THE TRENT, GLEN PARVA AND BLUE ANCHOR

FORMATIONS (UPPER TRIASSIC) OF THE EAST MIDLANDS

AND THEIR SULPHATE DEPOSITS.

by

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S. R. TAYLOR

ABSTRACT

Norian and Rhaetian (Upper Triassic) sediments from the Mercia Mudstone Group of the East Midlands are described. Four facies are recognized; (1) sheet flood, (2) red-bed saline mudflat, (3) lacustrine and (4) marginal. Sedimentation occurred on an extensive peneplain which had areas of minor relief. Interstitial sulphate precipitation from hypersaline brines formed widespread evaporitic gypsum and anhydrite horizons. Dolomitization of carbonate material and possibly extensive interstitial halite growth also occurred. Displacive sulphate masses in Fauld Mine are recognized to be diapiric structures associated with an identifiable source bed. Fault movement triggered flowage of water-saturated, source bed gypsum.

A new approach to correlating fine-grained, unfossiliferous red-beds utilizes variations in clay mineralogy and $\delta^{18}O$ dolomite. The Trent Formation is subdivided by this method into the Fauld Member and overlying Hawton Member. An isotopic study of the Trent Formation sulphates (S, Sr) and dolomites (C, O) indicates that during deposition of the Fauld Member, grabens and adjacent areas had a marine brine regime with minor continental input (up to 20% sulphate derived from continental sulphides). The horst area was strongly influenced by continental brines which derived sulphate from the exposed Carboniferous Hathern Anhydrite Series. The Hawton Member brine regime was predominantly continental with periodic marine influxes.

Two distinct clay suites are recognized. The first is characterized by a predominance of detrital illite and chlorite (= Hawton Member); the second is characterized by the presence of either Mg-rich smectite/chlorite and/or palygorskite and/or sepiolite with minor illite and chlorite (= Fauld Member). Sepiolite and palygorskite were neoformed, whilst smectite/chlorite was transformed from degraded detrital illite. Both processes occurred in an alkaline brine regime rich in Mg^{2+} and amorphous silica. The distribution of Mg-rich clays reflects availability of Mg^{2+}, Al^{3+} and Si within the brines.

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AN INTRODUCTION TO THE THESIS.

GEOGRAPHICAL SETTING.

The Triassic Mercia Mudstone Group of Britain forms a broad Y-shaped area with the easterly limb stretching from Teeside on the east coast to Sidmouth, Devon, on the south coast. The westerly limb of the Y runs from Cheshire, through northern Staffordshire into the East Midlands of Britain. The study area covers the central part of the Y, stretching eastwards from Burton-upon-Trent along the Trent Valley to Newark (Fig.1.1). Grantham lies on the eastern boundary, whilst the southern boundary lies south of Leicester. In the study area the Mercia Mudstones underlie much of the Trent, Dove and Soar river valleys (Fig. 1.1), forming flat low lying ground with very few natural exposures. Pre-Triassic and Jurassic strata tend to form areas of upstanding relief.

PREVIOUS WORK.

Previous work on the East Midlands Triassic commenced in 1869 when Hull subdivided the Triassic strata around Nottingham on a lithological basis into the "Bunter" and "Keuper". Bosworth (1913) gave a comprehensive account of the "Keuper Marl" as exposed in brickpits and quarries around Leicester and Charnwood Forest. Brief descriptions of the "Keuper Marl" lithologies are given by Fox-Strangeways (1900, 1903, 1905) and Stevenson and Mitchell (1955). Elliott (1961) subdivided the East Midlands Triassic into eight lithostratigraphic formations, six of which were subdivisions of the "Keuper Marl". Subsequent to Elliott (1961) further work on the stratigraphy of the area was carried out by Audley-Charles (1970a), Warrington (1970) and Wills (1970). Audley-Charles (1970a) correlated major lithostratigraphic units across Britain and constructed a series of maps depicting the palaeogeography throughout the Triassic (Audley-Charles, 1970b). Less detailed maps are included in Wills (1970) and Warrington (1970). Warrington et al. (1980) broadly replaced the Bunter by the Sherwood Sandstone Group and the Keuper Marl

Fig. 1.1.　　Location map and major stratigraphical units of the East

Midlands of Central England.　Geological linework based

on Institute of Geological Sciences geological maps.

PENARTH GROUP & JURASSIC

MERCIA MUDSTONE GROUP

SHERWOOD SANDSTONE GROUP

PALAEOZOIC

CHARNIAN + CALEDONIAN IGNEOUS

DANE HILLS SANDSTONE MEMBER

0   10 km.

R. TRENT

NEWARK

Hawton &
Jericho Quarries

Staunton
Quarry

Kilvington
Quarry

NOTTINGHAM

Cropwell Bishop
Quarry

GRANTHAM

Keyworth

Bunny Brickpit

R. DOVE

NEEDWOOD
BASIN

Fauld Mine

Gotham
Mine

East Leake Mine

BURTON

R. SOAR

MELTON
MOWBRAY

R. TRENT

Barrow upon Soar

N

LEICESTER

Gypsy Lane

Enderby

Croft

Stoney Cove
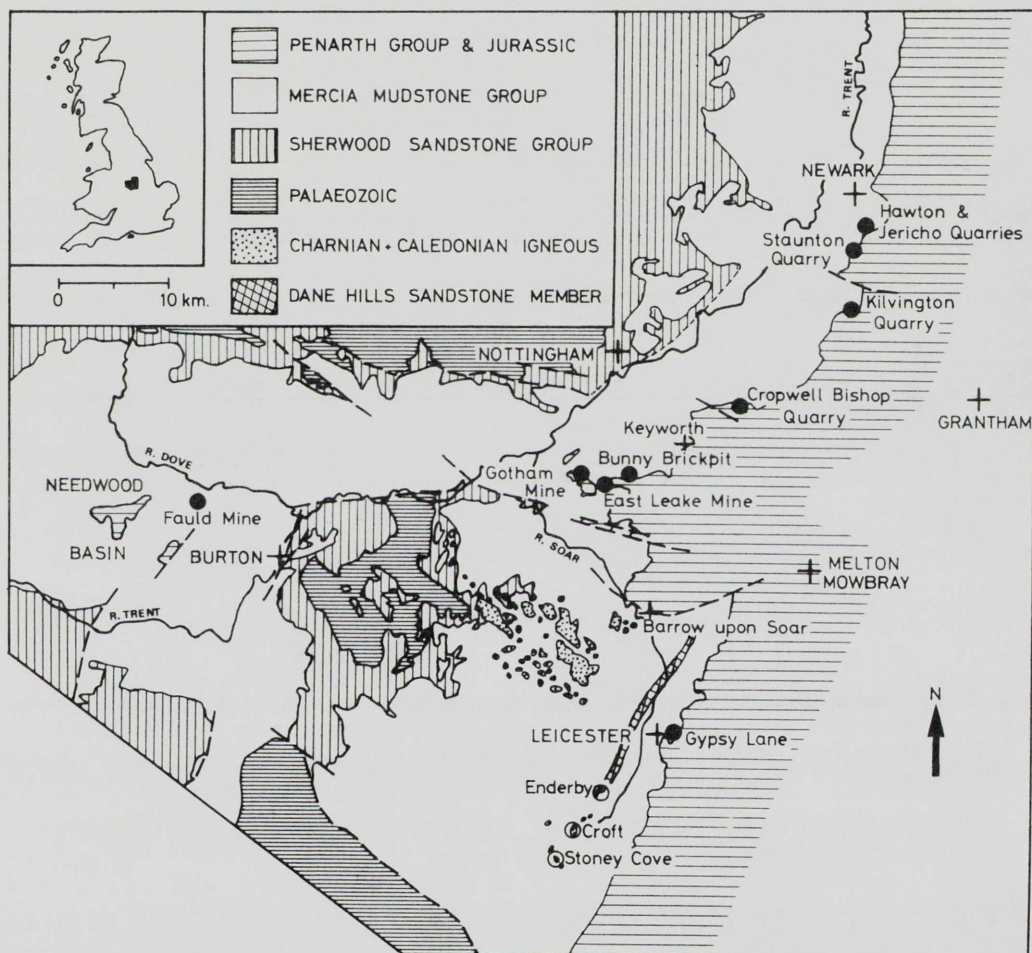
by the Mercia Mudstone Group, retaining seven of Elliott's (1961) formations and dividing the uppermost Parva Formation into the Glen Parva and Blue Anchor Formations (Fig. 1.2). The Penarth Group (Warrington et al., 1980) is equivalent to the Rhaetic and basal Liassic strata.

The Mercia Mudstone Group contains widespread deposits of evaporitic gypsum, anhydrite and halite. In the study area, British Gypsum Ltd. extensively mine and quarry gypsum from the Trent Formation (Fig. 1.1). Earlier work on the sulphate deposits of the area includes Trafford Wynn (1906), Smith (1918), Sherlock (1918), Sherlock and Hollingworth (1938), Richardson (1920, 1921), Firman (1964) and Taylor (1968). Aljubouri (1972) studied the geochemistry of both sediments and evaporites in the East Midlands Triassic, whilst Honeybourne (1951), Stephen and MacEwen (1951), Dumbleton and West (1966) and Jeans (1978) examined the clay mineralogy of the mudstones.

BASIS AND AIMS OF STUDY.

The present study is a C.A.S.E. project with British Gypsum Ltd. In recent years British Gypsum have drilled many boreholes in the East Midlands area, particularly around Newark, East Leake, Barrow-upon-Soar and Fauld (Fig. 1.1). Many of the boreholes passed through Liassic strata, the Penarth Group, and Blue Anchor and Glen Parva Formations and terminated within the Trent Formation (Fig. 1.2). The boreholes were, in general, drilled "open hole" with cores only taken from short, selected intervals within the Trent Formation. The National Coal Board kindly made available geophysical logs of many boreholes sunk between Melton Mowbray, Cropwell Bishop and Grantham (Fig. 1.1). These boreholes passed completely through Triassic strata. This study therefore concentrates upon the Trent Formation and overlying Triassic strata, as it is for this interval that most previously unutilized information is available.

Although a substantial amount of work had been carried out on the East Midlands Triassic prior to this study, the following points were

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Fig. 1.2.   Subdivisions and correlations of the East Midlands

Triassic to the standard Triassic sequence (after

Elliott, 1961; Warrington et al., 1980).

AREA	NEEDWOOD BASIN (Fauld)	SOUTH NOTTINGHAMSHIRE
LIAS	Lias	Lias

STANDARD SEQUENCE OF TRIASSIC STAGES		NEEDWOOD BASIN (Fauld)		SOUTH NOTTINGHAMSHIRE
	RHAETIAN	Penarth Group		Penarth Group
		Blue Anchor Fm.		Blue Anchor Fm.
		Glen Parva Fm.		Glen Parva Fm.
	NORIAN	Trent Fm.		Trent Fm.
				HOLLYGATE SSTS.
	CARNIAN	Edwalton Fm.		Edwalton Fm.
		HALITE		
		Harlequin Fm.		Harlequin Fm.
	LADINIAN	Carlton Fm.		Carlton Fm.
		Radcliffe Fm.		Radcliffe Fm.
				Colwick Fm.
	ANISIAN	Sandstones		Woodthorpe Fm.
				non sequence
	SCYTHIAN	Conglomerates		Sandstones and Conglomerates

(Column labels within South Nottinghamshire: MERCIA MUDSTONE GROUP)

Legend:
SHERWOOD SANDSTONE GROUP

unresolved:-

(1) correlation within the Trent Formation, especially with regard to individual sulphate horizons;

(2) detailed lithological/sedimentological studies of the British Gypsum core material and East Leake, Gotham and Fauld Mines;

(3) the origin of large displacive sulphate masses exposed in Fauld Mine;

(4) the origin of the sulphate-rich brines, i.e. were they marine, continental or mixed marine-continental;

(5) the clay mineralogy of the mudstones associated with major sulphate horizons in the Trent Formation;

(6) the relationship of sulphate distribution to palaeogeography.

The study is therefore aimed at solving these particular problems.

OUTLINE OF THESIS CONTENTS.

Chapter 2 describes a new approach to correlation of fine-grained unfossiliferous red-beds. Based upon dolomite oxygen isotope geochemistry and clay mineralogy, it enables one to subdivide the Trent Formation into two members correlatable over a distance of 70 km from a horst to graben sequence. The type sections and typical lithologies of the two new members are briefly described. Dolomite oxygen isotope geochemistry and clay mineralogy results are drawn from, and discussed more fully in, Chapters 5 and 7 respectively. Their inclusion in Chapter 2 is justified by the increase in ease of stratigraphical correlation. Extension of this correlation method to other areas of the British Triassic is also discussed.

Chapter 3 is divided into two parts, the first of which describes the detailed stratigraphy, lithologies and sedimentology of both clastic and evaporitic rocks for the upper part of the Edwalton Formation and the Trent, Glen Parva and Blue Anchor Formations. Extensive use is made of

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British Gypsum borehole data and core material, N.C.B. borehole data and sections exposed in mines and quarries. In the second part the various clastic and evaporitic lithologies are interpreted in terms of depositional environment. The chapter concludes with a discussion of changes in depositional environment from the top of the Edwalton Formation to the Penarth Group and the possible relationships between these changes and climatic variation.

The geology of Fauld Mine, only briefly mentioned in Chapter 3, is more fully described in Chapter 4. Of particular interest are large displacive sulphate masses which pass from the floor into the roof of the mine. A description of the morphology and distribution of the sulphate masses and associated features is followed by a discussion of their possible mode of origin.

Chapter 5 documents the results of a stable isotope study of East Midlands sulphate (^{34}S) and dolomites (^{13}C, ^{18}O). Samples from other areas of the British Triassic, as well as Permian and Carboniferous evaporite sequences, were analysed for comparative purposes. Interpretation of the analyses in terms of continental vs. marine brine sources is followed by a discussion of the palaeohydrology of the Trent Formation.

Strontium isotope measurements of calcium sulphates previously analysed for ^{34}S (Chapter 5), are described in Chapter 6. The results are related to the ^{34}S analyses and are interpreted in terms of the palaeo-hydrology with particular emphasis on the relative importance of marine and continental brines.

An X.R.D. study of the clay mineralogy of the upper part of the Edwalton Formation and Trent Formation is described in Chapter 7. The relative proportions of various clay minerals recorded in each sample are shown graphically. A discussion then relates their presence, relative amount and in many cases origin to various aspects, including original detrital clay mineralogy, palaeogeography, brine chemistry and diagenetic processes.

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CHAPTER 2.

CLAY MINERALOGY AND DOLOMITE OXYGEN ISOTOPE GEOCHEMISTRY;

A NEW APPROACH TO LITHOSTRATIGRAPHIC CORRELATION OF

FINE-GRAINED UNFOSSILIFEROUS RED BEDS.

This chapter formed the basis of a paper given at the 11th International

Congress on Sedimentology (International Association of Sedimentologists),

Hamilton, Ontario, Canada. It has also been submitted for publication

to Sedimentary Geology.

INTRODUCTION

The Trias sequence in the Midlands was first subdivided by Hull (1869) into a number of lithostratigraphic units (Table 2.1). He compared the Midlands Trias sequence with the Bunter, Muschelkalk and Keuper divisions of the German Trias. He stated that the Muschelkalk was absent in the Midlands and that it was represented by a regional unconformity. This led to the widely held belief that Hull's divisions had chronostratigraphic significance which could be applied to the rest of Britain. Warrington et al., (1980) in a review and revision of British Triassic stratigraphy have shown that this view was "fundamentally incorrect". They proposed dividing the Trias into three major lithological units. In ascending order these are, the Sherwood Sandstone Group (approximately equivalent to the Bunter and lower arenaceous part of the Keuper of previous authors), the Mercia Mudstone Group ("Keuper Marl") and the Penarth Group ("Rhaetic"). The terms in brackets are now discontinued. Rocks equivalent to all the standard Tethyan marine stages (Fig. 2.1) appear to be represented in the thicker British sequences (Warrington et al., 1980).

The Mercia Mudstone Group comprises predominantly argillaceous continental red beds. The Group shows considerable variations in thickness which reflect the horst and graben type structures present in Britain during the Triassic (Audley-Charles, 1970a; Warrington et al., 1980). The Mercia Mudstone Group in the East Midlands was divided into lithostratigraphic formations by Elliott (1961). These formations were retained by Warrington et al., (1980) with the exception of Elliott's Parva Formation. This was subdivided into the Glen Parva Formation and the overlying Blue Anchor Formation (Warrington et al., 1980). Figure 2.1 shows the complete new nomenclature for the North and East Midlands along with possible correlations with the standard Tethyan marine sequence (after Elliott, 1961 and Warrington et al., 1980). Correlation was carried out on the basis of scattered miospore assemblages and scanty macro-fossil data.

Correlation of the East Midlands lithostratigraphic formations with

Table 2.1   The classification of Triassic rocks in the English

Midlands, and correlatives in mainland Europe,

proposed by Hull (1869).

Fig. 2.1.   Subdivisions and correlations of the East Midlands

Triassic to the standard Triassic sequence (after

Elliott, 1961; Warrington et al., 1980).

English Midlands succession.	Foreign equivalents.
Rhaetic or Penarth Beds	
New Red Marl	Keuper
Lower Keuper Sandstone:	Letten Kohle (?)
Waterstones	
Building Stones	
Basement Beds	
("Wanting in England", i.e. stratigraphic gap)	Muschelkalk
Bunter Sandstone:	
Upper Red and Mottled Sandstone ⎫	
Pebble Beds ⎬	Bunter
Lower Red and Mottled Sandstone ⎭	Sandstein

STANDARD SEQUENCE OF TRIASSIC STAGES	AREA	NEEDWOOD BASIN (Fauld)		SOUTH NOTTINGHAMSHIRE
	LIAS	Lias		Lias
RHAETIAN		Penarth Group		Penarth Group
		Blue Anchor Fm.		Blue Anchor Fm.
		Glen Parva Fm.		Glen Parva Fm.
NORIAN		Trent Fm.	MERCIA MUDSTONE GROUP	Trent Fm.
				—HOLLYGATE SSTS—
CARNIAN		Edwalton Fm. HALITE		Edwalton Fm.
		Harlequin Fm.		Harlequin Fm.
LADINIAN		Carlton Fm.		Carlton Fm.
		Radcliffe Fm.		Radcliffe Fm.
				Colwick Fm.
ANISIAN		Sandstones		Woodthorpe Fm.
				non sequence
SCYTHIAN		Conglomerates		Sandstones and Conglomerates

▦ SHERWOOD SANDSTONE GROUP

surrounding areas is at best tenuous. Elliott (1961) based his correlation on both detailed analysis of sedimentary structures in closely spaced borehole cores and the presence of lithologically distinctive sandstone horizons. For example the Hollygate Skerries mark the Trent Formation – Edwalton Formation boundary. These sandstones frequently thin out away from horst areas which combined with variations in sedimentary structures makes recognition of the formation boundaries within adjacent basinal sequences difficult. Subdivision of the basinal sequences commonly relies on the presence of halite-bearing units. The Cheshire Basin sequence was subdivided in this manner by Pugh (1960) whereas halite units form important marker horizons in the Somerset, Central Dorset, Southern North Sea, Northern Irish, Needwood and Worcester Basins (Warrington et al., 1980). Halite units are absent on horst areas such as the East Midlands, hence difficulties are encountered when correlating basinal sequences to adjacent horst sequences. In short, lithostratigraphic correlations of the Mercia Mudstone Group have fallen into one of two broad categories and only rarely do the two meet.

Work on the Trent Formation in the North and East Midlands has resulted in a new method of dividing a sequence of lithologically monotonous unfossiliferous red bed mudstones. More importantly, the method enables the new units to be correlated from a horst to graben sequence over a distance of 70 km. A study of the clay mineralogy of the mudstones and oxygen isotope geochemistry of associated dolomite has shown that the Trent Formation can be divided into two members; the lower is here named the Fauld Member and is characterized by a suite of magnesium-rich transformational and neoformational clays with minor illite and chlorite. The upper member is here named the Hawton Member and is characterized by illite and minor chlorite. In addition the Fauld Member dolomites were formed in a predominantly marine brine regime whereas the Hawton Member dolomites were formed in a continental brine regime.

FAULD MEMBER.

The type section for the Fauld Member is at Fauld Mine (SK 181283)(Fig.2.2)

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Fig. 2.2.   Location map and major stratigraphical units of the East

Midlands. Geological linework based on Institute of

Geological Sciences geological maps.

PENARTH GROUP & JURASSIC

MERCIA MUDSTONE GROUP

SHERWOOD SANDSTONE GROUP

PALAEOZOIC

CHARNIAN + CALEDONIAN IGNEOUS

DANE HILLS SANDSTONE MEMBER

0   10 km.

R. TRENT

NEWARK

Staunton
Quarry

Hawton &
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Kilvington
Quarry

NOTTINGHAM

Cropwell Bishop
Quarry

GRANTHAM

R. DOVE

Keyworth

NEEDWOOD

Fauld Mine

Gotham
Mine

Bunny Brickpit

East Leake Mine

BASIN

BURTON

R. SOAR

R. TRENT

MELTON
MOWBRAY

Barrow upon Soar

LEICESTER

Gypsy Lane

N

Enderby

Croft

Stoney Cove

near Burton upon Trent, Staffordshire. The uppermost part of the member is exposed in the mine workings whilst a borehole in the mine floor cored the remainder of the member. It was also fully cored in the Institute of Geological Sciences borehole at Keyworth, Nottinghamshire (SK 62343074), (Fig. 2.2). The uppermost part of the member is exposed in East Leake Mine (SK 555280) and Glebe Mine (SK 538291) south of Nottingham (Fig. 2.2).

The Fauld Member is composed of massively bedded red-brown dolomitic mudstones and siltstones with rare thin sandy horizons (Windmill Skerries of Elliott, 1961) (Fig. 2.3). The sandy horizons are frequently massive and rarely exhibit sedimentary structures. The mudstones and siltstones contain small scattered pockets of angular to sub-angular quartz grains.

Gypsum or anhydrite is present in varying amounts throughout the member, occurring as nodules, near horizontal fibrous veins or vertical massive sulphate veins. A major development of sulphate, the Tutbury Sulphate Bed, previously called the Tutbury Gypsum (Stevenson and Mitchell, 1955) is present towards the top of the member (Fig. 2.3). The Tutbury Sulphate Bed is present in the Fauld area of the Needwood Basin (Fig. 2.2) and the area around East Leake. Farther east it gradually thins and passes into a nodular sulphate horizon. The thickness of the Tutbury Sulphate Bed varies between 1 and 3 m.

The Fauld Member extends to the base of the Trent Formation as defined by Elliott (1961). In the East Midlands horst area this boundary is marked by the top of the sandy Hollygate Skerries. This has been extended into graben areas with varying degrees of success (Warrington et al., 1980). The thickness of the Fauld Member varies between 25 and 31 m in the Needwood Basin whilst at Keyworth it is 20m thick.

HAWTON MEMBER

The type section for the upper 12.5 m of the Hawton Member is at the British Gypsum Hawton quarry (SK 805501) near Newark on Trent (Fig. 2.2) whilst the lowest 5 m have been cored in numerous adjacent British Gypsum boreholes (Fig. 2.4). Similar sections are also exposed in British Gypsum

Fig. 2.3. Lithological logs of the Fauld Member at Fauld Mine and

Keyworth. The former is a composite of the section

exposed in the mine and a borehole sunk in the mine floor.

The latter is from an I.G.S. borehole. Both complete

sections analysed for clay mineralogy.

FAULD

KEYWORTH

HAWTON MEMBER

FAULD MEMBER

TRENT FORMATION

EDWALTON FORMATION

KEY
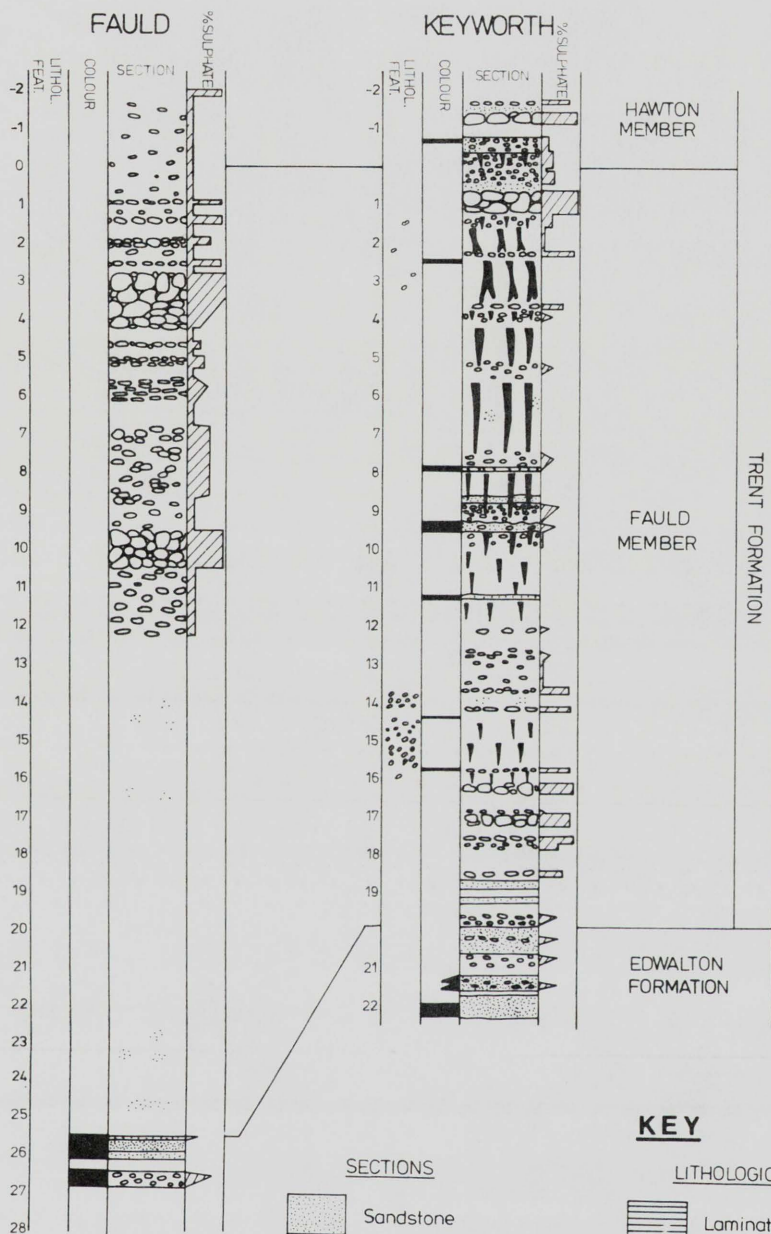
SECTIONS

	Sandstone
	Siltstone
	Mudstone
	Sulphate nodules Macro-Micro
	Quartz Grain 'Nests'
	Sulphate 'Veins'

COLOUR

	Reddish Brown
	Grey-Green
	Mixed Red-Brown & Grey-Green
	Minor Grey-Green

LITHOLOGICAL FEATURES

	Laminated Beds
	Dolomitized Horizon or Dolomitic Clasts
	Poorly laminated Beds
	Laminated beds cut by Siltstone
	Fish remains / fragments
	'Fish eyes' or reduction spots

SULPHATE %

30% 90%	Thickness in metres

quarries at Jericho (SK 808506), Staunton (SK 797448) and Kilvington (SK 797435). A section from the middle of the Hawton Member to within 1.5 m. of its base is exposed in Cropwell Bishop quarry (SK 675356) (Fig.2.2). The Hawton Member is named after Hawton quarry. Previous workers referred to part of this unit as the "Newark Gypsum" (Firman, 1964). This term was applied to any gypsiferous strata underlying the Penarth Group and as such is a generalized "bucket" term lacking in precise definition and has therefore been discarded.

In lithological terms the base of the Hawton Member cannot be defined; at Hawton and the surrounding area it lies within a sequence of red-brown muddy siltstones containing small scattered gypsum nodules, whereas at Keyworth it lies 20 to 80 cm. above a 65 cm. thick sulphate seam which possibly correlates with the Tutbury Sulphate Bed (Fig. 2.4). At East Leake and Fauld the boundary lies at varying distances above the top of the Tutbury Sulphate Bed. There is no distinct change in lithology to mark the boundary at any of these localities (Fig. 2.4).

The Hawton Member is predominantly composed of red-brown massively bedded dolomitic mudstones and siltstones with periodic sandier horizons. Two grey-green units present in the upper part of the member (Fig. 2.4) can be correlated over a distance of 25 km. These units contain laminated siltstones showing rhythmic graded bedding and fine grained sandstones which exhibit cross-bedding and disturbed, slumped bedding picked out by clay drapes. The units also contain rounded dolomitic clasts which were possibly derived from a pre-existing dolomite crust. By analogy with the work of Arthurton (1980) on the Cheshire Basin the grey-green units are interpreted by the author as lacustrine deposits which were subjected to periodic desiccation and erosion. The red-brown mudstones in the Trent Formation have been interpreted by numerous authors (Bosworth, 1913; Wills, 1970; Audley-Charles, 1970b), as desert red beds deposited either subaerially or subaqueously. The sulphate precipitated from hypersaline brines,which were also responsible for the formation of the dolomite.

The Hawton Member is overlain by the Glen Parva Formation (Warrington

Fig. 2.4.   Lithological logs of the Hawton Member at Fauld (composite

of British Gypsum borehole and I.G.S. Bagots Park borehole

(Stevenson and Mitchell, 1955)), Keyworth (I.G.S. borehole)

and Hawton near Newark (composite of quarry section and

British Gypsum borehole). Sections analysed for clay

mineralogy indicated by dotted line. Key as shown on

Fig. 2.3.

