically higher than the alluvial plain surface (Morison et al 1948). A strongly seasonal climate is not inconsistent with the sedimentology of the channel sand bodies in the Etruria Formation (see 10.5 .). The existence of such a climate during the Westphalian has been suggested, on independent sedimentological grounds, by Broadhurst et al (1980).

CHAPTER 9

Stratigraphy of the Etruria Formation

9.1 Introduction

The present knowledge of the Productive Coal Measure and Etruria Formation stratigraphy in Central England derives largely from data gained during mining operations in the exposed coalfields and their immediate concealed extensions. Between adjacent coalfields, especially in the south of the area, there are considerable differences in the thickness of these Formations and in the characteristic sequence of coal seams and other markers found in the Productive Coal Measures.

Previous workers on the regional stratigraphy (e.g. Trueman 1947) have implied that such changes in sequence between coalfields result from the regional southerly thinning of the whole succession between a depocentre situated in South Lancashire, North Staffordshire, and West Yorkshire (Wills 1956), and a stable area of low subsidence forming the southern edge of the depositional basin, usually referred to as the Wales - Brabant Massif, the Midland Barrier, or St. George's Land.

As exploration of the concealed sub-Triassic extensions of the exposed coalfields has proceeded, it has become apparent that the lateral transition between the typical sequences of the individual coalfields is not simply achieved by regional thinning of the sequence and amalgamation/splitting of the coal seams. Rather, the lateral transition and associated seam-splitting and thickness changes take place abruptly, usually within a belt of less than 10km width. Between these zones of rapid lateral change, the Productive Coal Measure sequence is characterised by a pronounced continuity of marker horizons, and fairly gentle changes in thickness. Similarly the base of the Etruria Formation is markedly more diachronous in the zones of rapid lateral change in the underlying Productive Coal Measures, than in the intervening areas. The areas of rapid lateral change in both grey and red bed facies usually correspond to mapped fault zones.

In the light of the above, the area under study has been divided up into 'depositional areas' for the purposes of stratigraphic descriptions. These are areas where the Productive Coal Measures have distinctive and consistent sets of coal seams and marker horizons in a sedimentary sequence showing similar lithostratigraphic characteristics and thickness trends over the whole area. They range in size from ca. 300 km^2 (South Staffordshire) to 1000 km^2 (Mid Staffordshire). In addition, for the sake of completeness, data are included from less thoroughly studied areas outside the area of detailed sedimentological study (North Wales, Lancashire, Yorkshire/Nottinghamshire, Lincolnshire).

In this Chapter, stratigraphic information on the occurrence of Etruria red bed facies is collated. Most of these data have been derived from the borehole and shaft records held by the National Coal Board (NCB). The locations of all boreholes and shafts studied are shown on Enclosure 1. Most of these data are unpublished: material cited in the text without a specified reference is derived from this source. Most of the boreholes have been drilled since 1947, and have thorough written logs in which marker horizons and coal seams have been reliably identified by NCB or Institute of Geological Sciences (IGS) personnel. In only a few cases have sections been re-correlated in the light of more recent exploration.

To simplify lithostratigraphic description, and in the absence of sedimentological descriptions for most sections, the sedimentary sequences are described in terms of four very loosely defined "facies assemblages". These are not the same as the Facies Associations described in Chapters 3 to 5. The informal assemblages are:

i) a 'paralic' facies assemblage, forming, in the centre of the Pennine Province, most of the Productive Coal Measures. This assemblage consists of non-marine and marine shales, deltaic and alluvial sandbodies, and coal seams.

ii) an 'alluvial swamp' assemblage, forming a varying, generally minor, proportion of the Productive Coal Measures. This assemblage is identical to that described as Facies Association I in this thesis (Chapter 3). It has been recognised in sequence descriptions from National Coal Board borehole logs in an arbitrary manner, based on the number and spacing of seat-earths. Seat-earth is one of the few facies which can be unambiguously recognised in non-sedimentological lithology descriptions. In the 'paralic' facies assemblage their spacing is irregular, between 4m and 30m, and they make up ca. 15% of the sequence (NCB data; Heward 1976), while in the 'alluvial swamp' assemblage their spacing is more regular, ca. 6m., and they form at least 40% and often more of the sequence (NCB data; see Chapter 3).

iii) a 'transitional' assemblage. This assemblage contains interdigitations of 'alluvial swamp' sediments (Facies Association I) and red beds (Facies Association II). Typical sequences of such interdigitations have been described in Chapter 6. Because of the difficulty in recognising the scale or type of such interdigitations in records of uncored boreholes, no attempt is made to distinguish between 'smaller' and 'larger' scale interbedding.

iv) a 'red-bed' assemblage, containing predominantly red sediments of Facies Associations II and III.

These facies assemblages cannot be referred to as Formations, as they do not conform to the recommendations of Harland _et al_ (1972) for lithostratigraphic nomenclature, lacking clearly defined boundaries, and being extensively interdigitated. However, as facies belts they do have a real lithostratigraphic significance, in allowing the sedimentological and tectonic evolution of the area to be mapped (see Chapter 10).

9.2 Westphalian biostratigraphy

Although the red beds of the Etruria Formation are largely unfossiliferous, the age of their base can be precisely determined from biostratigraphic study of the underlying, interbedded, and laterally equivalent coal bearing strata.

Biostratigraphic subdivision of the Westphalian is based on three major faunal and floral groups.

i) _Goniatites._ Throughout Europe the Westphalian rocks contain widespread marine intercalations in predominantly non-marine successions. The most important and extensive of these are characterised by goniatites of the genera _Gastrioceras_ and _Anthracoceras_, and are used to divide the lower part of the Westphalian into the stages A, B and C (Ramsbottom _et al_ 1978). The highest marine band in the British Westphalian is the _A. cambriense_ band, conventionally taken to divide the Lower Westphalian C from the Upper Westphalian C. Locally the marine bands may lack their characteristic goniatite, but may contain a distinctive marine fauna (Calver 1968), or be recognisable by their occurrence in a distinctive position with respect to other marker horizons, usually coal seams. The marine faunas tend, in all cases, to become impoverished towards the edges of the depositional basin.

In the following description of Westphalian stratigraphy, the identification of marine bands by NCB or IGS geologists has been used throughout: no inconsistencies have emerged. The nomenclature used corresponds to the standard nomenclature of English marine bands proposed by Ramsbottom *et al (op cit).* To facilitate comparison with existing literature, local marine band names specific to individual coalfields are added in brackets.

ii) *Non-marine bivalves.* A zonation of the English Westphalian based on the occurrence of bivalves of the genera *Anthraconaia, Carbonicola, Anthracosia,* and *Anthraconauta* was introduced by Trueman (1933), and has been refined by various workers, notably Weir (in Trueman and Weir 1946 pp. xxvi - xxxiv), Calver (1956), and Eagar (1956). The latter recognised that, within the bivalve zones, 'faunal belts' could be recognised on the basis of changes in the composition of the bivalve communities following widespread coal forming events.

The presently accepted bivalve zonation is sumarized in Fig. 6.

iii) *Miospores.* Miospore zonation of the British Westphalian is based on the comprehensive work of Smith and Butterworth (1967). Butterworth and Smith (1976) have also discussed miospore zonation with specific reference to the identification of the Westphalian D and Stephanian in Britain and on the continent.

In the present study, miospore zonation has been used to confirm the completeness of the Westphalian A to C succession in places where the marine band and bivalve zonation is not known, and to identify the base of the Westphalian D.

9.3 North Staffordshire depositional area

9.3.1 Location and structure

This depositional area (Fig. 180) occupies a south plunging syncline at the south western corner of the Pennine chain. It is bounded on the west by the Western Boundary and Red Rock faults, which have a large post-Triassic westward downthrow. The available information on Carboniferous rocks to the west of these faults is summarized in 9.10. The eastern boundary of the area is not recognisable, owing to post-Carboniferous erosion. The southern boundary, between the North and Mid-Staffordshire depositional areas lies between Stafford and Stone, and is further discussed below.

9.3.2 Stratigraphy of Etruria Formation and related facies

A complete Westphalian sequence is present in North Staffordshire, the Productive Coal Measures and Etruria Formation being conformably overlain by Westphalian D Newcastle and Keele Formations. In the Productive Coal Measures all of the major marine bands, and all of the non-marine bivalve zones are recognised (Ramsbottom et al 1978).

The relationship between the facies assemablages is strongly diachronous. Although the transition from the 'paralic' to 'swamp' assemblage is difficult to identify, it is probable that most of the succession above the Cambriense (Bay) Marine Band consists of seat-earth dominated mudstones with occasional alluvial channel sandstones. From the Great Row seam upwards, the oil shales and blackband ironstones, typical of Facies Association I in this area, are also present. The bases of the 'transitional' and 'red bed' assemblages are much easier to identify from borehole logs. The red intercalations forming the 'transitional' assemblage have already been described, from most of the area, in 6.1.2. The horizon of the first appearance of red intercalations exhibits strong diachronism to the south west, and to a lesser extent, to the south east (Figs. 181-184). Thus, the lowest red intercalation at Parkhouse No. 3 shaft, near the northern outcrop limit of the Etruria Formation, lies ca. 25m above the Blackband Coal (Figs. 182, 183). To the south and west, red intercalations appear at progressively lower horizons until, in the Radwood and Sidway Mill boreholes at the southwestern explored limits of the depositional area, the diachronism of the red beds is so pronounced that coal seams and red intercalations above the Winghay coal can no longer be correlated. In Sidway Mill borehole, the first red intercalation of the 'transitional' assemblage is 37m below the Winghay coal, 16m above the Edmonia (Rowhurst Rider) Marine Band. In this borehole and Radwood borehole the Cambriense (Bay) Marine Band is not present. In the south east of the area, the first red intercalation in the Kibblestone borehole occurs near the horizon of the Bassey Mine coal.

The base of the continuous 'red bed' assemblage sequence also shows marked diachronism, paralleling that of the 'transitional' assemblage. At Parkhouse No. 3 shaft it lies ca. 80m above the Blackband coal, at

Moddershall borehole at about the horizon of the Bassey Mine coal, at Stabhill borehole ca. 30m. above the Great Row coal, and at Sidway Mill ca. 40m above the Winghay coal.

The entire thickness of strata above the Cambriense Marine Band, including the whole of the Etruria Formation, has been referred to the late Westphalian C Phillipsi chronozone (Ramsbottom et al 1978). The inclusion of the Etruria Formation in this zone is based on the sporadic occurrence within the Formation of Anthraconauta phillipsi and Anthraconaia cf. saravana (Trueman and Weir 1946-1958). The extent to which A. phillipsi is facies controlled is not known. Palynological studies by Butterworth and Smith (1976) have suggested that much of the Etruria Formation in North Staffordshire is, in fact, very early Westphalian D in age. This conclusion is based on the occurrence of Thymospora obscura, T. pseudothiesseni, Cardiospora magna, and Schopfites dimorphus, all of which are diagnostic Westphalian D forms, in a coal seam sampled at the surface in a quarry in Chesterton and in coals near the base of the Formation in the Pie Rough borehole. The former coal, occurring near the base of the Formation is probably at the same horizon as a coal and ironstone referred to by Gibson (1901) as containing A. phillipsi. The occurrences in Pie Rough borehole are less reliable, as the samples were cutting samples, and the borehole was uncased (Millot et al 1946) rendering contamination by caving from coals in the base of the Newcastle Formation possible.

9.3.3 Facies associations present in the Etruria Formation

From the available borehole and outcrop data, it is apparent that the entire Etruria Formation consists of sediments of Facies Associations I and II.

9.3.4 Subsidence pattern

Because of the great thickness of the Westphalian A to C succession (including the possible earliest Westphalian D parts of the Etruria Formation), very few data exist as to thickness variations in the succession as a whole. The total thickness of the succession to the north of the Tunstall area is ca. 2100m. (Evans et al. 1968; Gibson 1905; NCB data). The thickness decreases to the south west and east, being less than 1200m to the south of Longton (NCB data), and thinning sharply onto the Western Anticline (Gibson 1905). Malkin (1961) suggested that, superimposed on the progressive thickening to the north, there was a centre of subsidence which ran more or less axially along what is now the axis of the main North Staffordshire coalfield syncline (Potteries syncline). This is supported by the isopach maps (Figs. 164, 165) which have been produced during the present study. As mentioned in 6.1.5, there is not always a correlation between subsidence rate, as recorded by thickness of sediment preserved, and the early occurrence of red bed facies. This is particularly noticeable in the Western Anticline area.

Amalgamation of coal seams is only recorded in two areas. Along the extreme western margin of the coalfield, the Yard, Ragman, and Rough Seven-Feet coals in the lower part of Westphalian B amalgamate to form a 'thick coal' (Evans et al 1968; NCB data from Bowsey Wood borehole), and the same group of seams also are amalgamated at Holt's Barn borehole.

The nature of the southern boundary of the North Staffordshire depositional area is not clear. However, a major change in the sequence occurs in the 7.5km gap between the southernmost proving of the North Staffordshire succession, at Moddershall borehole, and the northernmost proving of the Mid Staffordshire succession, at Enson borehole. Neither of these boreholes prove the Subcrenatum Marine Band (base Westphalian), and the Cambriense Marine Band is not present at Enson. The interval between the Aegirianum and Vanderbecki Marine Bands decreases by 40% from 464m at Moddershall to 289m at Enson. This thinning is accompanied by a change in the sequence of seams present between that of North Staffordshire, which is consistent for at least 25km to the north, and that of Mid Staffordshire which remains unchanged for a similar distance to the south. The upper seams of the North Staffordshire sequence are lost completely in this transition. At the same time the base of the Etruria Formation appears ca. 80m lower in the succession (150m above the Shafton (Priorsfield) Marine Band in Moddershall borehole; 70m above the Shafton (Sylvester's Bridge) Marine Band in Enson borehole).

It is assumed that this change in sequence results from extensive seam splitting and thinning in the Westphalian A and B, associated with a syndepositionally active fault. A similar control has been described in more detail at the boundary between the Mid and South Staffordshire areas, when seam splitting and thickness changes were controlled by the Bentley Fault (see 9.4 and 9.7). It is probable that the southern bounding fault of the North Staffordshire coalfield, the Swynnerton Fault, may have been the structure responsible for the North to Mid Staffordshire sequence change. This fault, which now has a southerly downthrow, is related to the Church Stretton Fault system, which was undoubtedly active during the Westphalian (see 9.9, 9.10).

9.4 Mid Staffordshire depositional area

9.4.1 Location and structure

In the Mid Staffordshire area (Fig. 185) the Productive Coal Measures contain the sequence of coal seams characteristically developed in the district around Cannock. The depositional area includes the whole of the Cannock Chase Horst, between the Swynnerton Fault near Stone and the Bentley Faults between Wolverhampton and Walsall. It also extends into part of the area below the thick Triassic fill of the Stafford Basin, and below the Trias to the east of the Cannock Chase structure in an area bounded, approximately, by lines stretching between Rugeley, Burton-on-Trent, and Sutton Coldfield. At all of the boundaries of the depositional area there is a rapid transition into the more condensed sequences typical of the Coalbrookdale, South Staffordshire, and Warwickshire areas, and into the thicker sequences of Leicestershire/ South Derbyshire and North Staffordshire. The latter transition has been described previously. The boundary between the Mid Staffordshire and Coalbrookdale areas is discussed in 9.9. That between the Mid and South Staffordshire areas occurs within a belt ca. 5km wide associated with the W-E trending Bentley Faults (see 9.7; Whitehead and Eastwood 1927; Mitchell 1945). The transition from the Mid Staffordshire to the Warwickshire depositional area occurs in the 6km. separating the boreholes at Whittington Heath and Comberford Lane, while that between the Mid Staffordshire and South Derbyshire depositional areas occurs in the 2km separating the Ryelands and Catton hall boreholes. The boundary between the Mid Staffordshire and both the South Derbyshire and Warwickshire areas occurs at the line of the Birmingham/Hints fault system, which at present forms the eastern boundary of a large horst extending between Tamworth and Rubery. The throw on this fault system has been reversed since the Westphalian C, when, to the south of Tamworth, it formed the western boundary of a large horst of pre Westphalian sediments, which occupied an area between Birmingham and the Warwickshire coalfield which is now infilled with a large thickness of Triassic sediments (see 9.6).

Between Burton-on-Trent and Rugeley the Westphalian subcrops beneath the Triassic. The original extent of the Mid Staffordshire depositional area in this tract is not known. However, at Chartley, 7km, north north-west of Rugeley, a borehole has shown the Westphalian to be absent, the Trias resting on Carboniferous Limestone. The Chartley borehole is only 4.5km north east of Parkhouse borehole, where a full Westphalian sequence is present. It seems from this, that the intervening eastern bounding fault of the Cannock Chase Horst - the Sandon Fault - locally forms a sub-Triassic tectonic boundary to the Westphalian basin. To the north and east of Chartley nothing is known of the sub Triassic structure.

9.4.2 Stratigraphy of Etruria Formation and related facies

A full Westphalian sequence is present in most of the depositional area, although the absence of many of the minor marine bands and some of the bivalve faunal belts (Ramsbottom *et al* 1978) suggests a more marginal position in the depositional basin than that occupied by the North Staffordshire area. The sequence is locally incomplete in two respects: i) to the south of Cannock, the basal Subcrenatum (unnamed) Marine Band is absent, and the Westphalian onlaps onto pre Carboniferous rocks; ii) throughout the depositional area there is a local unconformity beneath the Halesowen Formation, which has the effect of thinning or occasionally removing the Etruria Formation. Of the major marine bands, the Vanderbecki (Stinking) and Maltby (Sub-Brooch) Marine Bands in the Westphalian B, and the Aegirianum (Charles), Edmondia (Kendrick), and Shafton (Sylvester's Bridge) Marine Bands in the Westphalian C, are found over the whole area. The Cambriense (unnamed) Marine Band has a limited extent to the north of Cannock.

Grey coal measures extend up to and locally above, the Shafton (Sylvester's Bridge) Marine Band. The stratigraphy of these has been described in detail by Barnsley (1965), from whose description it is apparent that the majority of this sequence belongs to the 'paralic' facies assemblage. Seat - earth dominated 'swamp' assemblage sediments occur at two horizons, below the Mealy Grey coal, and above the Edmondia (Kendrick) Marine Band.

Below the Mealy Grey coal seam a 20m thick package of sediments ca. 20m above the base of the Westphalian consists almost entirely of seat earths , thin coals, and carbonaceous mudstones. Over much of the area there are red beds reported at this horizon, which may belong to the 'transitional' facies assemblage. Above the Edmondia (Kendrick) Marine Band, seat earths locally form 100% of the sequence, and have frequent brown colouration and sphaerosiderite associated with them.

The first appearance of the 'transitional' facies assemblage above the main Productive Coal Measures takes the form of red intercalations in the seat earth dominated 'swamp' assemblage sequences. The stratigraphy of the 'swamp' and 'transitional' assemblages and the position of the base of the continuous red beds is often not clear, as this part of the succession is seldom cored, and the old mining records in the southern part of the area are generally poor. Where the base of the red bed succession has been cored, as in Playground No. 8 borehole (Fig. 153) and Devils Dumble borehole (Fig. 188), it is evident that it is similar to the transitional sequence in North Staffordshire, with both 'larger' and 'smaller' scale red intercalations present (cf. 6.1.2).

In the north of the area (e.g. in Allotment No. 1 borehole; Fig. 186) the first appearance of red beds is some 100m above the Shafton (Sylvesters Bridge) Marine Band. In this borehole there is no 'transitional' assemblage, and elsewhere in the vicinity this assemblage is very thin. Over much of the area the base of the red beds occurs at this horizon, about 90m above the Shafton (Sylvester's Bridge) Marine Band around Cannock and Aldridge (Fig. 186), and 60m above it in the Rugeley/Lichfield area (Fig. 187).

A consistent pattern of diachronism is visible near large structures bordering the depositional area. This is most apparent in three areas:

i) In the eastern part of the Cannock Chase Horst, to the east of Stafford and to the north of Cannock, the bases of both 'transitional' and 'red bed' assemblages occur at progressively lower horizons towards the east (Figs. 188, 189). The lowest occurrences of the base of the transitional 'assemblages' (in Parkhouse and Bricklawn boreholes) are ca. 30m above the Aegirianum (Charles) Marine Band, and higher marine bands are not recognisable. The area of this strong diachronism corresponds to an area of marked thinning in the underlying Coal Measures. Between Blackheath and Bricklawn boreholes, the thickness of sediment between the base of the 'transitional' assemblage and the

Aegirianum (Charles) Marine Band drops from 140m to less than 30m in 2.5km. This very pronounced change might suggest the presence of an unconformity. The presence, however, of interbedded grey and red beds in the cored sequences of Brancote Gorse, Blackheath, Berryhill, and Devil's Dumble boreholes and their inferred presence from the cuttings descriptions of Parkhouse and Bricklawn boreholes, suggests that the relationship between grey and red beds is, at least in part, diachronous.