FAULD   KEYWORTH   HAWTON
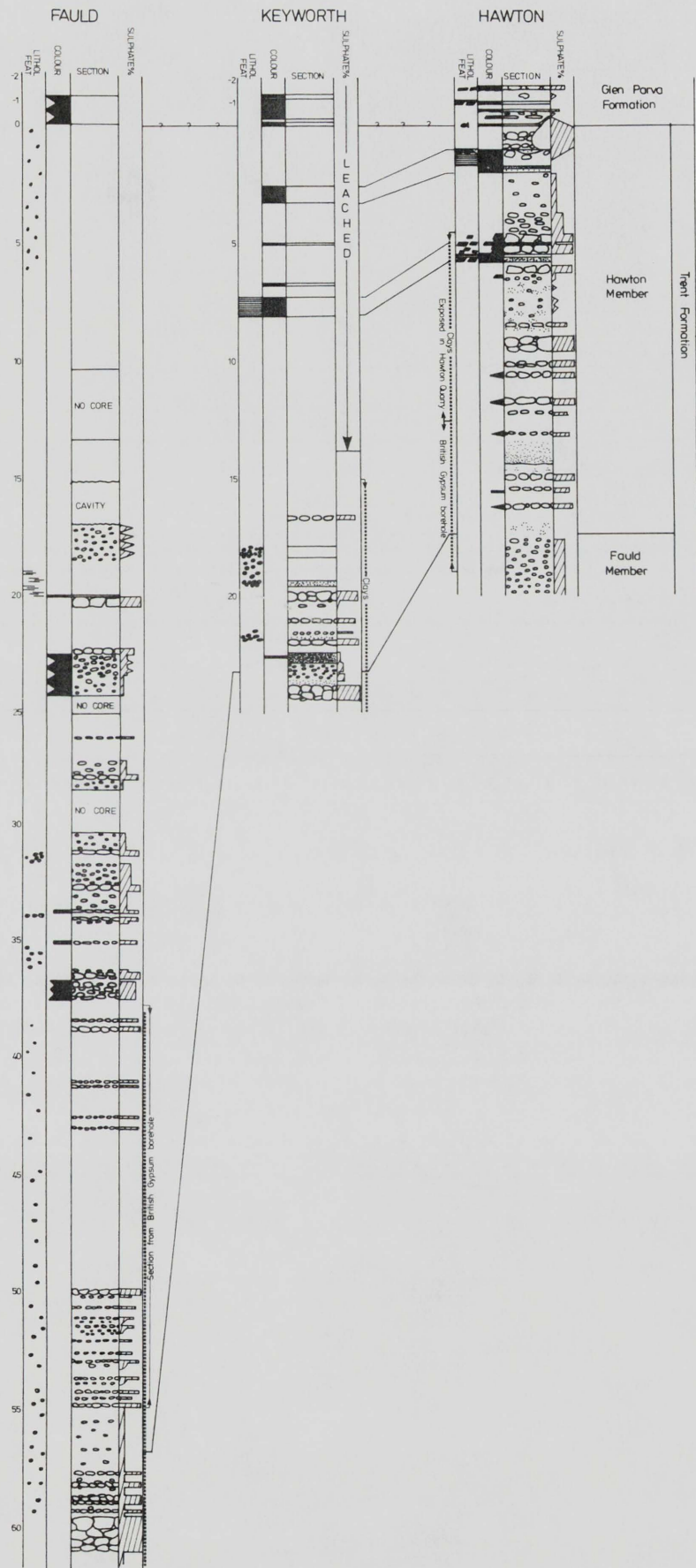
et al., 1980), the basal boundary of which is marked by the lowest fish-scale horizon. The nature of the boundary is variable; Elliott (1961) recorded "a sharp irregular surface" in the Fox Holes boreholes (338666), near Cropwell Bishop, whilst the author recorded an apparently conformable boundary in Jericho Quarry, Newark, with no evidence of an erosional surface.

CRITERIA FOR SUBDIVISION OF THE TRENT FORMATION.

The Trent Formation as previously described consists of a lithologically monotonous succession of dolomitic red-bed mudstones and siltstones which cannot be subdivided on either a biostratigraphical or conventional lithostratigraphical basis. However, the Trent Formation shows a distinct coincident variation in clay mineralogy and dolomite oxygen isotope geochemistry which enables one to define the two new members described above. Each member can be characterized in terms of clay mineralogy and, in most areas, $\delta^{18}O$ dolomite values.

CLAY MINERALOGY.

The clay fraction of mudstones and siltstones was analysed using standard X-ray diffraction techniques (Brindley and Brown, 1980). The clays were air-dried, glycerolated, heated to 440°C. (in some cases 550°C) and run on a Phillips diffractometer after each treatment. Semi-quantitative analysis of the results was carried out using the method described in Biscaye (1965) and Bradshaw (1976).

The Fauld Member contains a suite of transformational and neoformational magnesium-rich clay minerals with minor illite and chlorite. The magnesium-rich suite comprises sepiolite and palygorskite and irregular mixed-layer smectite/chlorite containing varying proportions of smectite. Smectite/chlorite occurs throughout the Fauld Member across the whole area, whereas the sepiolite and palygorskite have a restricted distribution (Fig. 2.5). The proportion of magnesium-rich clays in the Fauld Member varies between 15 and 80% with an average of 43.5%. Sepiolite and palygorskite precipitated from

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Fig. 2.5.   Clay mineralogy and associated $\delta^{18}O$ dolomite values of the

Hawton and Fauld Members. The boundary between the two

members is taken at 15% mixed-layer content which at Fauld

and Newark coincides with a marked change in $\delta^{18}O$ dolomite

values (see text). The Fauld Member is characterized by

the predominance of mixed-layer and Mg-rich clays, whilst

the Hawton Member is characterized by illite and chlorite.
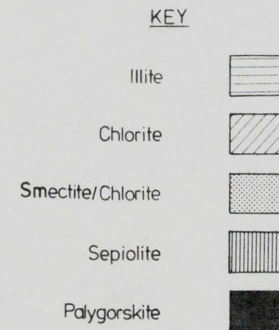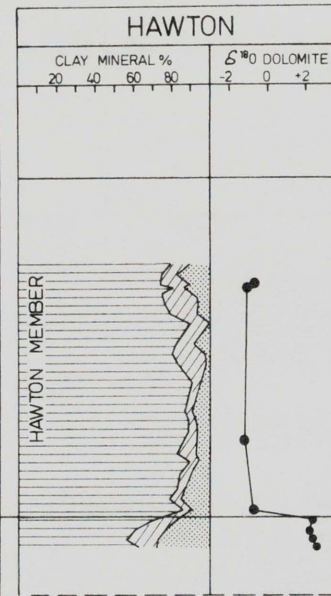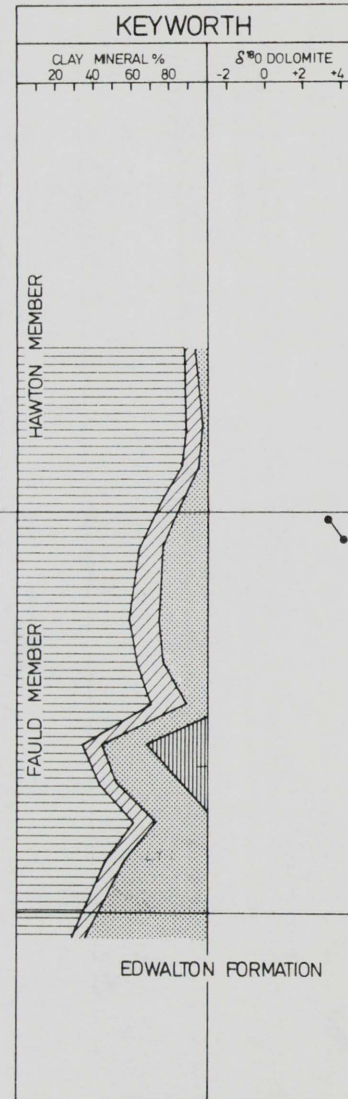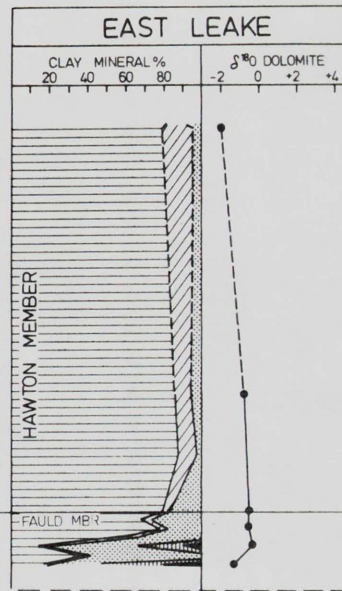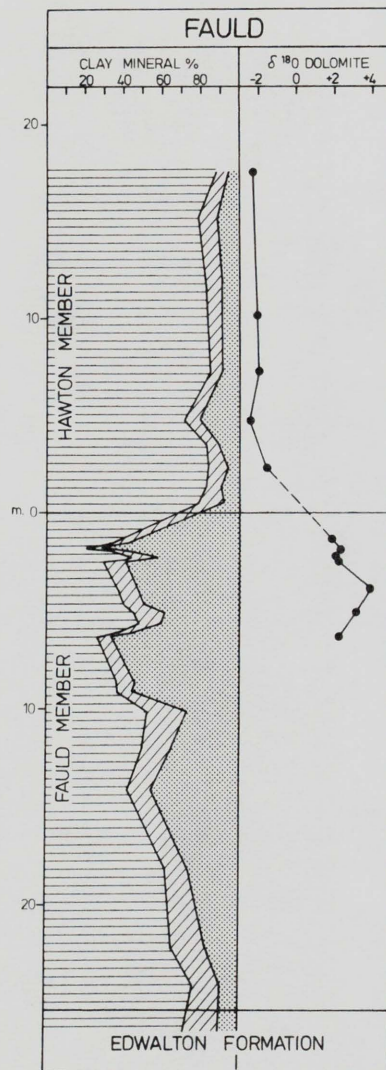
alkaline brines rich in Mg^{2+} and amorphous silica. Smectite/chlorite resulted from a transformational reaction between degraded detrital illite and alkaline Mg^{2+} rich brines. The Mg^{2+} and amorphous silica contents were enriched by the earlier precipitation of sulphate from the brines and continental input respectively (Chapter 7). The illite and chlorite are thought to be of detrital origin (Jeans, 1978).

The clay mineralogy of the Hawton Member consists predominantly of illite with minor chlorite (Fig. 2.5). Irregular mixed-layer smectite/chlorite is present in minor amounts (0 to 20%, average 8.5%). In contrast to the smectite/chlorite of the Fauld Member, the Hawton Member mixed-layer mineral contains very few smectite interlayers. Sepiolite and palygorskite were not recorded in the Hawton Member. The origins of the different clays are similar to those of the Fauld Member clays. The change in clay mineralogy is very marked, occurring within 1 to 2 m. throughout much of the area (Fig. 2.5). The Hawton Member base is arbitrarily fixed at the point where the average magnesium-rich content falls below 15%. The change in clay mineralogy probably reflects an alteration of brine chemistry, for example a reduction in Mg^{2+} , amorphous silica and/or pH. At Fauld and Newark the change in clay mineralogy is coincident with a change in $\delta^{18}O$ dolomite values.

DOLOMITE - OXYGEN ISOTOPES.

The carbon and oxygen isotopic ratios of dolomite within the Trent Formation mudstones were analysed at the Institute of Geological Sciences Geochemical Unit, London. It is the oxygen isotopic ratios which are of interest here. Results are expressed as parts per thousand (‰) relative to the P.D.B. standard (Chapter 5).

The Fauld Member dolomites are characterized by having $\delta^{18}O$ values which lie in two broad fields (Fig.2.5).

(1) $\delta^{18}O$ of +2‰ to +4‰.

(2) $\delta^{18}O$ of -1.5‰ to 0‰.

Analyses from Fauld, Keyworth and Newark lie within the former range of values whereas analyses from East Leake lie within the latter range. East

Midlands dolomites with $\delta^{18}O$ values between +2‰ and +4‰ are interpreted as having been formed from predominantly marine brines (Chapter 5). From the spatial distribution of these values it is postulated that marine brines were confined to graben areas such as Fauld and immediately adjacent horst areas such as Newark and Keyworth. Dolomites with $\delta^{18}O$ values between -1.5‰ and 0‰ are interpreted as having been formed from continental brines. It is inferred that continental brines were present in the East Leake area in the centre of the East Midlands horst block, close to the Charnwood Forest Massif (Chapter 5).

The Hawton Member dolomites have $\delta^{18}O$ values between -2.5‰ and 0‰ (Figure 2.5), indicating that the dolomite formed in a continental brine regime (Chapter 5). No dolomites of marine origin were identified in the Hawton Member.

Therefore at Newark and Fauld isotopically "heavy" marine dolomites of the Fauld Member are overlain by isotopically "light" continental dolomites of the Hawton Member. At Fauld, the Fauld - Hawton Members' $\delta^{18}O$ boundary lies within 4 m. of unanalysed sequence overlying the Tutbury Sulphate Bed (Fig. 2.5). At Newark, continental brine regime dolomites have been found 0.5 m. above marine brine regime dolomites (Fig. 2.5). Dolomites of mixed marine - continental origin have not been detected. This indicates that the alteration from a marine to a continental brine regime must have occurred fairly rapidly. The sequence at East Leake does not show any alteration of $\delta^{18}O$ dolomite as continental brines prevailed in both Fauld and Hawton Members. This indicates that factors such as Mg^{2+} concentration were of more importance for the formation of smectite/chlorite than the initial source of the brines.

The alteration in clay mineralogy and $\delta^{18}O$ dolomite coincides at Fauld and Newark (Fig. 2.5). At both localities this horizon is taken as the boundary between the Hawton Member and Fauld Member, whereas at East Leake the boundary is marked solely by the change in clay mineralogy (Fig. 2.5). As previously mentioned, the exact position of the boundary is arbitrarily fixed at the 15% magnesium-rich clay content.

20

DISCUSSION

Both clay mineralogy and dolomite oxygen isotope geochemistry are facets of brine chemistry in much the same way as the better known carbonate, sulphate and chloride evaporite sequence. The latter chemical sequence has been used for lithostratigraphic correlation purposes, such as in the Cheshire Basin (Warrington et al., 1980). The major difference between the evaporite sequence and clay mineral variations lies in ease of identification. Neither clay mineralogy or $\delta^{18}O$ dolomite can be determined in the field. However, this difficulty is more than compensated for by the fact that the clay mineral variations have been successfully correlated from a graben sequence to a horst sequence. This transition has been one of the major stumbling blocks of British Triassic stratigraphy. The inability to correlate from graben to horst stems from both sequences being subdivided on the basis of lithological features unique to that type of sequence. As previously discussed halite is confined to grabens and has been used extensively to subdivide graben sequences. Conversely horst sequences were subdivided using well developed sandstone units which are only poorly developed in graben sequences. Hence the basis of subdivision of horst sequences is effectively removed for graben sequences and vice versa. Now studies of clay mineralogy and $\delta^{18}O$ dolomite, as just described, may provide an alternative method of lithostratigraphic correlation within the Mercia Mudstone Group.

EXTENSION TO OTHER AREAS OF BRITAIN.

The test of any stratigraphical correlation comes with its extension to areas away from the study area. Jeans (1978) carried out an extensive survey of the clay mineralogy of the Mercia Mudstone Group. The Hawton and Fauld Members can be readily distinguished in the Clipston borehole (Fig.2.6) southeast of Nottingham (Jeans,1978). Good correlation was achieved for both members with the nearby Keyworth borehole. Samples of the upper part of the Trent Formation from brick pits near East Leake and Leicester contained the typical Hawton Formation assemblage of illite with minor chlorite (Jeans,1978).

Fig. 2.6. Comparison of X.R.D. traces from Fauld Mine (present study)

with traces from the Wilkesley, Clipston, Upton and Stowell

Park boreholes (Jeans, 1978). In the Clipston and Upton

boreholes and at Fauld there is a distinct increase in the

size of the glycerolated mixed-layer smectite/chlorite

peak (16 - 18Å) which is correlated with the top of the

Fauld Member. In the Wilkesley and Stowell Park boreholes

the boundary is less well defined.

WILKESLEY   FAULD   CLIPSTON   UPTON   STOWELL PARK

KEY

Blue Anchor Formation

E.F. Edwalton Formation

G.P.F. Glen Parva Formation

Hollygate Skerries/Arden Sandstone

Hawton Member ⎫ Trent
Fauld Member ⎭ Formation

Norian landmasses
☆ Borehole on horst
★ Borehole in basin
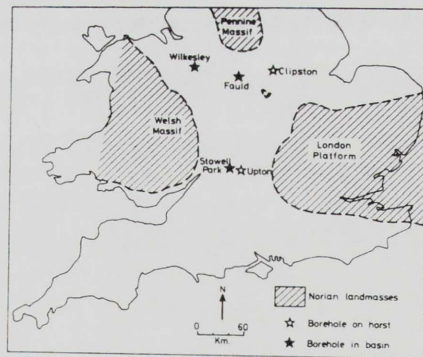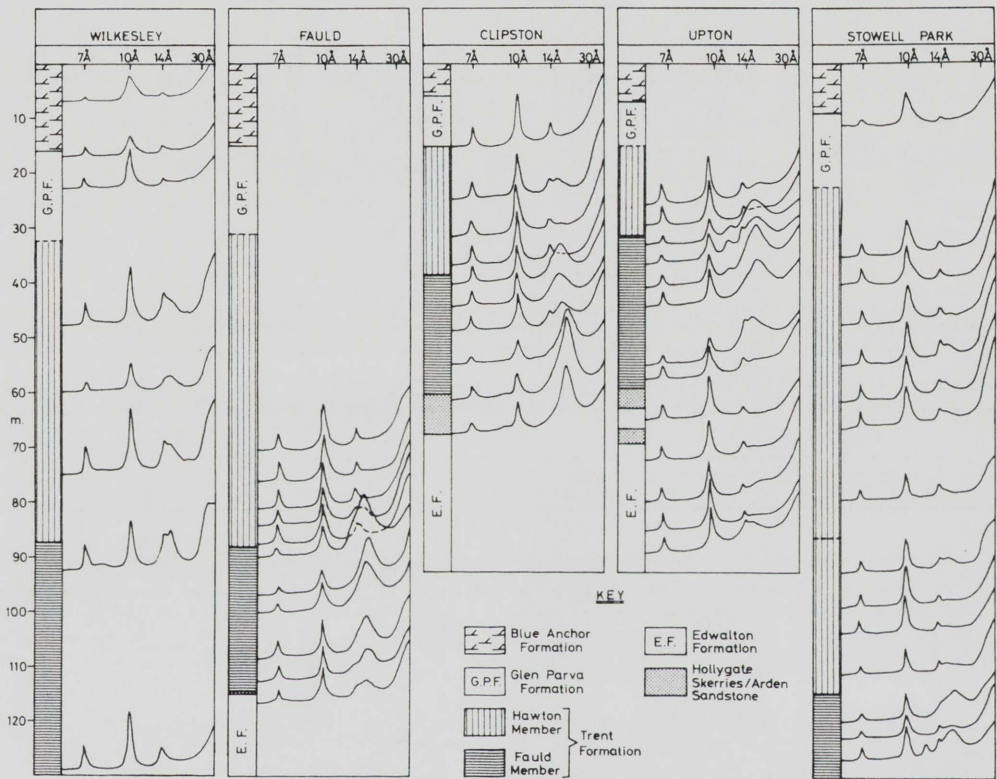
Samples from the Staunton and Cropwell Bishop gypsum pits also contained similar Hawton Member assemblages (Jeans, 1978).

Extension of the members into the Cheshire Basin is more difficult. Jeans (1978) analysed samples from the Wilkesley borehole (SJ 62864144) (Fig. 2.6) with an average sampling interval of 23 m. through mudstones overlying the Wilkesley Halite Formation (Warrington et al., 1980). The small number of samples combined with the large sampling interval hinder accurate correlation of this sequence with the East Midlands sequence. However, a distinct increase in corrensite content (a regular mixed-layer smectite-chlorite) occurs between two samples taken at 75 and 95 m. below the Penarth Group (Fig. 2.6). The Hawton Member - Fauld Member boundary may lie within this interval.

Similarly the Hawton Member - Fauld Member boundary is not obvious in the Stowell Park borehole (SP 084118) (Fig. 2.6),Gloucestershire. Careful examination of Jean's results show that the boundary may be marked by an increase in corrensite content which occurs between two samples taken at 80 and 90 m. below the Penarth Group (Fig. 2.6). Another strong possibility is that the boundary is marked by a large increase in magnesium-rich clay content between 111 and 119 m. below the Penarth Group. Sandy horizons 140 and 160 m. below the Penarth Group (Fig. 2.6) have been tentatively correlated with the Arden Sandstone Member (Warrington et al., 1980) and thence to the Hollygate Skerries (Audley-Charles, 1970a). Therefore the top sandy horizon may mark the base of the Fauld Member. By contrast, in the nearby borehole at Upton (SP 23151313) (Fig. 2.6), Oxfordshire, Jeans (1978) detected a very marked increase in magnesium-rich clay content 30 to 33 m. below the Penarth Group (Fig. 2.6). Samples between the Blue Anchor Formation and 30 m. below the Penarth Group contain illite with minor chlorite which, as previously stated, is typical of the Hawton Member (Fig. 2.6). Samples from 33 to 65 m. below the Penarth Group have a clay mineralogy typical of the Fauld Member. The thinning of the Penarth Group to Fauld Member interval reflects the transition from a graben sequence at Stowell Park to a horst sequence at Upton (Audley-Charles, 1970b) (Fig. 2.6).

It is not yet possible to state with certainty that the increases in
magnesium-rich clay mark the Hawton Member - Fauld Member boundary in Cheshire,
Gloucestershire and Oxfordshire, but the clay mineral distribution in those
areas closely resembles that found in the East Midlands Trent Formation.
Further work is needed to establish any possible positive correlation. This
would entail closely spaced sampling of the Wilkelsey and Stowell Park bore-
holes and sampling from boreholes more closely adjacent to the East Midlands.
Both clay mineralogy and $\delta^{18}O$ dolomite should be analysed to aid correlation
with the East Midlands sequence.

CONCLUSIONS

The lithologically monotonous Trent Formation red beds of the Upper
Triassic succession in the North and East Midlands have been divided into a
lower Fauld Member and an upper Hawton Member. The division is based on
coincident variations in clay mineralogy and oxygen isotope geochemistry of
associated dolomite. The clay mineralogy of the Fauld Member is characterized
by a suite of magnesium-rich transformational and neoformational clays
(greater than 15%) with minor illite and chlorite. The Hawton Member contains
predominantly detrital illite and chlorite with minor magnesium-rich clay
minerals (less than 15%). The $\delta^{18}O$ dolomite analyses indicate that during
the deposition of the Fauld Member a marine brine regime dominated in graben
areas and closely adjacent horst areas whereas the centre of the East Midlands
horst block had a continental brine regime. The Hawton Member was deposited
in a predominantly continental brine regime.

The Fauld and Hawton Members are recognized in both a graben sequence and
horst sequence for a total distance of 70 km. The ability to correlate from
a horst to graben sequence by using clay mineralogy and dolomite oxygen
isotope analyses has overcome one of the major difficulties of British
Triassic lithostratigraphy. Utilization of previously published work
(Jeans, 1978) indicates that similar variations in clay mineralogy are
widespread and occur in coeval strata in Cheshire, Gloucestershire and

Oxfordshire. Further work is needed to correlate accurately these variations with the East Midlands sequence. The study of clay mineralogy and $\delta^{18}O$ dolomite could be used as a powerful tool to aid the lithostratigraphic correlation of the Mercia Mudstone Group throughout the British Isles.

CHAPTER   3

The author regrets that chapter 3 has had to be

omitted from this copy of the thesis due to its

confidential nature. Anyone wishing to read this

chapter should contact:-

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CHAPTER 4.

GEOLOGY OF FAULD MINE, NEAR TUTBURY,

STAFFORDSHIRE.

INTRODUCTION.

The Upper Triassic Trent Formation of the East Midlands area consists of a sequence of lithologically monotonous continental red bed mudstones. The formation is noted for its economically important gypsum and anhydrite deposits (Elliott, 1961). One such deposit is the Tutbury Sulphate Bed which is worked at Fauld Mine (41813283), the entrance of which lies 3 km. west of Tutbury and 17 km. north west of Burton upon Trent. The mine workings extend over an area of 4 sq. km. to the south of the entrance (Fig. 4.1). The modern workings expose a 2.5 m. section through the Tutbury Sulphate Bed and overlying strata. Recently, a new drivage (mine passage) has exposed a 2.5 m. section below the base of Tutbury Sulphate Bed.

The Fauld Mine deposit was first described by Wynne (1906) with later brief accounts by Sherlock and Hollingworth (1938) and Stevenson and Mitchell (1955). All these accounts give short descriptions of the geology concentrating particularly on the Tutbury Sulphate Bed. The aim of this chapter is to redescribe the geology of the Tutbury Sulphate Bed and associated strata.

LITHOLOGY AND MINERALOGY.

The section which is consistently exposed throughout Fauld Mine shows the Tutbury Sulphate Bed overlain by brown-red mudstones which contain variable amounts of sulphate. This gives the mudstones a distinctive layered appearance (Plate 4.1, Fig. 4.2). The layering is remarkably consistent as individual layers have been correlated across the whole mine area with only minor variations in thickness.

SULPHATE.

The sulphate occurs as ovoid shaped nodules within the mudstone matrix. The diameter of the nodules varies from 0.5 to 30 cm. with a particular

Fig. 4.1.   Location map and geology of the Needwood Basin between

Burton-upon-Trent, Tutbury and Bagots Park. Also shown are

the Fauld Mine workings which extend southwards from the

mine entrance on the outcrop of the Tutbury Sulphate Bed.

R. DOVE

HATTON CREAMERY
● BOREHOLE

OUTCROP OF TUTBURY SULPHATE BED

TUTBURY

BAGOTS PARK
BOREHOLE

FAULD MINE
WORKINGS

NEEDWOOD
BASIN

BURTON
UPON
TRENT

R. TRENT

Penarth Group

Blue Anchor Formation

Mercia Mudstones Group,
pre–Blue Anchor Formation
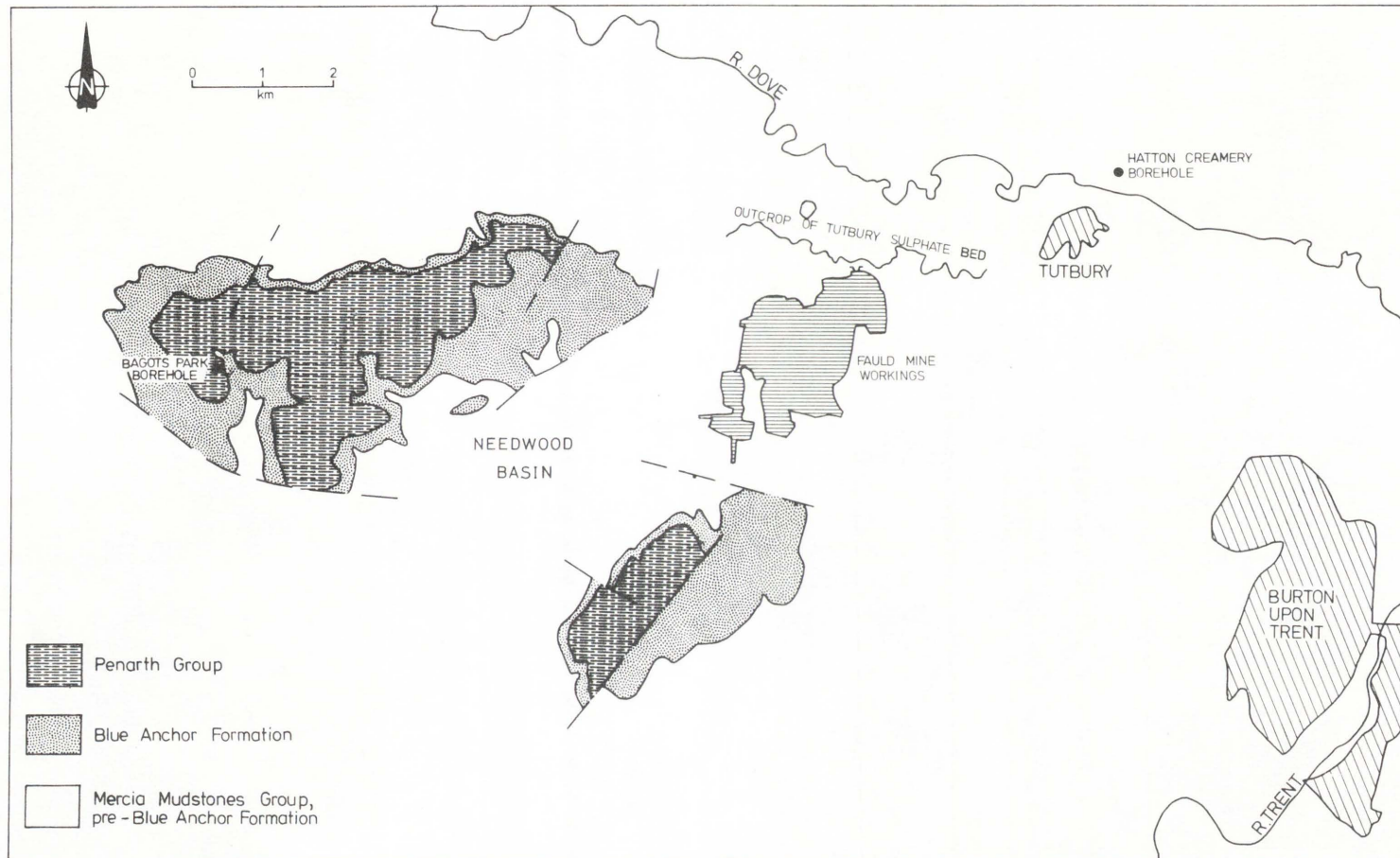
Plate 4.1.   Typical working face in Fauld Mine. The Tutbury

Sulphate Bed covers the lower half of the face

and comprises white coalescing gypsum nodules.

The Tutbury Sulphate Bed is overlain by mudstones

containing scattered sulphate nodules and a thin,

well-defined sulphate horizon. The darker grey

material is dust.

Fig. 4.2.   Lithologies exposed in normal mine workings and drivage

beneath Tutbury Sulphate Bed, Fauld Mine.   Also shown are

typical sulphate and dolomite contents.   The latter were

determined by X.R.D. studies and two sets of analyses are

shown.