In the area to the east of Stafford, red beds appear progressively earlier in the sequence towards the Sandon Fault, which locally forms the eastern boundary of the coalfield. If the relationship between grey and red beds is diachronous, this may suggest that this structure was active during deposition, and that it may have formed the north eastern boundary to the Mid Staffordshire depositional area. Possible confirmation of this is found in a slight northward diachronism in the base of the 'transitional' assemblage in the Rugeley area (Fig. 187), and in the northward deterioration of the Bottom Robins coal, and lateral passage into a seat-earth dominated sequence, in the area to the north of Lichfield (E.L. Boardman, personal communication).

ii) Near the western margin of the depositional area, the base of the red bed succession may occur at a lower horizon. The first appearance of red beds is at 52m above the Aegirianum Marine Band in Ashflats borehole (Fig. 188), and 24m above the same marine band in the Gravelly Way borehole (Fig. 190). In both cases the base of the red beds is further above the Aegirianum Marine Band in nearby sections. The Gravelly Way borehole is very close to the boundary between the Coalbrookdale and Mid-Staffordshire depositional areas, and it is possible that the succession proved in it could better be compared with that in the Coalbrookdale area (see 9.9).

iii) At the eastern edge of the area, 'transitional' assemblage red intercalations appear at progressively lower horizons to the east of Lichfield (Fig. 191). Thus at Sandyway borehole the first red beds are 110m above the Aegirianum (Charles) Marine Band, at Whittington Heath borehole ca. 70m above it (ca. 50m above the Bottom Robins coal), and at Bowman's Bridge borehole 27m above it (underlying the Bottom Robins coal).

The age of the bulk of the Etruria Formation in this area is not known. The only determinable fossils, plants in cores from Gravelly Way borehole (Crookall, in Mitchell 1945), do not allow distinction between Westphalian C and D ages (C. Cleal, personal communication).

9.4.3 Facies Associations present in Etruria Formation

In the northern part of this depositional area, in Allotment No. 1 borehole, the Etruria Formation consists entirely of rocks of Facies Association II. Further south, in exposures at Rosemary quarry, Redhurst Wood west quarry and Essington quarry, rocks of Facies Associations II and III are present. These are obviously interbedded at Redhurst Wood west quarry (Fig. 148), although the scale and extent of this interbedding, both in the area and time, is not known.

In the Aldridge area, the lower part of the exposed Etruria Formation sequence, in Atlas quarry, consists of Facies Association II sediments. The remainder of the sequence, in Empire and Utopia quarries, consists of Facies Association III, containing sheets of very immature conglomerate.

In both Atlas quarry and the Cannock area, the channel sandbodies of Facies Association II are much coarser than they are, for instance, in North Staffordshire.

No sedimentological information is available for the Etruria Formation to the east of the Cannock Chase Horst.

9.4.4 Subsidence pattern

Data are too scattered to allow construction of a reliable isopach map for the depositional area. It is apparent (Barnsley op cit, NCB data) that subsidence was fairly uniform over the whole area, showing a progressive increase to the north, the Productive Coal Measures thickening from 360m at Holly Bank Colliery to more than 750m at Sandon Bank borehole.

There are two major anomalies in this pattern, both associated with structures near the margins of the depositional area. One is in the thinning observed adjacent to the Sandon Fault, to the east of Stafford, which has already been mentioned. The second is in the area around Brinsford, where, in a very small area, the Productive Coal Measure sequence thins dramatically, and is intensely folded beneath the sub-Halesowen unconformity (Barnsley op cit, Hoare 1959). The stratigraphy of the Etruria Formation in this area is not known, as, firstly, it has probably all been removed by the pre-Halesowen denudation, and, secondly, the sub-Halesowen surface is often reddened, and has been mistakenly described as "Etruria Marl" in many old mining records. However, in the one nearby borehole, at Saredon Hill, that has penetrated a sequence extending above the Aegirianum (Charles) Marine Band, no red beds are present up to the sub-Halesowen unconformity, 25m above the Shafton (Sylvester's Bridge) Marine Band (Fig. 190). It thus appears that early red bed formation was not associated with the Brinsford 'high'.

9.5 Leicestershire and South Derbyshire depositional area

9.5.1 Location and structure

This area (Fig. 192) consists of an exposed coalfield centred on the town of Ashby de la Zouch, together with sub-Triassic extensions to the north, west and south. The area is divided into two synclinal basins separated by the NW-SE trending Ashby Anticline. The eastern basin forms the Leicestershire Coalfield, bounded on the east by the Thringstone Fault, a reverse fault throwing Precambrian to Namurian rocks against the Westphalian. The western basin forms the South Derbyshire coalfield.

In the Leicestershire coalfield the sequence above the Haughton (unnamed) Marine Band is not present. Over most of the South Derbyshire coalfield the succession above (approximately) the Maltby (Two Foot) Marine Band is absent. However, higher horizons, up to the Shafton Marine Band, occur in an outlier around Moira, and the Etruria, Halesowen, and Keele Formations have been encountered in a sub-Triassic downfaulted trough, 5 km. to the west of Moira.

The relationship between the Leicestershire/South Derbyshire depositional area and the adjoining Mid Staffordshire and Warwickshire areas is not known. As suggested in 9.4, the South Derbyshire/Mid Staffordshire transition occurs sharply at the line of the sub-Triassic continuation of the Hints Fault. Not enough of the succession is preserved at the western edge of the South Derbyshire area for this transition to be documented. However, the South Derbyshire succession, up to the Aegirianum (Overseal) Marine Band, appears to be thicker than the comparable succession in Mid Staffordshire (ca. 730m as opposed to ca. 550m: Ramsbottom et al 1978), although the 'transitional' and 'red bed' assemblages appear earlier in the succession in South Derbyshire (see below). Between the South Derbyshire and Warwickshire areas the Westphalian is absent over a sub-Triassic anticline (Trias resting on Cambrian in boreholes at Twycross and Market Bosworth). The nature of the boundary between the depositional areas is thus not known. The transition involved a thickening of the succession up to the Aegirianum Marine Band from ca. 260m in Warwickshire to ca. 730m in South Derbyshire (Ramsbottom et al. op cit), and must therefore have involved considerable differential subsidence, probably controlled by a major structure.

9.5.2 Stratigraphy of Etruria Formation and related facies

A complete Westphalian succession is present in the South Derbyshire coalfield, although varying amounts of the sequence above the Aegirianum Marine Band are locally absent, owing to an unconformity below the Moira Grits (= ?Halesowen Beds equivalent; Worssam et al 1971).

The major part of the Productive Coal Measures compares with the southern part of the Nottinghamshire and North Derbyshire coalfield (Greig and Mitchell 1955), and probably consists of 'paralic' assemblage sediments. Seat-earth dominated 'swamp' assemblage facies occur, in the Moira area, from ca. 4m beneath the Maltby (Two Foot) Marine Band (Fig. 193) to the base of the Etruria Formation (Worssam et al 1971; Worssam 1977). This part of the succession, known locally as the Pottery Clay Formation, consists almost entirely of seat-earths, with thin impersistent coal seams and lenticular southerly derived immature sandstones. (A. Knight, personal communication; Worssam 1977). In Hanginghill Farm borehole (Worssam op cit) brown colouration and sphaerosiderite are of common occurrence above the Shafton Marine Band. It is interesting to note that, despite this sequence consisting dominantly of subaerially exposed, probably alluvial sediments, all of the marine bands are present.

5km to the west, in the Church Flats and Grange Wood boreholes (Fig. 193), brown coloured sphaerosideritic seat earths occur from 10m. above the Aegirianum (Overseal) Marine Band, but it is not possible to identify the lowest occurrence of the 'swamp' assemblage from the borehole description. The base of the continuous red Etruria sequence occurs ca. 50m. above the Aegirianum Marine Band.

In view of the very limited occurrences of red beds in this depositional area, it is not possible to observe any diachronism in their occurrence. It is not known which Facies Associations are present in the Etruria Formation.

9.6 Warwickshire depositional area

9.6.1 Location and structure

This depositional area (Fig. 194) falls naturally into two distinct subdivisions, which will be discussed separately. These are: i) the area at present occupied by the Trias-filled graben lying between the South Staffordshire and the Warwickshire Horsts; and ii) the Warwickshire Horst. Both of these areas maintain their identity for some distance to the south, although their boundaries are not mappable at the surface to the south of the southern limit of Westphalian exposures.

In the area between the two horst structures, stratigraphic control of the Westphalian is very poor. Three boreholes have been drilled in the north of the area, between Sutton Coldfield, Tamworth, and Coleshill. One of these, Blythe Bridge borehole, failed to reach the bottom of a very much thickened Triassic sequence, stopping at 1066m. in Keuper Sandstone. Trickley Lodge borehole encountered the Halesowen Formation resting unconformably on Cambrian shales and Coton Hall Farm borehole found the Halesowen Formation resting unconformably on Dinantian limestone. The results of these boreholes suggest that the Westphalian A to C may be absent over this entire area, a view supported by the stratigraphy and distribution of conglomerates in the Etruria Formation in the adjacent Warwickshire coalfield (see 9.6.3; 10.1.1; 10.2). The lack of Westphalian in this area is borne out by the results of the Ultramar Netherton-1 borehole, situated ca. 40km to the south west (Enclosure 1), near Evesham, where the Trias rests directly on acid tuffs of either Lower Palaeozoic, or, more probably, Precambrian age.

From this limited stratigraphic control, it appears that the Trias filled graben between the South Staffordshire and Warwickshire Horsts is a classical inversion structure. The area was uplifted as a horst prior to the deposition of the Halesowen Formation, and became a branch of the Permo-Triassic Worcester Graben after the subsequent reversal of throws on the bounding faults. In subsequent discussions, the northern part of this structure, which is immediately adjacent to coalfields containing Etruria Formation sequences, will be termed the West Warwickshire Palaeohigh. Evidence for the existence of a Westphalian A to C cover to all or part of this horst is reviewed below.

The remainder of the Warwickshire depositional area occupies a horst, bounded by the Western Boundary and Polesworth Faults. Within this horst, the structure consists of a major southward plunging syncline, which passes beneath an unconformable Triassic cover to the south of Kenilworth. Westphalian A to C sediments are present as far south as Batsford, in Gloucestershire (Williams and Whittaker 1974), and to the east of Banbury (A.R.L. Jones, personal communication). To the east of the Polesworth Fault Cambrian subcrops beneath the Trias. The original extent of the depositional area in this direction is thus not known. The relationship with the Mid and South Staffordshire depositional areas is complicated by the presence of the West Warwickshire Palaeohigh. The relationship with the South Derbyshire area was discussed in 9.5.

9.6.2 Stratigraphy of Etruria Formation and related facies

In the north of the Warwickshire coalfield a complete Westphalian sequence is present, although locally much of the Etruria Formation and upper part of the Productive Coal Measures is absent owing to pre-Halesowen folding and denudation. To the south of Merevale and Dosthill, the basal Westphalian A is absent, progressive onlap occurring towards Coventry, where the basal Westphalian is referrable to the Communis zone (Ramsbottom et al 1978). The southerly onlap is accompanied by considerable thinning in the Westphalian A to C sequence, from ca. 260m at Tamworth, to ca. 140m at Coventry, and ca. 80m at Moreton Morrell. The thinning is accompanied by the amalgamation of the coal seams in the lower part of Westphalian B to form the Warwickshire Thick Coal. In the northern part of the coalfield only the Subcrenatum (unnamed), Listeri (unnamed), Vanderbecki (Seven Feet), Maltby (unnamed), and Aegirianum (Nuneaton) Marine Bands are present. The first two of these are absent to the south of the Thick Coal split, and the Vanderbecki Marine Band is absent south of Coventry. The occurrence of the Maltby Marine Band is not well documented, being limited to a few boreholes to the north west of Coventry, and a few occurrences in the north of the coalfield (I. Fulton, personal communication). The Aegirianum Marine Band occurs over the whole area, but is absent locally to the west of Coventry, where the base of the Etruria Formation is very low in the sequence.

In the northern part of the coalfield there are few reliable descriptions of the lithologies in the Productive Coal Measure sequence. In Comberford Lane borehole the basal 20m. of the Westphalian consist mainly of seat earths, with brown and sometimes red colouration. Red beds have also been recorded at the base of the Westphalian around Wilnecote and Tamworth (Barrow et. al. 1919; Mitchell 1942 p. 3). Many of the coal seams in the lower part of the Westphalian (below the Yard Coal of Tamworth - top Westphalian A) have thick fireclays associated with them (Mitchell op cit pp 4-5, 29). It thus appears that, in this district, there is an appreciable development of 'swamp' and 'transitional' facies assemblages in the Westphalian A.

The 'swamp' facies assemblage re-appears at about the horizon of the Aegirianum (Nuneaton) Marine Band and, in the boreholes at Amington Hall and Syerscote Barn, passes gradationally upwards into red beds. Mitchell (op. cit) described three exposures of 'swamp' assemblage above the Aegirianum Marine Band, recording that, at Tamworth Colliery, "most of the clays are grey, though a little red staining occurs, causing some resemblance to Etruria Marl". At Polesworth he noted red pigment developed as streaks in a seat earth, while at Albion Brickworks, Dordon, he recorded 60ft (18m) of seat-earths, ironstone nodules and coals, "the general aspect reminiscent of the Pottery Clays of South Derbyshire". The present writer has observed one red pigmented palaeosol at the latter locality.

The age of the base of the Etruria Formation at Wilnecote has been determined palynologically. Miospores extracted from a coal seam low in the red bed succession, ca. 10m above the top of the Productive Coal Measures, at Wilnecote Brickworks are of an assemblage characteristic of horizons between the Aegerianum and Cambriense Marine Bands (lowest Westphalian C - M.A. Butterworth, personal communication).

The relationship between the lithostratigraphy of the Tamworth district and that of the Thick Coal area to the south of Bedworth is not known, there being a lack of reliable shaft and borehole sections. In the Thick Coal area, and to the south, the sequence between the base of the Westphalian and the base of the Thick Coal consists dominantly of seat-earths, often with brown colouration ('swamp' or ?'transitional' assemblage). Between the Thick Coal and the Aegirianum (Nuneaton) Marine Band more varied lithologies are present, and it is not known to which facies assemblage these belong. Above the Aegirianum Marine Band, 'swamp' assemblage sediments predominate, with the transition into red bed facies occurring 50 and 100m. above the base of the Thick Coal (Figs. 195-198). Strong diachronism is present at the base of the Etruria formation (Figs. 195, 196), the base becoming, apparently, older towards the Western Boundary Fault of the coalfield. As in the Stafford area (9.4.2) this may in part be due to an unconformity, but the presence of interbedded 'transitional' grey and red beds suggest that lateral facies change also occurs in this very short distance.

In the boreholes to the south of Leamington the Thick Coal deteriorates, and the Etruria Formation ceases to be recognisable (Fig. 198), possibly as a result of pre-Halesowen erosion. The comparative lack of seat-earths between the Thick Coal and the Aegirianum Marine Band is still recognisable at Southam and Middle Road boreholes, but at Moreton Morrell borehole the entire Westphalian A to C sequence consists of seat-earth dominated facies. Red beds of the 'transitional' assemblage occur here in the basal 10m. of the succession, and from 4m. below the Aegirianum Marine Band upwards.

9.6.3 Facies Associations present in the Etruria Formation

Within the Warwickshire coalfield, two distinct Facies Association zones are recognisable in the Etruria Formation.

i) Within a belt some 5km. wide, adjoining the Western Boundary Fault of the coalfield, the Formation consists entirely of Facies Association III sediments, in which the dominant clast type is an indurated shale. This facies belt was first recognised, although not as such, by Eastwood et al (1923 pp. 69-70). Nearly the full thickness of the Formation, exposed in the Wilnecote and Stoneware quarries, contains sheet conglomerates composed dominantly of shale clasts, groups of such sheets being separated by palaeosols and mudstones. Further south, in a recently excavated drift tunnel at Daw Mill Colliery, the Formation was found to consist of two crudely fining upward cycles, each some 50m. thick, consisting of shale clast conglomerates, mudstones and palaeosols (I. Fulton, personal communication). The predominance of shale clast conglomerates continues in this belt as far south as the boreholes around Leamington.

ii) In the remainder of the coalfield shale clast conglomerates are absent, and the Etruria Formation consists of Facies Association II sediments.

9.6.4 Subsidence pattern

Isopach maps for various intervals in the northern and eastern parts of the coalfield are given by Mitchell (1942). These all show a consistent increase in thickness to the north and east, with an east-west oriented area of maximum subsidence near Tamworth. This pattern is not contradicted by more recent data, although the combination of sedimentary thinning, facies changes, and pre-Halesowen folding make it difficult to produce reliable isopach maps for the area south of Nuneaton.

9.6.5 The Westphalian A to C cover of the West Warwickshire palaeohigh

Three independent lines of evidence suggest that, at least during the Westphalian A and B, the West Warwickshire palaeohigh did not form a physical barrier between the southern part of the Mid Staffordshire depositional area and the Warwickshire depositional area. This evidence is largely circumstantial, but is none the less important in making palaeogeographical reconstructions.

i) The Westphalian A and early Westphalian B isopach pattern in North Warwickshire runs at right angles to the present Western Boundary fault (Fig. 199).

ii) The splits in the Staffordshire and Warwickshire Thick Coals are almost co-linear.

iii) The distribution of the ?Maltby Marine Band in Warwickshire strongly suggests that the main marine connection lay to the west (Fig. 194). If this is the case, the overlying Four Foot coal seam, which has a similar very restricted distribution to that of the ?Maltby Marine Band, may be correlatable with the Brooch coal seam in Mid Staffordshire.

9.7 South Staffordshire depositional area

9.7.1 Location and structure

The South Staffordshire depositional area (Fig. 200) occupies the southern continuation of the Mid Staffordshire horst structures. To the east and south east the area is bounded by the Birmingham Fault, a continuation of the Hints Fault. These faults form the western edge of the West Warwickshire palaeohigh (see 9.6), whose existence to the east of the South Staffordshire area may be confidently inferred by the occurrence of Westphalian D on pre-Carboniferous in Trickley Wood borehole and, in outcrop, at Rubery (Wills 1950), where the Halesowen Formation rests unconformably on sandstones of Llandovery age. To the west, the depositional area is bounded by a complex of faults, of which one forms the Western Boundary Fault of the coalfield. The transition between the South Staffordshire depositional area and the Wyre Forest depositional area to the west is discussed in 9.8. The northern boundary of the depositional area is formed by the Bentley Faults, running between Wolverhampton and Walsall. To the north of these faults, the Westphalian A to C productive coal measures are ca. 350m thick at Holly Bank Colliery and ca. 300m thick at Forge Mill borehole, while to the south of the faults, the productive sequence is ca. 130m thick at Baggeridge Colliery and ca. 180m thick at Hamstead No. 1

borehole. Some of this thinning is the result of diachronism in the base of the Etruria Formation (see below), but it is mainly due to thinning over the Bentley structure, which is accompanied by major seam splitting in the Westphalian A and lower Westphalian B. In a zone between 4 and 8 km wide, the Bottom coal of South Staffordshire splits into the Shallow and Deep coals of Mid Staffordshire, the New Mine coal into the Yard and Bass coals, the Heathen coal into the Upper and Lower Heathen coals, and the South Staffordshire Thick coal into the Benches, Eight Foot, and Park coals.

Within the horst, the structure of the coalfield is dominantly anticlinal, the axis consisting of a series of domes running from north west to south east through Dudley and Netherton.

9.7.2 Stratigraphy of Etruria Formation and related facies

In this area the Subcrenatum Marine Band is absent, and the Westphalian everywhere rests unconformably on pre-Carboniferous, mainly Silurian and Devonian rocks. None of the Westphalian A marine bands are present. The Vanderbecki (Stinking) and Maltby (Sub Brooch) Marine Bands occur over the whole area as far south as Dudley, but are apparently absent in the Halesowen area (Poole 1970). The Aegirianum (Charles) Marine Band is present at Hamstead and Baggeridge but is not present at, and to the south and east of, Dudley (Poole op. cit). The lack of marine bands, especially in the south of the area, reflects a persistently marginal position in the Pennine Basin, and is reflected in the paucity of the non-marine bivalve faunas. Of the twelve faunal belts in the Westphalian A and B, only four are present (Ramsbottom et al 1978).

In almost all parts of the depositional area there is a pronounced unconformity at the base of the Halesowen Formation. This causes the Etruria Formation to be locally thinned or on occasion absent. Locally extremely thick Etruria sequences are preserved in pre-Halesowen Formation synclines (e.g. Manor Colliery Halesowen, 240m of Etruria Formation).

The facies assemblages present in the depositional area have been documented by Poole (1970), based on data from Hamstead No. 1 borehole, in the north of the area, and site investigation boreholes around Dudley. The lack of recent exploration or colliery workings renders lithostratigraphic analysis difficult, especially as the Etruria Formation has been removed by erosion from the exposed coalfield area to the north of Dudley.

'Swamp' assemblage sediments, consisting of seat-earths and carbonaceous shales are prominent in Westphalian A over the whole depositional area, especially in the interval between the Bottom coal and the Vanderbecki (Stinking) Marine Band (Poole op cit). At the base of the sequence in the Dudley area red pigmented seat-earths occur in the basal 10m of the section. To the south, in the Amblecote and Stourbridge area the Westphalian A consists of seat-earths ("fireclays") without significant coal seams (Whitehead and Eastwood 1927), while at Wassel Grove Colliery (Fig. 201) the Vanderbecki Marine Band is probably absent, and most of the Westphalian A consists of 'transitional' and/or red bed facies (Poole op cit).