ROOF

% SULPHATE CONTENT
0   50   100

% DOLOMITE CONTENT
0   20   40   60   80   100

BLOCKY RED MUDSTONE

MANY MICRO NODULES
OR SCATTERED
10 cm DIAMETER NODULES

BLOCKY RED MUDSTONE

BLOCKY RED MUDSTONE

TUTBURY SULPHATE
BED

0

m

1

DRIVAGE SECTION

LENSE OF ELONGATED
ANHYDRITE NODULES

VARIABLE

NODULAR DOLOMITE
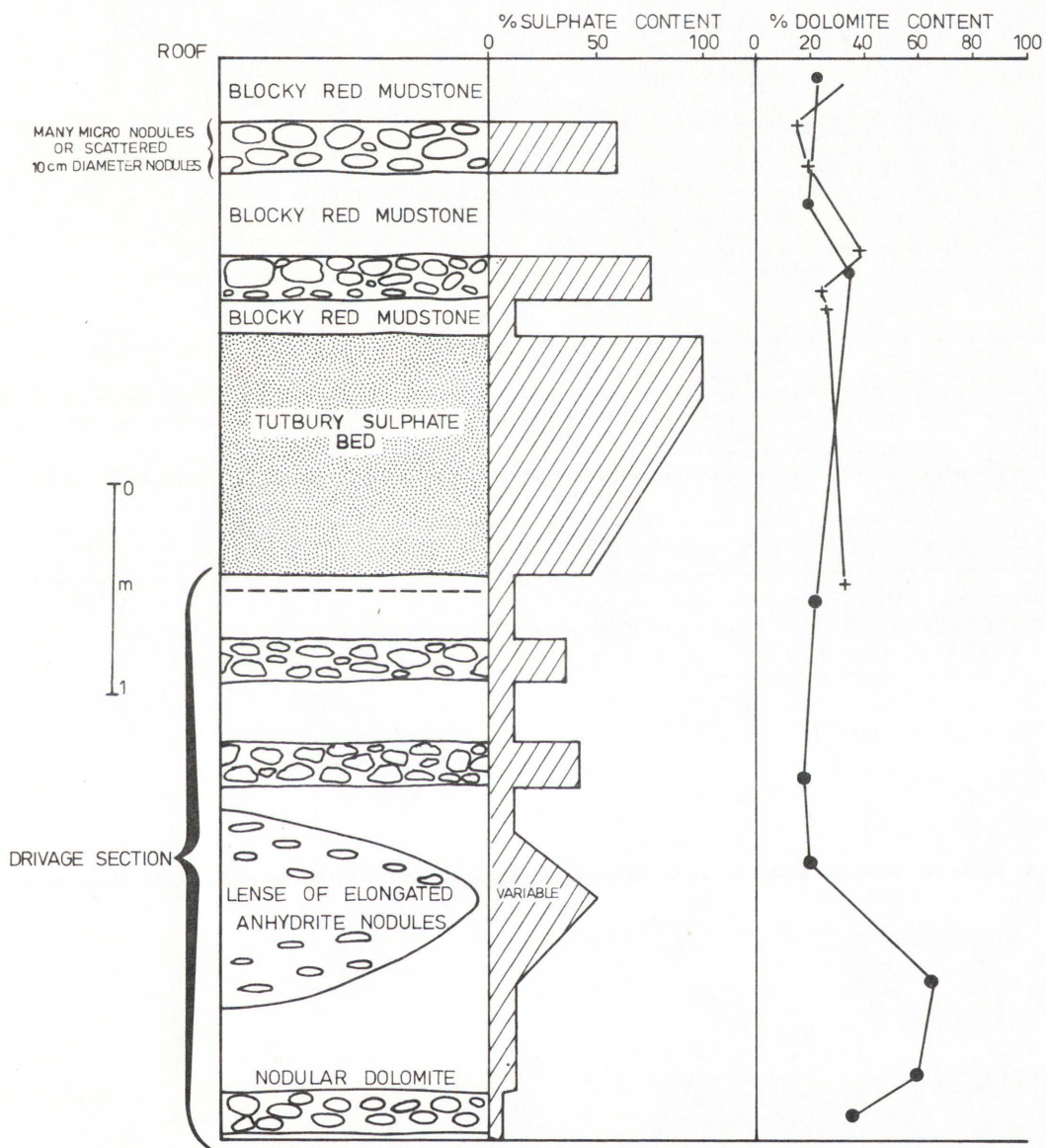
horizon being characterized by sulphate nodules of similar size. The
average diameter of nodules forming the Tutbury Sulphate Bed is 20 to 30 cm.
with nodule size and frequency decreasing towards the base of the bed,
i.e. the basal part of the bed contains a higher proportion of mudstone.
The nodules sometimes coalesce to give a chicken-wire texture (Shearman,1966)
which is picked out by trapped mudstone. Plate 4.2 illustrates the well
defined layering just above the massive top of the Tutbury Sulphate Bed.
A chicken-wire texture can be seen within both the sulphate horizons above
the Tutbury Sulphate Bed. The lighter red-grey is surface dust which is
partially obscuring the massive sulphate lithology of the Tutbury Sulphate
Bed. Both the Tutbury Sulphate Bed and overlying strata are transected
by large sub-spheroidal sulphate masses between 2 and 12.5 m. in diameter.
The morphology and origin of these masses is described later.

The sulphate occurs as either gypsum or anhydrite. In areas of the
mine close to the surface the hydration of anhydrite to gypsum is complete
whilst in the more deeply buried parts of the mine only anhydrite is
present. The anhydrite often has a schistose type texture with individual
anhydrite laths orientated parallel to each other. The laths pick out
minor folds within anhydrite nodules (Plate 4.3). In intermediate areas
the anhydrite has been partially hydrated to gypsum. Hydration first occurs
around the outer margins of the sulphate nodules leaving an anhydrite core.
The core is often cut by thin gypsum veinlets (Plates 4.4, 4.5), which
contain "rafts" of remnant anhydrite laths. The veinlets may have resulted
from water entering microfractures in the anhydrite. They probably mark the
site of present day hydration. The veinlets are either filled with large
plates of gypsum (Plate 4.4) or with "feathery" alabastrine gypsum
(Plate 4.5). The exact controls on the formation of the two crystal types
is uncertain.

The base of the Tutbury Sulphate Bed forms the mine floor so until
recently the exact nature of the underlying beds was unknown. However, a new
drivage under the seam (commenced May, 1981) exposed a further 2.5 m.

Plate 4.2A.   Layering of strata above Tutbury Sulphate Bed –
               a reflection of varying sulphate content. The
               top of the Tutbury Sulphate Bed is a well defined
               plane occurring just below the red 11. The grey
               is surface dust.

Plate 4.2B.   Close-up of above. The Tutbury Sulphate Bed top
               lies just below the red 1. The overlying sulphate
               horizons have a good chicken wire texture and
               planar upper boundaries. These are features
               indicative of interstitial sulphate growth and
               possible later erosion of the upper boundary
               (see text).

Plate 4.3A.   Folded anhydrite laths with trapped mudstone
               (PPL, x 10).

Plate 4.3B.   Folded anhydrite laths (XP, x 10).
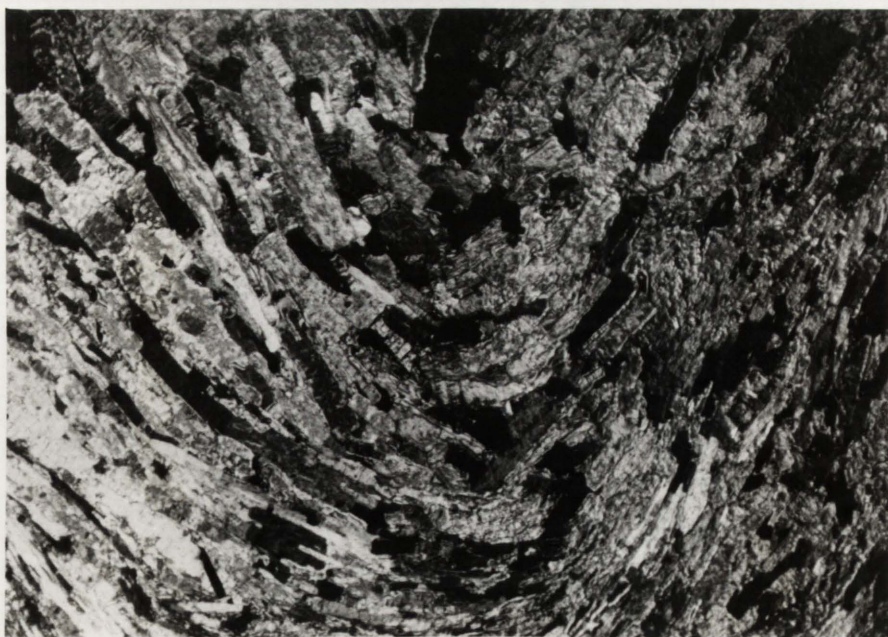
Plate 4.4A.   Hydration of anhydrite results in anhydrite (high

relief) laths or rafts floating in gypsum.

Anhydrite laths are fragmented by hydration

occurring along cleavage planes (PPL, x 10).

Plate 4.4B.   As above (XP, x 10). Note that the anhydrite is

enclosed by large optically continuous plates of

gypsum.
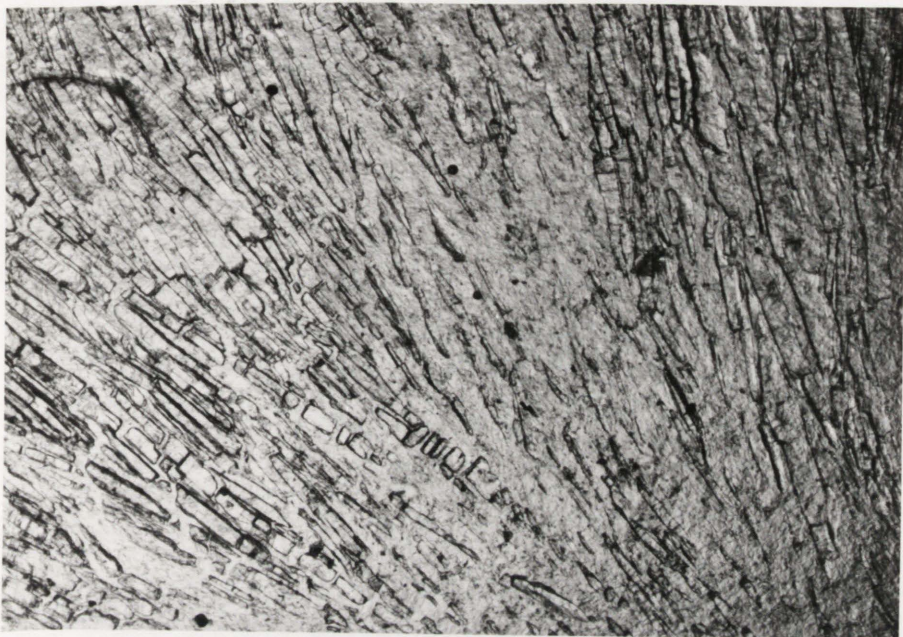
Plate 4.5A.   Hydration of anhydrite - anhydrite laths in

large plates of gypsum (XP, x 16).

Plate 4.5B.   Feathery gypsum vein cutting anhydrite

(XP, x 12.8).

sequence. Two sulphate rich horizons occur in this lower sequence (Fig.4.2), one of which merits further description. It contains many highly elongated anhydrite nodules which have an average horizontal diameter of 10 cm. and a thickness of 1 cm. (Plates 4.9, 4.10). The nodules are concentrated into lenses, 90 cm. high and 10 - 15 m. in diameter, between which are sulphate poor mudstones. The degree of flattening or elongation of the nodules can be seen by comparing their 10:1 axial ratio with the 2:1 axial ratio of a sulphate nodule from the Tutbury Sulphate Bed. The highly elongated anhydrite fabric is similar to that found in metamorphic rocks which have undergone crystal flowage (Spry, 1969). The anhydrite fabric therefore provides evidence that extensive horizontal sulphate flowage has occurred along this horizon. The importance of such extensive sulphate movement is discussed later.

MUDSTONE.

The mudstones are reddish-brown, frequently massive, poorly jointed and do not show distinct bedding structures. The only bedding which is distinguishable is due to either variation in sulphate content or the degree of fracturing of the mudstone.

In thin section the mudstones are seen to contain quartz grains between 0.175 and 0.030 mm. in diameter, with occasional plagioclase grains (Plate 4.6A and B). The grains are angular to sub-angular with rounded grains only rarely occurring (Plate 4.6C). The grains are randomly scattered throughout the dolomite-clay matrix with no indication of preferential orientation. In many of the mudstones the quartz grains are matrix supported (Plate 4.6C). When the quartz content increases the grains touch to form a clast supported rock (Plate 4.6A, B and Plate 4.7). The larger quartz grains commonly form "nests" where they accumulate together in a small area. The quartz content varies between 15% and 40%. Dolomite is evenly distributed throughout the mudstone and occurs as euhedral rhombs with an average diameter of 0.006 mm. (Plate 4.8). The largest observed rhombs had a diameter of 0.03 mm.

Plate 4.6A.   Mudstone containing a high proportion of

moderately rounded, predominantly quartz clasts

(PPL, x 16).

Plate 4.6B.   As above (XP, x 16). Note the unweathered

plagioclase and a (bright) carbonate clast on

the right.

Plate 4.6C.   Typical mudstone from Fauld Mine containing

angular to subangular quartz and tiny dolomite

rhombs (PPL, x 10).
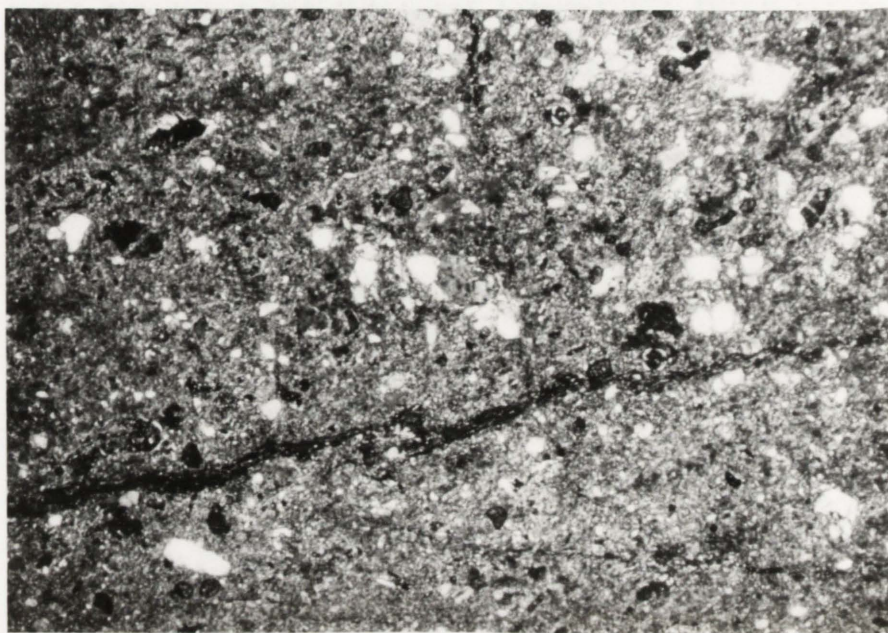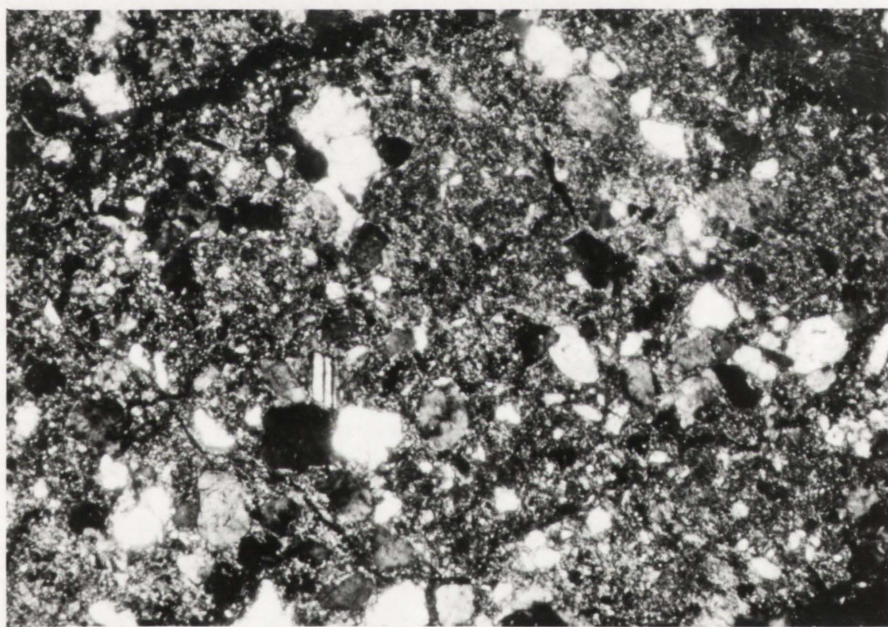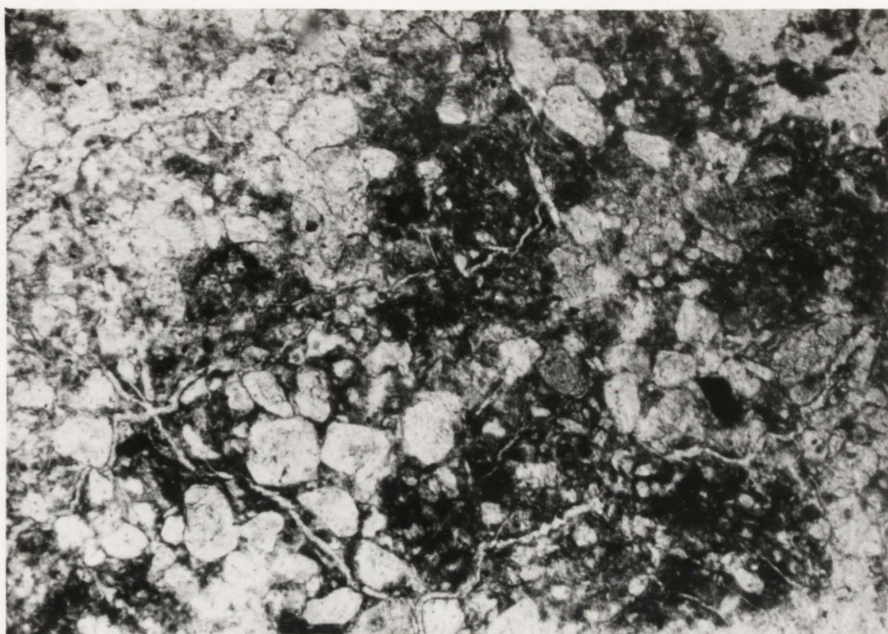
Plate 4.7A.   Sandier horizon in mudstone, from Fauld Mine

roof. Note angularity of some grains

(PPL, x 16).

Plate 4.7B.   As above, angular quartz grains with scattered

dolomite rhombs (PPL, x 64).
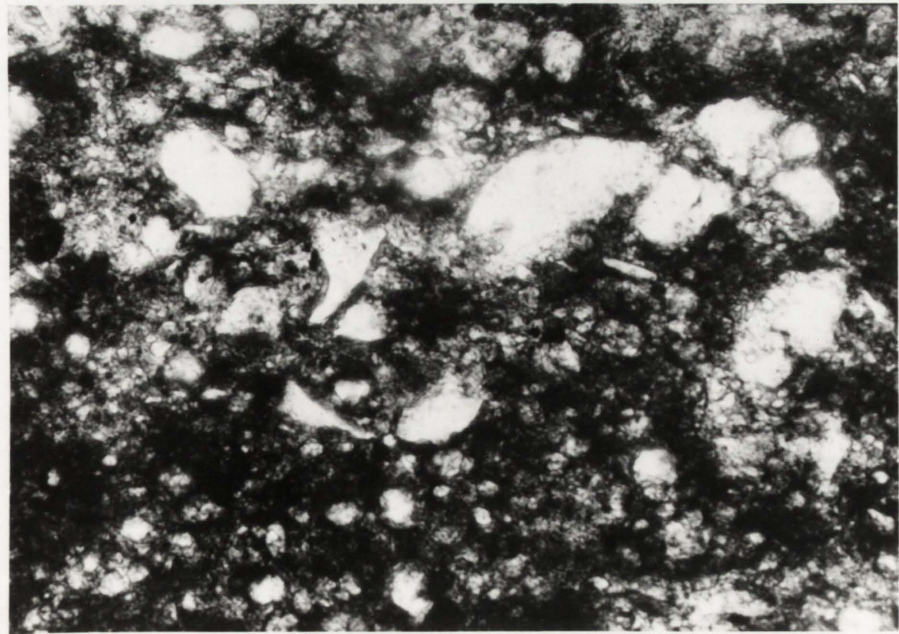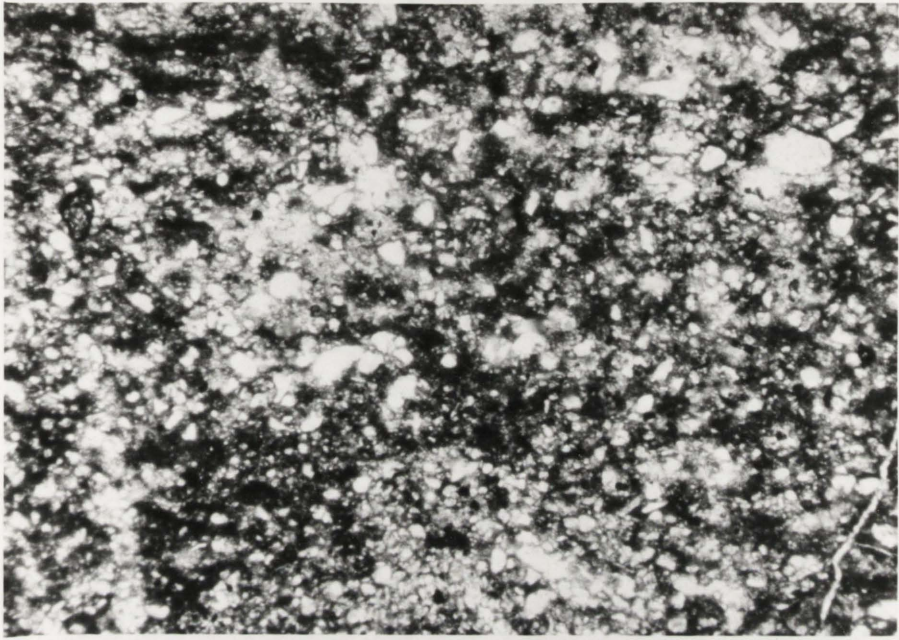
Plate 4.8A.   Mudstone containing dolomite rhombs (higher

relief) (PPL, x 64).

Plate 4.8B.   As above (PPL, x 205). Most of the dolomites

are anhedral with a few euhedral rhombs.
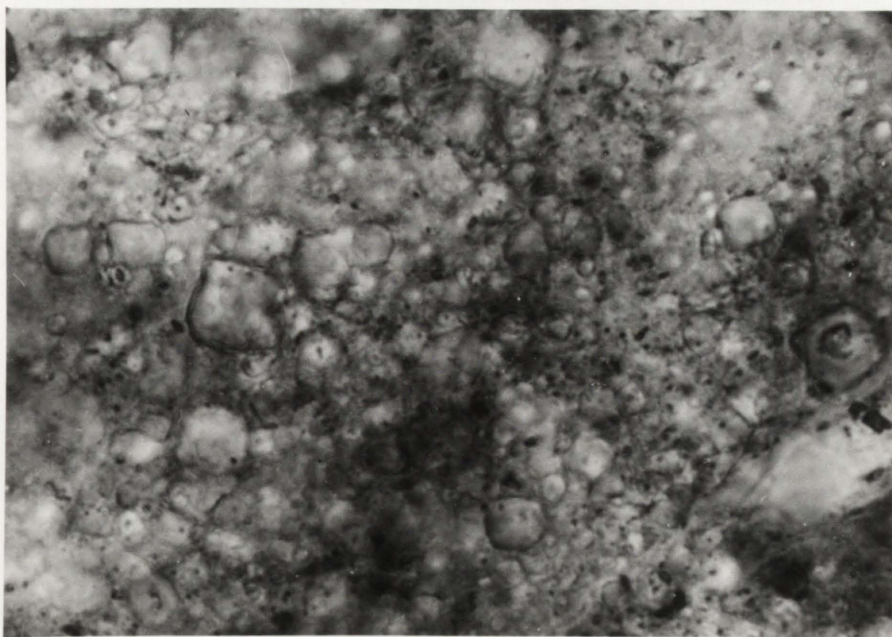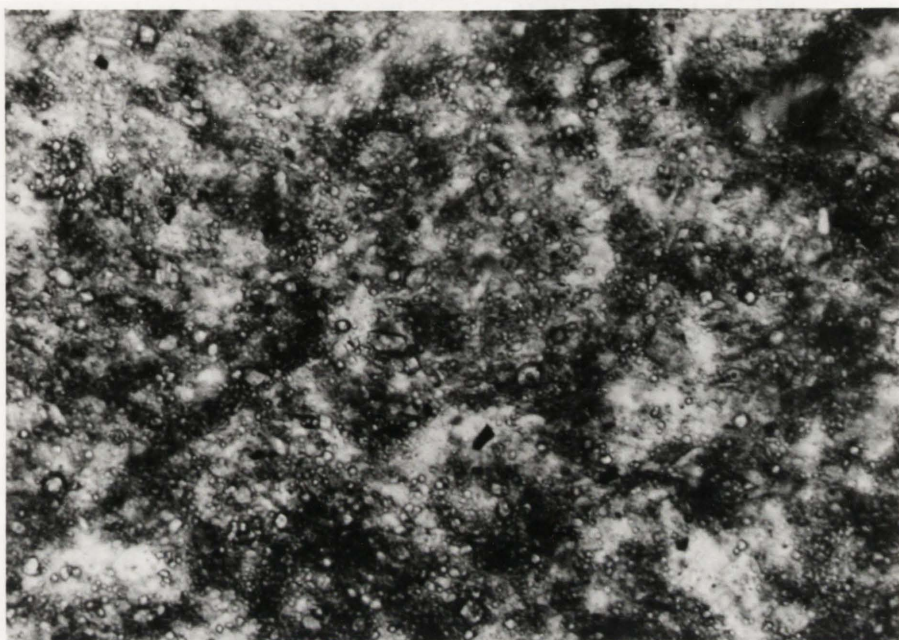
Plate 4.9.   Elongated anhydrite exposed in mine wall,

drivage beneath Tutbury Sulphate Bed.

Plate 4.10.   Close-up of elongated anhydrite.

Scale bar in cm.

X.R.D. studies show that dolomite content varies between 20 and 40% (Fig. 4.2). An exception to this is the extensively dolomitized mudstones between the elongated anhydrite lenses. Here the dolomite content rises rapidly to a peak of 60% towards the base of the elongated anhydrite horizon (Fig. 4.2). The dolomite within the inter-lens mudstone and the mudstones 50 to 75 cm. below this layer occurs as evenly distributed euhedral rhombs 0.002 to 0.018 mm. in diameter. Dolomite within underlying mudstones forms massive, well-cemented nodules up to 10 cm. in length (Fig. 4.2), which occasionally coalesce to form a chicken-wire texture (Plate 4.11). The nodules are composed of tightly packed poorly formed dolomite rhombs 0.005 mm. in diameter (Plate 4.12) Intranodular material comprises mudstone, sulphate and euhedral dolomite rhombs up to 0.025 mm. in diameter (Plates 4.11, 4.12).

Sulphate occurs as intranodular material and as scattered grains within the dolomite nodules. In all observed cases the boundary between sulphate and dolomite is distinct with the dolomite rhombs forming preferentially to the sulphate (Plate 4.13). There is no clear evidence of corrosion and replacement of the sulphate by dolomite, indicating that dolomite growth was displacive rather than replacive. The intra-nodular mudstones exhibit similar signs of having been displaced by the growth of the dolomite nodules (Plate 4.11B).

<u>DEPOSITIONAL MODEL.</u>

The precise depositional environment of the Trent Formation mudstones is debat able. One school of thought favours deposition of wind-blown dust on to a damp, essentially flat desert surface (Dumbleton and West, 1966; Wills, 1970), whilst another school argues that the sediment was deposited subaqueously, either in scattered playa lakes (Bosworth, 1913), larger hypersaline lakes (Audley-Charles, 1970b; Tucker, 1978), or an inland sea (Warrington, 1970).

The sediments at Fauld fail to yield any indication of whether they were

Plate 4.11A.   Nodules of fine-grained dolomite (dark) in a

matrix of coarser dolomite rhombs and sulphate

(PPL, x 10).