In the Dudley area, and to the north, the Westphalian B, up to the Maltby (Sub-Brooch) Marine Band, contains more varied lithologies, including the Thick Coal which is locally up to 14m thick. Seat earth dominated 'swamp' facies are again present at about the horizon of the Maltby (Sub Brooch) Marine Band. At Hamstead No. 1 borehole (Fig. 201) the thin, ca. 20m, interval between the Maltby and Aegirianum (Charles) Marine Bands is occupied by 'transitional' assemblage grey and red seat earths, and the base of the continuous 'red bed' assemblage lies ca. 5m above the Aegirianum Marine Band. At Dudley the Aegirianum Marine Band is not present, and the 'transitional' assemblage is found ca. 14m above the Maltby (Sub-Brooch) Marine Band, passing upwards into continuous red beds some 7m higher (data from Poole op cit).

To the south of Dudley seat-earths become dominant and, to the south of Halesowen, the Thick Coal splits into two leaves, which are separated by brown pigmented seat-earths. Most of the remainder of the Westphalian B in this area probably consists of 'swamp' or 'transitional' facies, although in the absence of marine bands and correlatable coals, biostratigraphic control is non existent. This succession, from the horizon of the Thick Coal upwards is typified by the Bogs Farm borehole, where the very precise core descriptions of the NCB allow the recognition of the facies present (Fig. 202).

Elsewhere in the coalfield the base of the 'red bed' assemblage occurs at approximately the horizon of the Aegirianum Marine Band, usually underlain by a 'transitional' assemblage succession similar to that seen in exposures at Ibstock Himley quarry (Fig. 154). Considerable diachronism is present in the area immediately to the south and west of Dudley, as witnessed by the behaviour of the base of the red beds with respect to the Two Foot and Upper Sulphur coal seams (Whitehead and Eastwood 1927; Whitehead and Pocock 1947). These coals are not, however, correlated reliably enough for the pattern of this diachronism to become apparent. Towards the west, more pronounced diachronism is recorded by the occurrence of red beds between the Stinking and New Mine coals at Corbyn's Hall Colliery, Pensnett (recorrelation of seams by R. Hamblin, personal communication). These occur at a horizon of pronounced 'swamp' facies development over the whole depositional area. Red beds are also probably present in this horizon at Claverley borehole, in the Wyre Forest depositional area (see 9.8).

No body fossils have been recorded from the Etruria Formation, but a list of fifty eight plant fossils from the 'Red Clay Series' is given by Arber (1916). Most of the species are long ranging 'Yorkian' (mid Westphalian A to mid Westphalian C) or 'Staffordian' (mid Westphalian C to early Westphalian D) types. Arber noted that some species are more characteristic of the 'Radstockian' (i.e. definite Westphalian D), notably Annularia sphenophylloides, Alethopteris serli, and Pecopteris oreopterida. The first two of these species, if correctly identified, suggest that part of the Etruria Formation in this area is of late Westphalian C age to, possibly, early Westphalian D age (C. Cleal, personal communication).

9.7.3 Facies Associations present in the Etruria Formation

Information on the Facies Associations present in the Etruria Formation in South Staffordshire comes from exposures, which are at present limited to the area around Gornalwood (Fig. 5), and, very indirectly, from previous publications.

The four quarries in the Gornalwood area expose a representative sample of the whole Formation, the base being exposed in the Ibstock Himley quarry, the highest part of the Formation below the sub-Halesowen unconformity at Ketley quarry, and the middle of the Formation at Himley Wood and Smithy Lane quarries. From these exposures it is evident that a large scale interdigitation of Facies Associations II and III is present, the Facies Association III facies tending to contain rather thin, distal sand and conglomerate sheets.

Most published references (e.g. Whitehead and Eastwood 1927; Robertson 1931) to the Etruria Formation concentrate on the areas to the east of Dudley, in the Old Hill, Oldbury, and Dudley Port areas, where clay extraction was concentrated in the last century and up to the 1930's. Here mention is frequently made of the presence of conglomerates. It is not known whether this marks a genuine difference in facies, or merely reflects a bias towards mentioning the more unusual and spectacular rock types present.

9.7.4 Subsidence pattern

The lack of recent and reliable data from large areas of the coalfield makes it possible only to draw very general conclusions regarding the subsidence pattern in the South Staffordshire area. This is compounded by the loss of substantial thicknesses of sediment below the pre-Halesowen unconformity.

Jukes (1859) and Whitehead and Eastwood (1927) both record a gradual but irregular decrease in thickness of the 'Middle Coal Measures' (i.e. the productive Westphalian A and B) from north to south. Some of this undoubtedly results from the earlier onset of the Etruria facies in the south. There is, however, a consistent thinning between various marker horizons, for instance, between the Brooch and Thick coals from 60m. at Tipton to 20m. at Stourbridge (Whitehead and Eastwood op cit).

Superimposed on this pattern, there are considerable local variations in thickness in the vicinity of 'banks' formed by sharply folded anticlines in the underlying Silurian floor, which, at the time of Westphalian deposition, formed loci of slow subsidence. Largely as a result of differential compaction, the Westphalian A to C thins over these structures (Whitehead and Eastwood op cit; Mitchell in Trueman 1954). However, as in the case of a similar structure in the Brinsford area in Mid Staffordshire (see 9.4.4) the stratigraphy of the base of the Etruria Formation does not seem to have been affected by these structures.

9.8 Wyre Forest depositional area

9.8.1 Location and structure

The Westphalian A to C rocks in this depositional area (Fig 203) form the heavily wooded Forest of Wyre, between Bewdley and Cleobury Mortimer. They also crop out in a strip along the west bank of the River Severn between Bewdley and Upper Arley, and in a narrow belt around the Devonian inlier at Trimpley. To the north-east, the Westphalian A to C has been encountered in boreholes underlying Westphalian D and Permo-Triassic rocks in the half graben whose eastern bounding fault forms the Western Boundary Fault of the South Staffordshire area. The Westphalian A to C everywhere rests unconformably on Devonian, and is cut out to the north and the south by the unconformity beneath the Highley (=Halesowen) Formation. The thinning of the Westphalian A to C sediments towards the south is not entirely due to this unconformity, resulting in part from a pronounced southwards onlap (Mitchell et al 1961).

Because of the complete removal of the pre-Halesowen Westphalian sequence to the north and south, the boundaries of this depositional area cannot all be accurately located. The eastern boundary is arbitrarily placed at the line of the Western Boundary Fault of the South Staffordshire coalfield horst. This line is not nearly as well defined as the boundary between other depositional areas, as there is not a marked thickness change between the two sequences, the change being mainly a facies change, involving the proportion and stratigraphy of red beds in the succession and the number and thickness of coal seams (see below). One prominent seam split is present at this boundary. The upper leaf of the South Staffordshire Thick coal splits from the main part of that seam in a westerly direction (Fig. 204), becoming known as the Flying Reed coal. The locus of this split lies some way to the east of the arbitrary depositional area boundary (Whitehead and Eastwood 1927). In a westerly direction, the lithostratigraphy of the Westphalian A to C sequence remains similar as far as the most westward outliers on Titterstone Clee Hill, although there is a marked change in thickness (Fig. 204). The Westphalian geology of this outlier is summarised in 9.8.5.

9.8.2 Stratigraphy of the Westphalian A to C

Only two marine bands are present in the Westphalian A to C succession. The Vanderbecki (Stinking) Marine Band has been recorded in the Alveley boreholes, and possibly in workings at Kinlet Colliery (Whitehead and Pocock 1947 p. 42). The Aegirianum (Eymore Farm) Marine Band is known at outcrop in the Eymore Farm railway cutting (Poole 1966) and has been recorded in the Alveley boreholes, in shallow boreholes near Eymore Farm (Poole *op cit*), and in the Kinlet borehole (Whitehead and Pocock *op cit* p. 42). The latter marine band contains a rich fauna of calcareous brachiopods.

The non-marine bivalve zones and faunal belts are not known. However a full sequence of miospore zones (Smith and Butterworth 1967) demonstrate that a continuous Westphalian A to C sequence is present. Spore floras also enable the correlation of the Highley Brooch coal with the Brooch coal of South and Mid Staffordshire and the Marquis/Fungus coal of Coalbrookdale, rather than with the South Staffordshire Thick coal as indicated by Poole (1970).

To the south, where the succession passes laterally into entirely red bed facies (see below), no biostratigraphic control is possible.

The existing lithostratigraphic nomenclature for the succession is inadequate. In the northern part of the depositional area, lying in Geological Survey map sheet No. 167 (Whitehead and Pocock 1947), the entire Westphalian A to C, up to the unconformable base of the Highley (= Halesowen) Formation, is described as the Kinlet Group, and divided into undifferentiated 'Middle Coal Measures' below the Aegirianum Marine Band and 'Etruria (Old Hill) Marl' above that marine band. This classification is adhered to in Geological Survey map sheet 182 (Mitchell et al 1961), notwithstanding the fact that, while some grey coal measures are present in the more northerly area, the entire succession in the southern area consists of red beds (see below). Further confusion has been added by Ramsbottom et al (1978), who, without any justification, have applied the name Kinlet Formation to describe only the red beds overlying the Aegirianum Marine Band.

To the north of Upper Arley the sequence can be divided into three parts: lower and upper dominantly red bed intervals, separated by a dominantly grey, coal bearing interval (Fig. 204). The Westphalian A consists of interbedded grey and red seat earths, giving rise to a succession similar to that found at Wassel Grove Colliery, at the extreme south of the South Staffordshire area (9.7.2). The Westphalian B contains a group of thin coal seams which correlate with the Thick and Brooch coals of South Staffordshire. Above these, the sequence up to the Aegirianum (Eymore Farm) Marine Band consists entirely of seat-earths, with two or three red bed intercalations. Above the marine band, the sequence consists predominantly of red beds. All horizons in the succession thus belong to the 'swamp', 'transitional', and 'red bed' assemblages recognised elsewhere. Their alluvial origin is reflected in the occurrence of locally derived immature sandstones at all levels.

To the south and west, the grey coal bearing interval in the middle of the succession contains increasing proportions of red beds. Westwards, this tendency is slight, and the principal coals can be correlated to Titterstone Clee Hill (Fig. 204). To the south (Fig. 205) the lateral transition into red beds is complete, the upper part of the grey unit containing abundant red and mottled intercalations in the Eymore Farm boreholes (Poole 1966). No reliable correlation is possible between these boreholes and the Alton No. 1 borehole, 3.5 km to the south west, where the entire Westphalian A to C sequence consists of "red or green or chocolate coloured clays. ... .sometimes mottled, sometimes sandy, with occasional thin bands of "Espley Grit" (Arber 1914). Plants collected by Arber from this borehole showed that this succession correlated unmistakably with the Westphalian A to B coal bearing succession to the north of Upper Arley. The observations of Cantrill (1902) on the Birmingham aqueduct trench, running from west to east across the Wyre Forest some 2km to the north of Alton No. 1 borehole, suggest that the succession encountered in this borehole is typical of that in the southern part of the Wyre Forest depositional area.

9.8.3 Facies Associations present in the "Etruria Formation"

In the only reasonable exposure seen in the Wyre Forest depositional area, at Bayton sawmill, the sediments consist dominantly of sandstone, with subordinate pebbly sandstone and mudstone. This exposure is interpreted to be a sand dominated example of Facies Association II (See 4.6.1, and Fig. 114).

It is not known to what extent this small exposure is typical of the depositional area as a whole. On the Geological Survey map (sheet 182), the Etruria facies rocks in the southern part of the area are mapped as consisting of laterally persistent sand bodies, which form mappable ridges, separated by mudstone dominated lithologies. The Bayton exposure is in an area mapped as mudstone. This exposure and the Geological Survey map, taken in conjunction, suggest that the succession in the Wyre Forest is, at least in the southern part, more arenaceous than the typical Etruria Formation sequences seen in Staffordshire.

The lack of thick conglomerate units in the cored boreholes at Eymore Farm (Poole 1966), Alton No. 1, and the Alveley boreholes, suggests that the sediments consist dominantly of Facies Association II.

9.8.4 Subsidence pattern

There are insufficient data to comment on the subsidence pattern in this area. What information is available is rendered questionable by the lack of biostratigraphic control, and effects of the sub-Halesowen unconformity.

If the (very dubious) correlation in Fig. 205 is accepted, there would appear to have been a slightly thicker succession deposited at Alton No. 1 borehole than in the Kinlet-Alveley area. To the south of Alton No. 1 borehole, the succession thins rapidly owing to onlap.

9.8.5 Clee Hills

The Clee Hills are included in the Wyre Forest area on account of the small extent of the outcrop, and its remoteness from the main Westphalian outcrop. The succession presents both similarities and differences to the Wyre Forest.

There are Westphalian outliers on both Brown Clee and Titterstone Clee Hills. The most complete section is on Titterstone Clee, and is recorded by Kidston _et al_ (1917 p 1068) (Fig. 206). The succession is ca. 420 m. thick, of which the top 30m can probably be correlated with the Westphalian D Highley (= Halesowen) Formation. The basal 30m of the succession is formed by the Cornbrook sandstone, which has been shown by Jones and Owen (1961) to be of Westphalian A to B age. The remainder of the succession consists predominantly of interbedded red and grey seat earths mudstones and contains immature, locally derived 'espley' sandstones throughout. A group of grey mustones, ca. 130m above the base of the succession, contains coal seams which can be correlated with those in the Wyre Forest (Whitehead and Pocock 1947) (Fig 204). The overlying sequence contains at least one marine band, which, on the basis of its fauna of calcareous brachiopods (Kidston _et al, op cit_ p. 1069) can almost certainly be correlated with the Aegirianum (Eymore Farm) Marine Band in the Wyre Forest area.

The succession in the Clee Hills is thus broadly similar in age and lithostratigraphy to that in the Wyre Forest. It is distinguished from the latter by its considerably greater thickness.

9.9 Coalbrookdale depositional area

9.9.1 Location and structure

The Coalbrookdale area (Fig. 206) consists of an exposed coalfield situated between Coalport and Lilleshall, and its eastern extension beneath the Triassic fill of the Stafford Basin. To the north west the area is bounded by the Lilleshall or Boundary Fault, beyond which Westphalian A to C rocks are in general absent (see 9.10). To the west the Westphalian A crops out, resting unconformably on Dinantian and Lower Palaeozoic rocks. South of Coalport the Westphalian A to C is cut out by the sub-Halesowen unconformity, while to the east it passes unconformably below Westphalian D and Triassic cover. The eastern extent of the characteristic Coalbrookdale sequence is known at Stretton borehole, 10 km to the north of Wolverhampton. To the north of this, the boundary between the Coalbrookdale and Mid Staffordshire areas passes to the west of Ashflats borehole. No information is available as to the relationship between the Coalbrookdale and the Wyre Forest and North Staffordshire areas.

The structure of the exposed coalfield is comparatively simple. The Westphalian, much faulted, dips gently to the south and east. Two south west to north east trending synclines, the Donnington and Madeley Wood synclines, are probably largely pre-Halesowen structures.

9.9.2 Stratigraphy of the Westphalian A to C

The basal Westphalian Subcrenatum Marine Band is not present. Otherwise a complete Westphalian sequence is present, on the evidence of miospores (Smith and Butterworth 1967), with two marine bands present in Westphalian A, probably the Listeri and Amaliae Bands

(Ramsbottom et al 1978), and only three marine bands present in the Westphalian B, the Vanderbecki (Pennystone), Maltby (Blackstone), and Aegirianum (Chance Pennystone) Bands.

The most recent description of the lithostratigraphy of this area is in the key to the 1:25,000 geological map of Telford New town (Geological Survey of Great Britain 1978). The Westphalian A consists of about 40m of sandstone overlain by 20m of seat earths with coal seams. This entire succession can be referred to the 'swamp' facies assemblage. The Westphalian B contains more varied lithologies, dominantly argillaceous, but with major sand bodies which locally erode up to 35m of the sequence. There are a number of thick coal seams; of these at least one, the Double coal, probably splits at the boundary with the thicker sequence in the Mid Staffordshire area. This seam is correlated with the South Staffordshire Thick coal.

It is not clear, from the available information, to which facies assemblage the main part of the Westphalian B belongs. The upward transition into red beds occurs above the Maltby (Blackstone) Marine Band, where seat earth dominated 'swamp' assemblages pass upwards rapidly into red beds. Continuous red beds are present from immediately above the Aegirianum (Chance Pennystone) Marine Band (Fig. 207). No pronounced pattern of diachronism is apparent.

Considerable confusion exists as to the extent of the Etruria Formation and its relationship to the underlying Productive Coal Measures and the overlying Coalport (= Halesowen) Formation. Three types of red beds are present in the succession overlying the Productive Measures. These are

i) undoubted Etruria facies red beds resting conformably on Productive Coal Measures in the Donnington syncline (Stonehouse 1951) and seen in exposure at Blockley's quarry and Donnington Wood quarry;

ii) a red weathering mantle developed on the eroded surface of the Productive Coal Measures at the unconformity below the Coalport (= Halesowen) Formation (Hoare 1959);

iii) red beds associated with laterally extensive sheet sandstones in the basal part of the Coalport (= Halesowen) Formation.

In the past all of these red beds have been grouped together and regarded as a basal unit of the 'Coalport Beds' (*sensu* Whitehead *et al* 1928), which were thought to lie unconformably on the Productive Coal Measures. Results of more recent exploration have shown that, in the Donnington syncline, the strata which had previously been regarded as a red basal unit of the 'Coalport Beds' were conformable with the Productive Coal Measures and similar in facies to the Etruria Formation. In recognition of this, these red beds have been mapped separately on the 1:25,000 geological map of Telford New Town (Geological Survey of Great Britain 1978), and given a new Formation name, the Hadley Formation. The remaining upper part of the 'Coalport Beds' is called the Coalport Formation, consisting of laterally extensive sand bodies, thin coals and limestones, and impersistent red beds.

Despite this re-definition, ambiguity still persists. On the new 1:25,000 map, the junction between the Hadley and Coalport Formations, taken at the Main Sulphur coal, is regarded as conformable, while a major unconformity is identified at the base of the Hadley Formation. This is entirely contrary to the succession found in the workings of Madeley Wood Colliery, where grey Coalport Formation facies rest unconformably on Westphalian B Productive Coal Measures. In the Madeley Wood boreholes (Hoare 1959) a similar sequence is present, the only red beds being those forming a weathering mantle on the surface below the unconformable base of the Coalport Formation. In the extreme south of the area, the Hadley Formation is mapped as resting unconformably on all horizons of the Westphalian A and B, even though the rocks beneath the Main Sulphur coal in this area are almost entirely grey, and bear no resemblance to the facies of the Etruria (= Hadley) Formation (Whitehead et al op cit p. 78). Such red beds as are present would not be untypical of the Coalport Formation elsewhere (Whitehead et al op cit p. 79).

The possibility of a local unconformity at the base of the Etruria (Hadley) Formation in this area cannot be disregarded, as to the west of the Western Boundary Fault, in the adjoining North Shropshire and South Cheshire area, the Wrekin Buildings borehole encountered a thin section of typical Etruria Formation lithologies below the Coalport Formation, resting unconformably on Precambrian (see 9.10). It seems likely, however, from the Madeley Wood data and from regional considerations, that the major unconformity in the succession is at the base of the Coalport Formation, and that this can be correlated with the sub-Halesowen unconformity in the Mid and South Staffordshire areas, as suggested by Hoare (1959).

9.9.3 Facies Associations present in the Etruria Formation

Both of the accessible exposures of the Etruria Formation contain sediments of Facies Association III only. It is not known to what extent these are representative.

9.9.4 Subsidence Pattern

The Westphalian A to C thickens progressively to the north east. No information is available as to the nature of the sequence beneath the Trias of the Stafford Basin, but at Stretton borehole the Westphalian A to C is extremely thin. Here erosion prior to the deposition of the Coalport/Halesowen Formation has cut deeply into the pre-Westphalian D sequence, a situation analogous with that in the adjacent Brinsford area. Within the exposed Coalbrookdale coalfield, R. Hamblin (personal communication) has documented similar thinning of the Westphalian A to C sequence in areas now occupied by pre-Coalport Formation anticlines, with complementary thickening in the Madeley Wood and Donnington synclines.

9.10 North Shropshire and South Cheshire area

This area (Fig. 208) occupies a horst which runs between Shrewsbury and Market Drayton, an area to the north of Shrewsbury where, in places, the Trias at outcrop rests unconformably on Lower Palaeozoic, and possibly the southern part of the Trias filled Cheshire basin. The boundary between this area and the Coalbrookdale area is formed by the Lilleshall/Western Boundary fault. The relationship with other depositional areas is not known, although it is likely that, in all cases, the boundaries are associated with faults which, like the Lilleshall fault, are associated with the Church Stretton Fault system.

Over most of the area Etruria Formation sediments are absent, the Halesowen Formation equivalents (Coalport and Coed-yr-Allt Formations) resting unconformably on Lower Palaeozoic to Precambrian in the Edgemond borehole (Wills 1956), in a group of boreholes to the north of Wellington (Kinley Farm, Hoo Hall, Lodge Farm, and Leegomery House Farm) (IGS data), and at the outcrop in the Shrewsbury area. In some of this area the Halesowen Formation equivalents have undergone post depositional oxidation, and have been incorrectly mapped as Keele Formation (R. Hamblin, personal communication). In the Ternhill and Stoke-on-Tern boreholes, the Halesowen Formation rests on Carboniferous Limestone (Wills 1956).

Etruria Formation sediments occur at two places: in the Wrekin Buildings borehole, resting unconformably on Precambrian (IGS data); and in a small outcrop near Woolaston (Earp and Hains 1971), where they rest unconformably on Lower Palaeozoic. At the latter place there appears to be a conformable relationship between the Etruria Formation (or Ruabon Formation: see 9.11) and the Coed-yr-Allt (= Halesowen) Formation.