Plate 4.11B.   Nodular dolomite, trapped intranodular

mudstone and parallel orientated anhydrite

laths (PPL, x 10).
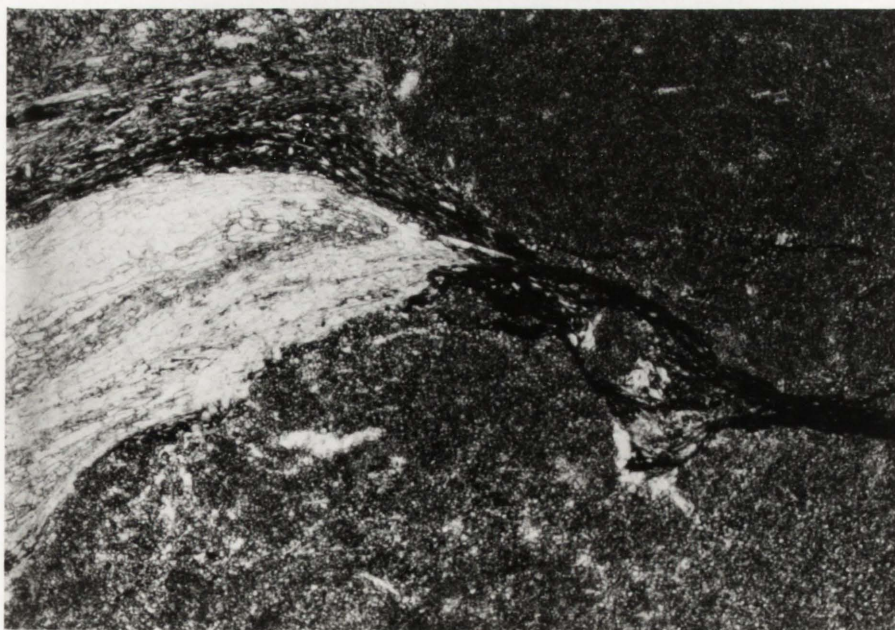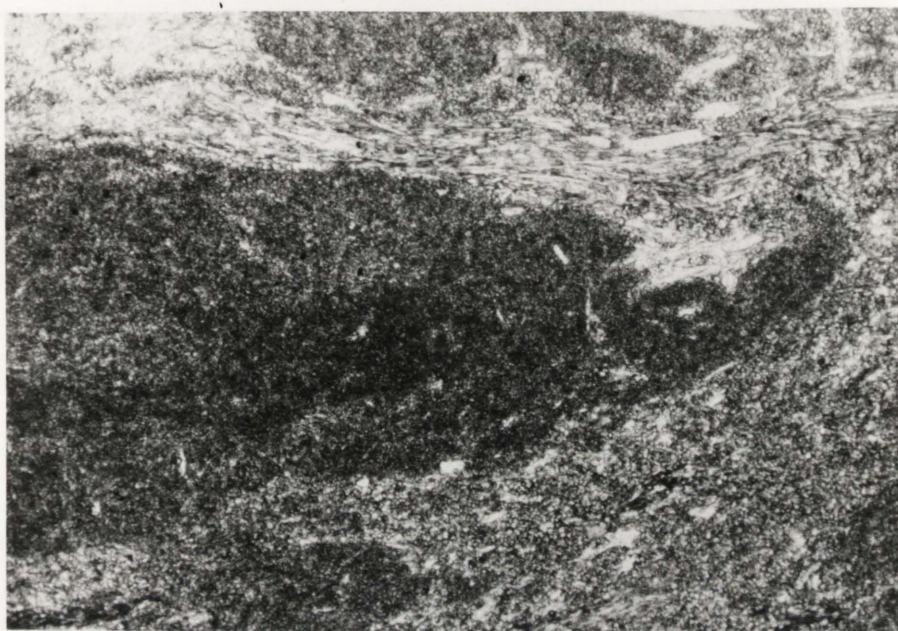
Plate 4.12A.   Edge of a dolomite nodule which is composed of

fine-grained anhedral dolomite. The matrix

comprises coarser grained euhedral dolomite and

minor sulphate (PPL, x 40).

Plate 4.12B.   As above (PPL, x 64). Note contrast in

crystal size and shape from nodule to matrix.
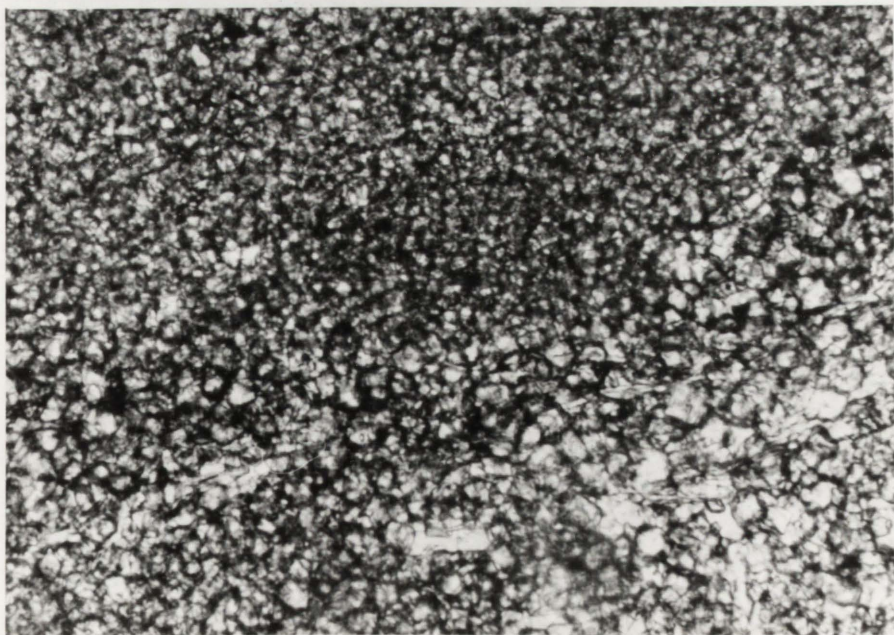
deposited subaerially or subaqueously. Any original bedding structures have very probably been destroyed by both aeolian reworking and the growth of salt within the sediment. The salt was dissolved by the next brine influx.

The nodular sulphate texture at Fauld is directly comparable with that found in modern sabkhas on the Trucial Coast (Shearman, 1966; Butler, 1970) and the Egyptian coast (West, Ali and Hilmy, 1979). Hence it is highly likely that the sulphate at Fauld was precipitated interstitially, close to the sediment surface, from concentrated brines. Stable isotope work indicates that the Fauld area brines were of predominantly marine origin with minor continental input (Chapter 5).

There is evidence at Fauld Mine of erosion of the sediments and the Tutbury Sulphate Bed. The top of the Tutbury Sulphate Bed is marked by a remarkably smooth, distinct, laterally continuous junction between mudstone and massive sulphate (Plate 4.1). This is contrary to the irregular surface which would result from interstitial sulphate growth. Therefore the original irregular top of the Tutbury Sulphate Bed was probably exposed, and thence eroded away resulting in the flat surface now seen.

A small channel 1.2 m. wide and 25 cm. deep cut into the top of the Tutbury Sulphate Bed provides further evidence of erosion (Plate 4.14). The channel was probably cut by surface waters during the period of exposure and erosion of the bed top. The channel is infilled with reddish-brown mudstones containing poorly developed sulphate. A 5 to 6 cm. thick mudstone layer overlies both bed top and channel infill, marking the resumption of sedimentation over the area.

The marked layering of the overlying strata reflects variations in the position of the water table relative to the sediment surface. A steady water table would result in the interstitial precipitation of sulphate across a wide area within a restricted horizon close to the sediment surface. A fall in the water table level would halt sulphate precipitation as the brines are subjected to decreasing evaporation effects with increasing depth from the sediment surface. Hence periods of high sediment input and/or low water

Plate 4.14.   Channel cut into the top of the Tutbury Sulphate

Bed, Fauld Mine. The channel provides evidence

of temporary exposure and erosion of the Tutbury

Sulphate Bed. The channel was infilled by

sediment prior to the top of the Tutbury

Sulphate Bed being covered.

table resulted in the formation of blocky mudstones whilst low sediment input or more importantly, high water table, gave rise to the gypsiferous horizons. The height of the water table would be directly controlled by the volume of brine input. The sediments at Fauld show that the brine input volume was not constant. The control on brine input has been attributed to fault movement (Tucker, 1978).

Dolomite formation associated with sulphate growth has been recorded from the Trucial Coast sabkhas (Kinsman, 1969; McKenzie, 1981). The Trucial Coast dolomite formed during early diagenesis by the interaction of Mg-rich brines with pre-existing calcium carbonate. A similar early diagenetic origin is postulated for the Fauld dolomites. The euhedral nature of the dolomite rhombs indicates that they formed over a period of time with degree of crystallinity improving with age (McKenzie, 1981).

The origin of the nodular dolomite beneath the Tutbury Sulphate Bed is problematical. Petrographic evidence points to a displacive origin for the nodules. As they are unlikely to be primary dolomite there must have been a precursor carbonate mineral. The nodules appear to have displaced the sulphate associated with them, implying that the nodules formed after the sulphate. One possibility is that the nodules formed during a period of slow sedimentation which was coincident with input of low salinity waters. The slow sedimentation rate confined the carbonate precipitation to a thin horizon resulting in the formation of a well defined nodular layer. The pre-existing sulphate was deposited during an earlier phase of higher salinity brine input, whilst dolomitization occurred during a later phase of high salinity brine input.

SULPHATE DIAPIRS.

Sherlock and Hollingworth (1938) and Stevenson and Mitchell (1955) briefly described the occurrence of large sulphate masses lying just above the Tutbury Sulphate Bed in Fauld Mine. The present author has recognized

that the sulphate masses also transect the Tutbury Sulphate Bed and under-lying strata. The morphology of the masses and their relationships to surrounding strata is described, followed by an explanation of their origin.

The sulphate masses occur throughout the mine with no observable trend to their distribution (Fig. 4.3). The separation between masses is variable; in the area mapped (Fig. 4.3) the maximum separation was 25 m. (centre - centre) but masses frequently coalesce to form large irregularly shaped bodies. The boundary between two coalesced masses is often marked by a band of trapped mudstone (Plate 4.15).

MORPHOLOGY.

The masses most commonly have a sub-spheroidal shape with diameters varying between 2 and 12.5 m. whilst the thickness can reach 4 m. In plan the masses appear near circular to sub-ovoid. The top of the masses pass into the mine roof; where the masses have been removed it can be seen that they extend above the roof for 1 to 1.5 m. and have very broad, shallowly domed convex tops (Fig. 4.4). Some masses pass into the mine floor, whilst the bases of others lie either within, or just on top of the Tutbury Sulphate Bed. The variation in mass base position is a function of the line of section through that mass, i.e. a section positioned towards the centre of a mass will result in the mass appearing to have its base within, or below, the Tutbury Sulphate Bed (Fig. 4.4A), whilst a section towards the edge of the mass will result in the mass appearing to lie above the Bed (Fig. 4.4B). Where the masses pass into the mine floor they often resemble a mushroom cap with part of the stalk attached (Fig. 4.4A). The nature of the bases of the sulphate masses below the mine floor was, until recently, problematical. However, in the new drivage beneath the Tutbury Sulphate Bed a single small sulphate mass was temporarily exposed (Plate 4.16). The top of the mass was level with the top of the Tutbury Sulphate Bed and it had a total height of 2.10 m. and a width of 2.4 m. The base of the mass is irregular and passes into the centre of one of the previously discussed lenses containing highly

Fig. 4.3. Distribution of sulphate masses as seen in part of Fauld

Mine roof.
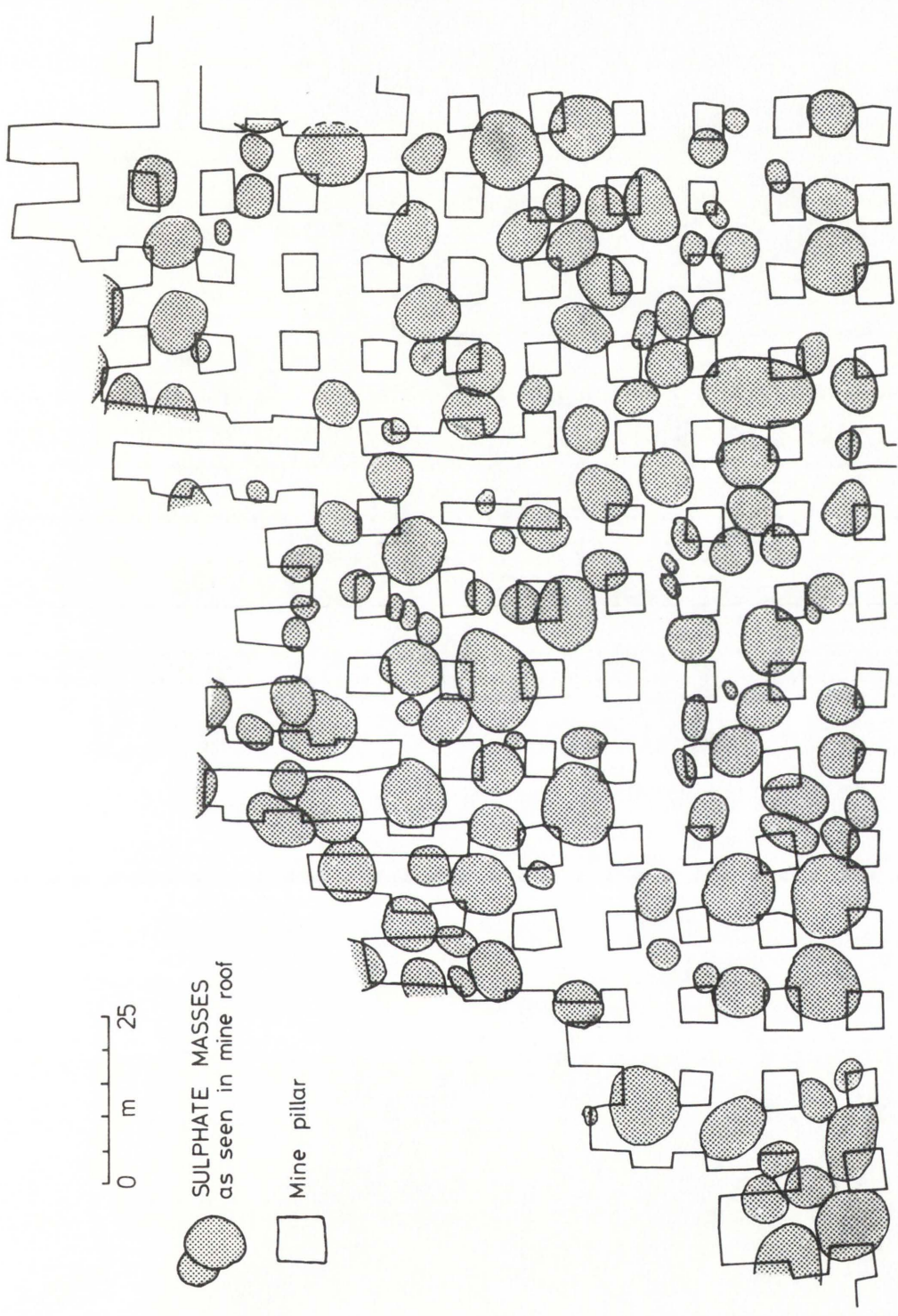
SULPHATE MASSES
as seen in mine roof

Mine pillar

0   m   25

Plate 4.15.   Mudstone trapped between two large sulphate

masses exposed in the mine roof. The mudstone

is some 0.3 m wide.
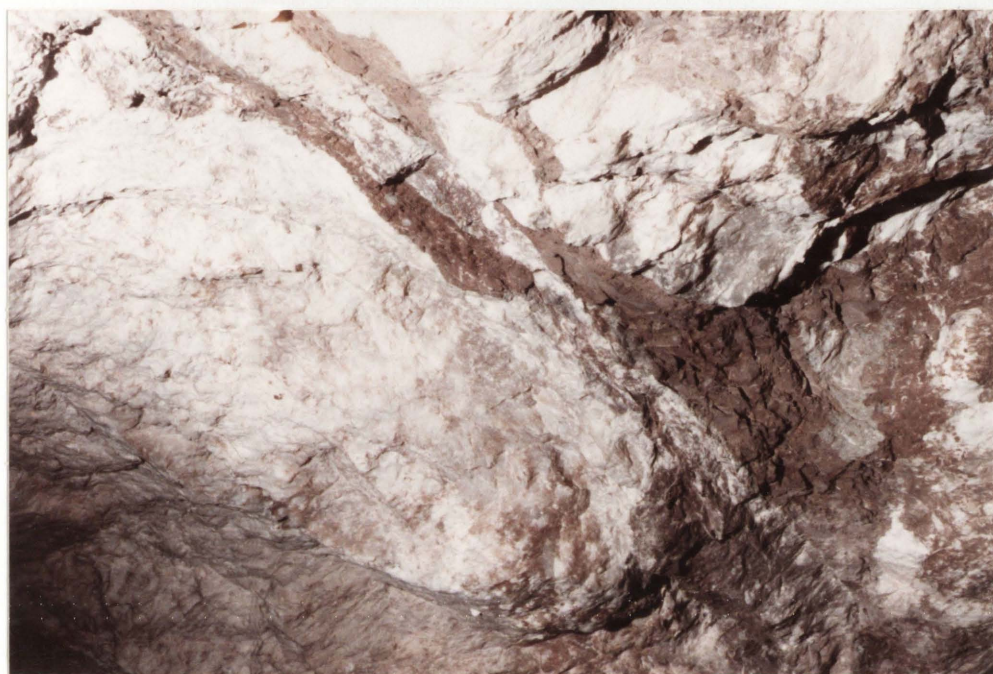
Fig. 4.4a.   Section through centre of displacive sulphate mass.

Structures associated with the sulphate mass are described

in the text.   Scale as on Fig. 4.4b.

Fig. 4.4b.   Section through edge of displacive sulphate mass.   This

section gives the appearance of the sulphate mass resting

just above the Tutbury Sulphate Bed, but as in Fig. 4.4a

displacement of thin nodular sulphate horizons occurs

against the margins of the mass.

A)

ROOF

TUTBURY
SULPHATE
BED

FISSURE

ZONE OF VERTICALLY
ELONGATED SULPHATE
NODULES

FLOOR

B)

ROOF

TUTBURY
SULPHATE
BED
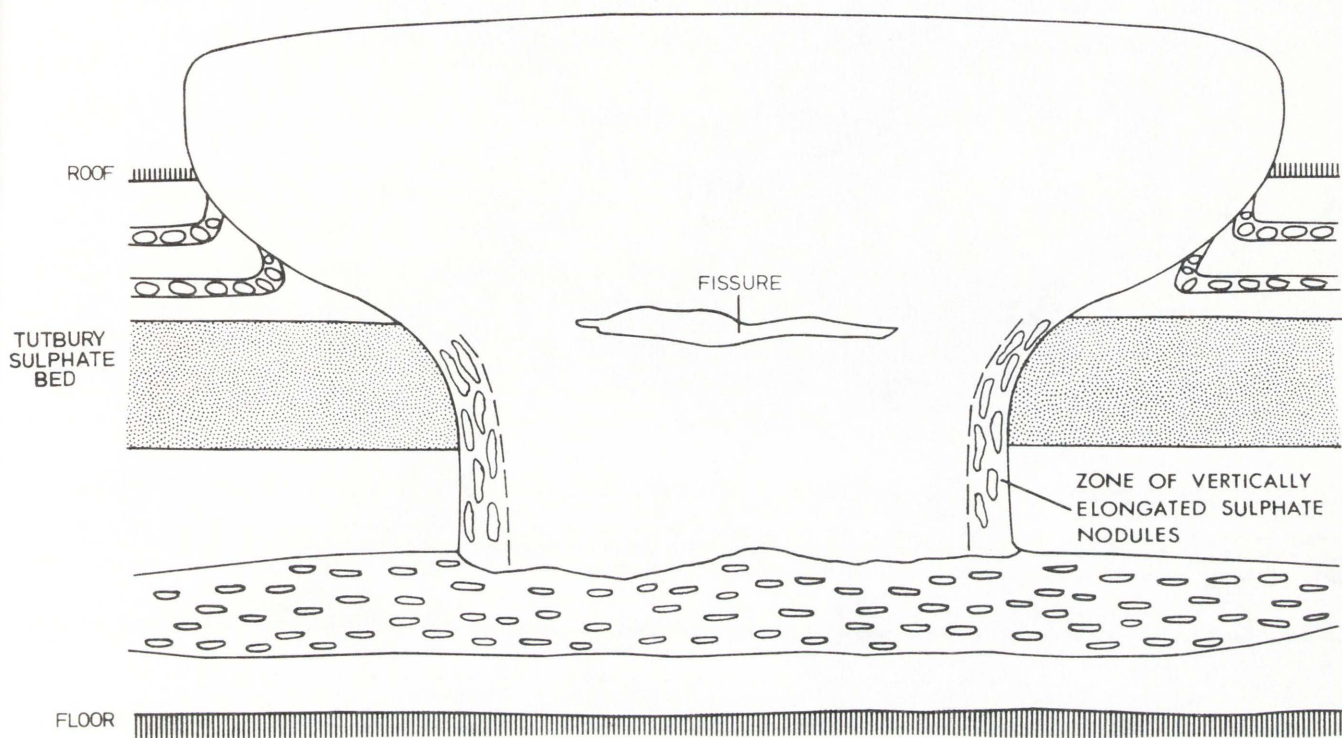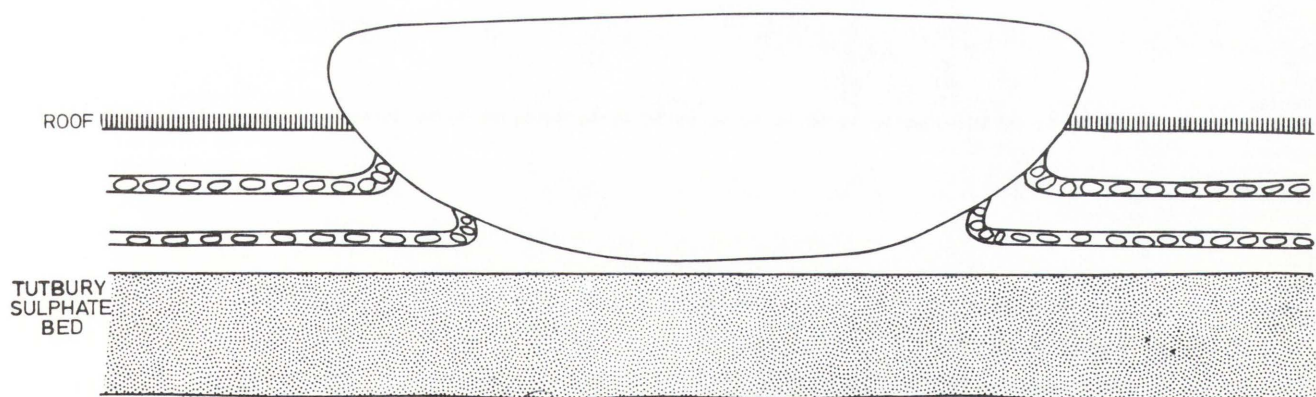
0   1   2 m
Vertical scale equals horizontal scal

PLAN OF SULPHATE
MASS

SECTION IN FIGURE 4a

SECTION IN FIGURE 4b
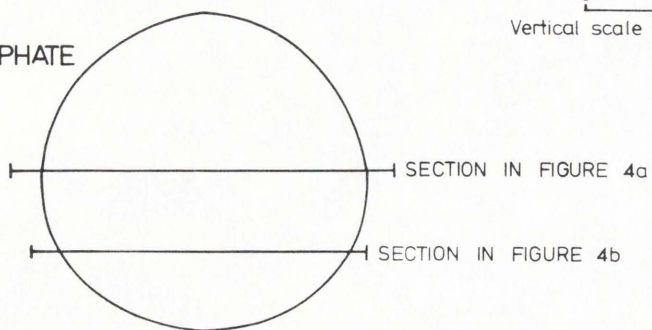
Plate 4.16A.   Small sulphate mass in drivage beneath Tutbury
                Sulphate Bed. Mass terminates at top of
                Tutbury Sulphate Bed, just below the picture
                top. Hammer on lower left margin of mass.

Plate 4.16B.   Closer view of left side of sulphate mass
                above. The deep blue colouration is typical of
                anhydrite. The nodular sulphates on the left
                of the mass are markedly displaced and in places
                are overturned. The mass is overlain by
                mudstones.

elongated anhydrite (Plate 4.10). The significance of this will be discussed later.

INTERNAL STRUCTURE.

The centres of the masses are composed of very pure sulphate and in general contain very little trapped mudstone. It is from the centre of hydrated masses that alabaster (equigranular gypsum) was extensively worked in the past. Minor extraction of alabaster blocks still occurs (Plate 4.17). The margins of the masses contain variable amounts of mudstone. The mudstone occurs as irregular patches or thin bands or stringers orientated parallel with, or sub-parallel to the margin of the mass. This is well illustrated by sections through the edge of the "mushroom stalk" which show ovoid sulphate nodules elongated in a vertical direction parallel to the side of the stalk (Fig. 4.4A, Plate 4.18). Many of the nodules have coalesced to form long sulphate "snakes", orientated vertically and separated from each other by a few centimetres of mudstone. The vertical nodule orientation is in marked contrast to the horizontal orientation seen in the main sulphate horizon. The textures seen around the stalk margins are interpreted as resulting from extensive sulphate movement or flowage in a vertical direction.

The uppermost 50 cm. of hydrated masses frequently have an exfoliated texture with 25 to 30 cm. thick foliations running parallel with the upper surface of the mass. This texture has been noted in sulphate nodules in borehole cores from the Trucial Coast sabkhas and it is thought to be a rehydration effect (J. Rouse, per. comm.).

In areas of the mine where anhydrite is prevalent the centres of the sulphate masses are frequently fissured. The fissures are generally located in the lower half of the "mushroom cap". They run horizontally for at least 2 m. and can be up to 20 cm. high with irregular margins. Fissures have not been found in gypsum masses.

As previously noted, the cap sulphate generally contains only minor amounts of mudstone, as thin stringers. Occasionally sinuous zones containing

Plate 4.17A.   Alabaster blocks in the process of being

extracted from a large hydrated sulphate mass.

Note the "flow texture" on the right side of

right hand block. Scale - shovel handle at

bottom of photograph.

Plate 4.17B.   Extracted alabaster block.

Plate 4.18.   Vertically orientated sulphate-mudstone partings

in the margins of a sulphate mass "stalk"

(Fig.4.4A). This texture is thought to result

from vertical sulphate movement. The sketch

outlines the major features in the photograph.

elongated sulphate and mudstone

sulphate

floor
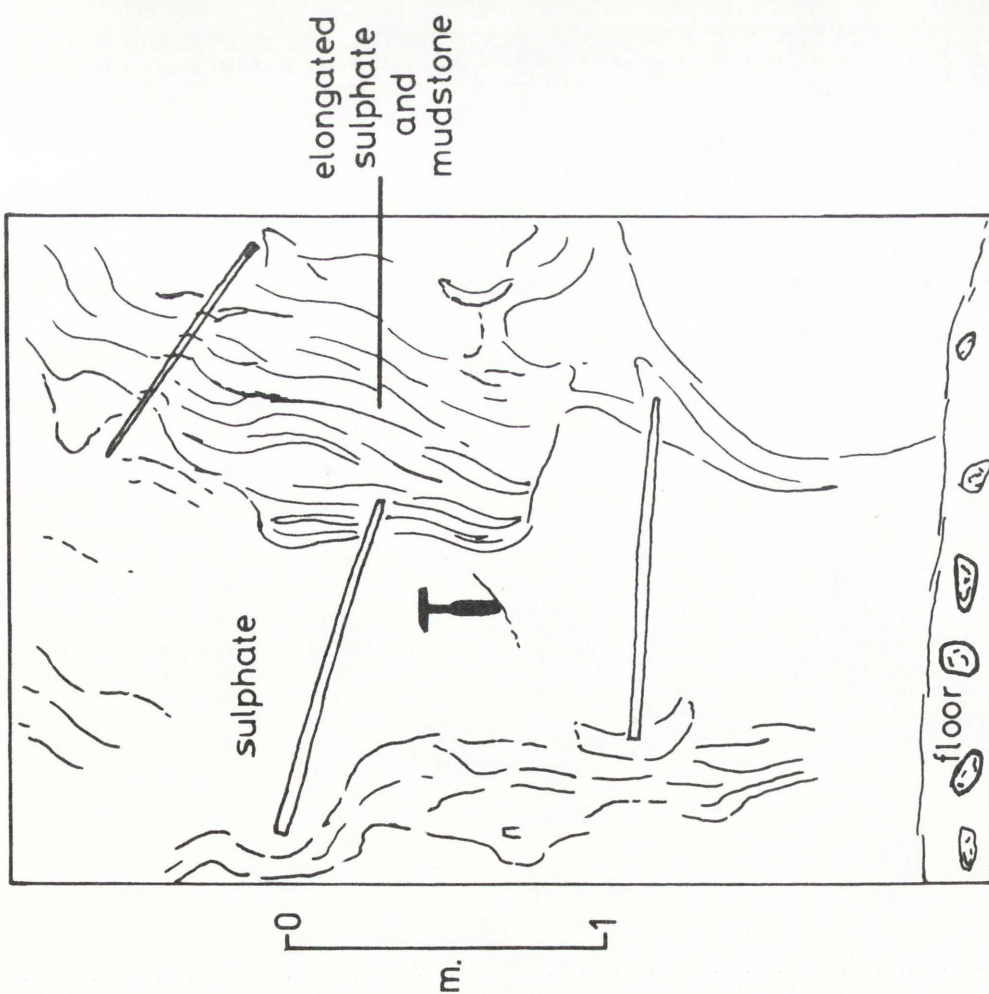
as much as 50% mudstone are seen cutting the sulphate masses. The maximum
width of a zone is 50 cm. In masses where these zones cut, or pass close to
fissures, the fissures contain halite cubes and fine acicular gypsum crystals
(Plate 4.19). Although the fissures have often been disturbed by blasting
the crystals appear to line the base of the fissures. No evidence has been
found for either halite or gypsum growing on a fissure roof. Gypsum crystals
up to 2 cm. in length and halite cubes up to 6 cm.[3] were found. Halite
cubes up to 10 cm.[3] are reported as being discovered by miners.

The halite cubes show bevelled edges but no indications of hopper type
faces. The cubes commonly coalesce and in one case form a cube chain
(Plate 4.20A). The cubes in the chain have perfect faces on each side
indicating that crystal growth was not initially impeded by sediment or
sulphate. Such growth could occur at the air-brine interface with the halite
forming floating rafts (Arthurton, 1973). Eventually the rafts would sink to
the floor of the void where growth continued on the crystal faces surrounded
by the brine. Many larger cubes exhibit one irregular surface lacking in
crystal form (Plate 4.20B). These are interpreted as being the surface which
rested on the void floor. If the cubes sunk in such a way so as to rest with
some lower faces away from the void floor, then crystal growth could continue
on these faces as well.

The origin of the fissures and their mineral content will be discussed
later.

SURROUNDING STRATA.

Strata closely adjacent to the upper portion of a sulphate mass,
i.e. strata above the Tutbury Sulphate Bed, frequently show partial or complete
recumbent folding (Plate 4.21). The overturning can be easily picked out by
following an individual horizon in towards a sulphate mass. The highest
visible horizon affected is a layer which shows a gentle doming over the top
of the sulphate mass.

Sulphate-rich mudstones form more competent horizons than sulphate-poor

Plate 4.19A.   Interpenetrative halite cubes from a void in

the centre of an anhydrite displacive mass.

Scale bar in cm.