The only control point in the southern part of the Cheshire Basin is the Prees no. 1 borehole, in which a thin, probably Westphalian D, red coal measure sequence lies unconformably on Lower Palaeozoic (Colter and Barr 1975). It is possible however that the Westphalian A to C may have been faulted out in this well by one of the major basin margin faults, with a throw of several thousand metres (N. Kusznir, personal communication).

9.11 Occurrence of Etruria Formation and Related Facies in the remainder of the Pennine Province

9.11.1 North Wales

Etruria Formation facies rocks are present throughout North Wales (Fig. 208). In the Denbighshire coalfield, between Wrexham and Oswestry, they are known as the Ruabon Formation. To the north, in the Flintshire coalfield and the Wirral the Ruabon Formation is usually very thin.

In the Wrexham area, the base of the red beds occurs at a fairly consistent horizon above the Bersham Yard coal (Wedd et al 1928). To the south, in the Oswestry area, the thickness of the Productive Coal Measures is much reduced, and some amalgamation of coals may have occurred (Wedd et al 1929). Correlation between Oswestry and the Wrexham area is not certain, but the base of the Ruabon Formation in the former area appears to be at a much lower horizon than in the latter (Fig. 210).

To the north of Wrexham, the presence of the Ruabon Formation is not documented in the exposed Flintshire coalfield. The highest Westphalian C present, the Buckley Fireclay Formation, probably contains 'transitional' and red bed facies (Thomas 1961), and Etruria facies with conglomerates are recorded by Wills (1956) in the Sealand No. 3 borehole. The base of these red beds occurs at widely varying horizons (Fig. 211) and it has been suggested that there may be an unconformity at their base (Wedd et al 1928, 1929; Wills 1956). The available data are inadequate to resolve this question.

9.11.2 Lancashire

Red beds of the Etruria Formation are undoubtedly present in the Lancashire coalfield. They are not exposed, and little has been published as to their stratigraphy. A degree of confusion exists as to their stratigraphic nomenclature. Above the Cambriense (Prestwich Top) marine Band there is a coal bearing sequence of Phillipsi zone age called the Bradford Coal Formation (Poole and Whiteman 1954). This Formation consists of seat earth dominated 'swamp' facies, and contains blackband ironstones (Poole and Whiteman op cit pp 297-299). It passes gradationally upwards into red beds, the initial appearance of red pigment being concentrated in the seat earths. The Bradford Coal Formation is overlain by the Ardwick Formation, which contains a lower unit of red and variegated mudstones, overlain by grey mudstones and sandstones containing thick fresh water limestones (Trotter 1953 a). By comparison with the succession in North Staffordshire, it is likely that the red beds in the lower part of the Ardwick Formation and in the upper part of the Bradford Coal Formation correspond to the Etruria Formation, while the grey upper part of the Ardwick Formation corresponds to the Newcastle (=Halesowen) Formation.

The extent of any diachronism present at the base on the Etruria Facies is not known, and is locally confused by the extent of penetrative weathering of the Productive Coal Measures beneath the Permian land surface (Trotter 1953 a).

9.11.3 Yorkshire, Nottinghamshire, North East Leicestershire

Red beds of Etruria Formation facies have been cored in boreholes near Farnsfield, 18 km north east of Nottingham (Gibson 1901, Edwards, 1951, 1967). Their presence is also recorded in the Vale of Belvoir, North East Leicestershire, and in south Yorkshire (Fig 212).

In South Yorkshire near Doncaster the base of the red beds is ca. 160m. above the Cambriense (Top) Marine Band (Smith et al 1973). To the south the base of the Formation is markedly diachronous, occurring ca. 50m above the Cambriense (Top) Marine Band at Farnsfield (Edwards 1967), 33m above the Maltby (Two Foot) Marine Band in the north west of the Vale of Belvoir, and below the Top Bright horizon, some 40m below the extrapolated horizon of the Maltby Marine Band, in the south eastern part of the Vale of Belvoir (E.L. Boardman, personal communication).

9.11.4 Lincolnshire

Observations by the present author on core material from four boreholes near Sleaford (Figs. 213, 213) show that there is an intercalation of conglomerates and red beds within the grey coal measure sucession. This succession is most complete in the Gables Farm borehole where the base of the red beds lies 20m above the Vanderbecki (Clay Cross) Marine Band. The red beds comprise some 30m of immature conglomerates and red mudstones, with some palaeosol horizons. They are overlain by ca. 10m of grey mudstones with two thin coal seams, which in turn are overlain by red sandstones similar in facies to the Keele Formation of the West Midlands. A similar sequence is present in the nearby Ruskington - 1 borehole (Falcon and Kent 1960).

In the four boreholes it is evident that there is a marked unconformity at the base of the conglomeratic red bed unit. The age of the red beds, and the overlying thin coals, is not at present known.

The overlying sequence of Keele Formation type sandstones resembles a sand dominated sequence encountered in the upper part of the Westphalian in oil and coal exploration boreholes to the south and west of Lincoln (Fig. 212, 213). Edwards (1951) suggested that there is an unconformity present at the base of this sequence in the Doddington borehole. The sequence clearly has an unconformable base between the Nocton and Dunston wells. At Nocton it rests unconformably on Namurian Millstone grit (Lees and Taitt 1946). Edwards correlated this unit with the Halesowen Formation.

While there is no evidence of an unconformity at the base of the 'Keele' type sandstones in the Sleaford boreholes, the grey succession beneath them has undergone extensive post depositional reddening. This implies a considerable hiatus in deposition at this horizon (cf. Hoare 1959).

9.12 Summary of the Stratigraphy

From the mass of data presented above, it is possible to distil a number of generalizations regarding the stratigraphy of the Etruria Formation, which will be of use in undertaking the palaeogeographic reconstruction attempted in Chapter 10.

i) Within the area under study, Westphalian A to C (or ? early D) sediments were deposited in a number of distinctive 'depositional areas'. Within these areas characteristic sequences of coal seams and marker horizons were deposited over a time span ranging from Westphalian A to late Westphalian B, and locally to late Westphalian C. Within individual areas, the spacing of these marker horizons demonstrates fairly uniform rates of subsidence. The boundaries between depositional areas, where known, are marked by rapid lateral changes of thickness and character in the coal bearing sequences. Such changes usually involve the amalgamation of separate coal seams and the disappearance of widespread marine marker horizons in the thinner sequence.

ii) The boundaries between depositional areas, where known, usually are associated with the line of a major fault. In some cases these faults are obviously related to the Church Stretton Fault system, which is a major basement fault of great antiquity (e.g. Earp and Hains 1971). Many of the other faults are related to major tectonic features such as the Worcester Graben or the Charnwood Forest Horst, and are probably of equal antiquity. The differing subsidence pattern of individual depositional areas thus probably reflects varying subsidence rates in pre-existing basement blocks.

iii) The total thickness of Westphalian A to C sediments, both grey and red beds, decreases southwards and southwestwards in the area studied, from the depocentre in the Lancashire, North Staffordshire area, to the feather edges of the sequence, which onlap onto the Wales - Brabant High.

iv) Within the Westphalian A to C, red beds occur at all stratigraphic horizons. In all areas there is a gradational sequence, both vertically and laterally, from grey Productive Coal Measures into Coal Measures dominated by palaeosol lithologies, whence into red beds. In the North and Mid Staffordshire, South Derbyshire, the northern parts of Warwickshire and South Staffordshire, and Coalbrookdale areas, red beds occur only above the Productive Coal Measures. In the Wyre Forest and the southern parts of South Staffordshire and Warwickshire, red beds occur both above and below the Productive Coal Measures, and the Westphalian A to C locally passes laterally into an all red bed facies.

v) The relationship between red and grey sediments is in all areas diachronous. The diachronism is more pronounced near the boundaries between depositional areas. Red beds appear earlier in the succession in the thinner sequences (Fig. 214).

vi) There is also strong diachronism between red and grey sediments in areas immediately adjoining horst structures upon which no Westphalian A to C is preserved, and which are known or inferred to have been denuded during the Westphalian A to C.

vii) In all areas, except North Staffordshire and North Shropshire/South Cheshire, there is an unconformity between the Etruria Formation and the overlying Halesowen Formation (or equivalents). On horsts where no Westphalian A to C is preserved, the Halesowen Formation rests unconformably on pre-Westphalian. In the Cheshire area, a thin Etruria Formation sequence lies unconformably on pre-Westphalian, and the Etruria/Halesowen boundary may be conformable. An unconformity may also be present at the base of the Ruabon (=

Etruria) formation in North Wales, although the evidence for this is unconvincing. The base of the Coed-yr-Allt (= Halesowen) Formation in this area may also rest unconformably on earlier rocks.

viii) The acme of coal formation, and the minimum extent of red bed facies, occurred at the time of thick coal seam formation, which was contemparaneous in several depositional areas during the early Westphalian B (Fig. 214).

9.13 Formal lithostratigraphical nomenclature

Up to this point, the formal lithostratigraphical nomenclature of the Westphalian A to C has been studiously avoided. It is apparent that: a) the overall lithostratigraphy of the entire region is fairly simple; and b) the existing lithostratigraphical classification is unnecessarily complex. A brief attempt to remedy this situation is therefore necessary.

With a few exceptions, it has been the general practice to name only coal seams, marker horizons, and major sand bodies in the British Westphalian. Exceptions to this rule are to be found in the late Westphalian, where local and occasionally regional Formation names have evolved for the Etruria and related facies and for the Westphalian D. However, for the bulk of the Westphalian no updating has been carried out to bring the lithostratigraphical nomenclature into line with the recommendations for stratigraphical nomenclature of Harland _et al_ (1972) and Hedberg (1976). Indeed Ramsbottom _et al_ (1978) deliberately, and commendably, refrained from applying strict Formation nomenclature to the British Silesian, on the grounds that the introduction of new Formation names would not result in any increase in clarity or usefulness. The only innovation that they introduced was to standardise the nomenclature of the Marine Bands, and to define them as having the status of Members.

Because the Etruria facies is so markedly different to the Productive Coal Measures, it obviously needs a Formation name which is consistent over the whole Pennine province. Owing to the intimacy of its relationship with the Productive Coal Measures, it is obviously desirable that the latter should also be formally named. To do so need not introduce the unnecessary nomenclature deplored by Ramsbottom et al (op cit). As, within the Coal Measures, several Formations have already been defined (e.g. the Bradford Coal Formation, the Blackband Formation), it follows that the Coal Measures have Group status, and that the Etruria facies, given its close relationship to the Coal Measures, should form a separate Formation within the Coal Measures Group. Since recent exploration has demonstrated the continuity of the Etruria facies between separate exposed coalfields, and the consistent nature of its relationship to the Coal Measures, there is no advantage to be gained from perpetrating the local names by which the Etruria facies is still known in some areas (Hadley Formation, Kinlet Formation, Old Hill Formation). As the section in North Staffordshire was the first to be formally described and named (Gibson 1901, 1905), the entire Formation should be formally called the Etruria Formation.

The argument above summarises the spirit in which the term 'Etruria Formation' has been used in the preceding chapters of this thesis. It remains to define the boundaries of this Formation. There are two principal problems: a) the nomenclature of red beds where these underlie the Productive Coal Measures; and b) the positioning of the boundaries of the Formation in places where grey and red beds are interdigitated.

The first of these questions is unusual in British lithostratigraphy, where Formations tend to be named in vertical type sections, and interdigitation through the migration of facies belts uncommon. It seems unnecessary to invent a new Formation name for the basal occurrence of red beds in the Midlands Westphalian, and it is proposed that this should be referred to as a 'tongue' of the Etruria Formation. This is an 'informal' classification (Hedberg 1976), but one which is often employed in analogous situations in North American lithostratigraphy (e.g. Braunagel and Stanley 1977, Fig 2).

The second question is one of practicality. The base or (in the case of the basal tongue) the top of the Etruria Formation, where it is interdigitated with Coal Measures, could be defined at the first (or last) appearance of red pigment in a vertical section, or at the base (or top) of a continuous sequence of red beds in a vertical section. Although the latter is at present employed in North Staffordshire, the only place where the base of the Formation is accurately defined (Gibson 1905), the present author prefers the former. Defining the Etruria Formation so as to include all significant red beds interdigitated with grey coal bearing sediments has two advantages. It groups into one Formation all the sediments in which reddening processes occurred, and it refers to the Etruria Formation thick sequences, such as those in the Wyre Forest, which are obviously predominantly of Etruria facies, but which contain regular grey intercalations.

9.14 Other occurrences of Etruria and related Facies

Red beds and other facies comparable to those found in, and associated with, the Etruria Formation have been recorded in three other localities in the British Isles.

i) South Wales. Red beds are intercalated with the Rhondda Beds near the east crop of the South Wales coalfield (Howell and Cox 1924; Downing and Squirrell 1965). Although these are the same age as much of the Etruria Formation, they occur in the lower part of the Pennant Group, which was, in the late Westphalian C, probably deposited in a separate basin. The South Wales and Midland basins were probably not linked until the early Westphalian D (Ramsbottom *et al* 1978).

Downing and Squirrell's description of these red beds shows that red pigment is concentrated in seat earth horizons. Their suggested mechanism of red bed formation is very similar to that suggested in this thesis, (see 8.3), with the exception that no penetrative oxidation seems to have occurred.

Red beds of probably similar facies have also been described from the late Westphalian D in the Swansea area (Archer 1965).

ii) Bristol. In the Bristol area red beds are present locally beneath the Aegirianum (Crofts End) Marine Band, and more extensively in the lower part of the Downend Group, between the Cambriense (Winterbourne) Marine Band and the base of the Pennant Sandstone (Kellaway 1970, and personal communication). The red beds contain coarse lenticular conglomerates, composed partly of locally derived detritus. It is possible that there may be a local unconformity at their base. The local derivation of such conglomerates is consistent with the stratigraphy of the area, Westphalian A to C (pre-Pennant) rocks being absent from the nearby Lower Severn Axis and from the Oxfordshire coalfield area to the north east. Kellaway (op cit) has suggested that tectonic rejuvenation of the Lower Severn structure was contemparaneous with the block movements which gave rise to the sub-Halesowen unconformity in the Midlands.

Red beds are extensively developed in the Westphalian D of the Somerset, Gloucestershire, and Oxfordshire coalfields (Kellaway op cit; Poole 1975). It is not known whether these are of Etruria facies, but the correlation of the Oxfordshire sequence with the Halesowen and Keele Formation in Warwickshire, where the red beds are not of Etruria facies, suggests that those in Oxfordshire and Gloucestershire are also not of Etruria facies.

iii) Passage Group, Midland Valley of Scotland. Red beds of Etruria facies and seat earth dominated 'swamp' facies occur in the predominantly fluvial Passage Group, of late Namurian age, in the Midland Valley of Scotland (Read 1969 and personal communication, Read and Dean 1978). This fluvial red bed facies passes laterally into a coal bearing sequence characterised by the presence of very thick coal seams, which occur in a tectonically defined trough (Brand et al 1979).

The occurrence of facies similar to those in the Etruria Formation in the Namurian of a separate basin is of interest, in that it demonstrates that such facies may have had a widespread development, given suitable palaeogeographic settings, rather than being an oddity resulting from a particular set of circumstances limited to the southern part of the Pennine province.

CHAPTER 10

Westphalian palaeogeography of Central England

10.1 Methods employed for palaeogeographic reconstruction

10.1.1 Palaeocurrent and provenance patterns

The nature of the exposures of both the Productive Coal Measures and the Etruria Formation has made collection of palaeocurrent data extremely difficult. In the present study seventy four palaeocurrent directions have been measured or estimated from cross bedding and sole structures, and the orientations of eighteen channel features have been estimated. These data are shown on Fig. 215.

With such a small collection of data, only generalized interpretations may be made, especially as many of the flow directions are measured from Facies whose palaeocurrents are likely to have been controlled by small scale and local sedimentary features rather than by the regional palaeoslope. It is however clear that sediment transport was from south to north consistently throughout deposition of the Etruria Formation.

The distribution of detrital lithic components also sheds some light on sediment transport paths. There is considerable variation in the types of lithic material present in the Etruria Formation sandstones in different parts of the area under study. These presumably reflect the varying geology of the various local source areas.

The principal detrital components, other than monocrystalline quartz grains, are as follows (Fig. 216):

North Staffordshire : ?basic igneous, acid igneous, occasional granite, metamorphic quartzite.

Shropshire : varied assemblage of igneous and metamorphic types, vein quartz.

South Staffordshire (Dudley area) : ?basic igneous, metamorphic, prominent muscovite, quartzite, sandstone.

Mid Staffordshire, and Aldridge area : basalt, quartzite, sandstone, indurated shale; abundant acid volcanic detritus, locally forming the sole detrital component.

Warwickshire : indurated shale (sole component along western margin of area), quartzite, granite.

This list is probably not exhaustive.

The relationship between detrital components and source areas is best demonstrated in Warwickshire, where conglomerates dominated by clasts of indurated shale are limited to the western part of the depositional area, adjoining the West Warwickshire palaeohigh. This structure was probably denuded during the Westphalian (see 9.6), and contains Cambrian shale which subcrops below the Halesowen Formation. The distribution of shale clast conglomerates thus represents local derivation of debris from this high into the Warwickshire area. A similar pattern of local derivation may be inferred for the immature conglomerates in the Aldridge area on the west side of the palaeohigh, which contain a variety of Cambrian and ?Precambrian debris, and for the very immature conglomerates in the Coalbrookdale depositional area.

The inferred local and regional transport patterns of lithic detritus are shown in Fig. 216.

10.1.2 Lithostratigraphic synthesis

By comparing the stratigraphic distribution of red beds with the distribution of marker bands in the Westphalian sequence, a fairly accurate palaeogeographic reconstruction can be made of the progressive spread of the red bed facies during the Westphalian.

The marker bands may in themselves give palaeogeographic information. This is particularly important in two areas : the disappearance of marine bands in marginal areas of the basin; and the splitting of coal seams.

In central areas of the Pennine Basin there are nineteen marine bands in the Westphalian. These can be divided into minor bands, which are not present in most other areas, and major bands, which are present even in very thin and condensed sequences near the margins of the depositional basin. Ramsbottom (1979) regards the varying distribution of the marine bands as the result of their having been deposited during pulsed mesothemic transgressions which were eustatically controlled. This view is discussed in 10.4, in the light of the red bed occurrences.

In the marginal areas of Shropshire, South Staffordshire, and Warwickshire the major marine bands in the Westphalian A are absent, and the only marine bands present are the three major bands of the Westphalian B - the Vanderbecki, Maltby, and Aegirianum. The presence of these marine incursions in predominantly alluvial deposits, sometimes containing red beds, in the Wyre Forest, South Staffordshire, and Warwickshire suggests that these intercalations are genuinely of eustatic origin. In some areas, however, these major marine bands are anomalously absent in areas of apparently continuous deposition. The main examples involve the Aegirianum Marine Band in the area south of Dudley and in the area around Daw Mill colliery in Warwickshire, and the Maltby Marine Band in much of Warwickshire. Such absences may result from the existence of depositional topography as a result of the deposition of Etruria facies prior to the marine incursion (see 10.2.6; 10.2.7).

Seam splitting occurs in two well define settings. Amalgamated seams are developed near the margins of the depositional basin, and split towards the centre of the basin in zones of thickening associated with the synedepositionally active faults which define the boundaries of individual depositional areas (cf. Broadhurst et al 1968). The sediments between split seams in such a setting are of typical Productive Coal Measures facies. Splits also occur towards the edges of the depositional basin in the extreme marginal areas of south Warwickshire, south Staffordshire, and the Wyre Forest. In this case, the sediment separating the split seams consists almost entirely of type 1 palaeosols (seat-earths) in successions of 'alluvial swamp' type (see 9.1, and Fig. 202). Towards the basin margins the coal seams affected by such splits deteriorate in a very small distance, and die out laterally. The origin of these splits is different to that of the 'down to the basin' fault related splits. During deposition of the thick peats making up the amalgamated coal seams, clastic sediment input from the hinterland must have been very low. The vegetation near the basin margin acted as a clastic trap for such sediment, which led to splitting and deterioration of the coal seam towards the edge of the basin. Such clastic trapping may locally have been controlled by syndepositional faulting, for instance in the case of the very pronounced split and deterioration of the South Staffordshire Thick Coal westwards into the Wyre Forest area (Fig. 204).

The two types of seam split, and their environmental interpretation, are summarized in Fig. 217.

10.2 Palaeogeography of the Westphalian

10.2.1 Introduction

The palaeogeographic evolution of Central England during the Westphalian is summarized in a series of eleven maps and seven sections (Figs. 219-232). The facies types illustrated on these maps correspond to the 'facies assemblages' defined in 9.1 and used throughout Chapter 9. In addition marine facies (in marine bands) and areas of 'thick coal' deposition are illustrated. The maps are largely self explanatory, and only a brief discussion is included. The legend for these maps is given in Fig. 218.

10.2.2 Westphalian A (pre-Mealy Grey coal) (Figs. 219,226)

The widespread Subcrenatum Marine Band marking the base of Westphalian A, and the Listeri Marine Band are absent in the Coalbrookdale and Wyre Forest areas, and to the south of a line between Cannock and Nuneaton. To the north of this line facies present in the Westphalian A are not dissimilar to those in the underlying late Namurian, consisting of the deposits of deltaic, lagoonal/lacustrine and alluvial complexes, with fairly frequent marine incursions.