4.19B.   Fine acicular gypsum crystals covering the irregular

surface of some nodular gypsum. From void in centre

of an anhydrite displacive mass. Scale bar in cm.
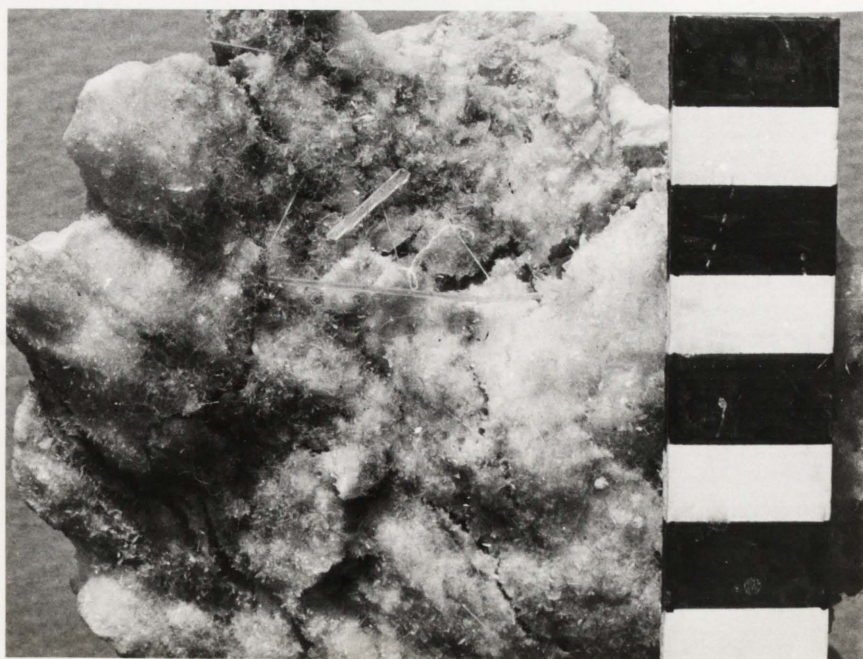
Plate 4.20A.   Halite cubes, the smallest of which form an

interpenetrating chain. Void in centre of

anhydrite displacive mass. Scale bar in cm.

Plate 4.20B.   As above; note the irregular surface on one

side of the cubes. This is the surface which

rested on the cavity floor during the formation

of the cubes.

Plate 4.21A.   Overturning of strata overlying the Tutbury
                Sulphate Bed adjacent to a sulphate mass. The
                Tutbury Sulphate Bed is the white band at the
                base of the picture. The blue colouration of
                the mass is indicative of anhydrite.

Plate 4.21B.   Strata overlying the Tutbury Sulphate folded
                vertically against the margin of a sulphate
                mass. Tutbury Sulphate Bed top lies just
                above "STOP".
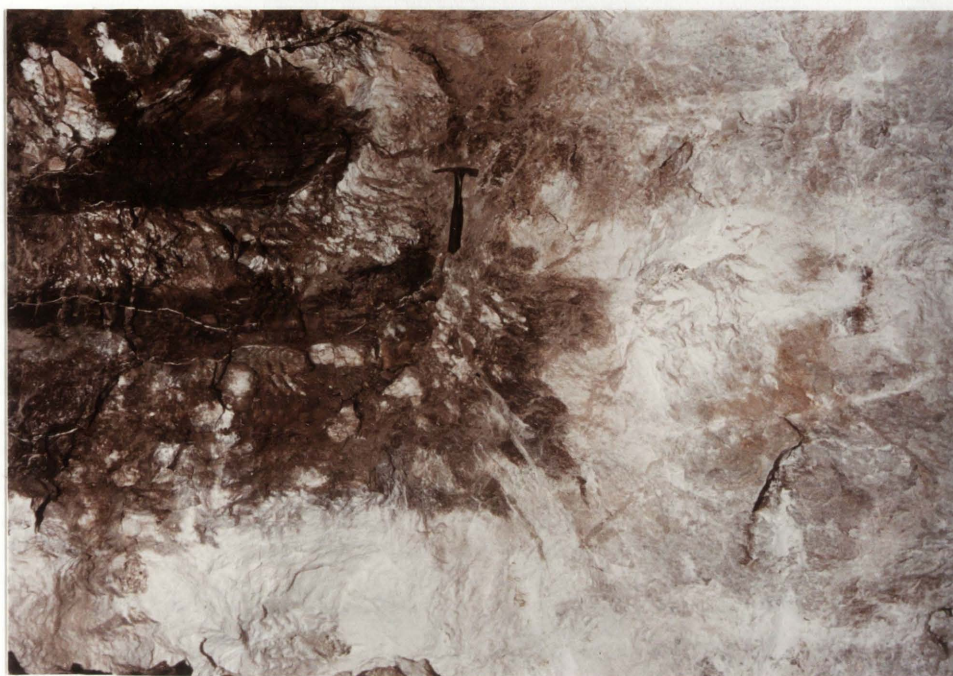
mudstones. This is well illustrated in Plate 4.22 where a sulphate-poor mudstone immediately above the Tutbury Sulphate Bed shows a pronounced thickening in the core of the fold. On the upturned limb the mudstone is rapidly pinched out between a thin sulphate horizon and the sulphate mass. More competent horizons continue upwards for varying distances until truncated by the sulphate mass (Plate 4.21B).

Sulphate nodules in the fold core are frequently deformed. Their long axes show a fan type arrangement around the fold core, a feature due to compressive folding and deformation (Plate 4.22B). Nodules on either limb are undeformed with their long axes parallel to bedding.

The strata adjacent to the sulphate mass exposed below the Tutbury Sulphate Bed showed similar overturning with deformed nodules across the fold core (Plates 4.23, 4.16B).

GENESIS OF THE SULPHATE MASSES.

The sulphate masses may be either primary depositional features or have a secondary displacive origin.

A primary sulphate mass would act as a relatively incompactible body whereas newly deposited mudstones would be readily compactible. On burial, the differential compaction would result in folding of the mudstones over and around the sulphate masses. However, a "compactional model" will not satisfactorily account for the following observed features:-

i) recumbent folding. Differential compaction will only give rise to open folds and variations in bed thickness across the sulphate mass (Fig. 4.5);

ii) truncation of strata;

iii) vertical orientation of sulphate nodules around the margins of the "stalk".

Both (ii) and (iii) imply that some degree of sulphate movement or flowage has occurred, indicating that the masses are more likely to have a secondary displacive origin.

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Plate 4.22A.   Sulphate horizon and mudstones folded vertically

against a sulphate mass (on extreme right).

Note thickening of strata in fold core.

Plate 4.22B.   Close-up of fold core above. Note the

thickening of mudstones in the fold core and

the fan type arrangement of sulphate nodules

around the fold axis.
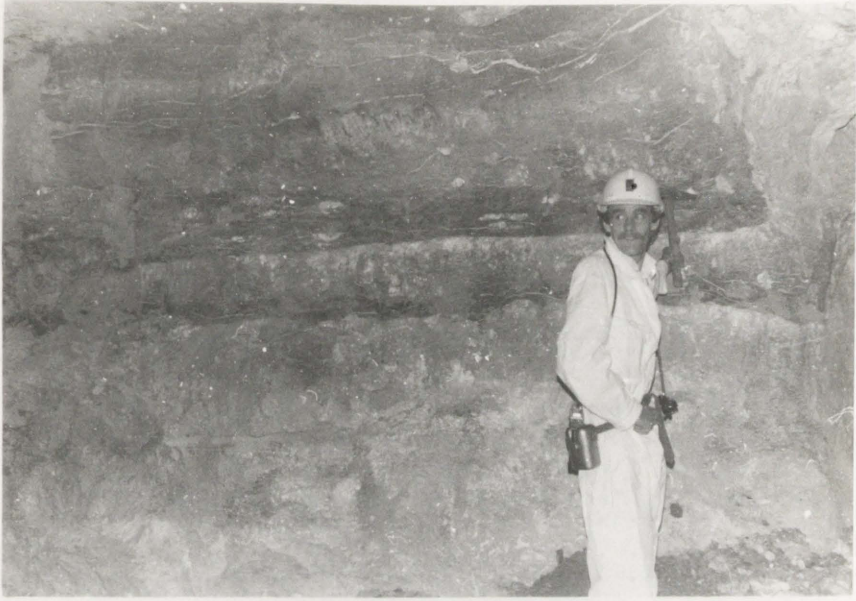
Plate 4.23.   Overturning and re-orientation of anhydrite

                nodules and overlying sulphate rich mudstones.

                Note fan-type arrangement of nodules around fold

                core. Sulphate mass, drivage beneath Tutbury

                Sulphate Bed (as on Plate 16).

Fig. 4.5.   Illustration of compaction effects over an incompactable

mass. The beds have been compacted to half their original

thickness, resulting in open folds in overlying sediments

and sediments adjacent to the mass.

1

2

COMPACTION

$\frac{1}{2}a$

$\frac{1}{2}b$

$\frac{1}{2}c$
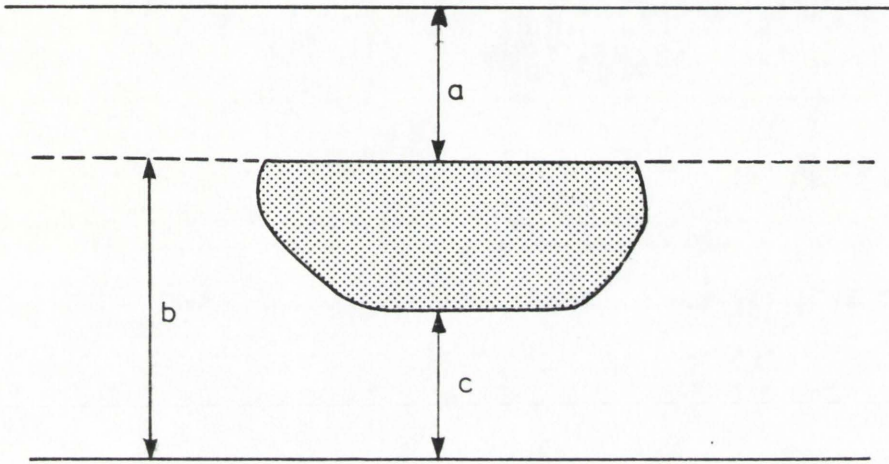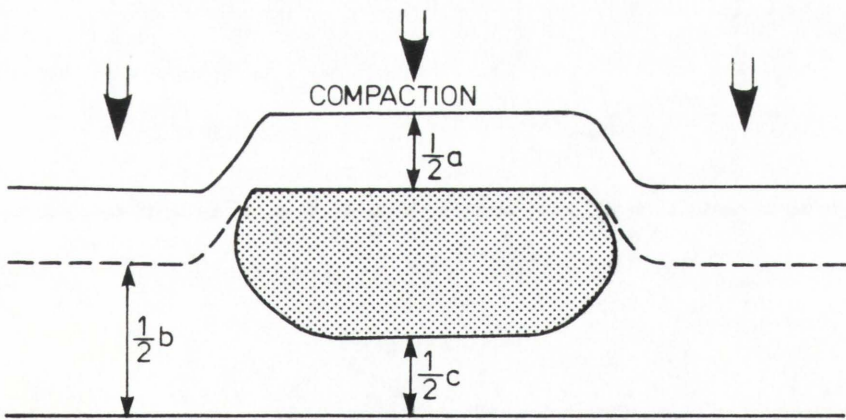
In the case of halite, the formation of secondary displacive masses or diapirs commonly occurs when thick halite beds are deeply buried (Murray,1968; Kent,1970; Sanneman, 1968). Halite diapirs are often of a very large size, e.g. Gulf Coast diapirs reach heights of 6000 m. and cover an area of 40 sq. km. (Murray, 1968).

Sulphate diapiric structures are very much rarer. Large scale sulphate piercement structures in the Canadian Artic Islands have been described by Hoen (1961, 1964) and Gould and de Mille (1964). Many of these have an areal extent of 2 sq. km. but some cover 25 sq. km. Gould and de Mille (1964) postulated that some of the sulphate diapirs were due to the movement of underlying and as yet unproved halite, whilst the rest resulted from movement of closely associated faults.

The sulphate masses being studied are very much smaller than this, but their morphology and internal structure can be favourably compared to that of the larger halite diapirs.

The mushroom like shape of the sulphate masses is very similar to that of halite diapirs, which frequently have a relatively smooth convex domed top and a poorly defined irregular base (O'Brien, 1968). Both features have been described from the sulphate masses.

Emplacement of a halite diapir results in adjacent strata being markedly displaced, often to a vertical dip. Complete overturning or recumbent folding of the strata associated with salt structures is known in the Gulf of Mexico Basin (Murray, 1968). The folding is thought to be due to the drag of the halite against adjacent strata (Murray, 1968). Truncation of strata is another feature frequently associated with diapiric structures (Gussow, 1968).

The internal structure of halite diapirs is complex, consisting of large scale rheomorphic folds which are characteristic of strong plastic flow (Hill, 1972; Kupfer, 1968). Rheomorphic folds have not been observed in the Fauld Mine sulphate masses. However, the previously described sinuous "sulphate snakes" around the stalk margins very possibly formed by plastic

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flow of large volumes of sulphate with mudstone being incorporated from the surrounding strata.

The author believes that the folding and truncation of the strata adjacent to a sulphate mass in Fauld Mine is a direct result of the emplacement of that mass. The similarities between the morphology and structure of halite diapirs and the features associated with the Fauld Mine sulphate masses lead the author to the conclusion that the latter have a diapiric origin.

THE SULPHATE SOURCE BED.

Halite diapirs may be connected with a source bed or "mother" bed at the base, or may form disconnected masses above the "mother" bed. The "mother" bed may have thinned as on the Gulf of Mexico Coast, or it may have been squeezed completely into the diapirs as in the Paradox Basin (O'Brien, 1968). By analogy with the halite structures, a sulphate "mother" bed should lie at or below the base of the diapirs. Until August, 1981, the diapir bases were not exposed in the mine and borehole cores failed to yield any conclusive evidence. However, the new drivage (1981) under the Tutbury Sulphate Bed exposed a diapir base which, as previously described, passed into a lens of highly elongated anhydrite nodules indicative of horizontal sulphate movement. The nodule distribution at the diapir base, combined with the evidence of sulphate movement, implies very strongly that we are seeing the sulphate "mother" layer. The sulphate lenses are interpreted as relics of the original sulphate layer. Lateral migration of its sulphate resulted in gradual thinning of the layer until only the lenses were left (Fig. 4.6).

The mudstones surrounding the lens also provide evidence of the previous existence of a sulphate horizon. Petrographic studies on the mudstones revealed the presence of fragmented anhydrite laths disseminated evenly throughout the dolomitic mudstone matrix (Plate 4.24). The laths are matrix supported and rarely touch. The maximum observed lath length was 0.25 mm.

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Fig. 4.6.   Origin of the Fauld Mine sulphate diapirs and elongated

anhydrite lenses. (1) Sequence prior to any sulphate move-

ment. (2) Lower sulphate source layer starts to migrate

laterally and form "swells". (3) Swells enlarge by

continued sulphate migration and form tall stalk-type

structures. Sulphate source layer pinches out.

(4) Sulphate source layer now pinched out into individual

lenses each containing elongated sulphate nodules. The

displacive sulphate masses now have the typical mushroom

type structure seen in the mine today.

1

TUTBURY SULPHATE BED

SULPHATE SOURCE LAYER,
ORIGINAL DEPOSITIONAL THICKNESS

NODULAR DOLOMITE HORIZON

1m

2

3

4

DISPLACIVE SULPHATE DIAPIR

DISPLACIVE SULPHATE DIAPIR
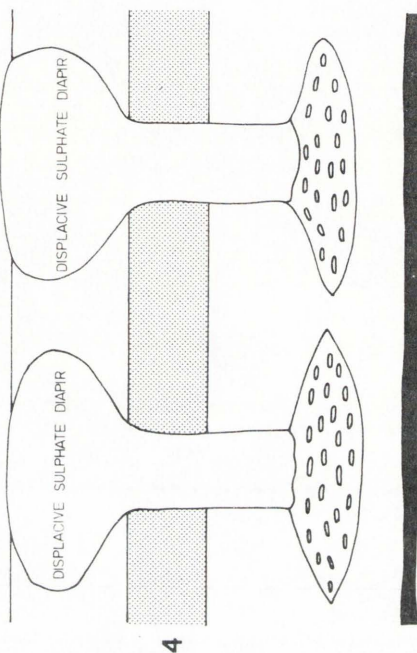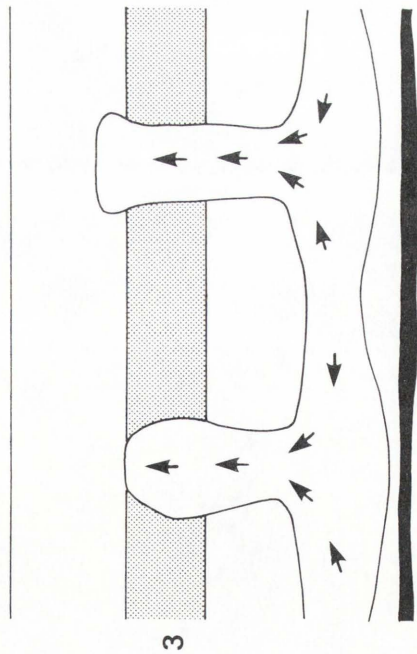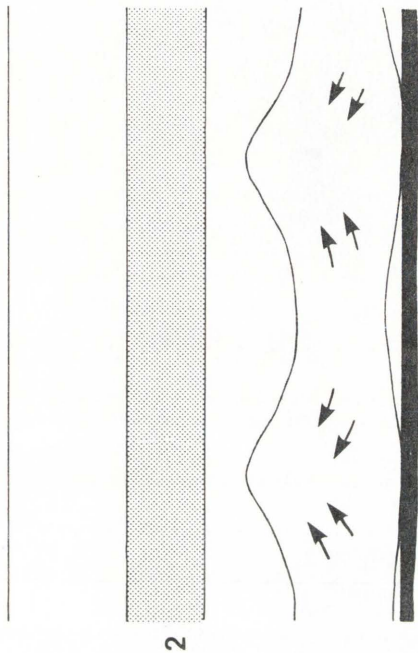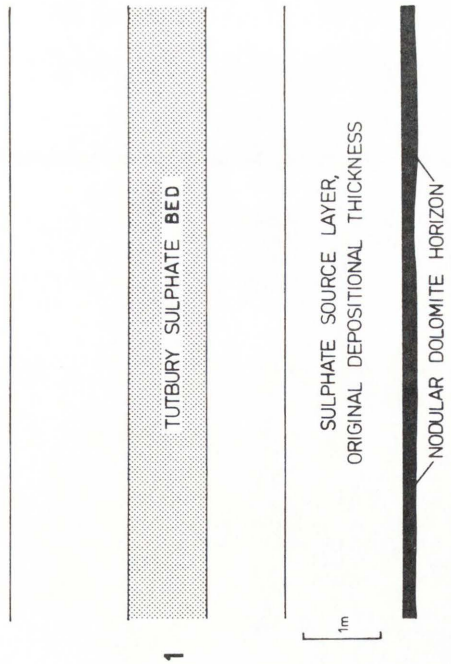
Plate 4.24A.　Sub-horizontal elongate anhydrite laths and

　　　　　　　　fragments in dolomitized mudstone from between

　　　　　　　　lenses of elongated anhydrite nodules (Plate 9).

　　　　　　　　Exposed in drivage beneath Tutbury Sulphate Bed.

　　　　　　　　(PPL, x 12.8).

Plate 4.24B.　As above (PPL, x 64).　Subhorizontal anhydrite

　　　　　　　　laths and fragmented laths can be seen.　Texture

　　　　　　　　results from lateral movement, brecciation, and

　　　　　　　　finally migration of sulphate to form diapirs;

　　　　　　　　the anhydrite laths are residual sulphate

　　　　　　　　(see text).
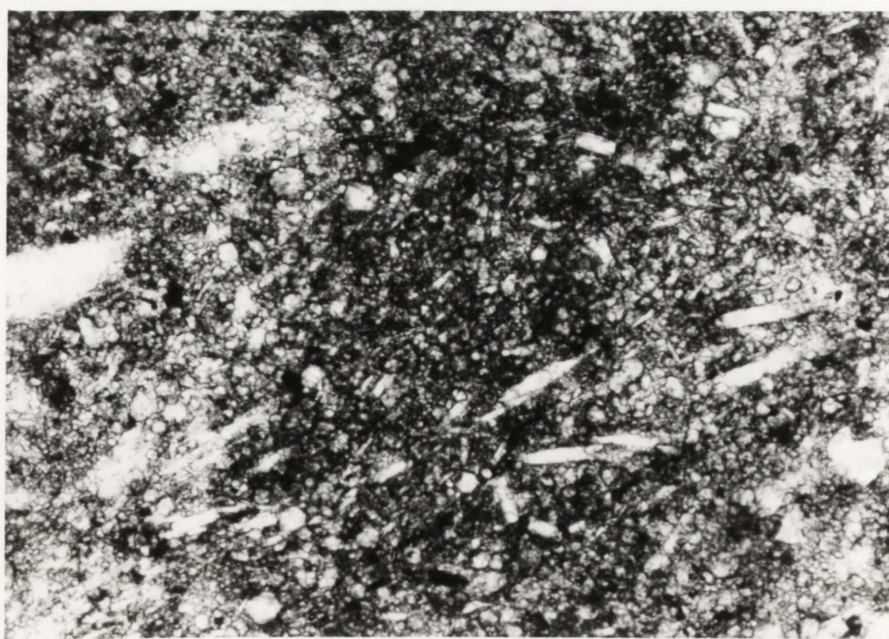
with the average length lying between 0.05 and 0.02 mm. Many of the laths are orientated parallel or sub-parallel to the bedding.

The original depositional fabric of the Trent Formation sulphate is as discrete nodules of various shapes and sizes. The disseminated and fragmentary nature of the sulphate laths indicates that they have suffered brecciation; i.e. it is a tectonic fabric rather than a depositional fabric. Mass lateral movement in the sulphate "mother" layer would initially result in the brecciation and fragmentation of much of the sulphate, which on migration would give rise to the observed fabrics of small disseminated residual sulphate laths in a mudstone matrix.

THE ORIGINAL THICKNESS OF THE "MOTHER" LAYER SULPHATE.

The following calculation applies to the area of the mine which was mapped (Fig. 4.3).

Diapiric sulphate forms 40% of roof area. Total roof area is 15,000 m.^{2} Diapir area therefore 6000 m.^{2}

Average diapir height is 4 m.

Assumption 1. Diapirs are cylinders covering 6000 m.^{2}

gives total visible diapir volume of $\underline{24,000\ m.^3}$

Diapir stalks are 1 m. high.

Assumption 2. Stalks cover $\frac{1}{4}$ of total diapir area.

Stalks cover 1500 m.^{2} and have a volume of $\underline{1500\ m.^3}$

Assumption 3. Elongated anhydrite remaining in "mother" layer covers same area as diapirs and it forms $\frac{1}{2}$ of remaining volume in "mother" layer.

Elongated anhydrite covers 6000 m.^{2}

Total present day "mother" layer thickness is 1 m. Therefore anhydrite thickness is 50 cm.

Total volume of elongated anhydrite is 3000 m^3

Total volume of sulphate required is 24,000 m.^{3} + 1500 m.^{3} + 3000 m.^{3} = $\underline{28,500\ m.^3}$

If "mother" layer sulphate covered the whole of the 15,000 m.^{2} area then its thickness would have been:

28,500/15000 m. = <u>1.9 m.</u>

Assuming the diapirs are cylinders covering the total area, then the "mother" layer sulphate is estimated to have had a thickness of 1.9 m.

A more realistic assumption is to say that the diapirs can be represented by cylinders covering 2/3 of the total diapir area (Fig. 4.7).

The diapir cylinder area is 2/3 of 6000 m.^{2} = 4000 m.^{2}

Total height of 4 m.

Volume of 4000 m^2 x 4 m. = <u>16,000 m.^{3}</u>

Utilizing same stalk and remnant "mother" layer volumes as before:

Total sulphate volume is 16000+1500+3000 m.^{3} = <u>20,500 m.^{3}</u>

Therefore original "mother" layer thickness would be:-

20,500 / 15,000 = <u>1.37 m.</u>

It is estimated that the "mother" sulphate layer had an original average thickness in the order of 1.4 m. The scattered nature of the diapirs (Fig. 4.3) perhaps indicates that the "mother" layer sulphate was not of a uniform thickness throughout the area. Zones with a higher proportion of diapiric material may reflect areas of thickening in the "mother" layer sulphate.

CAN SULPHATE FORM DIAPIRS?

For a material to form diapirs it must fulfil two major conditions. The most important of these is that the diapiric material has the ability to flow (O'Brien, 1968; Tanner and Williams, 1968; Gussow, 1968; Ode, 1968). The second is that the diapiric material is of equal or preferably lower density than the surrounding rock (Nettleton, 1934).

Fig. 4.7. Representation of diapir as a cylinder covering two-thirds of total measured diapir area. Radius decreases by one-fifth. Lower part of stalk represented by separate cylinder 1 m high.

Fig. 4.8. Assembled creep curves of alabaster. Seven specimens are in distilled water, at the same temperature, but subjected to different loads. The lower line is for alabaster in dry conditions at 205 kg/cm^2. Note the lack of deformation and compare to the deformation recorded at the same load under wet conditions (after Griggs, 1940).

DISPLACIVE SULPHATE MASS
(Cylinder 1)

4m

1m

(Cylinder 2)

RUPTURE

250kg/cm²

RUPTURE

205kg/cm²

300kg/cm²

181kg/cm²

150kg/cm²

125kg/cm

100kg/cm²

CREEP OF ALABASTER
IN H₂O - 24°C

DRY 205 kg/cm²

DEFORMATION %

TIME - DAYS
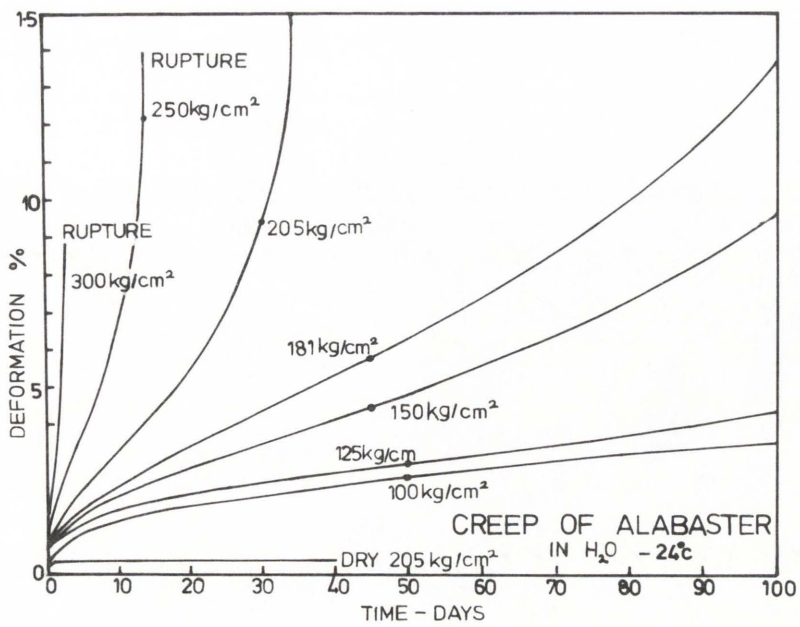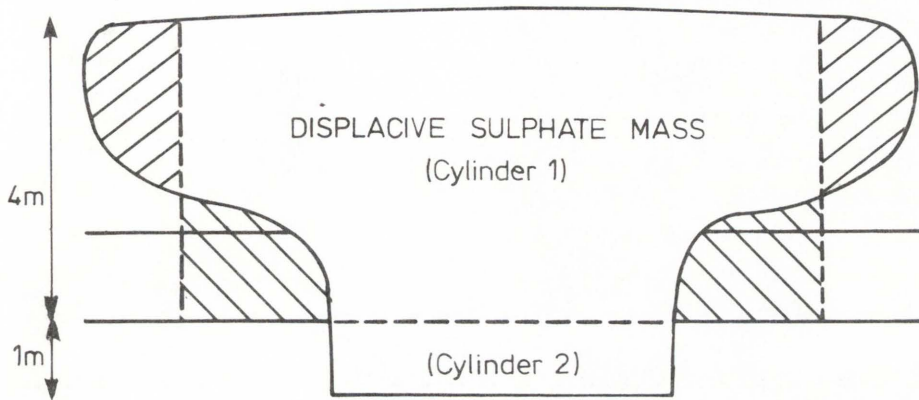
ABILITY OF DIAPIRIC MATERIAL TO FLOW.

The major characteristic of diapiric material is plasticity, that is, its ability to flow. Plasticity is better expressed as "equivalent viscosity". Equivalent viscosity is used for materials which do not exhibit Newtonian viscosity, e.g. metals and rocks. Equivalent viscosity indicates the viscosity of a Newtonian fluid which would exhibit the same strain rate under identical loading as the material used (Ode, 1968). Equivalent viscosity is a function of temperature, confining pressure, differential stress and factors such as the presence of water (Griggs, 1940; Ode, 1968).

All published work has used gypsum in experiments to study changes in equivalent viscosity with variation in physical conditions. In general terms the following occur: -

Temperature. As temperature increases the equivalent viscosity decreases, i.e. the sulphate will flow more easily at higher temperatures.

Confining pressure. Is equivalent to lithostatic pressure in which all three principle stresses are equal (Hobbs, Means and Williams, 1976). Lithostatic pressure increases with increasing depth of burial. At room temperature an increase in confining pressure has little effect on the equivalent viscosity. However, at higher temperatures (100 to 200°C) an increase in confining pressure appreciably raises the equivalent viscosity of dry rock (Ode, 1968).

Axial stress. Is equivalent to geostatic pressure and is produced by variations in overburden thickness; hence the vertical stress component is not equal to the two horizontal stress components. Increases in axial stress produce a decrease in equivalent viscosity.

Water. The presence of water with gypsum results in a saturated solution of gypsum being formed. This in turn leads to a remarkable reduction in the equivalent viscosity of gypsum (Griggs, 1940). Figure 4.8 (after Griggs,1940) shows the effect of the presence of water on the creep rate of alabaster. Griggs (1940) concluded that "In alabaster, where an inconsequential amount of deformation may be produced when the specimen is dry, the same specimen

wet will flow at a relatively rapid rate." Griggs (1940) also concluded

that flowage occurred by a process of solution and recrystallization and

that as the stress is increased the rate of flow increases exponentially

(Fig. 4.8).