A comparatively small proportion of the succession consists of coals and seat-earth palaeosols, representing deposition in near emergent conditions.

To the south of the Cannock-Nuneaton line a larger proportion of seat earth palaeosols are present. In the Coalbrookdale, Wyre Forest, South Staffordshire and Warwickshire areas alluvial swamp conditions prevailed for most of this time, with interbedded red beds of the transitional association present over most of the Wyre Forest, to the south of Dudley and to the south of Leamington. Continuous red beds were probably deposited in the extreme south west of the Wyre Forest area, although close biostratigraphic control is lacking. Environments in the southern part of the Mid Staffordshire area probably fluctuated. While seat-earth dominated sequences are present immediately below the Mealy Grey seam, more varied lithologies below and above this may indicate the presence of lagoonal/lacustrine and/or deltaic deposits similar to those present further north.

In the basal Westphalian A, red beds are locally present in North Warwickshire.

After the formation of the Mealy Grey seam, the seat-earth dominated swamp conditions in Mid and South Staffordshire were reduced in area, being restricted to the Wyre Forest and the area south of Dudley. At this time several extensive coal seams were deposited in the northern part of the South Staffordshire area, and in Mid Staffordshire. Swamp conditions, however, persisted in the Coalbrookdale and Warwickshire area, and in the Wyre Forest, where the whole of the Westphalian A contains red beds.

10.2.3 Vanderbecki Marine Band (Figs. 219, 226)

The Vanderbecki marine incursion reached all parts of the southern margin of the Pennine Basin. Its exact extent is only accurately known in Warwickshire. The marine band has not been recorded in the extreme south of the Wyre Forest, nor south of Dudley, but there are no recent data in either of these areas. The limit marked in Fig. 219 is therefore speculative. The orientation of the Vanderbecki shoreline in Warwickshire is not known. The marine band is generally absent to the south of Coventry, and the maximum extent of the incursion has been arbitrarily placed more or less parallel to the trend of the previously existing facies belts to suggest that the coastline was determined by the topography of the Midland high.

Red bed deposition probably continued in the southern part of the Wyre Forest during this incursion (Alton. No. 1 borehole).

10.2.4 Early Westphalian B (Period of 'Thick Coal' formation) (Figs. 220, 227)

This period, representing approximately the first half of Westphalian B time, encompasses the acme of coal formation in the southern part of the Pennine Basin. A number of economically important coal seams were deposited in the coalfield areas near the depocentre, while, towards the margins of the basin, these seams amalgamate progressively, forming one thick seam in the extremely marginal areas. The best documented examples of such amalgamation are in the Thick Coals in South Staffordshire and Warwickshire (see 9.6 and 9.7). However, similar amalgamation is known in the probably slightly younger Yard, Ragman, Rough Seven Foot group of seams in North Staffordshire, and in seams of the same age in Yorkshire and Nottinghamshire (Edwards 1951).

On the map (Fig. 220) the line of amalgamation of the Thick Coals of Warwickshire and South Staffordshire, and the Double Coal of Coalbrookdale are shown. Of more palaeogeographic significance, where it can be mapped, is the line where these thick seams split and deteriorate towards the basin margin, passing laterally into seat earth dominated 'fireclay' facies. This line closely parallels the inferred shoreline of the Vanderbecki marine incursion. From this it may tentatively be suggested that, in the south, facies patterns were still being determined by onlap onto the topographic feature formed by the Midland High. Contemparaneous block movement is however implied by the marked seam splits and thickness variation within and between the South-, Mid-, and North Staffordshire areas.

In the extreme south west of the basin, 'transitional' and red bed facies assemblage rocks were still deposited in the south and west of the Wyre Forest during the formation of the Thick Coals.

10.2.5 Mid Westphalian B (Maltby marine Band and Brooch Coal) (Figs. 221, 228)

The Maltby Marine Band is present in all parts of the North Staffordshire, Coalbrookdale, and Mid Staffordshire areas. In South Staffordshire it is present to the north of Dudley, while in the Wyre Forest it may be present at Claverley borehole (Fig. 204), but is absent elsewhere. In these areas the inferred position of the Maltby incursion shoreline is thus similar to that for the Vanderbecki incursion (Fig. 219).

In the Warwickshire depositional area, the restricted occurrence of the ?Maltby marine Band around Coventry (see 9.6.5) suggests that the incursion entered this area from the west. The shoreline of the Maltby incursion probably turned markedly to the north east in the area to the west and north of Coventry. This is in contrast to the general trend of previous facies belts controlled by onlap onto the Midland High, and suggests that, by the time of the Maltby incursion, some local tectonic activity was controlling facies distribution in this area.

The lateral passage of the Brooch Coal into seat earth-dominated 'swamp' facies can be mapped around the southern margin of the Pennine basin. In Warwickshire the disappearance of this coal seam corresponds approximately to the position of the Maltby incursion shoreline, although, as with the Maltby Marine Band, there are considerable doubts as to the correlation of the coal seam (see 9.6.5). In South Staffordshire the maximum extent of the Brooch Coal is similar to that of the Thick Coal. However, in the Wyre Forest, the Brooch Coal (known locally as the Highley Brooch) is well developed, and correlates well with the Great Coal of Titterstone Clee. At the time of formation of the Brooch Coal, 'swamp' assemblage sediments and red beds were still being deposited in the southern part of the Wyre Forest. Poole (1966) has correlated the one thin coal encountered in the Eymore Farm boreholes with the Brooch Coal, and it is tempting to suggest (without any evidence) that the one thin coal recorded in Alton No. 1 borehole also correlates with this horizon. These thin coaly horizons are, however, exceptional in the seat earth and red bed sequence which predominates in this area.

The south westward extension of the Brooch Coal, and of the thin coals occurring at the same horizon, marks another departure from the hitherto prevailing subsidence pattern, and again suggests the beginning of control of facies distribution by local tectonics rather than by onlap onto a stable slope formed by the Midland High.

10.2.6 Late Westphalian B (Aegirianum Marine Band, and immediately preceding period) (Figs. 222,228)

Red beds and 'transitional' facies are found between the Brooch Coal and the Aegerianum Marine Band throughout the Wyre Forest. Red beds become dominant in this interval at Kinlet and Eymore Farm, and, presumably, to the south west.

In the Dudley area red beds are present shortly above the Brooch Coal, and the Aegerianum Marine Band is absent. To the south of this the absence of both the Brooch Coal and the Aegerianum Marine Band makes it impossible to correlate the succession. It may, however, be assumed that red beds and 'transitional' facies were forming.

Red intercalations and seat-earth dominated sequences are present between the Brooch Coal and Aegirianum Marine Band as far north as Hamstead No. 1 borehole.

In the Coalbrookdale area 'swamp' assemblage sediments are dominant in the interval between the Marquis (= Brooch) Coal and the Aegirianum Marine Band. Occasionally improved drainage is indicated by the presence of brown pigmented seat earths (e.g. in Brickkiln Plantation borehole).

In Warwickshire, seat-earth dominated 'swamp' facies occur below the Aegerianum Marine Band over the whole depositional area. In the extreme south, in Morton Morrell borehole, red beds occur below this band, but their exact time relationship to sequences further to the north is not known.

The Aegirianum Marine Band has a much wider extent that any other in the Warwickshire area. It is, however, noticeably absent in the area around the Berryfields Farm and Kimberleys Grove boreholes. Two explanations for this are possible: i) that red beds of Facies Association II or III were deposited in this area prior to the Aegirianum incursion, and that a topographically elevated alluvial lobe was thus formed, which was not covered by the marine incursion; or ii) that, after the marine incursion, local uplift and erosion occurred in this area, giving rise to a local unconformity. The former explanation is marginally favoured by the occurrence of a red bed intercalation below the Aegirianum Marine Band in Ram Hall borehole, and the resulting facies pattern is illustrated in Fig. 222. In either case, it is apparent that the sequence preserved marks some form of local tectonic activity, continuing a trend possibly initiated before the Maltby marine incursion. This could have involved initial local uplift of the West Warwickshire palaeohigh, shedding proximal alluvium into a small part of the basin before the Aegirinum incursion, or folding within the basin itself resulting in a local unconformity.

In South Staffordshire the Aegirianum Marine Band is present at Hamstead but absent at Dudley. This may similarly result from the presence of alluvial topography in the area around and to the south of Dudley, but stratigraphic control is insufficient to test this hypothesis.

If the correlation of the Aegirianum Marine Band in the Wyre Forest with that recorded from Titterstone Clee is correct, the extent of the Aegirianum Marine Band in this area is similar to that of the Brooch Coal.

Although detailed lithostratigraphic analysis has not been carried out in North Wales and north east Leicestershire, it is probable that red beds were forming in the marginal parts of these areas before the Aegirianum marine incursion. The presence of these facies has been included in Fig. 222 for completeness.

10.2.7 Early Westphalian C (Period immediately after Aegirianum marine incursion) (Figs. 223, 229)

After the Aegirianum marine incursion red bed facies spread rapidly across the southern part of the study area.

'Transitional' facies with intercalated red beds were developed over the whole of Warwickshire and South Derbyshire. To the south of Coventry, continuous red beds were deposited, while alluvial fans probably formed along the entire western margin of the Warwickshire area, sourced by the West Warwickshire high which was, by now, extensively uplifted.

In the South Staffordshire, Wyre Forest, and Coalbrookdale areas, red beds were deposited over an extensive area, with large alluvial fans being sourced from the West Warwickshire high and from the North Shropshire Horst. It is possible that, during this period, uplift of the area between the Wyre Forest and Coalbrookdale areas began. There is, however, no biostratigraphic or lithostratigraphic control as to the age of this uplift.

In South Staffordshire, the northward extent of red beds seems to have been controlled during this period by differential subsidence associated with the Bentley Faults (see 9.7.1). To the south of this structure, red beds begin immediately above the Aegirianum Marine Band, while immediately to the north of it, some 80m. of grey sediments with coal seams are present between this Marine Band and the base of the red beds. An unconformity might be suspected to be present at the base of the red beds, in view of such a marked change in the position of their base. However, the presence of an interbedded red/grey sequence on both sides of the fault, and the uniform stratigraphic horizon of the base to the south of the faults (see 9.7.2) suggests that there is not an unconformity present, and that the position of the base of the red beds was genuinely controlled by differential subsidence across the fault complex. The presence of red beds below the Bottom Robins Coal in Bowman's Bridge borehole, in the eastern part of Mid Staffordshire, suggests that progradation of well drained facies into this area was starting at this time.

During this period, the advance of red bed facies continued in both north east Leicestershire/South Nottinghamshire and North Wales.

10.2.8 Mid Westphalian C (Bottom Robins Coal and Cambriense Marine Band) (Figs. 224, 229)

The rapid advance of red bed facies, which began immediately after the Aegirianum marine incursion, continued during this period. In the Coalbrookdale, South Staffordshire, and Warwickshire areas the northern limit of continuous red bed deposition advanced. The most marked palaeogeographical change was, however, the probable emergence of a high in the area now occupied by the Trias filled Needwood Basin.

There is no direct evidence for the existence of this structure. Although Westphalian A to C rocks are absent in Chartley borehole, the lithological descriptions of this very old borehole do not enable identification of the presence or absence of Westphalian D Halesowen or Keele Formation sediments, which, if present, would enable confident identification of an intra-Westphalian high. Identification of uplift in this area during deposition of the Productive Coal Measures and Etruria Formation thus relies on facies changes in the adjoining Mid Staffordshire area.

At the horizon of the Bottom Robins coal in Mid Staffordshire lateral facies changes are apparent in a northerly direction in the Lichfield area, and towards the north east in the Stafford area. To the north of Lichfield, the Bottom Robins coal deteriorates and a seat earth dominated 'swamp' facies is present. To the east of Stafford, red beds are present at the Bottom Robins horizon in the Bricklawn and Parkhouse boreholes. As in other areas where a rapid change in the horizon of the base of the red beds is found, it is possible that an unconformity may be present between these and the Productive Coal Measures. Unfortunately neither of these boreholes cored the base of the red beds. However, the cuttings description suggests that an interbedded red and grey sequence is present in the lowest 20 to 30m of the Etruria Formation. This, together with the occurrence of an interbedded succession at a slightly higher horizon in the cored sequences at Devils Dumble, Berryhill and Brancote Gorse boreholes, suggests that the relationship is truly diachronous (Figs. 188, 189). In turn, this supports the suggestion that uplift to the north east of the Mid Staffordshire area led to the establishment of an embayment in the Mid Staffordshire area, which was progressively filled by alluvial sediments derived from the north east, as well as from the south.

After formation of the Bottom Robins Coal, this embayment remained linked to the main Pennine basin depocentre for long enough to allow the incursion of the Shafton Marine Band. After this, red bed facies became established over this area quite rapidly. The area around Cannock may have become temporarily isolated as a swamp basin during this period, as there is locally a thick (up to 90m) coal bearing sequence between the Shafton Marine Band and the base of the Etruria Formation. There is a complex and poorly understood pattern of diachronism in this area between red and grey beds.

The formation of grey sediments continued in North Staffordshire throughout this period, although the encroachment of red beds was beginning at the margins of the depositional area. The first red beds appear in the sequence at Sidway Mill borehole between the Edmondia and Shafton Marine Bands. Both here and in the nearby Radwood borehole the Cambriense Marine Band is absent, and its usual horizon, immediately above the Winghay Coal, is occupied by red beds.

By the time of the Cambriense incursion the Cannock embayment was occupied by red bed facies or 'swamp' assemblage sediments which were not flooded by the marine incursion. The marine band is not found to the south of Stafford. It is, however, present in south Derbyshire, implying that this area was still connected to the north with the depositional area of Nottinghamshire.

It is assumed that the advance of red bed facies continued in North Wales and North East Leicestershire/South Nottinghamshire at this time. What is uncertain is the behaviour of the area between South Derbyshire and North East Leicestershire, the area now occupied by the horst of

Precambrian rocks forming Charnwood Forest. As, by this time, uplifted blocks were forming in association with basement faults throughout the Midlands, it is reasonable to assume that the Charnwood Horst was similarly uplifted. There is, as yet, no evidence to substantiate this.

10.2.9 Late Westphalian C/Early Westphalian D (Figs. 225, 229-232)

Red bed sedimentation became established over the whole of the southern part of the Pennine Basin during this period, at least as far north as a line running from Liverpool through Manchester to Doncaster. To the north of this there is, as yet, no evidence.

In the areas where 'swamp' facies persisted after the Cambriense incursion, progradation of red bed facies was probably rapid. In areas of rapid subsidence, 'wedges' of red beds formed, best documented in the North Staffordshire area (see 6.1.2).

The palaeogeographical evolution of more marginal areas is not confidently known, owing to the lack of biostratigraphical control. It is, however, probable that the block movement in these areas, which had been occurring sporadically since the mid-Westphalian B, culminated in a phase of regional deformation during the late Westphalian C/Early Westphalian D, giving rise to the regional unconformity between the Halesowen Formation (and equivalents) and the Etruria Formation, Productive Coal Measures, and pre-Westphalian, which is present in all areas except North Staffordshire. The initial pulse of sediment produced by this event was probably responsible for the rapid progradation of red bed facies in the areas near the centre of the Pennine basin. Shortly after this event, tectonic activity ceased, and the whole of the marginal area was progressively peneplaned prior to the deposition of the Halesowen Formation. The products of this peneplanation accumulated as the thick Etruria Formation sequence in North Staffordshire. The extent of the local continuation of deposition of Etruria facies in the areas undergoing deformation is not known. The preservation of thick Etruria Formation sequences in pre-Halesowen synforms (e.g. at Manor Colliery, Halesowen) may imply continued local deposition. A side effect of the regional tectonic event was to allow the onlapping of the North Shropshire high by Etruria Formation sediments, which are apparently suceeded conformably by the Coed-yr-Allt (≡Halesowen) Formation. This again supports local deposition of Etruria Formation sediments during the final peneplaning of the southern margin of the Pennine basin.

The regional tectonic event was accompanied by a phase of alkali basalt intrusion in the Wyre Forest, South- and Mid Staffordshire. These intrusions may have had a surface expression as volcanic eruptions, whose lavas may have contributed to the detritus forming the Etruria Formation further to the north.

10.3 Tectonic evolution

It is not proposed to devote extensive discussion to the tectonic evolution of Central England during the Westphalian. In this section the subject is discussed only insofar as it is related to the origin of the Etruria Formation.

10.3.1 Description of tectonic history

Four phases can be recognised in the tectonic development of Central England during the Westphalian.

i) Progressive onlap of the Wales Brabant high - Westphalian A to mid Westphalian B (Figs 226, 227)

During this period, there was fairly rapid southward onlap of the Wales Brabant high during the early Westphalian A. This onlap was accompanied by a southward shift in facies belts. The swamp and red bed assemblages, which extended into Mid Staffordshire during early Westphalian A, had a very restricted development in South Staffordshire by the end of Westphalian A.

Coal formation was not widespread during the early Westphalian A (Lenisulcata zone) but became so during the later Westphalian A and early Westphalian B (Communis and Modiolaris zones), reaching its acme with the formation of the South Staffordshire/Warwickshire Thick Coals in late Modiolaris zone time.

Throughout this period normal movement on the basement faults separating depositional areas led to the development of seam splits, and pronounced thickness changes. All of the major seams in the Communis and Modiolaris zones in South Staffordshire (Bottom, New Mine, Heathen, and Thick Coals) split into the Mid Staffordshire area, across the Bentley Fault.

ii) Development of local horst structures - mid Westphalian B to late Westphalian C (Figs 228-230)

The development of horsts can be dated by the first appearance of significant red beds above the Thick Coal and correlative horizons. This occurred locally immediately after deposition of the Thick Coal, and on a regional scale, thoughout the Warwickshire, South Staffordshire, Wyre Forest, and Coalbrookdale areas, immediately after the Aegirianum marine incursion. Red beds were produced by the development of alluvial facies in areas adjoining the horsts, which were progressively stripped of pre-existing Upper Palaeozoic sediments, Lower Palaeozoic sediments and igneous rocks, and Precambrian igneous rocks.

Within the subsiding areas, the pattern of subsidence changed. Although subsidence was still greater in the north than the south, the lack of seam splitting after the Thick Coals suggests that differential subsidence across basement faults became less marked. These fault zones did still, however, show sufficient differential movement to control the distribution of red bed facies. The differential subsidence is reflected in the relative thicknesses of sediment deposited. The Westphalian sequence underlying the Yard/Ragman (?correlative of the Benches coal = top of the Thick Coal) in North Staffordshire is ca. 2.7 times thicker than the correlative unit in Mid Staffordshire, whereas the sequence between the Yard/Ragman and the Edmondia Marine Band in the former area is only 2 times as thick as in the latter area.

The progressive uplift of the horsts flanking the depositional basin is charted by the speed with which red beds spread over the basin. From the red bed stratigraphy, it is evident that two major pulses of uplift occurred. The first of these occurred immediately after the Aegirianum incursion, giving rise to a rapid spread of red beds in the Coalbrookdale, South Staffordshire and Warwickshire areas. The second of these occurred between the Shafton and Cambriense incursions, resulting in an almost instantaneous change to red bed deposition in Mid Staffordshire.

iii) Regional deformation of the southern part of the area - late Westphalian C to early Westphalian D (Fig. 231)

At some time after the Shafton Marine incursion a futher phase of deformation occurred in Mid and south Staffordshire, involving the Westphalian fill of this part of the basin. In a zone of ca. 5km width, stretching from Cannock to Stourbridge, the Productive Coal Measures were folded into a series of tight SW to NE trending anticlinal domes and denuded. These domes are best documented in the Brinsford area, where borehole and mining data have shown them to have a wavelength of 0.5 to 1km, and an amplitude of at least 160m. The folds are arranged in an en echelon pattern, both with respect to one another, and to the Bushbury Fault, which locally forms the boundary of the post Triassic Cannock Horst and the hinge line between the Coalbrookdale and Mid Staffordshire depositional areas. To the west of the Bushbury Fault, folding of similar intensity is present, the basal Halesowen Formation resting on Etruria Formation in the Gravelly Way borehole, and on a horizon just above the New Mine coal (ca. Vanderbecki Marine Band) in the Stretton borehole, 4km to the west.

Folding of similar intensity is present in the southern part of the Telford area (Fig. 233), where the folds have a wavelength of 1-2km, an amplitude of c. 180m, and trend SW to NE. These folds are parallel to the Limestone and Madeley Faults which are branches of the Western Boundary Fault of the coalfield. The Limestone fault appears to cause a lateral offset of the two limbs of a dome shaped anticline (Whitehead et al 1928 : Plate V) Similar folds have also been reported trending NNW to SSE in the area to the south of Tamworth (Kellaway 1970).

Away from these fairly limited areas of intense deformation, gentler folding is apparent beneath the base of the Halesowen Formation in the northern part of the Coalbrookdale area, throughout Mid Staffordshire, and throughout Warwickshire. In these areas a variable thickness of the Etruria Formation is usually preserved.

During this period rapid subsidence continued in North Staffordshire, and a thick sequence of red beds accumulated, sourced in part by the denudation of earlier Westphalian sediments in the southern flanks of the basin.

iv) Cessation of deformation; peneplanation (Fig. 232)

In the early Westphalian D deformation ceased throughout the area, and following peneplanation, a uniform subsidence regime was established throughout the area, including the former horsts. At this time the Wales-Brabant ridge ceased to form the southern margin of a distinct Pennine province, and arkosic fluvial coal bearing sediments were deposited over broad areas of Somerset (Mangotsfield Formation), Oxfordshire (Arenaceous Coal Formation), and the Midlands (Halesowen Formation and equivalents) (Kellaway 1970; Ramsbottom *et al.* 1978).