The presence of water also has a marked effect on gypsum's equivalent

viscosity when the confining pressure is increased. As the confining

pressure increases, the equivalent viscosity drops (Griggs, 1940), an effect

opposite to that seen in dry halite and gypsum.

To summarize; to aid gypsum flowage the equivalent viscosity of gypsum

has to be as low as possible. This can be achieved by adding a saturated

solution of gypsum and raising the temperature, confining pressure

(lithostatic pressure) and axial stress (geostatic pressure). On the basis

of the published experiments it seems that gypsum has the ability under these

circumstances to flow as easily as halite.

DENSITY OF MATERIAL.

Typical diapiric materials are sylvite (density 1.98 g/cm.^{3}) and halite

(density 2.16 g/cm.^{3}). Gypsum and anhydrite have densities of 2.31 g/cm.^{3}

and 2.90 g/cm.^{3} respectively. The density of sediments varies with lithology

and degree of compaction, which is a function of depth of burial.

Nettleton (1934) and Dickinson (1953) produced curves of average shale

density against depth for Gulf Coast sediments (Fig. 4.9). Newly deposited

Gulf Coast sediments have an average density of 1.6 g/cm.^{3} but this rapidly

increases with burial as compaction expels pore fluids. The shale depth –

density curve gradually approaches the approximate shale mineral density

which is theoretically the point at which shale becomes slate. An average

density of 2.32 g/cm.^{3} was obtained for three mudstone samples from Fauld

Mine. In view of the fact that the depositional density of the mudstone is

unknown, as is its behaviour on compaction, it is unwise to attempt an

estimate of the maximum depth of burial of these rocks.

Unlike other sedimentary material, evaporites do not show increases in

Fig. 4.9.   Buoyancy of salt in relation to height of intrusive column,

Gulf Coast. Before diapirism, buoyancy is zero; it

increases gradually with height of the dome above base.

Also shown is an average shale density - depth curve and

the densities of gypsum and salt (after Nettleton, 1934;

Dickinson, 1953 and Ode, 1968).

HEIGHT OF TOP OF DOME ABOVE BASE (METRES)

7220 6000 4880 3660 2440

Differential Hydrostatic Pressure (Bars) at Base of Salt Dome (8440 metres)

175
140
106
70
36

SHALE DENSITY (GM/CC)

1·6

2·0

2·4

2·8

SALT SEDIMENT DENSITY CONTRAST

+0·6
+0·5
+0·4
+0·3
+0·2
+0·1
0·0
-0·1
-0·2
-0·3

Avg. Salt Density 2.20

Avg. Shale Density Depth Relationship

Avg. Gypsum Density 2·32

Approx Shale-Mineral Density

DEPTH (METRES)

1220 2440 3660 4880 6000 7220

density until buried to great depths (Ode, 1968).

The average density of shale exceeds that of halite when the shale is buried below 914 m. (Fig. 4.9). Similarly average shale density exceeds that of gypsum at 2133 m. These are the depths where the evaporite – sediment density contrast becomes negative, i.e. the evaporite becomes more buoyant than the overlying sediment. The effectiveness of the density contrast is dependant on the height of the diapir. Before intrusion begins no part of the halite (or gypsum) is more buoyant than the rest. As the height of the diapir increases the effect of buoyancy gradually increases, eventually becoming extremely powerful (Nettleton, 1934). The small height of the Fauld diapirs means that any buoyancy-aiding effect between gypsum and the sediment was negligible.

The depths at which shale becomes denser than halite and gypsum have been noted. Shale will be metamorphosed to slate long before its density equals that of anhydrite (2.90 g/cm.^{3}). Consequently anhydrite will not readily act as a diapiric material in non-orogenic areas as it is always denser than the surrounding sediment. It may form piercement structures in orogenic zones (Gould and de Mille, 1968). Fauld Mine does not lie in an orogenic zone and the mudstones have a density well below that of anhydrite. Therefore it is thought unlikely that the Fauld diapirs were formed by the movement of anhydrite. Consequently one must conclude that the diapirs were formed by the movement of gypsum. The presence of gypsum-saturated waters would greatly aid gypsum flowage.

Observations made in the mine enable one to assess independently the hydration state of the sulphate when it flowed. The presence of cavities within anhydrite diapirs in Fauld Mine has already been described. It is important to note that all the gypsum in Fauld Mine has been hydrated from anhydrite. There is no evidence of primary gypsum in the mine.

A volume increase may occur when anhydrite is hydrated (Pettijohn,1957). The sulphate will expand in the direction of least resistance which will initially be into the voids, resulting in their infilling and absence in

the gypsum diapirs.

FORMATION OF THE VOIDS.

As already stated, gypsum occupies a greater volume per unit weight than anhydrite. Therefore if the diapirs were originally composed of gypsum, dehydration would cause a volume decrease and hence voids would be formed in the anhydrite masses. The fact that the voids are seen implies that the diapirs are close to their original size and that the volume contraction has been accommodated by internal shrinkage rather than by contraction of the outer surface of the diapir. Therefore the voids provide evidence that the sulphate diapirs were emplaced as gypsum. With increasing depth of burial the gypsum dehydrated to anhydrite which is now undergoing hydration close to the land surface.

ORIGIN OF THE HALITE AND GYPSUM CRYSTALS.

The occurrence of gypsum and halite crystals in the voids has been described. The origin of the crystals may have been as follows.
The Fauld Mine diapirs were emplaced as gypsum. When the gypsum dehydrated substantial volumes of water were released, most of which escaped through the surrounding strata. However, the voids forming at the centre of the diapirs acted as traps for water released from the surrounding sulphate. The waters contained sulphate dissolved from the diapir and more infrequently, halite. The close association of halite crystals with diapirs containing trapped mudstone indicates that the mudstone was the source of sodium chloride. Aljubouri (1972) obtained a positive correlation between sodium and chlorine in the Trent Formation mudstones. He concluded that the two elements were present as discrete sodium chloride disseminated in trace amounts throughout the mudstone.

Another source of sodium is the clay fraction of the mudstones which are known to contain up to 60% mixed-layer smectite chlorite (Chapter 7). Smectite has a cation exchange capacity of 80 to 150 m.eq./100 gm.

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(Deer, Howie and Zussman, 1962) with sodium and calcium being the common exchangeable cations. On release of the dehydration waters, a process of cation exchange with a sodium-rich smectite-chlorite could occur resulting in a sodium-rich brine. Mudstones intersecting or passing close to a void also had their halite leached out by dehydration waters, which were then trapped in the void. The resulting brine could not rapidly escape and with increasing burial the brine was subjected to higher temperatures. The brine suffered gradual evaporation and concentration until gypsum and eventually halite crystallized out. It is not known whether any of the voids contained residual brines when breached by blasting.

Summarizing the preceeding discussion, both field evidence and considerations of the physical properties of the sulphate and mudstone indicate that the diapirs were formed by the flowage of gypsum. The ability of gypsum to flow was greatly increased by the presence of gypsum-saturated waters. Obviously the conclusion that sulphate emplacement occurred in the hydrated form places restrictions on the pressure-temperature conditions at the time of emplacement.

Gypsum dehydrates with increased temperature and pressure which can be related to the depth of burial. The stability fields of gypsum and anhydrite as functions of temperature and pressure were calculated by McDonald (1953) and subsequently modified by Berner (1971) (Fig. 4.10) to include surface temperature data of Hardie (1967). The fields were calculated as two solid phases under lithostatic pressure (gypsum and anhydrite) and the fluid phase under hydrostatic pressure. The most important parameters controlling the relative stabilities of anhydrite and gypsum are salinity and temperature. As can be seen in Fig. 4.10 increases in salinity of solutions associated with the sulphate markedly lower the temperature at which anhydrite becomes stable (with constant pressure). Also shown is a geothermal gradient from the evaporite-sedimentary terrain at Grand Saline Dome, Texas (McDonald, 1953). This gradient agrees with gradients measured in other sedimentary basins (Graf et al., 1966; Jones and Wallace, 1974).

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Fig. 4.10. Stability fields of gypsum (G) and anhydrite (A) as a function of temperature and pressure (depth). The G - A curves with negative slope correspond to cases where pressure on the solids is lithostatic and that on the surrounding water is hydrostatic. The G - A curve with positive slope is for the pure lithostatic situation. The geothermal gradient is that measured over the Grand Saline Dome, Texas. (After MacDonald, 1953; Hardie, 1967 and Berner, 1971). In this study the G - A curve used is that of seawater evaporated to gypsum saturation. Using this curve it can be seen that when gypsum is buried it is stable to a depth of 860 m (intersection of G - A curve and geothermal gradient curve).
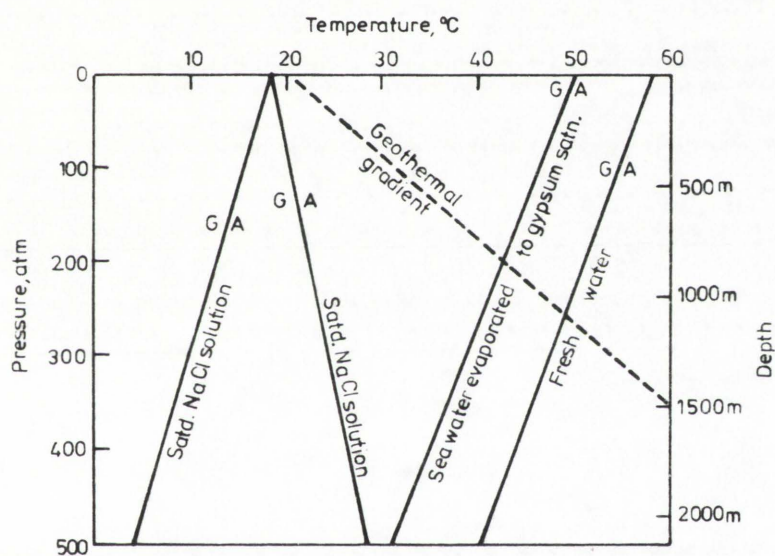
If one assumes that the Mesozoic geothermal gradient of the Triassic rocks in the Fauld area was of a similar magnitude, then gypsum with associated fresh water fluids would have been stable to a depth of 260 bars of lithostatic pressure. This is equivalent to 1120 m. if the overlying rocks have a density of 2.32 g./cm.^{3}. Below this depth gypsum would start to dehydrate to anhydrite. As the salinity of the fluids increases the gypsum-anhydrite transition depth decreases until with halite saturated solutions anhydrite is stable at the surface.

As previously discussed, the viscosity of gypsum is markedly lower when saturated solutions of gypsum are present. This situation corresponds to the gypsum-anhydrite equilibrium curve for sea-water evaporated to gypsum saturation (Fig. 4.10). One can argue that because movement of gypsum to form diapirs is aided by the presence of gypsum saturated waters this is the best curve to represent the burial of the Fauld sequence. On this curve gypsum dehydrates at a depth of 860 m.

Therefore based on the following assumptions,

a) the geothermal gradient was in the region of 25 to 30°C/km.

b) saturated solutions of gypsum were present to lower the viscosity of the gypsum,

it is estimated that the maximum depth at which gypsum flowage could have occurred is 860 m. At depths below this anhydrite is the stable phase. The estimated temperature at that depth is 43°C and the lithostatic pressure is 200 bars.

POSSIBLE TRIGGERS OF SULPHATE MOVEMENT.

Many ideas on the causes of diapir formation invoked a situation of gravitational instability (Dobrin, 1941; Nettleton and Elkins, 1947; Parker and McDowell, 1951). However, whilst all these model studies were able to generate diapiric structures, they all had to initiate the movement of the diapirs by artificial means, such as dropping glass beads onto the source layer. Parker and McDowell (1955) initiated domes by deliberately

varying the overburden thickness and also by placing mounds of source material on to the source layer. Gussow (1968) proposed that variations in geostatic load resulted in salt flowing to areas of "low" pressure where a diapir subsequently formed. If one is to invoke this type of model for the origin of the Fauld diapirs, then it follows from the distribution of the diapirs (Fig. 4.3) that the overburden must have been of either a highly irregular thickness and /or rapidly varying density. It is felt that both of these are unlikely.

More recently a model study by Tanner and Williams (1968) has shown that diapiric growth can be initiated by tension. They also concluded that "source-bed viscosity, rather than density differential, is the critical parameter in determining which beds respond to tension by producing diapiric structures." A similar conclusion has already been reached with regard to the Fauld diapirs due to their small height.

The Triassic in Britain was a time of large scale extensional or tensional tectonic activity. This resulted in the formation of extensive north-south trending grabens such as the Irish Sea and Cheshire-Worcestershire grabens (Audley-Charles, 1970b). Extensive fault activity continued through-out the Lower and Middle Triassic (Audley-Charles, 1970b; Tucker and Tucker, 1981) and decreased in the Upper Triassic. There is no known evidence of fault movement in Penarth Group times (Audley-Charles, 1970b).

Fauld lies within the Needwood Basin close to the Basin's eastern boundary. The Needwood Basin is thought to be an easterly extension of the Cheshire graben and, like the Cheshire graben, to have originated by tensional tectonic activity. Whereas the eastern boundary of the Needwood Basin could have been marked by a steep unfaulted slope, Audley-Charles (1970b) postulated that the boundary was marked by a major fault zone, in agreement with the tensional tectonic activity hypothesis (Fig. 4.11). It is difficult to locate the exact position of the fault zone. A borehole 4 km. east-north-east of Fauld Mine at the Hatton Creamery near Tutbury, penetrated a Triassic sequence (Stephens, 1929) very much thinner than the equivalent basinal

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Fig. 4.11.   Section across the eastern margin of the Needwood Basin

illustrating the faulted boundary close to Fauld Mine.

Note the thickening of Triassic strata in the graben area.

W

BAGOTS PARK
BOREHOLE
089269

PENARTH GROUP

BLUE ANCHOR FORMATION

HANBURY

TUTBURY SULPHATE BED

TUTBURY CASTLE

HATTON
CREAMERY
BOREHOLE
218197

TUTBURY SULPHATE BED

FAULD MINE

MERCIA
MUDSTONE
GROUP

HALITE

PALAEOZOIC

SHERWOOD
SANDSTONE
GROUP

FAULT ZONE

PALAEOZOIC

E

100

0 — O.D.

-100

-200

METRES

-300

-400

-500

-600

-700

0   1   2km

Vertical exaggeration ×10

CHESHIRE BASIN

PENNINE
MASSIF

0   25
Km.

Newark ●

EAST MIDLANDS HORST BLOCK

NEEDWOOD
BASIN
Fauld Mine

● Nottingham

CHARNWOOD
MASSIF

N

WELSH
MASSIF

● Leicester

sequence found in the Bagots Park borehole (Stevenson and Mitchell, 1955) (Fig. 4.11). The thinned sequence at the Hatton Creamery resulted from sedimentation on a Triassic "high", so the fault zone must lie within 4 km. of the mine entrance.

It is proposed that post-Triassic movement along this adjacent fault line resulted in the movement of the "mother" layer sulphate to form a random pattern of diapirs. The lack of surface evidence for later fault movement may be due to either:-

i) the very poor surface exposure combined with the difficulty of detecting faults at surface in the Mercia Mudstones or

ii) the faulting was confined to the deeper part of the sequence not exposed. The overlying Upper Triassic strata would have responded by forming a shallow monocline over the fault.

This latter case is more applicable to the Tanner and Williams (1968) tensional model as beds over the fault would be subjected to increased tension. The increased tension initiated gypsum flowage within a sulphate layer containing gypsum saturated water. Gypsum flowage need not be confined to a zone directly over the fault and probably occurred some distance either side of the monoclinal stress area.

POSSIBLE REASONS FOR THE ABSENCE OF
TRIASSIC HALITE DIAPIRS IN BRITAIN.

If gypsum can form diapirs, albeit small ones, then one might reasonably expect that halite in the same sedimentary sequence would also form diapirs. An obvious example is the halite beds present in the Cheshire Basin, yet there is no published account of extensive halite movement or diapirism having occurred there.

From studying Gulf Coast diapirs Gussow (1968) concluded that salt diapirism was initiated at a minimum depth of 7600 m. when the salt becomes plastic and mobile as a result of the increased temperature. Unlike the

previously discussed effect of water on the viscosity of gypsum, the effect of water on the viscosity of halite is minimal (Ode, 1968).

The two halite-bearing units present in the Cheshire Basin are the Northwich Halite Formation (Warrington et al., 1980) and the Wilkelsey Halite Formation (Warrington et al., 1980). They are overlain by 925 m. and 200 m. of Triassic strata respectively. For these units to form diapirs they would have had to have been buried beneath some 6600 to 7000 m. of Jurassic and Cretaceous strata. It is unlikely that this thickness of sediment was deposited in the Cheshire Basin during the Jurassic and Cretaceous. Therefore the reason for the absence of halite diapirism in the Cheshire Basin is very probably that the sequence has not been buried deeply enough to heat the halite to a temperature sufficient to promote halite flowage.

CONCLUSIONS.

The desert red bed mudstones at Fauld do not yield any indication of whether they were deposited subaerially or subaqueously. Any original bedding was probably destroyed by aeolian reworking and interstitial salt growth. The nodular nature of the sulphate in the bedded horizons is comparable with textures found in modern sabkhas indicating that sulphate precipitated from concentrated brines close to the sediment surface. There is evidence that the top of the Tutbury Sulphate Bed was exposed and eroded to a flat uniform surface. Small channels were cut into the top of the Tutbury Sulphate Bed during this period. The marked layering of the over-lying strata reflects variations in the depth to the water table, which in turn implies that the brine input volume was not constant.

The mudstones contain euhedral dolomite rhombs which are interpreted as the product of an early diagenetic reaction between Mg-rich brines and carbonate. A nodular dolomite horizon below the Tutbury Sulphate Bed initially formed as displacive carbonate nodules during a period of slow sedimentation and low salinity brine input. Dolomitization occurred when

high salinity brines returned to the area.

Large sulphate masses transecting the Tutbury Sulphate Bed and its surrounding strata are recognized to be diapiric structures. The mushroom-like shape of the masses with their smooth convex tops and irregular bases is characteristically found in halite diapirs. The emplacement of the sulphate diapirs resulted in partial or complete overturning and truncation of adjacent strata.

The sulphate "mother" bed is marked by lenses of horizontally elongated anhydrite nodules in a dolomitized mudstone containing minute fragments of sulphate. The original thickness of this bed is estimated at 1.4 m. The sulphate diapirs were emplaced as gypsum which was subsequently dehydrated giving rise to voids in the diapir cores. The voids frequently contain halite cubes and gypsum crystals. The halite was dissolved from mudstones trapped in the diapir by the dehydration waters. The sulphate was derived from the diapir.

The presence of gypsum-saturated waters within the "mother" layer greatly increases the ability of gypsum to flow. It is estimated that the maximum depth at which gypsum flowage could have occurred is 860 m. Movement along a nearby fault line increased the tension within the adjacent rocks and initiated gypsum flowage within the "mother" layer sulphate.

CHAPTER 5.

A STABLE ISOTOPE STUDY OF THE MERCIA MUDSTONES

(KEUPER MARL) AND ASSOCIATED SULPHATE HORIZONS

IN THE ENGLISH MIDLANDS.

INTRODUCTION

The origin of the evaporite horizons within the frequently dolomitic Mercia Mudstones (Triassic Keuper Marl of the English Midlands) has been debated for the past seventy years. In this time three main models have been proposed, each of which is similar in terms of palaeogeomorphology, the major differences between them being the source of the brines from which the evaporites were deposited.

(a) The Playa Lake Model.

After studying the area around the Charnwood Massif, Bosworth (1913) proposed that the evaporites, in this case gypsum, were precipitated from shallow ephemeral lakes scattered across an extensive peneplain. The water was derived from continental run-off, often in the form of sheet floods after torrential rainstorms.

Tucker (1977, 1978) interpreted the Mercia Mudstones of the South Wales coast as representing the coarse marginal facies of a giant playa. Wave-cut platforms eroded into the adjacent and underlying Carboniferous Limestone provide evidence for the periodic existence of a substantial body of water which Tucker thought formed after extended periods of wet climate. Evaporation of this lake led to the formation of nodular sulphate horizons in marginal sabkhas.

(b) Hypersaline Sea Model.

In this scenario the evaporites were deposited from a hypersaline sea cut off from the ocean by a sill (Sherlock, 1928; Sherlock and Hollingworth, 1938). Evans et al. (1968) invoked a marine origin for the brines to account for the massive quantities of halite in the Staffordshire - Cheshire - Irish Sea Basin. Warrington (1970) favoured the Mercia Mudstones being deposited from a shallow hypersaline epeiric sea deriving its water from the Tethys Ocean.

(c) Marine Flooding of Peneplain.

Wills (1970) proposed that a vast flat peneplain was periodically

flooded by ocean waters with some input by rivers and storm waters. Drying

out of these waters resulted in evaporite deposition with halite being

concentrated in the deeper basinal areas such as Cheshire. A modern analogue

for the large scale marine flooding of a peneplain is the Ranns of Kutch,

India, (Glennie and Evans, 1976). Arthurton (1973, 1980) concluded that the

Cheshire Basin halite accumulated in shallow lakes left after episodes of

marine flooding. He interpreted non-evaporitic laminated horizons within the

Cheshire sequence as lacustrine deposits with the water being derived from

continental run-off. Desiccation of the lakes initially formed by continental

run-off gave rise to minor amounts of evaporitic sulphate.

In the present work part of the Mercia Mudstone sequence was examined

in the English Midlands between Newark and Burton-upon-Trent (Fig. 5.1).

Sulphur isotope analyses and carbon and oxygen isotope analyses were carried

out on sulphate minerals and dolomitic mudstones respectively in an attempt

to:

(i) resolve the controversy of evaporite origin by seeing if the brines

were of either marine or continental origin or a mixture of both, and

(ii) construct a palaeogeographic and palaeohydrologic model for the area

of particular interest in this study.

Further samples from equivalent stratigraphic horizons elsewhere in

Central England were analysed for comparative purposes. As the dissolution

of pre-existing evaporite can be a major source of sulphate in continental

groundwaters, samples from major pre-Triassic evaporite sequences in or

close to the study area were also analysed.

LOCAL STRATIGRAPHY AND SEDIMENTOLOGY.

Elliott (1961) defined a sequence of formations for the Mercia Mudstones

of South Nottinghamshire (Fig. 5.2). The Trent Formation contains

particularly high concentrations of gypsum and anhydrite. The former

existence of minor halite is indicated by salt pseudomorphs found

Fig. 5.1. Location map and major stratigraphical units of the East
Midlands of Central England. Geological linework based on
Institute of Geological Sciences geological maps, Nos. 125,
126, 127, 140, 141, 142, 154, 155, 156, 169, 170. Numbers
on map of Great Britain indicate Yorkshire Basin (1),
Cheshire Basin (2) and Somerset Basin (3).

Legend

- Penarth Group & Jurassic
- Mercia Mudstone Group
- Sherwood Sandstone Group
- Palaeozoic
- Charnian / Caledonian Igneous

NEWARK
Quarries

NOTTINGHAM

Cropwell
Bishop Quarry

Gotham Mine

East Leake Mine

RIVER DOVE

Fauld Mine

BURTON

RIVER TRENT

MELTON
MOWBRAY

RIVER SOAR

N

LEICESTER

0   10 km.

MARKET
HARBOROUGH
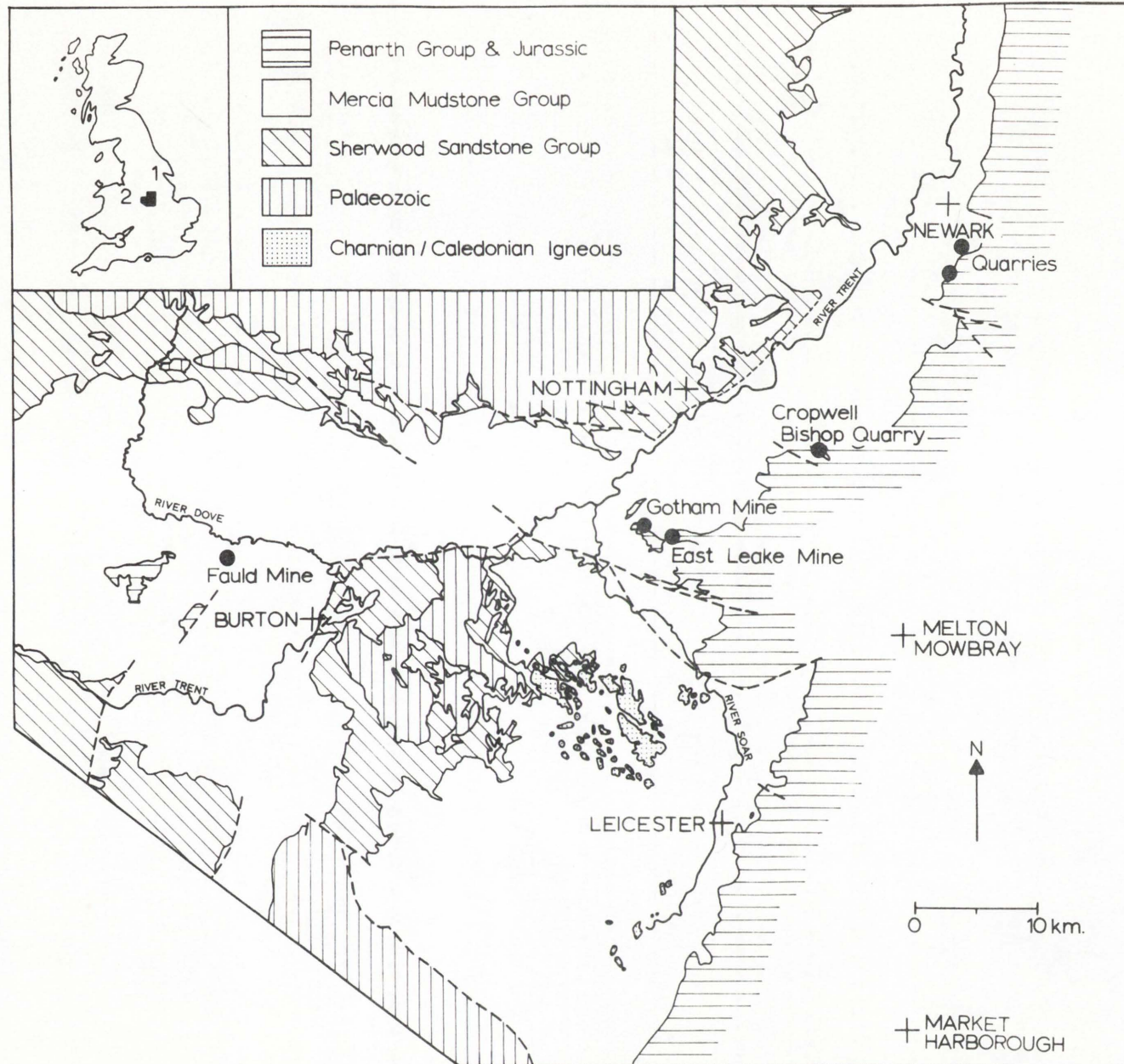
Fig. 5.2.   Subdivisions and correlations of the East Midlands Triassic
            to the standard Triassic sequence (after Warrington et al.,
            1980 and Elliott, 1961).

AREA	NEEDWOOD BASIN (Fauld)	SOUTH NOTTINGHAMSHIRE
LIAS	Lias	Lias

STANDARD SEQUENCE OF TRIASSIC STAGES		NEEDWOOD BASIN (Fauld)		SOUTH NOTTINGHAMSHIRE
	RHAETIAN	Penarth Group	MERCIA MUDSTONE GROUP	Penarth Group
		Blue Anchor Fm.		Blue Anchor Fm.
		Glen Parva Fm.		Glen Parva Fm.
	NORIAN	Trent Fm.		Trent Fm.
				—HOLLYGATE SSTS.—
	CARNIAN	Edwalton Fm.		Edwalton Fm.
		HALITE		Harlequin Fm.
		Harlequin Fm.		
	LADINIAN	Carlton Fm.		Carlton Fm.
		Radcliffe Fm.		Radcliffe Fm.
				Colwick Fm.
	ANISIAN	Sandstones		Woodthorpe Fm.
				non sequence
	SCYTHIAN	Conglomerates		Sandstones and Conglomerates

SHERWOOD SANDSTONE GROUP

predominantly on the underside of thin sandstone horizons. The sediments studied lie within the Trent Formation which is divided into the Fauld Member and the overlying Hawton Member (Fig. 5.3). Three samples from the later Blue Anchor Formation were also analysed (Fig. 5.3). The detailed stratigraphy and correlations are shown in Fig. 5.3.

FAULD MEMBER OF TRENT FORMATION.

Reddish-brown dolomitic siltstones and mudstones are exposed at Fauld Mine near Burton-upon-Trent, and at Gotham Mine and East Leak Mine south of Nottingham (Fig. 5.1). They are also known from boreholes between East Leake and Newark. Thin sandy horizons are locally present. In Fauld Mine the bedding is well picked out by slight changes in lithology and sulphate content. In the eastern part of the area (East Leake Mine) the strata are lithologically uniform and massively bedded. The Fauld Member contains a major nodular sulphate seam, the Tutbury Sulphate Bed, which in the western half of the area (Fig. 5.1) is 2 - 3 m. thick (Fig. 5.3). Between East Leake and Cropwell Bishop the bed gradually passes into a poorly developed nodular sulphate horizon.

HAWTON MEMBER OF TRENT FORMATION.

The Hawton Member is exposed in quarries around Newark and Cropwell Bishop and has been recognized in boreholes from Fauld Mine and East Leake. It contains nodular sulphate seams (0.3 - 1.5 m. thick) separated by sulphate-poor reddish-brown dolomitic mudstones lithologically similar to those of the Fauld Member. It also contains laterally persistent grey-green siltstone or sandy siltstone horizons which exhibit cross-bedding, graded bedding, planar lamination, ripples, salt pseudomorphs and desiccation cracks. By analogy with the work of Arthurton (1980) on lithologically similar horizons in the Cheshire Basin these beds are interpreted as temporary lake deposits.

Fig. 5.3.   The correlation of the Upper Triassic successions of the

Needwood Basin, East Leake and Newark. The Needwood Basin

succession is taken from the Institute of Geological

Sciences' Bagot's Park borehole (Stevenson and Mitchell,

1955). The East Leake and Newark sections are taken from

British Gypsum boreholes and quarries. The thickening of

the sequence represents a transition from horst to graben.

NEEDWOOD
BASIN

EAST
LEAKE

NEWARK

Penarth Group

Blue Anchor
Formation

Glen Parva
Formation

Trent Formation

Hawton
Member

Fauld
Member

nodular gypsum

NO CORE

NO CORE

CAVITY

NO CORE

NO CORE

GYPSIFEROUS
MUDSTONE

TUTBURY
SULPHATE

GYPSIFEROUS
MUDSTONE

TUTBURY
SULPHATE

GYPSIFEROUS
MUDSTONE

10m

Black shale

Grey-green beds

Red beds

Sulphate
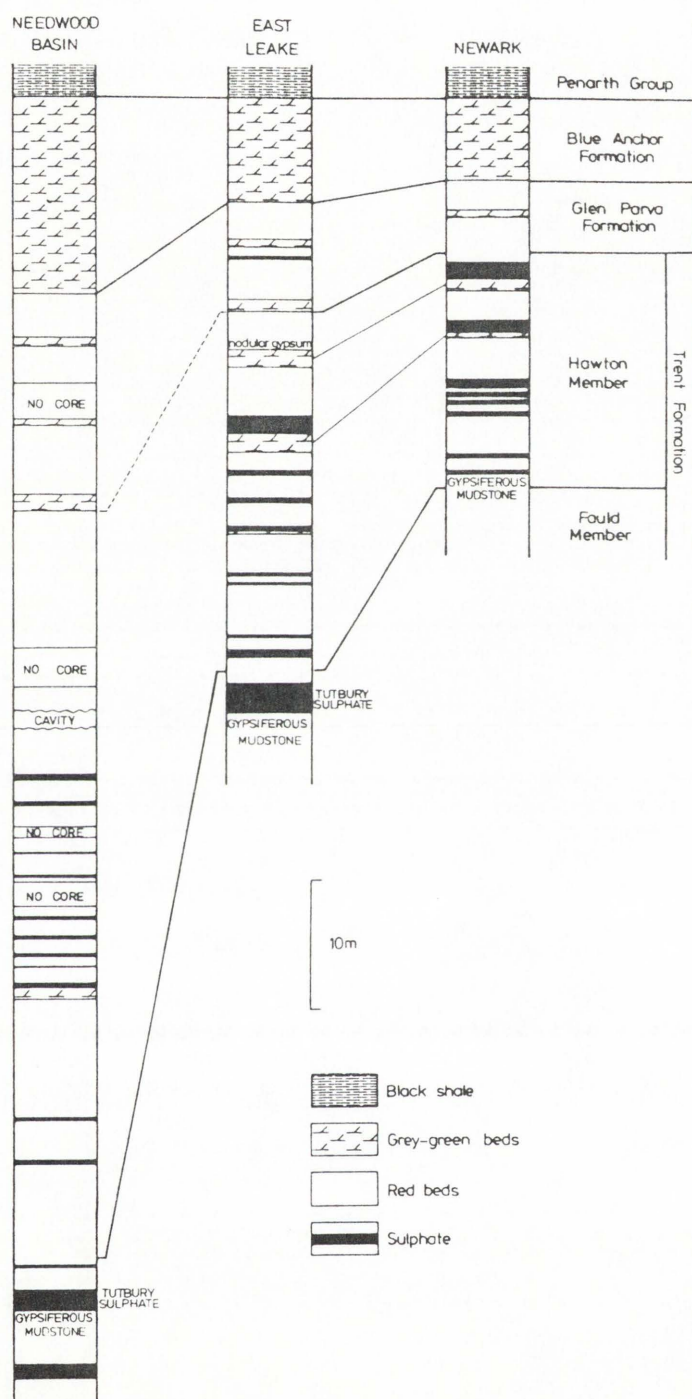
BLUE ANCHOR FORMATION.

The Blue Anchor Formation, temporarily exposed near East Leake Mine, consists of grey-green dolomitic mudstones and siltstones which have a predominantly massive appearance. Occasional planar lamination, cross lamination and graded bedding are seen. Desiccation cracks in the top of a heavily dolomitized mudstone are infilled by overlying dolomitic shaley mudstone. Many of the mudstones contain scattered but locally abundant fish remains. The fish remains and the similarity of the sedimentary structures to those of the lacustrine horizons in the Hawton Member and Cheshire Basin (Arthurton, 1980) indicates deposition in shallow, semi-permanent low salinity or freshwater lakes. Increasing salinity resulted in dolomitization and eventual desiccation. The absence of sulphate or halite minerals in the Blue Anchor Formation of the East Midlands supports the brine regime having been of overall lower salinity as suggested by Audley-Charles (1970b). The decrease in salinity may reflect a climatic change prior to the major marine transgression at the base of the Penarth Group (Audley-Charles, 1970b; Mayall, 1981).

Triassic sedimentation in Britain was controlled by penecontemporaneous block faulting (Audley-Charles, 1970b). Thick sequences accumulated in grabens, for example the Cheshire graben contains 3000 m.+ whilst horst blocks such as the East Midlands area were covered by relatively thin sequences (0 - 500 m.) (Audley-Charles, 1970a). The western portion of the study area is located within the Needwood Basin (Stevenson and Mitchell, 1955; Audley-Charles, 1970a) (Fig. 5.4), an easterly extension of the Cheshire Basin (Audley-Charles, 1970a). The steep (fault bounded?) eastern margin of the Needwood Basin passes close to Burton-upon-Trent (Audley-Charles, 1970a, b). Fauld Mine and the Institute of Geological Science's Bagot's Park borehole (Stevenson and Mitchell, 1955) (Fig. 5.3) both lie within the Needwood Basin. The central portion of the area from Burton-upon-Trent to Newark is situated on an extensive horst block (Fig. 5.4). The Derbyshire Massif, Charnwood Massif and the London Platform were areas of upstanding

Fig. 5.4.   Major palaeographic and structural elements present in the English Midlands during the Norian (adapted from Audley-Charles, 1970b; Wills, 1970).

CHESHIRE BASIN

PENNINE
DERBYSHIRE DOME
MASSIF

YORKSHIRE
BASIN

GENTLE SLOPE

NEWARK●

INTO

●NOTTINGHAM

YORKSHIRE

BASIN

NEEDWOOD
BASIN

EAST MIDLANDS
HORST BLOCK

FAULD MINE ●

CHARNWOOD
MASSIF

WELSH
MASSIF

●LEICESTER

WORCESTER GRABEN

NORTHAMPTON ●

LONDON   PLATFORM

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Km.

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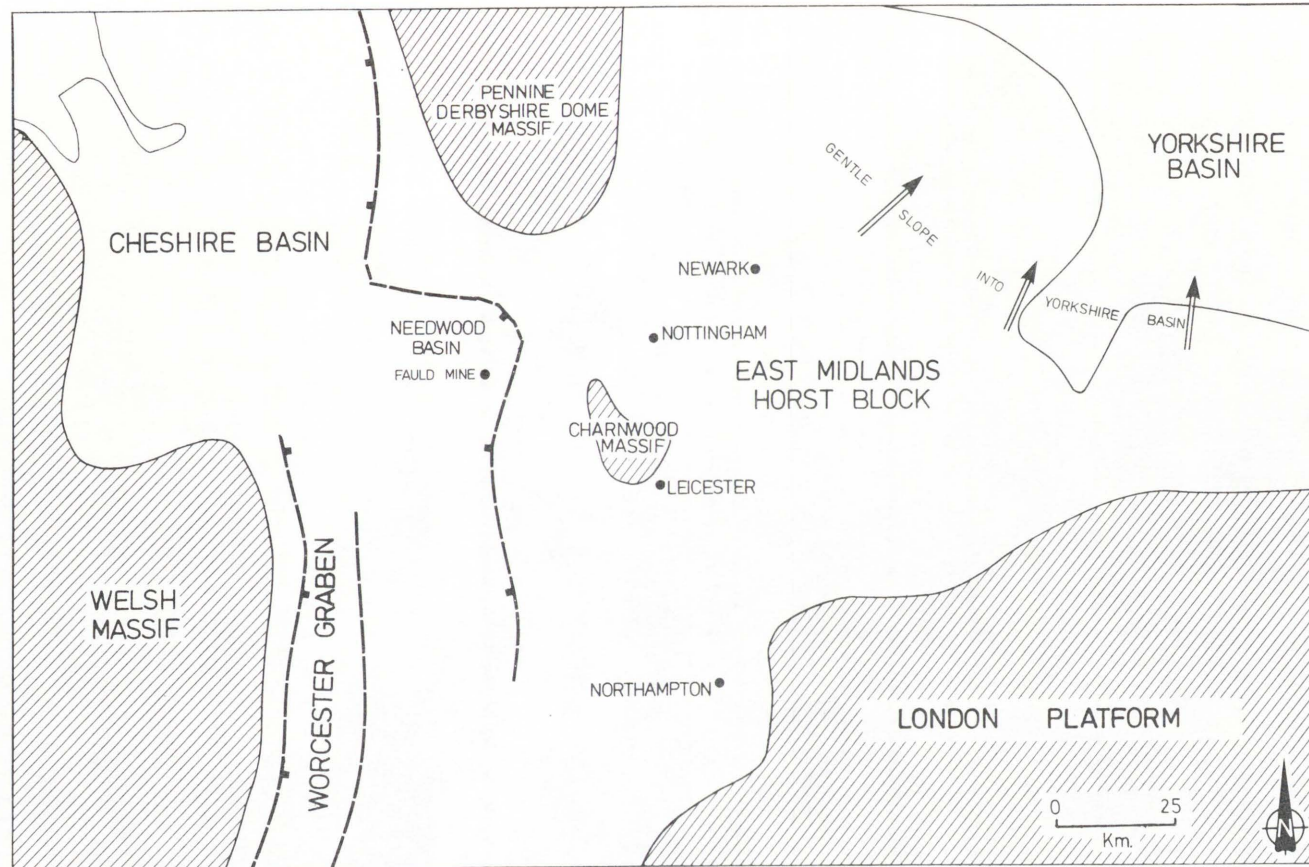
Areas exposed during Norian

Basin boundary

Faulted basin boundary

Present coast line

relief (Fig. 5.4) (Audley-Charles, 1970b; Wills, 1970). The gently
sloping southern margin of the Lincolnshire-Yorkshire Basin lies to the
north of Newark (Fig. 5.4) (Audley-Charles, 1970b). Borehole data show that
the Hawton Member (Fig. 5.3) and Fauld Member steadily thin from East Leake
Mine towards Newark indicating a possible palaeo-depositional slope direction.

Two major pre-Triassic evaporite sequences present in or close to the
study area are:

(a) Permian Zechstein sequence.

The Zechstein Basin extended as far south as Nottingham, although the
sulphate bearing horizons die out around Doncaster (Taylor and Colter, 1975)
80 km. north of the study area. It is uncertain whether Zechstein sulphates
were exposed during the Triassic.

(b) Hathern Anhydrite Series.

This is a Lower Carboniferous (Tournaisian) sequence of anhydrite-rich
sabkha cycles (Llewellyn and Stabbins, 1970). The lateral equivalent of
these beds crops out along the northern margin of the Charnwood Massif
(Monteleone, 1973). They are unconformably overlain by Mercia Mudstones and
contain abundant carbonate pseudomorphs of anhydrite chicken-wire structures
(Monteleone, 1973). The extent of the replacement zone is unknown, but
there were no indications of replacement textures in the Hathern borehole
cores drilled by British Petroleum in 1954 (Llewellyn and Stabbins, 1970).

MUDSTONE MINERALOGY.

Mudstones associated with the sulphate horizons contain 5 to 50% of
dolomite, finely disseminated throughout the silici-clastic material. A
single 12 cm. thick horizon of nodular dolomite occurs just below the Tutbury
Sulphate Bed at Fauld Mine.

Mudstones from the Fauld Member and Hawton Member were analysed by
X.R.D. Samples were taken from a borehole at Newark, East Leake Mine, Fauld
Mine and a borehole at Fauld. The analyses revealed the presence of two

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distinct clay mineral suites:-

(1) Predominantly mixed-layer smectite/chlorite containing a high percentage of smectite with sepiolite, palygorskite, illite and chlorite. This suite characterizes the Fauld Member and in all three areas grades up into suite 2.

(2) Predominantly illite and chlorite with minor mixed-layer chlorite/smectite containing a very small percentage of smectite. The suite characterizes the Hawton Member.

Sepiolite and palygorskite are Mg-rich neoformational clays which form from Mg-rich alkaline brines (Velde, 1977; Jeans, 1978). Smectite/chlorite forms by transformational aggradation of highly degraded illite in Mg-rich environments (Lucas, 1962; Chapter 7). The illite and chlorite represent detrital input (Jeans, 1978). Jeans (1978) detected, but did not interpret, a similar change of Mg-rich clays overlain by detrital illite and chlorite in a borehole at Cropwell Bishop. The horizon at which this change occurs correlates with the junction of the Fauld Member and Hawton Member. Therefore it appears that a major change in the clay mineralogy occurs at the junction of the Fauld Member and Hawton Member across a large part of the East Midlands. This can perhaps be explained by a change in either the brine composition and/or the pH of the brines. During the deposition of the Fauld Member the brines were Mg-rich with a high pH, resulting in transformation of degraded detrital illite and the neoformation of Mg-rich clay minerals in the detrital clay. Then a decrease in the Mg and/or Si content of the brines halted both transformational and neoformational processes (Velde, 1977) leading to a predominance of detrital clays in the Hawton Member (Chapter 7). The variations in brine chemistry could reflect a major change in the brine source.

SAMPLING.

The majority of Mercia Mudstone sulphate and dolomitic mudstone samples

were obtained from quarries, mines and borehole cores in the East Midlands area. For comparative purposes selected samples were analysed from the Mercia Mudstones of Yorkshire, Cheshire and the Somerset Basin, and from the Carboniferous Hathern Anhydrite Series and the Permian Upper Anhydrite, Sherburn Mine, Yorkshire.

EXPERIMENTAL PROCEDURE.

SULPHATES.

All samples except those from the Hathern Anhydrite Series were taken from massive or nodular gypsum. A 2 g. sample was obtained from each hand specimen by using a dentist's drill. All samples were drilled out from the centre of the gypsum block or nodule to try to avoid bacterial alteration which can occur on nodule surfaces.

There is very little core now remaining of the Hathern Anhydrite Series but chippings collected from the British Petroleum borehole were available, so representative samples of these were analysed. Gypsum samples were ground and then dehydrated by heating at 500° C. for thirty minutes. The high temperature was used to prevent later rehydration. The resulting powders were analysed by X.R.D. and they showed peaks characteristic of anhydrite. Anhydrite samples were treated in a similar fashion.

A 6 mg. sample of each sulphate powder was ground with 125 mg. Cu_2O and 375 mg. SiO_2 and then heated at $1,120^{\circ}$ C. under vacuum (Coleman and Moore, 1978). This resulted in the sulphate being reduced directly to SO_2 which was then analysed on a "Micromass 602" Spectrometer. Reproducibility of the overall procedure is 0.2‰.

CARBONATES.

Dolomitic mudstone samples were ground and then reacted with 100% phosphoric acid at 25° C. (McCrea, 1950). Sample weight varied between 15 mg. and 80 mg. according to the percentage of dolomite contained in the

sample. This was determined by X.R.D. As CO_2 is liberated more slowly from dolomite than calcite, the samples were left to react for 4 - 5 days. Isotopic ratios were determined on a "Micromass 602" Spectrometer and corrected using standard procedures (Craig, 1957). Reproducibility of the overall procedure is 0.05‰.

All analyses were carried out at the Geochemical Unit of the Institute of Geological Sciences, London.

RESULTS.

The $\delta^{34}S$ results are shown in Table 5.1 and are expressed in terms of the Canon Diablo standard. The $\delta^{13}C$ and $\delta^{18}O$ results are shown in Table 5.2 and are expressed in terms of the P.D.B. standard. As the oxygen-18 fractionation between CO_2 liberated by phosphoric acid from ideal ordered dolomite differs from that of calcite, -0.83‰ was added to the $\delta^{18}O$ values found for dolomite (Sharma and Clayton, 1965).

SULPHATE - SOURCES AND SULPHUR ISOTOPES.

The sulphate which is present in gypsum deposits could have been derived from either or both of two main sources:

(a) MARINE SOURCE.

Various sulphur isotope age curves for marine sulphates have been published (Holser and Kaplan, 1966; Nielsen, 1978; Claypool et al. 1980). It is well established that the Triassic was a period of extremely rapid change in the sulphur isotope values of marine sulphates. The low values of +10‰ to +12‰ for the Upper Permian are followed in the Lower Triassic by a steep rise in values with localized overshooting (Röt event of Holser, 1977). The published estimates of the $\delta^{34}S$ value of marine sulphate in the Upper Triassic are +16.5‰ (Claypool et al. 1980) and +19‰ (Nielsen, 1978, 1979). It is generally agreed that the Middle and Upper Jurassic period

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Table 5.1   Sulphur isotope data of calcium sulphates.

Sample No.	Locality and horizon	$\delta^{34}S\%_{\circ}$ (CDT)
17	Newark, Hawton Mbr, 13.5 m below P.G.	+14.1
11	Newark, Hawton Mbr, 17.8 m below P.G.	+14.2
3	Newark, Hawton Mbr, 23.4 m below P.G.	+14.4
32	Cropwell Bishop, Hawton Mbr, 24.1 m below P.G.	+14.2
33	Cropwell Bishop, Hawton Mbr, 28.3 m below P.G.	+14.3
67	East Leake, Hawton Mbr, 27.1 m below P.G.	+12.8
30	N. Leicestershire, Hawton Mbr, 24.0 m below P.G.	+14.4
29a	N. Leicestershire, Fauld Mbr, Tutbury Sulphate	+13.8
19	East Leake, Fauld Mbr, Tutbury Sulphate	+13.8
20	East Leake, Fauld Mbr, Tutbury Sulphate	+13.6
46	East Leake, Fauld Mbr, Tutbury Sulphate	+12.8
91	Chellaston, Hawton Mbr?	+13.4
21a	Fauld, Fauld Mbr, Tutbury Sulphate	+13.2
24	Fauld, Fauld Mbr, Tutbury Sulphate	+13.7
47	Fauld, Fauld Mbr, Tutbury Sulphate	+13.6
72	Fauld, Fauld Mbr, Tutbury Sulphate	+12.3
73	Fauld, Fauld Mbr, Tutbury Sulphate	+12.2
74	Fauld, Fauld Mbr, Tutbury Sulphate	+12.8
75	Fauld, Fauld Mbr, Tutbury Sulphate	+12.6
76	Fauld, Fauld Mbr, Tutbury Sulphate	+12.8
77	Fauld, Fauld Mbr, Tutbury Sulphate	+12.7
92	Fauld, Fauld Mbr, Tutbury Sulphate	+12.0
25	Fauld, Fauld Mbr, Tutbury Sulphate	+12.3
49	Fauld, Fauld Mbr, Tutbury Sulphate	+11.5
78a	Fauld, Fauld Mbr, 2.0 m below Tutbury Sulphate	+13.1
79a	Fauld, Fauld Mbr, 1.0 m above Tutbury Sulphate	+13.0
80	Fauld, Hawton Mbr, 3.5 m above Tutbury Sulphate	+12.2
81	Blyton, Yorkshire, 50.0 m below P.G.	+14.6
56	Whitchurch, Cheshire, 78.0 m below P.G.	+13.6
57	Whitchurch, Cheshire, 125.0 m below P.G.	+12.5
58	Whitchurch, Cheshire, 126.0 m below P.G.	+13.1
RT1	Winsford, Cheshire, Northwich Halite Fm.	+15.9
53	Lavernock Pt., S. Wales, 18.0 m below P.G.	+11.8
54	Lavernock Pt., S. Wales, 18.0 m below P.G.	+12.9
55	Lavernock Pt., S. Wales, 18.0 m below P.G.	+13.2
62	Somerset, 58.0 m below P.G.	+15.0
64	Somerset, Blue Anchor Fm. 15 m below P.G.	+16.6
71	Somerset, Blue Anchor Fm. 24 m below P.G.	+15.0
59	N. Yorkshire, 22 m below P.G.	+16.6
61	N. Yorkshire, 43.3 m below P.G.	+14.7
Y1	Sherburn, Yorkshire, Zechstein Upper Anhydrite	+12.0
Y2	Sherburn, Yorkshire, Zechstein Upper Anhydrite	+9.7
Y3	Sherburn, Yorkshire, Zechstein Upper Anhydrite	+8.2
H1	Hathern borehole, depth 1770 ft., Hathern Anhydrite	+17.6
H2	Hathern borehole, depth 1810 ft., Hathern Anhydrite	+19.6
H3	Hathern borehole, depth 1875 ft., Hathern Anhydrite	+21.4
H4	Hathern borehole, depth 1937 ft., Hathern Anhydrite	+19.3
H5	Hathern borehole, depth 2038 ft., Hathern Anhydrite	+19.6

Horizon key:-   P.G. = Penarth Group

Table 5.2   Carbon and oxygen isotope data of dolomites.

Sample No.	Horizon	Lithology	$\delta^{13}C‰$ (PDB)	$\delta^{18}O‰$ (PDB)
HH/TGM/1	Blue Anchor Fm.	Dolomite	-0.31	1.45
HH/TGM/2	Blue Anchor Fm.	Mudstone	-1.20	0.87
HH/TGM/3	Blue Anchor Fm.	Lamn siltst.	-0.53	0.07
EL97/7389	Hawton Mbr, 24.6 m below P.G.	Red mudst.	0.45	-0.74
EL97/6035	Hawton Mbr, 38.2 m below P.G. (6.7 m above T.S.)	Lamn shale	-1.78	-1.98
EL86	Hawton Mbr, 1.5 m above T.S.	Red mudst.	0.43	-0.42
EL85	Fauld Mbr, 0.5 m above T.S.	Red mudst.	0.49	-0.58
EL84	Fauld Mbr, within T.S.	Red mudst.	0.87	-0.32
EL83	Fauld Mbr, base of T.S.	Red mudst.	0.53	-1.28
N690	Hawton Mbr, 18.0 m below P.G.	Grey siltst.	-1.41	-0.63
N717	Hawton Mbr, 18.3 m below P.G.	Grey siltst.	-1.58	-1.02
N1500	Hawton Mbr, 26.1 m below P.G.	Red mudst.	0.01	-1.19
N1850	Hawton Mbr, 29.6 m below P.G.	Red mudst.	-0.10	-0.72
N1900	Fauld Mbr, 30.1 m below P.G.	Red mudst.	0.94	2.35
N1950	Fauld Mbr, 30.6 m below P.G.	Red mudst.	1.22	2.29
N2000	Fauld Mbr, 31.1 m below P.G.	Red mudst.	0.98	2.38
N2040	Fauld Mbr, 31.5 m below P.G.	Dolm.nods.	1.15	2.47
KY3000	Fauld Mbr, 35.0 m below P.G.	Red mudst.	1.41	3.38
KY3100	Fauld Mbr, 36.0 m below P.G.	Red mudst.	1.84	4.12
F193/7853	Hawton Mbr, 67.3 m below P.G.	Red mudst.	-0.27	-2.56
F193/8595	Hawton Mbr, 74.7 m below P.G.	Red mudst.	0.51	-2.08
F193/8870	Hawton Mbr, 77.5 m below P.G.	Red mudst.	0.54	-1.90
F193/9120	Hawton Mbr, 80.0 m below P.G.	Red mudst.	0.58	-2.38
F193/9366	Hawton Mbr, 82.4 m below P.G. (2.6 m above T.S.)	Red mudst.	0.67	-1.59
F010/D	Fauld Mbr, 1.5 m above T.S.	Red mudst.	2.50	1.92
F060	Fauld Mbr, 1.0 m above T.S.	Red mudst.	2.19	2.38
F100/D	Fauld Mbr, 0.6 m above T.S.	Red mudst.	2.23	2.14
F160	Fauld Mbr, top of T.S.	Red mudst.	1.97	2.19
F260/D	Fauld Mbr, base of T.S.	Red mudst.	3.01	3.85
F260/D	Duplicate analysis of above.		2.92	3.86
F380/D	Fauld Mbr, 1.2 m below T.S.	Red mudst.	1.73	3.13
F500/D	Fauld Mbr, 2.4 m below T.S.	Dolm. nod.	1.00	2.25
F500	Fauld Mbr, 2.4 m below T.S.	Dolm. nods.	0.83	3.66
B3294	31.7 m below P.G.	Red mudst.	-4.06	-1.14
B3436	33.2 m below P.G.	Dolomite	-4.30	-0.94
B3500	33.8 m below P.G.	Grey mudst.	-4.00	-0.72
B3640	35.2 m below P.G.	Dolomite	-3.68	-2.98
B5120	50.0 m below P.G.	Grey mudst.	-4.02	-1.81

Sample localities given by prefix on sample number.
B = Blyton, Yorkshire;   EL = East Leake;   F = Fauld;
HH = Hotchley Hill, nr. East Leake;   KY = Keyworth;
N = Newark.

Horizon Key:- P.G. = Penarth Group;   T.S. = Tutbury Sulphate Bed.

had a $\delta^{34}S$ value of about +16.5‰.

The author produced an age curve for the Triassic and Jurassic of Europe based on previously published analyses (Fig. 5.5). This suggests that the $\delta^{34}S$ value for Norian marine sulphate lies between +17‰ and +19‰ and that very possibly the $\delta^{34}S$ value was increasing steadily throughout the Norian.

(b) CONTINENTAL SOURCE.

The $\delta^{34}S$ of modern freshwater sulphate is close to zero whilst many continental saline basins (Dead Sea, Caspian Sea, Great Salt Lake) range between +5‰ and +15‰ (Holser and Kaplan, 1966).

Sulphate in continental waters can be derived from the following sources; the $\delta^{34}S$ ranges are shown in brackets (Holser and Kaplan, 1966):

(1) Dissolution of earlier evaporites (+10‰ to +30‰)

(2) Springs and fumaroles of volcanic origin (-10‰ to +15‰)

(3) Oxidation of pyrite from shales, hydrothermal veins and igneous rocks (-40‰ to +5‰)

(4) Precipitation and fallout from the atmosphere (+3‰ to +15‰).

(1) Dissolution of earlier evaporites.

Two evaporite sequences in or close to the area of study, through which Norian groundwaters could have passed, are:

(a) Permian Zechstein evaporites.

Three analyses carried out on samples from the Upper Anhydrite, Sherburn, Yorkshire, gave a mean value of +10.0‰. This is in agreement with the well documented $\delta^{34}S$ value of +10.5‰ for the Upper Permian (Claypool et al., 1980).

(b) Hathern Anhydrite Series.

Analysis of anhydrite chippings from the Hathern Anhydrite Series gave a mean $\delta^{34}S$ value of +19.5‰ . It is possible that sample H3 has been subjected to bacterial reduction. If sample H3 is discarded the mean $\delta^{34}S$ value is +19.0‰. Although the $\delta^{34}S$ value for marine sulphate in the Lower Carboniferous is poorly known, this mean is close to the value of +18‰

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Fig. 5.5.   Sulphur isotope date for European evaporitic sulphates of Triassic and Jurassic age. The heavy line represents the qualitative best estimate for δ^{34}S of sulphate minerals in equilibrium with the world ocean sulphate of that time. The dashed line represents an alternative for δ^{34}S during the Anisian if the Röt data is regarded as only representing a localized event. The stratigraphic time scale is based on the Geological Time Table (Van Eysinga, 1975). Data sources:- 1, Claypool et al. (1980); 2, Müller et al. (1966); 3, Nielsen (1968); 4, Nielsen (1979); 5, Pilot et al. (172).
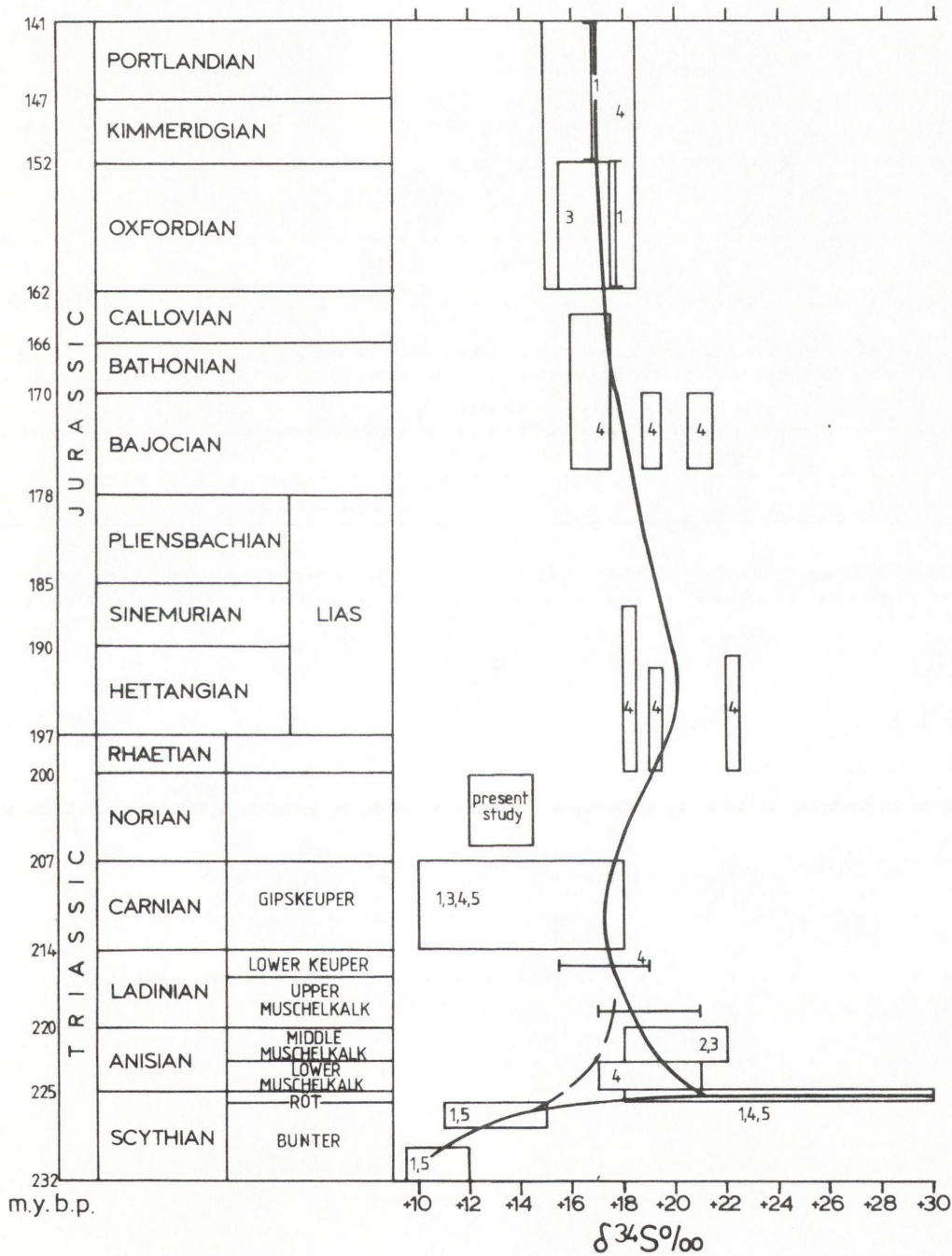
±2‰ of Claypool et al. (1980).