10.3.2 Discussion

The most recent analysis of the subsidence history of the Pennine province is that of Leeder (1982). He suggests that the subsidence pattern of this area during the Carboniferous follows that predicted by the model of McKenzie (1978). In this model, sedimentary basins evolve in two steps. Initially regional tension gives rise to crustal thinning and localised rifting. The crustal thinning causes hot asthenosphere to be uplifted. The cooling of this material gives rise to a regional downwarping, which follows the initial development of rift basins, and is referred to as the 'sagging' phase of basin development.

In the case of the Carboniferous of the Pennine province Leeder (op cit) suggests that the block and basin subsidence pattern developed in the Dinantian of northern and central England represents a rifting phase of basin development. This was followed by more widespread sagging, and the collapse of the blocks, during the Namurian and Westphalian. Leeder suggests that the rejuvenation of the Wales - Brabant ridge during the Namurian (shedding protoquartzite sediments into the North Staffordshire basin) represents the development of a bulge marginal to the Pennine basin in response to the transitionn from localized rift subsidence to regional flexural subsidence.

The decreasing importance of the Wales - Brabant ridge as a sediment source during the early Namurian, and its near eclipse in the late Namurian and early Westphalian was accompanied by progressive onlap of the ridge (Ramsbottom et al 1978; this thesis). It seems that the Wales - Brabant marginal bulge (if such it was) had largely decayed by this time.

The pattern of sedimentation in the area studied in this thesis conformed in its first phase (Westphalian A to mid B) to that described by Leeder. The progressive onlap of the Wales - Brabant ridge was largely completed by the end of Westphalian A, with very thin sequences accumulating as far south as Moreton-in-the-Marsh in Gloucestershire, well to the south of the present erosional limit of the bulk of the Westphalian A to C sequence in the West Midlands. The subsidence pattern at this time was complicated by the rejuvenation of Caledonian and pre-Caledonian faults as normal faults throwing downwards towards the depocentre, leading to extensive seam splitting and pronounced thickness changes.

The early Westphalian onlap on the Wales - Brabant ridge was accompanied by a progressive decrease in subsidence rate, due to the slowing and cessation of the sagging phase. This slowing is witnessed by the progressive increase in the number of and thickness of the coal seams present in the late Westphalian A throughout the Pennine province (Calver 1969). Coal formation reached an acme in the early Westphalian B, with the extensive formation of amalgamated thick coals. The generally emergent conditions of sedimentation which prevailed during this period prevented penetration of the depositional basin by all but the most widespread marine incursions. Thus only one marine band, the extensive Vanderbecki Band, is present in the coal rich interval of the late Westphalian A and early Westphalian B, in contrast to the abundant marine bands in the late Namurian and early Westphalian A.

After the deposition of the Thick coals, subsidence in the Pennine province was increasingly dominated by apparently independent movement between adjoining stable blocks. Initially this produced anomalous thickness variations, and allowed more frequent marine intercalations in the more rapidly subsiding areas. After the Aegirianum incursion, the rapid uplift of blocks in North Shropshire, Warwickshire, and elsewhere, led to local erosion, the deposition of red beds, and the progradation of alluvial swamp and red bed facies belts over the entire area. In the later stages of this block movement the Westphalian fill of the southern part of the depositional basin was affected by regional folding and denudation.

The effects of this were particularly apparent near the boundaries of the individual depositional area blocks, where locally tight folds were formed, sometimes en echelon to each other and to the major faults. Denudation of these structures locally removed the entire Carboniferous sequence. The folding was accompanied by the intrusion of alkali basalts in the Wyre Forest, South Staffordshire, and Mid Staffordshire areas. In early Westphalian D time tectonic activity ceased, and the area was peneplaned.

It is obvious that this phase of structural evolution, often described as the 'First Malvernian movements' (after Wills 1956) did not form part of the post-rift sagging phase in the Pennine province. Nor was it limited to that area. Uplift and local denudation occurred at the same time in the Bristol area (Kellaway 1970) and in the eastern part of the South Wales coalfield (Downing and Squirrel 1965). An unconformity is also reported at the base of the Westphalian D in the southern North Sea (Ramsbottom <u>et al.</u> 1978), in the Kish Bank basin (Jenner 1981), and possibly in the Kent coalfield (Bisson <u>et al</u> 1967).

Determination of the nature and cause of this deformation does not fall within the scope of this thesis. Only tentative suggestions can be made, which will need to be re-examined when a detailed structural study has been carried out.

A number of features suggest, but do not demonstrate, that the block faulting from mid Westphalian B onwards accompanied slight wrench movement on inherited Precambian and Lower Palaeozoic fault lines.

Such features include: the presence of folds with an 'en echelon' relationship to major fault lines; the rapid thickening and changes of facies across major faults; and the apparently unpredictable locations of uplifted blocks bounded by probably inherited fault structures.

The style of deformation resembles the Laramide block and lineament controlled style of folding and localized basin development in the Rocky Mountain foreland area (Thomas 1974). In this area the apparently random orientation of Laramide structures, when compared with contemporary structures in the main Rocky Mountain belt, is regarded as having been due to rejuvenation of several conjugate sets of pre-existing basement fractures in reponse to the contemporary phase of compressional deformation in the Rocky Mountain belt to the west.

A similar origin has been inferred for the Permian to Cretaceous subsidence pattern of the Sole Pit aea of the southern North Sea by Glennie and Boegner (1981). In this case movement was constrained by several intersecting sets of Variscan and earlier basement faults. The linear offset of geological features is negligible, and the effects of the wrench movement are limited to changes in the sense of throw along the lines of individual faults, rapid changes in sediment thickness across individual fault lines, the formation of antithetic faults, and the occurrence of faults which splay upwards to form so-called 'flower structures'. Faults of this kind, regarded as being characteristically formed during strike-slip movement, have been interpreted from seismic lines shot across the Coalbrookdale and Mid Staffordshire areas by Shell UK Limited (J. Haremboure, personal communication).

In the case of the Sole Pit area, crustal stretching due to the wrench movement was insufficient to allow the development of a basic volcanic complex (Glennie and Boegner op cit). In the Westphalian deformation in Central England, localised crustal stretching was sufficient to allow the development of intrusive, and possibly extrusive, alkali basalts in a zone stretching from the Wyre Forest to Cannock Chase.

On a regional scale, the 'First Malvernian' phase may be broadly regarded as representing the early part of the Leonian phase of deformation (Dvorak et al 1977). During this phase the first significant uplift of the Rheno Hercynian zone took place with the resulting northward shedding of lithic alluvium from mid Westphalian C onwards.

In south west England the start of this uplift can be established fairly accurately. It presumably occurred after deposition of the Bude Sandstone Formation, the highest preserved horizon of which is of earliest Westphalian C age (Ramsbottom et al 1978). The earliest appearance of the southerly derived 'Pennant' lithology is at a horizon just below the Cambriense Marine Band in the Somerset coalfield (Kellaway 1970). The rapid northward spread of this detritus into South Wales testifies to rapid uplift immediately to the south. Shackleton et al (1982) interpret this late Westphalian deformation in south west England to have involved the emplacement of a series of nappes, during which the deformation front moved progressively northwards. The resulting thrust loading gave rise to rapid subsidence in the South Wales and Somerset basins (Dewey 1982). This belt of rapid subsidence did not, however, extend continuously along the northern margin of the deformation front. In the Bristol Channel area the Lower Severn Axis, a pre Caledonian basement structure, was uplifted at this time (Kellaway 1970), presumably owing to wrench faulting in response to the approximately north south directed compression.

In the Midlands block movement of similar type had commenced locally within the Wales-Brabant Massif during the late Westphalian B. More intense deformation, with local uplift recorded by the rapid northward spread of the Etruria facies, began in the early Westphalian C, and reached its peak during late Westphalian C or early Westphalian D. It is tempting to relate this phase to the uplift in southern England, which began in the early Westphalian C and continued until the early Westphalian D, recorded by the occurrence of sand dominated Pennant sequences in South Wales and elsewhere.

Uplift in the Midlands was probably mainly in response to wrench along pre-existing structures, in a manner analogous to that previously described in the Rocky Mountain foreland Laramide structures (Thomas 1974). It is also possible that uplift of the Wales-Brabant Massif may have taken place as a marginal bulge, in response to the thrust loading which caused the rapid subsidence of the South Wales basin.

The regional structural evolution is illustrated schematically in Figs. 234 and 235.

10.4 Concluding remarks

The palaeogeographic synthesis made in this Chapter does not differ significantly, in its broad outlines, from that presented by Wills (1956). Much of Wills' reconstruction was, as he admitted, speculative. Subsequent exploration has justified many of his interpretations, particularly the importance of contemporary block movements on the patterns of subsidence and facies development during the deposition of the Etruria Formation. The greatly increased data base now available has allowed the stratigraphy of the Formation to be documented more accurately than could be achieved by Wills, and this in turn has led to the creation of more detailed palaeogeographical maps.

Three areas merit short concluding comments: climate, the relationship between subsidence rate and the occurrence of red beds, and the question of eustatic versus tectonic control of Westphalian facies.

Climate. The climate that prevailed throughout the deposition of the Etruria Formation was characterised by sufficient rainfall to maintain the water table at the surface in low lying floodplain areas. This allowed the formation of peats which have locally been preserved as coals or coaly horizons within the red bed sediments. Strong fluctuation in fluvial discharge is indicated by the dominance of lateral accretion sets in fluvial point bar deposits, and by the presence of graded 'single event' conglomerate sheets in the alluvial fan Facies Association. The presence of post depositionally oxidized coals and ironstones in the alluvial overbank deposits may be due to a marked fluctuation in water table, which in turn may also indicate a strongly fluctuating rainfall. In all of these instances sedimentological indications of rainfall fluctuation probably relate to the climate of the fluvial catchment areas.

It has often been claimed that the transition from coal measure to red bed sedimentation resulted from a change from a humid to a more arid climate (e.g. Wills 1956; Hains and Horton 1969). Hedemann and

Teichmüller (1971) argued that this change from grey to red beds forms part of a secular change, which occurred earliest in Scotland (early Westphalian C) and later in the Midlands (late Westphalian C) and south Wales (late Westphalian D), and reflects the southerly migration of an arid climatic zone.

The occurrence of red beds at all horizons in the Westphalian of Central England and the northward younging of the base of the main part of the Etruria Formation shows clearly that Hedemann and Teichmüller's interpretation is incorrect. Whether there was a local climatic change is less clear. It is possible that the growth (during the Westphalian) of the Variscan mountain chain to the south east led to a change from a humid tropical to a seasonal savanna climatic belt in the rain shadow (see Fig. 12). The climate inferred to have prevailed during the deposition of the Etruria Formation is similar to that found in modern savanna areas (Tricart 1972 pp. 7-14). If such a change occurred it might be invoked as a cause for the spread of red beds. However Broadhurst _et al_ (1980) have suggested that a seasonal climate already existed in the Westphalian A, before the peak of coal formation. The distribution of grey and red beds thus seems more likely to have been controlled by the water table than by the prevailing climate.

Role of subsidence. Another long held belief about the origin of the Etruria Formation is that the increase in red beds in the upper part of the Westphalian A to C sequence reflects the progressive 'silting up' of the depositional basin (e.g. Boardman 1978). This implies that the occurrence of red beds is closely linked either to subsidence rate or the rate of sediment input.

While it has been shown (Chapters 8, 9) that areas of low subsidence favoured the formation of red beds, the lack of extensive red beds associated with the thick coals, which formed during a period of very low subsidence rate, shows that subsidence alone was not a sufficient control to give rise to extensive red bed formation.

It was not until local sediment source areas were uplifted, from the late Westphalian B onwards, that widespread formation and progradation of red beds occurred. The Etruria Formation can thus be regarded as the very localised "molasse", derived from a phase of intra-Westphalian tectonism.

Eustacy and tectonics. The relative importance of eustatic and tectonic controls on sedimentary facies in the Dinantian and Namurian has been widely discussed (Ramsbottom 1973, 1977; George 1978). Ramsbottom (1979) extended his concept of pulsed, eustatically controlled mesothemic transgressions to include the Westphalian. He identified a repeated alternation of broad and restricted extensions in the marine incursions of the Pennine basin, grouping these into ten mesothems in which the extent of the later marine bands is consistently greater than that of the earlier ones. Similar mesothems were claimed to be present in South Wales, and in northern France, Belgium, and West Germany. On this basis the mesothems and individual marine bands were regarded as being of eustatic origin.

No direct evidence has been adduced during the present study which either confirms or disproves Ramsbottom's model. The presence of the Aegirianum Marine Band in very condensed dominantly alluvial sequences at the basin margins does, however, support the concept of eustatically controlled transgressions, at least in the cases of the more extensive Marine Bands.

Among aspects of Ramsbottom's model which have been questioned is the apparent lack of sedimentological record of the postulated regressive episodes. Regressive episodes are only locally recorded by channel incision and, in one case, a facies change from "distal deltaic" to fluvial facies (Read, in discussion of Ramsbottom 1979). Ramsbottom regarded this lack of "regressive" facies as indicating that the falls in sea level were less than the rises.

The presence of red beds in the Westphalian succession in marginal areas of the depositional basin should give a very delicate control on base level changes, as the initial appearance of red pigment was controlled by the position of the water table. It might be expected that, if major lowerings of sea level occurred, they would have caused an improvement of drainage in areas near the edges of the basin, giving rise to red pigmented soil development.

In three cases, the development of red beds may fulfil this criterion. These are: i) the very rapid and extensive onset of red beds in Mid Staffordshire after the Shafton incursion; ii) the very rapid and extensive onset of red beds in Coalbrookdale, South Staffordshire and Warwickshire after the Aegirianum incursion; and iii) the extensive development of 'swamp' and 'transitional' assemblage red beds below the Mealy Grey coal in Mid Staffordshire.

The first two examples have already been interpreted, in 10.3.1, as being due to pulses of tectonic activity. That this is to an extent the case is shown by the rapid spread in Warwickshire and

Coalbrookdale of locally derived alluvial fan sediments following the Aegirianum marine incursion. A similar spread of locally derived sediment followed the Shafton incursion in Mid Staffordshire. However, in the Wyre Forest, apparently not reached by alluvial fan sediments, there is also a rapid onset of red beds above the Aegirianum Marine Band. This suggests that such a facies change may have been partly caused by a drop in base level.

The red beds below the Mealy Grey coal are not obviously related to a Marine Band. They may have formed in response to base level change, but the progradation of well drained facies remains a more attractive explanation. Red intercalations of considerable lateral extent occur elsewhere (Blackband Formation, N. Staffordshire - see Chapter 6; below Aegirianum marine Band in S. Staffordshire; Westphalian A in Wyre Forest). In none of these cases can the red intercalation be linked confidently to a marine band.

Red beds do not, thus, appear to contribute much to the eustatic trangression/regression model. It is of interest, however, to note that two inferred tectonic pulses occur immediately after marine incursions. This raises the possibility that the marine incursions themselves may have been in part tectonically controlled. This question deserves more careful study.

REFERENCES

ABBOTT, P.L., MINCH, J.A. & PETERSON, G.L. 1976. Pre Eocene paleosol south of Tijuana, Baja California, Mexico. J. sediment. Petrol. 46, 355-361.

ALLEN, J.R.L. 1963. The classification of cross stratified units, with notes on their origin. Sedimentology, 2, 93-114.

ALLEN, J.R.L. 1964. Studies in fluviatile sedimentation: six cyclothems from the Lower Old Red Sandstone, Anglo-Welsh Basin. Sedimentology, 3, 163-198.

ALLEN, J.R.L. 1965. A review of the origin and characteristics of recent alluvial sediments. Sedimentology, 5, 89-191.

ALLEN, J.R.L. 1970. Studies in fluviatile sedimentation: a comparison of fining upwards cyclothems with special reference to coarse-member composition and interpretation. J. sediment. Petrol. 40, 298-323.

ALLEN, J.R.L. 1974a. Studies in fluviatile sedimentation: implications of pedogenic carbonate units, Lower Old Red Sandstone, Anglo-Welsh outcrop. Geol. J. 9, 181-208.

ALLEN, J.R.L. 1974b. Sedimentology of the Old Red Sandstone (Siluro-Devonian) in the Clee Hills area, Shropshire, England. Sediment. Geol. 12, 73-167.

ALLEN, J.R.L. & WILLIAMS, B.P.J. 1979. Interfluvial drainage on Siluro-Devonian alluvial plains in Wales and the Welsh Borders. J. geol. Soc. Lond. 136, 361-366.

ARBER, E.A.N., 1914. On the fossil floras of the Wyre Forest, with special reference to the geology of the coalfield and its relationships to the neighbouring coal measure areas. Phil. Trans. Roy. Soc. Lond. Series B, 204, 336-445.

ARBER, E.A.N. 1916. On the fossil floras of the Coal Measures of South Staffordshire. Phil. Trans. Roy. Soc. Lond., Series B, 208, 127-155.

ASHLEY, D.A. & PEARSON, M.J. 1978. Mineral distributions in sediments associated with the Alton marine band near Penistone, South Yorkshire. In M.M. Martland and V.C. Farmer, (Eds.) International Clay Conference. Developments in Sedimentology, 27, 311-321, Elsevier, Amsterdam.

BARNSLEY, G.B. 1965. The stratigraphy and structure of the Cannock Chase coalfield. Ph.D. Thesis, University of London. Unpublished.

BARROW, G., GIBSON, W., CANTRILL, T.C., DIXON, E.E.L. & CUNNINGTON, C.H. 1919. The Geology of the Country around Lichfield. Mem. geol. Surv. G.B., H.M.S.O. London.

BASU, A., YOUNG, S.W., SUTTNER, W., JAMES, C., & MACK, G.H. 1975. Re-evaluation of the use of undulatory extinction and poly crystallinity in detrital quartz for provenance interpretation. J. sediment. Petrol. 45, 873-882.

BEADLE, L.C. 1974. The inland waters of tropical Africa. Longman; London.

BENNISON, G.M. & WRIGHT, A.E. 1969. The geological history of the British Isles. Edward Arnold, London.

BERNER, R.A. 1971. Principles of chemical sedimentology. Magraw-Hill, New York.

BERSIER, A. 1958. Sequences detritiques et divagations fluviatiles. Eclog. geol. Helv. 51, 854-893.

BESLY, B.M. & TURNER, P. in press. Origin of red beds in a moist tropical climate (Etruria Formation, Upper Carboniferous, U.K.). In: R.C.L. Wilson (Ed.), Residual Deposits, Spec. Publ. geol. Soc. Lond.

BIRKELAND, P.W. 1974. Pedology, weathering and geomorphological research. Oxford University Press.

BISSON, G., LAMB, R.K. & CALVER, M.A. 1967. Boreholes in the concealed Kent coalfield between 1948 and 1959. Bull. geol. Surv. G.B., 26, 99-166.

BLAKE, D.H. & OLLIER, C.D. 1971. Alluvial plains of the Fly River, Papua. Z. Geomorphol. NF Suppl. 12, 1-17.

BLANCHE, J.B. & WHITAKER, J.H. MACD. 1978. Diagenesis of part of the Brent Sand Formation (Middle Jurassic) of the northern North Sea Basin. J. geol. Soc. Lond. 135, 73-82.

BLATT, H. 1967. Original characteristics of clastic quartz grains. J. Sediment. Petrol. 37, 401-424.

BLATT, H. & CHRISTIE, J.M. 1963. Undulatory extinction in quartz of igneous and metamorphic rocks and its significance in provenance studies of sedimentary rocks. J. Sediment. Petrol. 33, 559-579.

BLATT, H., MIDDLETON, G. & MURRAY, R. 1972. Origin of sedimentary rocks. Prentice-Hall, New Jersey.

BLESS, M.J., BOUCKAERT, J., CALVER, M.A., GRAULICH, J.M. & PAPROTH, E. 1977. Palaeogeography of Upper Westphalian deposits in NW Europe with reference to Westphalian C north of the mobile Variscan belt. Meded. Rijks. geol. Dienst. N.S. 28, 101-147.

BLOOMFIELD, C. 1964. Mobilisation and immobilisation phenomena in soils. In, A.E.M. Nairn (Ed.) Problems in palaeoclimatology. 661-665. Wiley, New York.

BOARDMAN, E.L. 1978. The Blackband Ironstones of the North Staffordshire coalfield. North Staffordshire J. Field Studies, 18, 1-13.

BOARDMAN, E.L. 1981. The genesis of Coal Measure Blackband Ironstones. Unpublished thesis, University of Aston.

BOULTON, W.S. 1917. The clays of South Staffordshire and its borders. Trans. Ceramic Soc. 16, 237.

BRAND, P.J., ARMSTRONG, M. & WILSON, R.B. 1980. The Carboniferous strata of the Westfield Opencast Site, Fife, Scotland. Rep. Inst. geol. Sci. London, 79/11.

BRAUNAGEL, L.H. & STANLEY, K.O. 1977. Origin of variegated redbeds in the Cathedral Bluffs Tongue of the Wasatch Formation (Eocene), Wyoming. J. Sediment. Petrol. 47, 1201-1219.

BREWER, R. 1964. Fabric and mineral analysis of soils. Wiley, New York. 470pp.

BRIDGES, R. & LEEDER, M. 1976. Sedimentary model for intertidal mudflat channels with examples from the Solway Firth, Scotland. Sedimentology, 23, 533-552.

BRINKMAN, R. & PONS, L.J. 1968. A pedo-geomorphological classification and map of the Holocene sediments in the coastal plain of the three Guianas. Soil Survey Institute, Wageningen. Soil survey paper 4.

BROADHURST, F.M., SIMPSON, I.M. & WILLIAMSON, I.A. 1968. Seam splitting. Ming. Mag. London, 119, 455-463.

BROADHURST, F.M., SIMPSON, I.M. & HARDY, P.G. 1980. Seasonal sedimentation in the Upper Carboniferous of England. J. Geol. Chicago, 83, 639-651.

BULLARD, E.C., EVERETT, J.E. & SMITH, A.G. 1965. The fit of continents around the Atlantic. Phil. Trans. Roy. Soc. Lond. A258, 41-51.

BUTTERWORTH, M.A. & SMITH, A.H. 1976. The age of the British Upper Coal Measures with reference to their miospore content. Rev. Palaeobot. Palynol. 22, 281-306.

BUURMAN, P. 1975. Possibilities of palaeopedology. Sedimentology, 22, 289-298.

BUURMAN, P. 1980. Palaeosols in the Reading Beds (Paleocene) of Alum Bay, Isle of Wight, U.K. Sedimentology, 27, 593-606.

BUURMAN, P. & JONGMANS, A.G. 1975. The Neerepen Soil, an early Oligocene Podzol with a frangipan and gypsum concretions, from Belgian and Dutch Limburg. Pedologie, 25, 105-117.

CALVER, M.A. 1956. Die stratigraphische Verbreitung der nichtmarinen Muscheln in den penninischen Kohlenfeldern Englands. Z. Deutsch. geol. Ges. 107, 26-39.

CALVER, M.A. 1968. Distribution of Westphalian marine faunas in northern England and adjoining areas. Proc. Yorks. geol. Soc. 37, 1-72.

CALVER, M.A. 1969. Westphalian of Britain. C.R. 6me Cong. int. Strat. Geol. Carb. (Sheffield 1967) 1, 233-254.

CAMERON, K.L. & BLATT, H. 1971. Durabilities of sand sized schist and 'volcanic' rock fragments during fluvial transport, Elk Creek, Black Hills, South Dakota. J. Sediment. Petrol. 41, 565-576.

CANTRILL, T.C. 1902. In: Summ. Prog. geol. Surv. G.B. 1901. p.63.

CARRIGY, M.A. & MELLON, G.B. 1964. Authigenic clay mineral cements in Cretaceous and Tertiary sandstones of Alberta. J. Sediment. Petrol. 34, 461-472.

COLEMAN, J.M. 1966. Ecological changes in a massive freshwater clay sequence. Trans. Gulf Coast Assoc. geol. Soc. 16, 159-174.

COLLINSON, J.D. 1978. In: H.G. Reading (Ed.), Sedimentary Environments and Facies, Blackwells, Oxford, 15-60.

COLTER, V.S. & BARR, K.W. 1975. Recent developments in the geology of the Irish Sea and Cheshire Basins. In: A.W. Woodland (Ed.). Petroleum and the continental shelf of North West Europe, 1, - Geology, 61-75. Applied Science Publ. Barking.

CUMMINS, W.A. 1962. The greywacke problem. Liverpool Manchester geol. J. 3, 51-72.

CURTIS, C.D. & SPEARS, D.A. The formation of sedimentary iron minerals. Econ. Geol. 63, 257.

CURTIS, C.D., PEARSON, M.J. & SOMOGYI, V.A. 1975. Mineralogy, chemistry and origin of a concretionary siderite sheet (clay-ironstone band) in the Westphalian of Yorkshire. Mineralog. Mag. London, 40, 385-393.

DAVIES, D. 1877. The relation of the Upper Carboniferous Strata of Shropshire and Denbighshire to beds usually described as Permian. Q.J. geol. Soc. London, 33, 10.

DICKINSON, W.R. 1970. Interpreting detrital modes of greywacke and arkose. J. sediment. Petrol. 40, 695-707.

DIX, E. & TRUEMAN, A.E. 1937. The value of non-marine lamellibranchs for the correlation of the Upper Carboniferous. Congress. Strat. Carb. Heerlen. 1935. 185-201.

DOBSON, E. 1850. A rudimentary treatise on the manufacture of bricks and tiles. Reprinted as - J. Ceramic History 5, George Street Press, Stafford. 1971.

DOWNING, R.A. & SQUIRREL, H.C. 1965. On the red and green beds in the Upper Coal Measures of the eastern part of the South Wales coalfield. Bull. geol. Surv. Gt. Britain, 23, 45-56

DUNNING, F.W. 1975. Precambrian craton of Central England and the Welsh Borders. In: A.L. Harris and others, A correlation of Precambrian rocks in the British Isles. Spec. Rep. geol. Soc. London, 6.

EAGAR, R.M.C. 1956. Additions to the non-marine fauna of the Lower Coal Measures of the north-midlands coalfields. L'pool Manchr. geol. J. 1, 328-369.

EARP, J.R. 1961. Exploratory boreholes in the North Staffordshire coalfield. Bull. geol. Surv. G.B. 17, 153-191.

EARP, J.R. & HAINS, B.A. 1971. British regional geology. The Welsh Borderland. H.M.S.O., London (3rd Edition).

EASTWOOD, T., GIBSON, W., CANTRILL, T.C. & WHITEHEAD, T.H. 1923. The geology of the country around Coventry. Mem. geol. Surv. G.B., H.M.S.O., London.

EDWARDS, W. 1951. The concealed coalfield of Yorkshire and Nottinghamshire. Mem. geol. Surv. G.B., H.M.S.O., London.

EDWARDS, W.N. 1967. The geology of the country around Ollerton. Mem. geol. Surv. G.B., H.M.S.O., London.

ELLIOT, R.E. 1968. Facies, sedimentation successions, and cyclothems in the Productive Coal Measures in the East Midlands, Great Britain. Mercian Geologist, 2, 351-372.

ELLIOT, T. 1974. Interdistributary bay sequences and their genesis. Sedimentology, 21, 611-622.

ETHERIDGE, F.G. & SCHUMM, S.A. 1978. Reconstructing paleochannel morphologic and flow characteristics: methodology, limitations and assessment. In: A.D. Miall (Ed.) Fluvial Sedimentology. Mem. Can. Soc. Petrol. Geol. 5, 703-721.

EVANS, W.B., WILSON, A.A., TAYLOR, B.J. & PRICE, D. 1968. Geology of the country around Macclesfield, Congleton, Crewe, and Middlewich. Mem. geol. Surv., G.B., H.M.S.O., London.

FALCON, N.F. & KENT, P.E. 1960. Geological results of petroleum exploration in Britain, 1945-1957. Mem. geol. Soc. Lond. 2, 56pp.

FISK, H.N., 1944. Geological investigation of the alluvial valley of the Lower Mississippi River. Mississippi River Comm., Vicksburg, Miss., 78pp.

FISK, H.N. 1947. Fine grained alluvial deposits and their effects on Mississippi River activity. Mississippi River Comm., Vicksburg, Miss., 82pp.

FITZPATRICK, E.A. 1971. Pedology. A systematic approach to soil science. Oliver and Boyd, Edinburgh.

FLEET, W.F. 1925. The chief heavy detrital minerals in the rocks of the English Midlands. Geol. Mag. 62, 98.

FREYTET, P. 1971. Paléosols résiduels et paléosols alluviaux hydromorphes associés aux dépôts fluviatiles dans le Crétacé Supérieur et l'Éocène Basal du Languedoc. Rev. Geogr. phys. Geol. dyn. Paris, 13, 245-268.

FRIEND, P.F. 1966. Clay fractions and colours of some Devonian red beds in the Catskill Mountains, U.S.A. Q.J. geol. Soc. London, 122, 273-292.

FRIEND, P.F. 1978. Distinctive features of some ancient river systems. In A.D. Miall (Ed.) Fluvial Sedimentology. Canadian Society of Petroleum Geologists Memoir 5, 531-542.

FRIEND, P.F., SLATER, M.J. & WILLIAMS, R.C. 1979. Vertical and lateral building of river sandstone bodies, Ebro Basin, Spain. J. geol. Soc. Lond. 136, 39-46.

FRIEND, P.F., MARZO, M., NIJMAN, W. & PUIGDEFABREGAS, C. 1981. Fluvial sedimentology in the Tertiary South Pyrenean and Ebro Basins, Spain. In: T. Elliott (Ed.), Field guides to modern and ancient fluvial systems in Britain and Spain, pp. 4.1-4.50. International Conference on Fluvial Sediments, University of Keele, U.K.

GALLOWAY, W.E. 1974. Deposition and diagenetic alteration of sandstone in North East Pacific arc-related basins: implications for greywacke. Bull. geol. Soc. Am. 85, 379-390.

GEOLOGICAL SURVEY OF GREAT BRITAIN. 1978. Telford. - Geological map at 1:25000. Institute of Geological Sciences.

GEORGE, T.N. 1978. Eustasy and tectonics: sedimentary rhythms and stratigraphical units in British Dinantian correlations. Proc. Yorks. geol. Soc. 42, 229-262.

GERASIMOV, T.P. 1971. Nature and originality of palaeosols. In: D.H. Yaalon (Ed.). Palaeopedology. Int.Soc. Soil Sci. and Israel Univ. Press, Jerusalem, pp.15-28.

GIBSON, W. 1901. The character of the Upper Coal Measures of North Staffordshire, Denbighshire, South Staffordshire and Nottinghamshire, and their relation to the Productive Series. Q.J. geol. Soc. Lond. 57, 251-265.

GIBSON, W. 1905. The Geology of the North Staffordshire coalfields. Mem. geol. Surv. G.B., H.M.S.O., London.

GILE, L.H. & HAWLEY, J.W. 1966. Periodic sedimentation and soil formation on an alluvial-fan piedmont in southern New Mexico. Proc. Soil Sci. Soc. 30, 261-268.

GILE, L.H., PETERSON, F.F. & GROSSMAN, R.B. 1966. Morphological and genetic sequences of carbonate accumulation in desert soils. Soil Sci. 101, 347-360.

GLENNIE, K.W. & BOEGNER, P.L.E. 1981. Sole Pit inversion tectonics. In: L.V. Illing & G.D. Hobson (Eds.). Petroleum Geology of the Continental Shelf of north-west Europe, 110-120. Institute of Petroleum, London.

GOLE, C.V. & CHITALE, S.V. 1966. Inland delta building activity of the Kosi River. Journal of the Hydraulics Division. Proc. Am. Soc. Civil Engineers. 92, 111-126.

GOUDIE, A. 1973. Duricrusts in tropical and subtropical landscapes. Clarendon Press, Oxford. 174pp.

GREIG, D.C. & MITCHELL, G.H. 1955. The western extension of the Leicestershire and south Derbyshire coalfield. Bull. geol. Surv. G.B., 7, 38-67.

GUION, P.D. 1971. A sedimentological study of the Crawshaw Sandstone (Westphalian A) in the East Midlands coalfield area. M.Sc. Thesis, University of Keele. Unpublished.

HAANTJENS, H.A. (Ed.). 1968. Lands of the Wewak - Lower Sepik area, Territory of Papua and New Guinea. CSIRO Australia Land Research Series, 22.

HAINS, B.A. & HORTON, A. 1969. British regional geology. Central England. H.M.S.O., London (3rd Edition).

HALLSWORTH, E.G. & COSTIN, A.B. 1953. Studies in pedogenesis in New South Wales. IV: the ironstone soils. J. Soil Sci., 4, 24-47.

HARLAND, W.B. and others. 1972. A concise guide to stratigraphical procedure. J. geol. Soc. Lond. 128, 295-305.

HAWKINS, P.J. 1978. Relationship between diagenesis, porosity reduction, and oil emplacement in late Carboniferous sandstone reservoirs, Bothamstall Oilfield, E. Midlands. J. geol. Soc. Lond. 135, 7-24.

HEDBERG, H.D. 1976. International stratigraphic guide. Wiley, London and New York.

HEDEMANN, H.A. & TEICHMÜLLER, R. 1971. The palaeogeographical development of the Upper Carboniferous. Fortschr. Geol. Rheinland Westfalen, 19, 132-145 (English section).

HEWARD, A.P. 1976. Sedimentation patterns in coal bearing strata: northern Spain and Great Britain. D. Phil. thesis, University of Oxford. Unpublished.

HEWARD, A.P. 1978a. Alluvial fan and lacustrine sediments from the Stephanian A and B (La Magdalena, Cinera-Matallana, and Sabero) coalfields, northern Spain. Sedimentology, 25, 451-488.

HEWARD, A.P. 1978b. Alluvial fan sequence and megasequence models: with examples from Westphalian D - Stephanian B coalfields, Northern Spain. In: A.D. Miall (Ed.). Fluvial Sedimentology. Mem. Can. Soc. Petrol. Geol. 5, 669-702.

HO, C. & COLEMAN, J.M. 1968. Consolidation and cementation of recent sediments in the Atchafalaya Basin. Bull. geol. Soc. Am. 80, 183-192.

HOARE, R.H. 1959. Red beds in the Coal Measures of the West Midlands. Trans. Instn. Ming. Eng. 119, 185-198.

HOLDRIDGE, D.A. 1959. Compositional variation in Etruria Marls. Trans. Brit. Ceram. Soc. 58, 301-328.

HOLMES, D.A. & WESTERN, S. 1969. Soil texture patterns in the alluvium of the Lower Indus plains. J. Soil Sci., 20, 23-37.

HOWELL, A. & COX, A.H. 1924. On a group of red measures or coloured strata in the east Glamorgan and Monmouthshire coalfield. Proc. S. Wales Instn. Eng. 43, 139-174.

HUDDLE, J.W. & PATTERSON, S.H. 1961. Origin of underclays and related rocks. Geol. Soc. Am. Bull. 72, 1643-1660.

HULL, E. 1869. The Triassic and Permian Rocks of the Midland Counties of England. Mem. geol. Surv. G.B., H.M.S.O., London.

HULL, E. 1881. The Coal-fields of Great Britain. 4th Edition. Stanford, London.

IRVING, E. 1977. Drift of the major continental blocks since the Devonian. Nature, 270, 304-309.

IXER, R.A., TURNER, P. & WAUGH, B. 1979. Authigenic iron and titanium oxides in Triassic red beds: (St. Bees Sandstone), Cumbria, Northern England. Geol. J. 14, 179-192.

JACKSON, R.G. 1978. Preliminary evaluation of lithofacies models for meandering alluvial streams. In: A.D. Miall (Ed.), Fluvial sedimentology, Mem. Can. Soc. Petrol. Geol. 5, 543-576.

JENNER, J.K. 1981. The structure and stratigraphy of the Kish Bank Basin. In: L.V. Illing and G.D. Hobson (Eds.). Petroleum Geology of the Continental Shelf of north west Europe, 426-431. Institute of Petroleum, London.

JOHNSON, A.M. 1970. Physical processes in geology. Freeman, Cooper & Co., San Francisco. 577pp.

JONES, D.G. & OWEN, T.R. 1961. The age and relationships of the Cornbrook Sandstone. Geol. Mag. 98, 285-294.

JUKES, J.B. 1859. The South Staffordshire Coalfield. 2nd edition. Mem. geol. Surv. G.B.

KEELING, P.S. 1961. The examination of clays by IL/MA. Trans. Brit. Ceram. Soc. 60, 217-

KEELING, P. & HOLDRIDGE, D. 1963. The mineralogical and chemical composition of the Etruria Marl from North Staffordshire. Proc. Int. Clay Conference, 2, 409-425.

KELLAWAY, G.A. 1970. The Upper Coal Measures of South West England, compared with those of South Wales and the Southern Midlands. C.R. 6 eme Cong. Int. Strat. Geol. Carb. (Sheffield 1967), 3, 1039-1055.

KIDSTON, R. 1894. The various divisions of British Carboniferous Rocks, as determined by their fossil flora. Proc. Roy. phys. Soc. Edinburgh, 12, 183-257.

KIDSTON, R. 1905. The divisions and correlation of the Upper Portion of the Coal Measures with special reference to their development in the Midland Counties of England. Q.J. geol. Soc. Lond. 61, 308- 323

KIDSTON, R., CANTRILL, T.C. & DIXON, E.E.L. 1917. The Forest of Wyre and the Titterstone Clee Hill coal fields. Trans. Roy. Soc. Edinb. 51, 999-1084.

KLIMEK, K. 1974. The structure and mode of sedimentation of the flood-plain deposits in the Wisloka Valley (South Poland). Studia Geomorphologica Carpatho-Balcanica, 8, 135-151.

KNIGHT, M.J. 1975. Recent crevassing of the Erap River, Papua New Guinea. Aust. Geogr. Studies 13, 77-84.

KRYNINE, P.D. 1936. Geomorphology and sedimentation in the humid tropics. Am. J. Sci., 32, 297-306.

KRYNINE, P.D. 1949. The origin of red beds. Trans. N.Y. Acad. Sci., 2, 60-68.

KRYNINE, P.D. 1950. Petrology, stratigraphy, and origin of the Triassic sedimentary rocks of Connecticut. Bull. Connecticut State Geol. Nat. Hist. Surv. 73, 247pp.

KULBICKI, G. & VETTER, P. 1955. Sur la presence d'argiles bauxitiques dans le Stephanien de Decazeville. C.R. Acad. Sci. Paris, 240, 104-106.

LARSEN, V. & STEEL, R.J. 1978. The sedimentary history of a debris-flow dominated Devonian alluvial fan - a study of textural inversion. Sedimentology, 25, 37-59.

LAVEINE, J.P. 1977. Report on the Westphalian D. In: V.M. Holub & R.H. Wagner (Eds.), Symposium on Carboniferous stratigraphy, 71-83. Geological Survey, Prague.

LEEDER, M.R. 1974. Lower Border Group (Tournaisian) fluvio-deltaic sedimentation and palaeogeography of the Northumberland Basin. Proc. Yorks. geol. Soc. 40, 129-180.

LEEDER, M.R. 1982. Upper Palaeozoic basins of the British Isles - Caledonide inheritance versus Hercynian plate margin processes. J. geol. Soc. London, 139, 479-491.

LEEDER, M.R., HASZELDINE, R.S. & HEWARD, A.P. 1981. Coastal plain fluvial facies of Carboniferous and Jurassic age in north-east England and southern Scotland. In: T. Elliot (Ed.), Field Guides to modern and ancient fluvial systems in Britain and Spain, 2.1-228, International Conference on Fluvial Sediments, University of Keele, U.K.

LEES, G.M. & TAITT, A.H. 1946. The geological results of the search for oil in Great Britain. Q.J. geol. Soc. Lond. 101, 255-317.

LÖFFLER, E. 1977. 1977. Geomorphology of Papua New Guinea. C.S.I.R.O. Australia and Australian National University Press. 196pp.

MALKIN, A.B. 1961. Contributions to the study of the Stratigraphy of the North Staffordshire coalfield. M.Sc. Thesis, University College of North Staffordshire. Unpublished.

McBRIDE, E.F. 1974. Significance of color in red, green, purple, olive, brown and grey beds of Difunta Group, north eastern Mexico. J. sediment. Petrol. 44, 760-773.

McKENZIE, D. 1978. Some remarks on the development of sedimentary basins. Earth planet. Sci. Lett. 40, 25-32.

MEYER, R. 1976. Continental sedimentation, soil genesis, and marine transgression in the basal beds of the Cretaceous in the east of the Paris Basin. Sedimentology, 23, 235-253.

MIALL, A.D. 1977. A review of the braided river depositional environment. Earth Sci. Rev. 13, 1-62.

MIALL, A.D. 1978. Lithofacies types and vertical profile models in braided river deposits: a summary. In: A.D. Miall (Ed.), Fluvial Sedimentology Mem. Can. Soc. Petrol. Geol. 5, 597-604.

MIDDLETON, G.V. & HAMPTON, M.A. 1976. Subaqueous sediment transport and deposition by sediment gravity flows. In: D.J. Stanley & D.J.P. Swift (Eds.), Marine Sediment Transport and Environmental Management, Wiley, New York, 197-218.

MILLOT, J. O'N., COPE, F.W. & BERRY, H. 1946. The seams encountered in a deep boring at Pie Rough, near Keele, N. Staffordshire. Trans. Instn. Min. Eng., 105, 528-

MITCHELL, G.H. 1942. The Geology of the Warwickshire coalfield. Geol. Surv. G.B. Wartime Pamphlet, 25.

MITCHELL, G.H. 1945. The Geology of the northern part of the South Staffordshire coalfield. Geol. Surv. G.B. Wartime Pamphlet, 43.

MITCHELL, G.H. 1954. The Whittington Heath borehole. Bull. geol. Surv. G.B., 5, 1-60.

MITCHELL, G.H., POCOCK, R.W. & TAYLOR, J.H. 1961. The Geology of the Country around Droitwich, Abberley, and Kidderminster. Mem.geol. Surv. G.B., H.M.S.O., London.

MOHR, E.J.C., VAN BAREN, F.A. & VAN SCHUYLENBORGH, J. 1972. Tropical soils. Mouton, The Hague.

MORISON, C.G.T., HOYLE, A.C. & HOPE-SIMPSON, J.F. 1948. Tropical soil-vegetation catenas and mosaics. A study in the south western part of the Anglo Egyptian Sudan. J.Ecol. 36, 1-84.

MOSSOP, G.D. 1978. Epsilon cross strata in the Athabasca oil sands (Abstract only). In: A.D. Miall (Ed.), Fluvial Sedimentology, Mem. Can. Soc. Petrol. Geol. 5, 854-855.

MUKERJI, A.B. 1976. Terminal fans of inland streams in the Sutlej-Yamuna Plain, India. Z. Geomorphol. NF, 20, 190-204.

MURCHISON, R.I. 1839. The Silurian System, founded on geological researches in the counties of Salop, Hereford, Radnor, Montgomery, Carmarthen, Brecon, Pembroke, Monmouth, Gloucester, Worcester and Stafford; with descriptions of the coal-fields and overlying formations. Murray, London, i-xxxii, 1-768 (2 Vols.).

MYKURA, W. 1960. Replacement of coal by limestone, and reddening of coal measures in the Ayrshire Coalfield. Bull. geol. Surv. G.B. 16, 69-109.

NAGTEGAAL, P.J.C. 1966. Scour and fill structures from a fluvial piedmont environment. Geol. Mijnbouw, 45, 342-354.

NAGTEGAAL, P.J.C. 1969. Sedimentology, and diagenesis of post-Hercynian continental deposits in the South Central Pyrenees. Leid. Geol. Meded. 42, 143-238.

NAGTEGAAL, P.J.C. 1978. Sandstone-framework instability as a function of burial diagenesis. J. geol. Soc. Lond. 135, 101-105.

NAMI, M. 1976. An exhumed Jurassic meander belt. Geol. Mag., 113, 47-52.

NIJMAN, W. & PUIGDEFABREGAS, C. 1978. Coarse grained point bar structure in a molasse-type fluvial system, Eocene Castisent Sandstone Formation, South Pyrenian basin. In: A.D. Miall (Ed.), Fluvial Sedimentology, Mem. Can. Soc. Petrol. Geol 487 - 510

NOLTIMIER, H.C. & ELLWOOD, B.B. 1977. The coal pole:palaeomagnetic-results from Westphalian B, C and D coals, Wales. Trans. Am. geophys. Union (Eos) 58, p.375.

ORTLAM, D. 1967. Fossile boden als leithorizonte für die Gliederung des höheren Buntsandsteins im nördlichen Schwarzwald und südlichen Odenwald. Geol. Jahrb. 84, 485-590.

ORTLAM, D. 1971. Palaeosols and their significance in stratigraphy and applied geology in the Permian and Triassic of Germany. In: D.H. Yaalon (Ed.), Palaeopedology. Int. Soc. Soil. Sci. and Israel Univ. Press, Jerusalem. 321-328.

PAIJMANS, K., BLAKE, D.H., BLEEKER, P. & McALPINE, J.R. 1971. Land resources of the Morehead-Kiunga Area, Territory of Papua and New Guinea. C.S.I.R.O., Australia, Land Research Series, 29.

PAIN, C.F. 1972. Some soils and surficial deposits in the Kokoda Valley, Papua and New Guinea. Pacific Science, 26, 335-345.

PEARSON, M.J. 1979. Geochemistry of the Hepworth Carboniferous sediment sequence, and origin of the diagenetic iron minerals and concretions. Geochim. cosmochim Acta, 43, 927-941.

PLOTT, R. 1686. The natural history of Staffordshire. Oxford.

POLLARD, J.E. & WISEMAN, J.F. 1971. Algal limestone in the Upper Coal Measures (Westphalian D) at Chesterton, North Staffordshire. Proc. York. geol. Soc., 38, 329-342.

POOLE, E.G. 1966. Trial boreholes on the site of a reservoir, at Eymore Farm, Bewdley, Worcestershire. Bull. geol. Surv. G.B. 24, 131-156.

POOLE, E.G. 1970. Trial boreholes in Coal Measures at Dudley, Worcestershire. Bull. geol. Surv. G.B. 33, 1-41.

POOLE, E.G. 1975. Correlation of the Upper Coal Measures of Central England and adjoining areas, and their relationship to the Stephanian of the continent. Bull. Soc. belge Geol. 84, 57-66.

POOLE, E.G. & WHITEMAN, A.J. 1954. Exploratory boreholes in the Prestwich area of the south east Lancashire coalfield. Trans. Instn. Min. Eng. 114, 291-317.

PUIGDEFABREGAS, C. 1973. Miocene point bar deposits in the Ebro Basin, N. Spain. Sedimentology, 20, 133-144.

PUIGDEFABREGAS, C. & VAN VLIET, A. 1978. Meandering stream deposits from the Tertiary of the Southern Pyrenees. In: A.D. Miall (Ed.), Fluvial Sedimentology, Mem. Can. Soc. Petrol. Geol. 5, 469-485.

PULLAN, R.A. 1967. A morphological classification of lateritic ironstones and ferruginised rocks in Northern Nigeria. Nigerian Journal of Science, 1, 161-173.

RADKE, B.M. & MATHIS, R.L. 1980. On the formation and occurrence of saddle dolomite. J. sediment. Petrol. 50, 1149 - 1168.

RAMSBOTTOM, W.H.C. 1973. Transgressions and regressions in the Dinantian: a new synthesis of British Dinantian stratigraphy. Proc. Yorks. geol. Soc., 39, 567-607.

RAMSBOTTOM, W.H.C. 1977. Major cycles of transgression and regression (mesothems) in the Namurian. Proc. Yorks. geol. Soc. 41, 261-291.

RAMSBOTTOM, W.H.C. 1979. Rates of transgression and regression in the Carboniferous of NW. Europe. J. geol. Soc. Lond. 136, 147-153.

RAMSBOTTOM, W.H.C., CALVER, M.A., EAGAR, R.M.C. & others. 1978. A correlation of Silesian rocks in the British Isles. Spec. Rep. geol. Soc. Lond. 10, 82pp.

RAWITSCHER, F. 1948. The water economy of the vegetation of the 'Campos Cerrados' in Southern Brazil. J. Ecol. 36, 237-268.

READ, W.A. 1969. Fluviatile deposits in Namurian rocks of Central Scotland. Geol. Mag. 106, 331-347.

READ, W.A. & DEAN, J.M. 1978. High alumina fireclays in the Passage Group of the Clackmannan Syncline, Scotland. Rep. Inst. geol. Sci., London, No. 78/12.

READING, H.G. 1978. Facies. In: H.G. Reading (Ed.), Sedimentary Environments and Facies, Blackwells, Oxford, 4-14.

REINECK, H.E. & SINGH, I.B. 1973. Depositional sedimentary environments. Springer-Verlag, Berlin, 439pp.

REINER, E. & ROBBINS, R.G. 1964. The Middle Sepik Plains, New Guinea. Geogrl. Rev. 54, 20-44.

RETALLACK, G.J. 1976. Triassic palaeosols in the Upper Narrabeen Group of New South Wales. I: Features of the palaeosols. J. geol. Soc. Aust. 23, 383-399.

RETALLACK, G.J. 1977. Triassic palaeosols in the Upper Narrabeen Group of New South Wales. II: Classification and reconstruction. J. geol. Soc. Aust. 24, 19-36.

REYNDERS, J. J. 1964. A Pedo-Ecological Study of Soil Genesis in the Tropics from Sea-level to eternal snow, Star Mountains, Central New Guinea. E.J. Brill, Leiden.

RICHARDSON, G.D. & FRANCIS, E.H. 1970. Fragmental clayrocks (FCR) in coal-bearing sequences in Scotland and north east England. Proc. Yorks. geol. Soc. 38, 229-259.

ROBERTSON, T. 1931. The origin of the Etruria Marl. Q.J. geol. Soc. Lond. 87, 13-29.

ROBINSON, P.L. 1973. Palaeoclimatology and continental drift. In: D.H. Tarling & S.K. Runcorn (Eds.), Implications of Continental drift to the earth sciences, Academic Press, New York, 451-476.

ROESCHMANN, G. 1971. Problems concerning investigations of palaeosols in older sedimentary rocks, demonstrated by the example of Wurzelboden of the Carboniferous System. In: D.H. Yaalon (Ed.) Paleopedology. Int. Soc. Soil Sci. and Israel Univ. press, Jerusalem, 311-320.

ROY, J.L. 1977. La position stratigraphique determinée paleomagnetiquement de sediments carbonifères de Minudie Point, Nouvelle Ecosse, a propos de l'horizon repère magnetique du Carbonifère. Can. J. Earth Sci. 14, 1116-1127.

RUST, B.R. 1978. Depositional models for braided alluvium. In: A.D. Miall (Ed.), Fluvial Sedimentology. Mem. Can. Soc. Petrol. Geol. 5, 605-625.

RZÓSKA, J. 1974. The Upper Nile swamps, a tropical wetland study. Freshwat. Biol. 4, 1-30.

SAPPER, K. 1935. Geomorphologie der Feuchten Tropen. Leipzig.

SCHILLER, H-J. 1980. X-ray textural analysis of clays and compacted shales. Internat. Assoc. Sedimentologists 1st Europ. Mtg., Bochum 1980, Abstr., 17-20.

SCHOLLE, P.H. 1979. A color illustrated guide to constituents, textures, cements, and porosities of sandstones and associated rocks. Mem. Am. Assoc. Petrol. Geol., 28.

SCOTESE, C.R., BAMBACH, R.K., BARTON, C., VAN DER VOO, R. & ZIEGLER, A.M. 1979. Paleozoic base maps. J. Geol. Chicago, 87, 217-277.

SCOTT, A.C. 1978. Sedimentological and ecological control of Westphalian B plant assemblages from West Yorkshire. Proc. Yorks. geol. Soc., 41, 461-508.

SCOTT, A.C. 1979. The ecology of Coal Measure floras from northern Britain. Proc. Geol. Ass. 90, 97-116.

SHARP, R.P. & NOBLES, L.H. 1953. Mudflow of 1941 at Wrightwood, Southern California. Bull. geol. Soc. Am. 64, 547-560.

SINGH, I.B. 1972. On the bedding in the natural levee and the point-bar deposits of the Gom ti River, Uttar Pradesh, India. Sediment. Geol. 7, 309-317.

SLAGER, S. & VAN SCHUYLENBORGH, J. 1970. Morphology and geochemistry of three clay soils of a tropical coastal plain (Surinam). Agricultural Research Report, 734, (Wageningen).

SLÁNSKÁ, J. 1976. A red bed formation in the South Bohemian basins, Czechoslovakia. Sediment. Geol. 15, 135-164.

SMITH, A.H.V. & BUTTERWORTH, M. 1967. Miospores in the coal seams of the Carboniferous of Great Britain. Special Pap. Palaeont, 1.

SMITH, E.G., RHYS, G.H. & GOOSSENS, R.F. 1973. Geology of the country around East Retford, Worksop, and Gainsborough. Mem. geol. Surv. G.B., H.M.S.O., London.

SMITH, N.D. 1972. Flume experiments on the durability of mudclasts. J. Sediment. Petrol. 42, 378-384.

SOIL SURVEY STAFF. 1960. Soil classification, a comprehensive system (7th Approximation). U.S. Dept. Agriculture, Soil Conservation Service. 265p.

SPEARS, D.A. & KANARIS-SOTIRIOU, R. 1979. Geochemical and mineralogical investigation of some British and other European tonsteins. Sedimentology, 26, 407-425.

SPEIGHT, J.G. 1965. Flow and channel characteristics of the Angabunga River, Papua. J. Hydrol. 3, 16-36.

SPENCER, E. 1925. Occurrences of spherulitic siderite and other carbonates in sediments. Q.J. geol. Soc. Lond. 81, 667-705.

STALDER, P.J. 1975. Cementation of Pliocene-Quaternary fluviatile clastic deposits in and along the Oman Mountains. Geol. Mijnbouw 54, 148 - 156

STEEL, R.J. 1974. Cornstone (fossil caliche) - its origin, stratigraphic and sedimentological importance in the New Red Sandstone, western Scotland. J. Geol. Chicago, 82, 351 - 369.

STEWART, D.J. 1981. A meander-belt sandstone of the Lower Cretaceous of Southern England. Sedimentology, 28, 1-20.

STONEHOUSE, T.H. 1951. Recent provings of concealed extensions of the coalfields of Shropshire and South Staffordshire. Trans. Instn. Min. Eng. 110, 289-

STUBBLEFIELD, C. & TROTTER, F. 1957. Divisions of the Coal Measures on Geological Survey Maps of England and Wales. Bull. geol. Surv. G.B., 13, 1-5.

TAYLOR, G. & WOODYER, K.D. 1978. Bank deposition in suspended load streams. In: A.D. Miall (Ed.), Fluvial Sedimentology, Mem. Can. Soc. Petrol. Geol. 5, 257-275.

TERUGGI, M.E. & ANDREIS, R.R. 1971. Micromorphological recognition of paleosolic features in sediments and sedimentary rocks. In: Yaalon (Ed.), Paleopedology. Int. Soc. Soil Sci. and Israel Univ. Press, Jerusalem. 161-172.

THOMAS, G.E. 1974. Lineament-block tectonics: Williston-Blood Creek Basin. Bull. Am. Assoc. Petrol. Geol. 58, 1305-1322.

THOMAS, T.M. 1961. The mineral wealth of Wales and its exploitation. Oliver and Boyd.

TRICART, J. 1972. The Landforms of the Humid Tropics, Forests and Savannas. Longman, London.

TRICART, J. 1977. Types de lits fluviaux en Amazonie brésilienne. Annales de Geographie, 84, 1-54.

TROTTER, F.M. 1953a. Exploratory borings in south west Lancashire. Trans. Instn. Min. Eng. 112, 261-283.

TROTTER, F.M. 1953b. Reddened beds of Carboniferous age in north west England and their origin. Proc. Yorks. geol. Soc. 29, 1-20.

TROTTER, F.M. 1954. Reddened beds in the Coal Measures of South Lancashire. Bull. geol. Surv. G.B. 5, 61-80.

TRUEMAN, A.E. 1933. A suggested correlation of the Coal Measures of England and Wales. Proc. S. Wales Instn. Eng. 49, p.63.

TRUEMAN, A.E. 1947. Stratigraphical problems in the coalfields of Great Britain. Q.J. geol. Soc. Lond. 103, lxv-civ.

TRUEMAN, A.E. 1954. The coalfields of Great Britain. Arnold, London.

TRUEMAN, A.E. & WEIR, J. 1946-1958. A monograph on British Carboniferous non-marine Lamellibranchia. Palaeontogr. Soc. Monogr.

TURNER, P. 1974. Origin of red beds in the Ringerike Group (Silurian) of Norway. Sediment. Geol. 12, 215-235.

TURNER, P. & TARLING, D.H. 1975. Implications of new palaeomagnetic results from the Carboniferous system of Britain. J. geol. Soc. Lond. 131, 469-488.

TURNER, P. & WHITAKER, J.H. MACD. 1976. Petrology and provenance of late Silurian fluviatile sandstones from the Ringerike Group of Norway. Sediment. Geol. 16, 45-68.

VAN HOUTEN, F.B. 1968. Iron oxides in red beds. Bull. geol. Soc. Am. 79, 399-416.

VAN HOUTEN, F.B. 1972. Iron and clay in tropical savanna alluvium, Northern Columbia: a contribution to the origin of red beds. Bull. geol. Soc. Am. 83, 2761-2772.

VAN STRAATEN, L.M. 1951. Texture and genesis of Dutch Wadden Sea sediments. Proc. 3rd Internat. Congress Sedimentology, Netherlands, 225-255.

WALKER, R.G. & CANT, D.J. 1979. Sandy fluvial systems. In: R.G. Walker (Ed.), Facies Models, Geoscience Canada, Reprint Series 1, 23-31, Geological Association of Canada.

WALKER, T.R. 1967a. Formation of red beds in modern and ancient deserts. Bull. geol. Soc. Am. 78, 353-368.

WALKER, T.R. 1967b. Color of recent sediments in tropical Mexico: a contribution to the origin of red beds. Bull. geol. Soc. Am. 78, 917-919.

WALKER, T.R. 1974. Formation of red beds in moist tropical environments: a hypothesis. Bull. geol. Soc. Am. 85, 633-638.

WALKER, T.R. 1976. Diagenetic origin of continental red beds. In: H. Falke (Ed.). The continental Permian in Central, West and South Europe. D. Riedel, Dodrecht. 240-282.

WALKER, T.R., WAUGH, B. & CRONE, A.J. 1978. Diagenesis in first-cycle desert alluvium of Cenozoic age, southwestern United States and northwestern Mexico. Bull. geol. Soc. Am. 89, 19-32.

WATTS, N.L. 1976. Paleopedogenic palygorskite from the basal Permo-Triassic of northwest Scotland. Am. Mineral. 61, 299-302.

WEDD, C.B., SMITH, B. & WILLS, L.J. 1928. The geology of the country around Wrexham. Part II. Mem. geol. Surv. G.B., H.M.S.O., London.

WEDD, C.B., KING, W.B.R. & WRAY, D.A. 1929. The geology of the country around Oswestry. Mem. geol. Surv. G.B., H.M.S.O., London.

WHITEHEAD, T.H. & EASTWOOD, T. 1927. The geology of the southern part of the South Staffordshire coalfield. Mem. geol. Surv. G.B., H.M.S.O., London.

WHITEHEAD, T.H. & POCOCK, R.W. 1947. Dudley and Bridgenorth. Mem. geol. Surv. G.B., H.M.S.O., London.

WHITEHEAD, T.H., ROBERTSON, T., POCOCK, R.W. & DIXON, E.E.L. 1928. Geology of the Country between Wolverhampton and Oakengates. Mem. geol. Surv. G.B., H.M.S.O., London.

WHITTEN, D.G.A. & BROOKS, J.R.V. 1972. A dictionary of Geology. Penguin.

WILLIAMS, B.J. & WHITTAKER, A. 1974. The geology of the country around Stratford-on-Avon and Evesham. Mem. geol. Surv. G.B., H.M.S.O., London.

WILLIAMS, G.E. 1971. Flood deposits of the sand-bed ephemeral streams of Central Australia. Sedimentology, 17, 1-40.

WILLIAMSON, W.O. 1946. Some grits and associated rocks in the Etruria Marl of N. Staffordshire. Geol. Mag. 83, 20-32.

WILLS, L.J. 1935. An outline of the palaeogeography of the Birmingham Country. Proc. Geol. Ass. 46, 211-246.

WILLS, L.J. 1950. The palaeogeography of the Midlands. 2nd Edition. University Press of Liverpool

WILLS, L.J. 1956. Concealed coalfields. Blackie, London, 208pp.

WILSON, M.D. & PITTMAN, E.D. 1977. Authigenic clays in sandstones: recognition and influence on reservoir properties and palaeoenvironmental analysis. J. Sediment. Petrol. 47, 3-31.

WILSON, M.J. 1965. The origin and geological significance of the South Wales underclays. J. Sediment. Petrol. 35, 91-99.

WORKING GROUP ON THE ORIGIN AND NATURE OF PALEOSOLS. 1971. Criteria for the recognition and classification of paleosols. In: D.H. Yaalon (Ed.). Paleopedology, Int. Soc. Soil Sci. & Israel Univ. Press, Jerusalem. 153-158.

WORSSAM, B.C. 1977. The Hanginghill Farm (1970) Borehole, South Derbyshire coalfield. Bull. geol. Surv. G.B. 63, 17pp.

WORSSAM, B.C., CALVER, M.A. & JAGO, G. 1971. Newly discovered marine bands and tonsteins in the South Derbyshire coalfield. Nature, Lond. 32, Phys. Sci. 121-122.

YAALON, D.H. 1971. Soil forming processes in space and time. In: D.H. Yaalon (Ed.). Paleopedology. Int. Soc. Soil Sci. and Israel Univ. Press, 29-40.

YOUNG, A. 1976. Tropical soils and soil survey. Cambridge University Press.

YOUNG, S.W. 1976. Petrographic textures of detrital polycrystalline quartz as an aid to interpreting crystalline source rocks. J. Sediment. Petrol. 46, 595-603.

ZIEGLER, A.M., SCOTESE, C.R., McKERROW, W.S., JOHNSON, M.E. & BAMBACH, R.K. 1979. Paleozoic paleogeography. Ann. Rev. Earth Planet. Sci. 7, 473-502.

ADDENDUM

ARCHER, A. 1965. Red beds in the Upper Coal Measures in the Western part of the South Wales Coalfield. Bull. Geol. Surv. G.B. 23, 57 - 66.

GLENNIE, K.W. 1981. Early Permian - Rotliegendes. In: Introduction to the Petroleum Geology of the North Sea. JAPEC Course Notes 1, Geol. Soc. London

ADDENDUM (continued)

DEWEY, J.F. 1982. Plate tectonics and the evolution of the British Isles. J. geol. Soc. London, 139, 371-412.

DVORAK, J., MIROUSE, R., PAPROTH, E., PELHARTE, A., RAMSBOTTOM, W.H.C., & WAGNER, R.H. 1977. Relations entre la sedimentation eodevono-carbonifère et la tectonique varisque en Europe centrale et occidentale. In: La chaine varisque d'Europe moyenne et occidentale. Coll. intern. CNRS, Rennes, 243, 241-273.

SHACKLETON, R.M., RIES, A.C., & COWARD, M.P. 1982. An interpretation of the Variscan structures in SW England. J. geol. Soc. London, 139, 533-541.