Sulphate derived from the Zechstein Basin and Hathern Anhydrite Series would have $\delta^{34}S$ values of +10.0‰ ±2.0‰ and +19.0‰ ±1.0‰ respectively.

(2) Springs and fumaroles of volcanic origin.

There is no indication of volcanic activity in the British Isles or Northern Europe during the Norian. Late Triassic volcanicity occurred in S. Europe, the Aquitaine Basin and eastern N. America (Hallam and Sellwood, 1976). Therefore the sulphur input from this source was probably negligible.

(3) Oxidation of pyrite.

As there were substantial areas of land exposed to weathering during the Norian, such as the London Platform, there could have been a significant input of isotopically light sulphur derived from oxidized sulphides. Holser and Kaplan (1966) give an average value of $\delta^{34}S$ -12‰ for biogenic sulphide and a value close to zero for igneous and hydrothermal sulphide. As sediments were the major rock type exposed on the Norian landmasses (Wills, 1971) the input of oxidized biogenic sulphide probably surpassed that of the latter sulphides.

(4) Precipitation and fallout.

The atmospheric inflow of sulphur is likely to be of minor importance when compared to the total sulphur input of sources 1 and 3.

To summarize, the major sources of sulphur in Norian continental waters would have been the dissolution of earlier evaporites and the oxidation of sulphides. The isotopic composition of these waters could theoretically have varied between $\delta^{34}S$ +19‰ and $\delta^{34}S$ -40‰ depending on the predominant rock types exposed in the catchment area. However, by analogy with modern continental waters it is likely that Norian continental waters would have had $\delta^{34}S$ values between -12‰ and 0‰ when the sulphate was derived from the oxidation of pyrite. The $\delta^{34}S$ values would have increased as the content of "heavy" dissolved evaporite increased. The amount of lateral variation preserved in the evaporite record would depend on the $\delta^{34}S$ values of streams feeding the sabkha and the degree of brine mixing within the sabkha area.

MIXING OF MARINE AND CONTINENTAL WATERS.

Towards the landward margin of modern sabkhas continental run-off mixes with marine brines (Butler, 1973). The mixing of the two brines gives rise to a wide range of possible $\delta^{34}S$ values depending on the ratio of marine to continental input and the isotopic composition of the continental brines.

A mixing graph (Fig. 5.6) illustrates the possible variations in the input ratios of the four Norian sulphate sources which when combined give a single $\delta^{34}S$ value.

DEHYDRATION - HYDRATION OF CALCIUM SULPHATE.

Holser and Kaplan (1966) concluded that repeated hydration-dehydration of calcium sulphate does not significantly alter its $\delta^{34}S$ value.

DOLOMITES - OXYGEN AND CARBON ISOTOPES.

OXYGEN ISOTOPES.

It is well documented that calcite precipitated from normal salinity marine waters of $\delta^{18}O$ 0.0‰ SMOW will have a $\delta^{18}O$ of -1‰ PDB at 25° C. This value will decrease with increasing temperature (Epstein, 1959), but in a strongly evaporitic environment temperature effects are very rapidly masked by the preferential evaporative loss of $H_2^{16}O$ which leaves residual brines enriched in $H_2^{18}O$ (Lloyd, 1966). Lloyd (1966) concluded that the maximum enrichment in oxygen isotopic composition of brines evaporated under arid conditions would be 6‰. Calcite precipitated from these enriched brines at 25° C. would have a $\delta^{18}O$ of between +1.5‰ PDB and +2.5‰ PDB (Land, 1980). Dolomite replacing these carbonates in a hypersaline environment will be enriched by 3‰ ±1‰. (Land, 1980; McKenzie, 1981) resulting in a range of values from +3.5‰ to +6.5‰ . Holocene dolomites of the marine sabkhas on the Persian Gulf, where dolomitization of lagoonal carbonate is occurring, have $\delta^{18}O$ values of between +1.3‰ and +4.0‰ (McKenzie, 1981).

Fig. 5.6. A sulphur isotope mixing diagram with the three possible

sources of Upper Triassic sulphate. The evaporite source

shows Permian input (dotted line $\delta^{34}S$ +10‰) and

Carboniferous input (solid line, $\delta^{34}S$ +18.5‰).
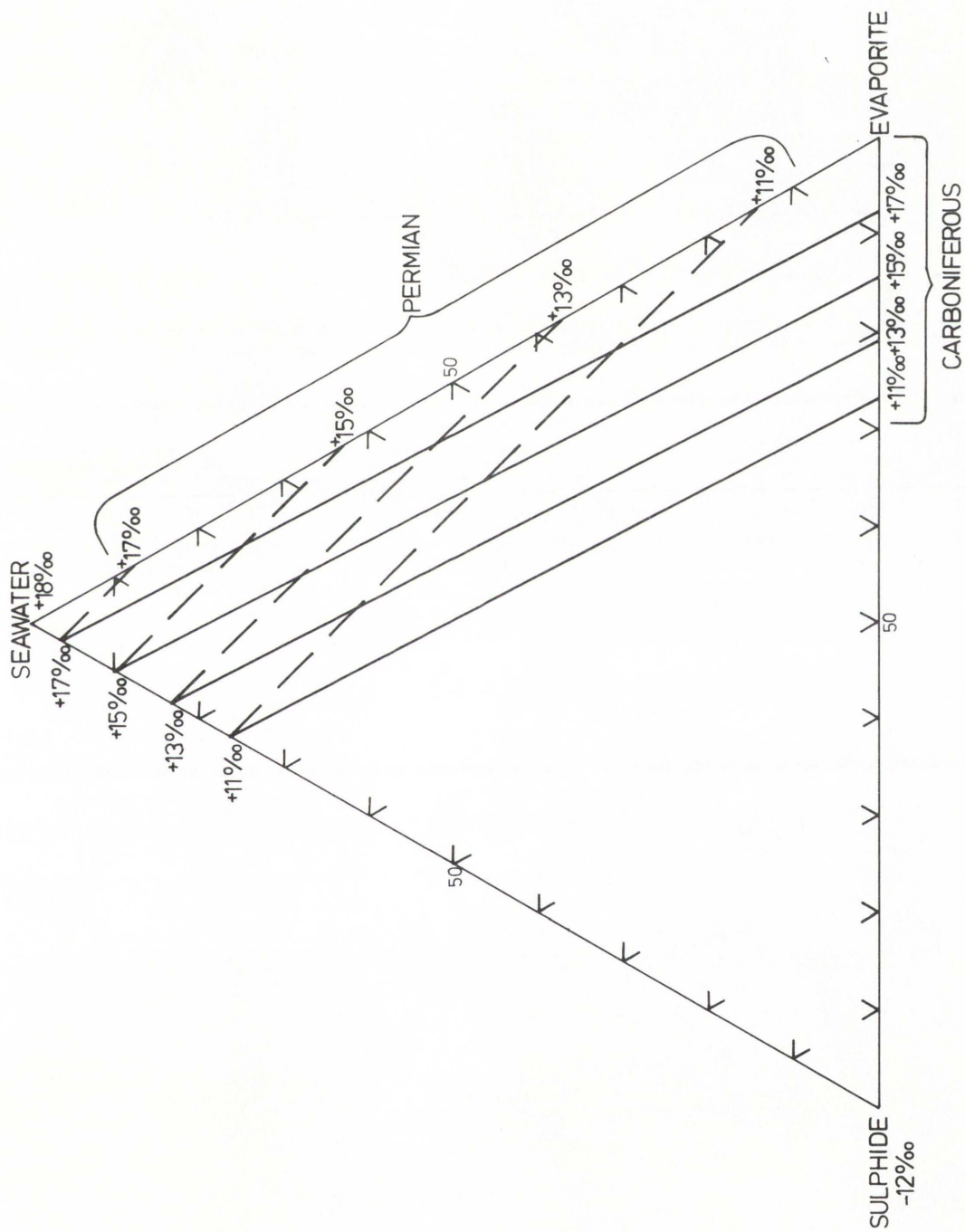
Continental waters are generally isotopically lighter than marine waters with ^{18}O depletion increasing towards the poles (Dansgaard, 1964). Assuming a palaeolatitude for Norian Britian of around $30^{o}N$ (Smith and Briden, 1977) and a similar $\delta^{18}O$-latitude trend then as now, then rainfall in the area would have had a $\delta^{18}O$ of about $-6\%_o$ SMOW. By the Upper Triassic most of Britain and Northern Europe had been reduced to a peneplain at or close to sea level (Audley-Charles, 1970b). Therefore the $\delta^{18}O$ values of the precipitation are unlikely to have been affected by altitude. Evaporation of these waters could ultimately lead to the formation of dolomites with a $\delta^{18}O$ range of $-4\%_o$ to $0\%_o$. This range is equivalent to that found in modern continental evaporitic dolomites at similar latitudes (Rouse, 1981).

CARBON ISOTOPES.

The $\delta^{13}C$ of dolomite from the Trucial Coast sabkha ranges between $+2\%_o$ and $+4\%_o$ PDB (McKenzie, 1981). Continental brine dolomites have $\delta^{13}C$ values between $-1.5\%_o$ and $+2.0\%_o$ PDB (Rouse, 1981). The carbon isotope composition of dolomites reflects the carbon isotope composition of both the original carbonate and the dolomitizing solutions (Land, 1980; Mattes and Mountjoy, 1980). Organic carbon and organically produced CO_2 are isotopically light with an average $\delta^{13}C$ of $-25\%_o$ (Degens et al., 1968; Deuser, Degens and Guillard, 1968). Incorporation of light carbon into the carbonate equilibria system via CO_2 (aq.) will result in a lowering of $\delta^{13}C$ in both solution and any carbonate being precipitated. A major source of organic material in modern marine sabkhas is algal mats (Shearman, 1966; Kinsman, 1969) the decay of which could release large amounts of light CO_2 into the groundwaters. Algal mats are rarely documented in continental sabkhas; instead organic carbon is derived from salt tolerant plants (Hunt et al.,1966) and from material washed in from surrounding landmasses. The Trent Formation mudstones in the Midlands have not yielded any traces of algal mat material. However minor organic input may be indicated by the presence of 'fisheyes' - a term given to small spherical areas of reduced mudstone with a possibly

organic nucleus (Aljubouri, 1972).

DISCUSSION OF RESULTS.

Having discussed the possible sulphate sources during the Norian and
the factors controlling the C/O isotope ratios, the analyses from each area
will now be considered.

EAST MIDLANDS SULPHATES.

Samples were analysed from Newark, East Leake, Chellaston and Fauld.
Figure 5.7 shows the results plotted against depth from the base of the
Penarth Group. Samples from the Hawton Member show a close grouping with a
mean $\delta^{34}S$ of +14.3‰. The Fauld Member at East Leake and North Leicester-
shire, and at Fauld shows a wide spread of values with respective means of
$\delta^{34}S$ +13.5‰ and $\delta^{34}S$ +12.7‰ . All three mean values are lighter than the
assumed Norian sea water sulphate value of $\delta^{34}S$ +18.0‰. This implies that
some sulphate was derived from the weathering of pre-existing pyrite and
evaporites exposed on adjacent landmasses, i.e. the sulphate is either wholly
or partially of continental origin.

A t-test showed that the means for the Hawton Member and Fauld Member
are significantly different. Perhaps more surprisingly it also showed that
the means for the Fauld Member in the East Leake area and in the Fauld area
are significantly different. These differences point towards changes in the
sulphur isotope composition of the original brines. The simplest interpre-
tation of these differences is that the Fauld Member sulphate of the Fauld
area had a higher input of isotopically light continental sulphate than the
East Leake area. Similarly the Hawton Member sulphate shows a decrease in
the level of light sulphate input. It is very difficult, if not impossible,
to estimate solely on the basis of $\delta^{34}S$ values which was the predominant
"heavy" sulphate source, i.e. seawater or Carboniferous evaporite, and
whether the "light" sulphate was derived from pyrite or Permian evaporite.

Fig. 5.7.   Sulphur isotope data for Upper Triassic sulphates from the

East Midlands, Somerset, Yorkshire, Cheshire Basin and the

South Wales coast.   Analyses are plotted against depth below

the base of the Penarth Group, not on a stratigraphic time

scale.
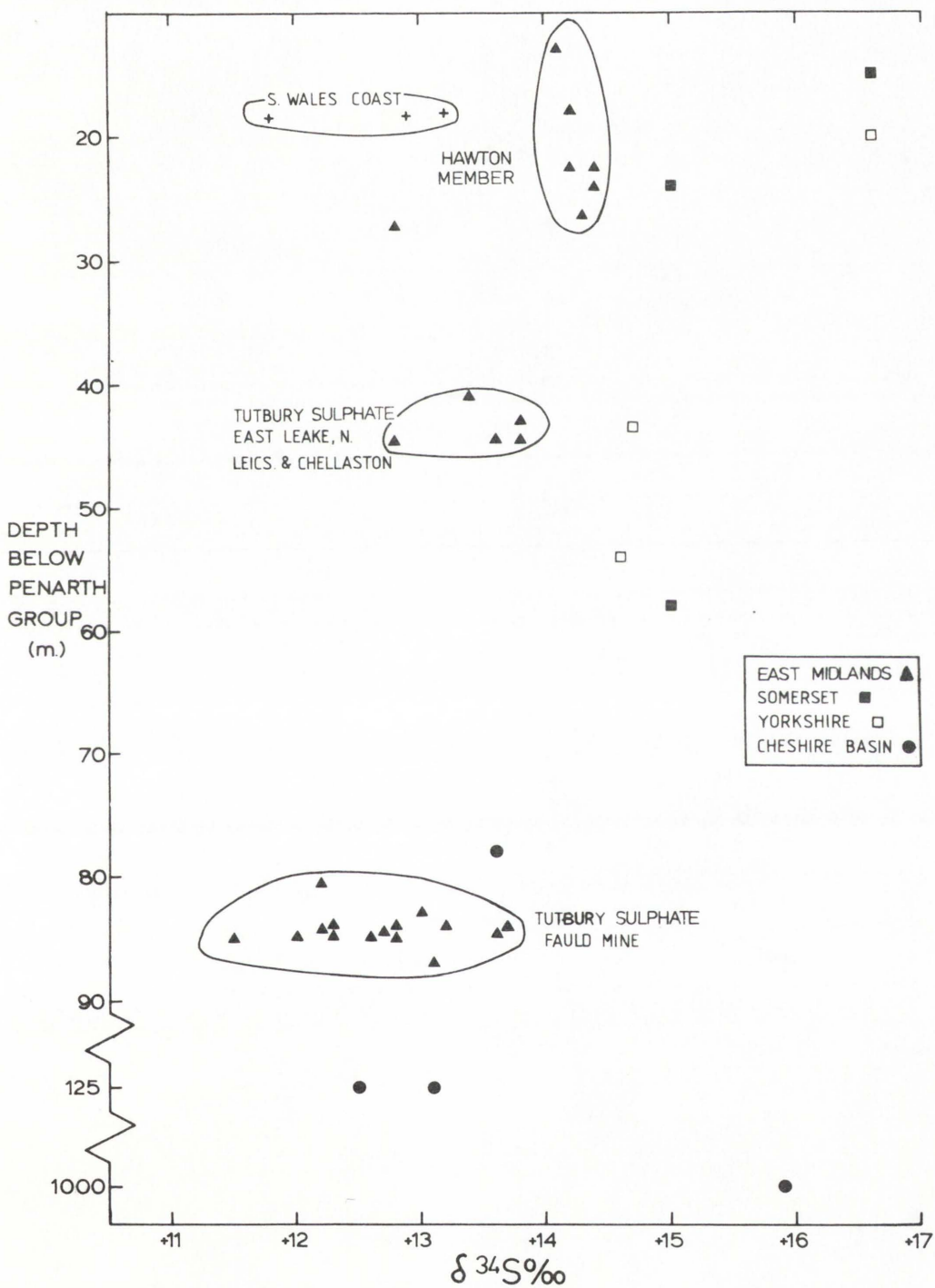
The $\delta^{18}O$ analyses of the associated dolomites help to resolve this problem.

EAST MIDLANDS AREA CARBONATES - OXYGEN ISOTOPES.

The $\delta^{18}O$ values for the East Midlands samples fall into three distinct groups (Fig. 5.8).

(1) Values range from +2‰ to +4‰. These samples came from the Fauld Member at Fauld, Newark and Keyworth (8 km. N.E. of East Leake Mine). The values can be interpreted as representing either brines of predominantly marine origin or brines of continental origin which have been enriched in ^{18}O by evaporation.

(2) Values range from -3‰ to 0‰. These samples came from Blyton, Yorkshire, the Hawton Member and the Fauld Member at East Leake. The light values are interpreted as indicating a continental origin for the brines.

(3) Values range from 0‰ to +1.5‰. These samples came from the Blue Anchor Formation at East Leake. The values can be interpreted as representing either continental brines heavily enriched in ^{18}O by evaporation, marine brines or a mixed marine-continental brine source.

Fauld Member at Fauld, Keyworth and Newark.

Fauld is situated in a Triassic graben, the Needwood Basin, whereas East Leake and Newark lie on a horst block and exhibit attenuated sequences (Audley-Charles, 1970a). Continental ground waters would have flowed into the topographically low lying basinal areas. If isotopic enrichment through evaporation occurred within the brines $\delta^{18}O$ carbonate would increase in the direction of brine movement. This could account for the East Leake - Fauld variation in the Fauld Member, with brines of continental origin flowing downslope from East Leake to Fauld. However, it is difficult to account for the Newark - Keyworth - East Leake variation with this model as Newark and Keyworth were situated "upslope" of East Leake. Therefore it is unlikely that the variations are solely due to isotopic enrichment of continental

Fig. 5.8.   Carbon and oxygen isotopic data for Upper Triassic
dolomites from the East Midlands and Blyton, Yorkshire. The
results predominantly lie in the evaporitic marine brine
dolomite field ($\delta^{18}O$ +1‰ to +5‰) or in the evaporitic
continental brine dolomite field ($\delta^{18}O$ - 4‰ to 0‰). The
Blue Anchor Formation dolomite is of lower salinity mixed
marine continental origin (see text).
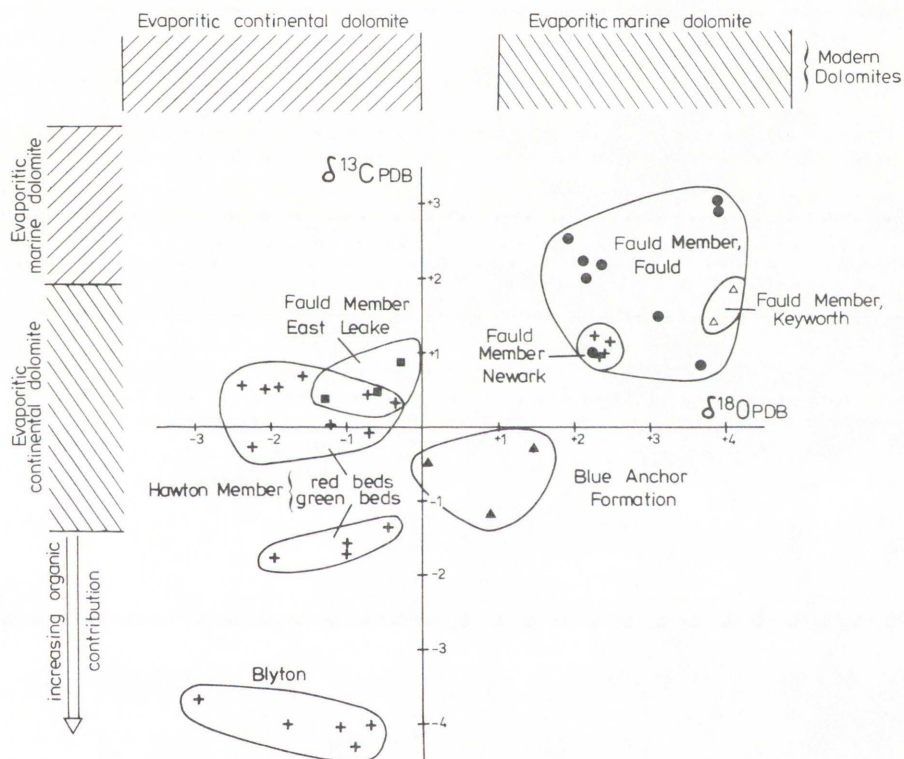
ground waters. Topographical lows would be the first areas to be influenced by marine incursions. It is proposed that marine brines flowed from the adjacent Irish Sea - Cheshire Basin into the Needwood Basin (Fig. 5.9).

The Newark and Keyworth area, although structurally higher than the East Leake area, lay close to the south-west corner of the Lincolnshire - Yorkshire Basin (Audley-Charles, 1970b). The gradient of the Lincolnshire - Yorkshire Basin margin is far shallower than the eastern margin of the Needwood Basin (Audley-Charles, 1970b). It is possible the marine brines were present in the Lincolnshire - Yorkshire Basin coeval with the deposition of the Fauld Member in the Needwood Basin. The $\delta^{18}O$ values for the Fauld Member at Newark and Keyworth suggest that this area was subject to encroachment of marine brines from the Yorkshire Basin.

There was some degree of continental input into the Needwood Basin as evidenced by the $\delta^{34}S$ values obtained from Fauld mine. If the rivers and streams flowed over the pyrite-rich Upper Carboniferous Coal Measures exposed during the Norian to the north on the flanks of the Derbyshire Dome, then the continental waters flowing into the Needwood Basin could have had a $\delta^{34}S$ value close to -12‰ (Holser's pyrite average). Figure 5.6 shows that a marine source with a $\delta^{34}S$ value of +18.0‰ contributed 82.5% of the sulphate at Fauld, whilst continental waters with a $\delta^{34}S$ value of -12‰ contributed the remaining 17.5‰.

Modern sea water contains 2710 ppm sulphate at 35‰ salinity whilst the average sulphate concentration in continental waters is only 11 ppm (Livingstone, 1963). However, in areas of restricted rainfall, e.g. Rio Grande, New Mexico and Death Valley, California, the sulphate concentration of continental waters can rise as high as 5000 ppm and commonly lies between 1000 - 2000 ppm (Livingstone, 1963). Evaporated marine brines in the Trucial Coast sabkhas have sulphate concentrations of 4000 ppm to 12,000 ppm in the intertidal and inner flood recharge zones, which decrease to between 1000 and 2000 ppm in the outer flood recharge zone and continental input zone (Butler, 1969).

Fig. 5.9.   The palaeohydrology of the East Midlands area during the

deposition of the Tutbury Sulphate Bed. Marine brines

predominated in the Needwood Basin and around Newark. The

horst area between was strongly influenced by continental

brines. Dissolution of the Hathern Anhydrite provided a

localized sulphate source on the horst area.

Known extent of Tutbury Sulphate

Possible outcrop of Hathern Anhydrite Series

10 km

minor marine input from Yorkshire Basin
+ NEWARK

runoff from Derbyshire Dome

NOTTINGHAM +

poorly developed nodular sulphate

marine input from Cheshire Basin

Fauld Mine

Needwood Basin boundary

sulphate rich

East Leake Mine

runoff from London Platform

CHARNWOOD MASSIF

sulphate poor

poorly developed nodular sulphate

+ LEICESTER

DERBYSHIRE DOME

YORKSHIRE BASIN

CHESHIRE BASIN

WELSH LANDMASS

LONDON PLATFORM

If it is assumed the marine brines flowing into the Needwood Basin were similar to those on the Trucial Coast and contained around 8000 ppm sulphate (i.e. 3x normal marine concentration), then the continental sulphate concentration required to provide 17.5% of the total sulphate input can be calculated (Table 5.3). A further variable to be considered is the relative volumes of marine and continental waters flowing into the basin as the marine input volume probably exceeded the continental input volume. This variable is expressed as a ratio of relative input volumes (Table 5.3). The concentrations of continental sulphate required for varying input volume ratios are calculated (Table 5.3). The initial pre-evaporitive continental sulphate concentration is calculated by assuming that the continental brines had been concentrated by a factor of 3 (Table 5.3). Comparison of Table 5.3 with present day sulphate concentrations in arid environments (Livingstone, 1963) indicates that the marine-continental input volume ratio for the Needwood Basin may have lain between 3:1 and 4:1.

Fauld Member, East Leake Area.

The $\delta^{18}O$ values of carbonates in the marginal East Leake area imply that they were precipitated from brines of predominantly continental origin. The $\delta^{34}S$ values, which are relatively heavy for sulphate precipitated from continental brines, indicate that much of the sulphate was derived from the Hathern Anhydrite Series. Zechstein sulphates are isotopically lighter than the Tutbury Sulphate Bed sulphate, a fact which argues against the latter being exclusively derived from the former. The areal distribution of the Tutbury Sulphate Bed in the East Leake area supports the idea of the sulphate being derived from the Hathern Anhydrite Series. The Tutbury Sulphate Bed is present only to the north and east of the Charnwood Massif. The seam thins to the east, passing into a thin, poorly-bedded nodular sulphate horizon (Fig. 5.9). The Tutbury Sulphate Bed has not been recognized by the author around Leicester, but it is thought to be represented by a poorly developed nodular sulphate horizon.

Table 5.3 Calculations of sulphate concentration of continental input such that the continental input provides 17.5% of total available sulphate at varying ratios of marine to continental volume inputs. Also calculated, the sulphate concentration of the continental brines prior to 3x evaporitic concentration.

Concentration of marine sulphate contained in 1 volume of marine input. (A)	Ratio of relative input volumes, marine: continental. (B)	Total marine sulphate available. (A x B)	Sulphate concentration of continental input to provide 17.5% of total sulphate reservoir (ppm)	Sulphate concentration of continental input prior to 3x evaporitic concentration (ppm)
8000	1:1	8000	1700	570
8000	2:1	16,000	3400	1130
8000	3:1	24,000	5100	1700
8000	4:1	32,000	6800	2270
8000	5:1	40,000	8500	2830

Table 5.4 Calculations of sulphate concentration of continental input such that the continental input provides 12.5% of total available sulphate at varying ratios of marine to continental input. Also calculated is the sulphate concentration of the continental brines prior to 3x evaporitic concentration.

Concentration of marine sulphate contained in 1 volume of marine input. (A)	Ratio of relative input volumes, marine: continental. (B)	Total marine sulphate available. (A x B)	Sulphate concentration of continental input to provide 12.5% of total sulphate reservoir (ppm)	Sulphate concentration of continental input prior to 3x evaporitic concentration (ppm)
8000	1:1	8000	1140	380
8000	2:1	16,000	2280	760
8000	3:1	24,000	3420	1140
8000	4:1	32,000	4560	1520
8000	5:1	40,000	5700	1900
8000	6:1	48,000	6840	2280

It is suggested that the Charnwood Massif controlled the ground water flow by acting as a hydrological barrier. The sulphate-rich waters which had passed over the Hathern Anhydrite Series had to flow to the north and east where the sulphate was reprecipitated as a localized deposit (Fig. 5.9). The sulphate content of the brines decreased away from the original source which resulted in a thin, poorly developed sulphate horizon surrounding the area of major sulphate development.

Hawton Member.

The $\delta^{18}O$ carbonate values for the Hawton Member indicate that continental brines were predominant in the whole area shortly after the deposition of the Fauld Member. The change in brine sources from the Fauld Member to the Hawton Member is reflected by the previously discussed change in clay mineralogy. The $\delta^{34}S$ values for the sulphate horizons can be interpreted as either purely continental brines deriving "heavy" pre-existing sulphate from exposed evaporites, or as marine brines which have been diluted by continental input. The former is unlikely as the widespread nature of individual horizons would have meant that a large evaporite source would have. to have been available to supply such vast quantities of sulphate. Such a large source of evaporitic sulphate is unknown (the Permian evaporites are too light to give sulphates of +14‰, +15‰). It is proposed that there was a steady continental input into the area which resulted in the formation of the carbonates. Periodically, there were influxes of marine water which gave rise to the sulphate horizons. The close grouping of the $\delta^{34}S$ values suggests that the degree of dilution of marine brines by continental brines was nearly constant for a long period of time. Assuming that the continental input had a $\delta^{34}S$ value of -12‰ it is estimated that 87.5% of the sulphate was of marine origin, whilst the remaining 12.5% was of continental origin (Fig.5.6). As for Fauld Mine, there is a wide range of possible sulphate concentrations in the unevaporated continental waters, (Table 5.4), e.g. 380 ppm sulphate for equal input volumes rising to 1520 ppm sulphate when marine input is 4x

continental input. Comparison with present day sulphate concentrations of waters in arid environments indicates that the marine - continental volume input ratio for the Hawton Member at times of marine influx probably lay between 4:1 and 5:1.

A single analysis of a micronodular gypsum horizon shows a significant drop in δ^{34}S (Sample 67). This could imply that the marine sulphate contribution has fallen by 5% to 82.5%. The micronodular nature of the gypsum suggests that it was deposited from brines with lower sulphate concentrations than those brines which deposited the main sulphate horizons and consequently it is interpreted as representing a minor marine influx. To illustrate the variations in brine source the δ^{34}S sulphate analyses obtained from the Hawton and Fauld Members are plotted against associated δ^{18}O carbonate analyses (Fig. 5.10). The expected field of sulphate and carbonate precipitated from evaporated Norian marine brines is shown. Also shown are the origins and mixing histories of the brines which supplied the sulphate and/or carbonate to each field of analyses.

Blue Anchor Formation.

The δ^{18}O carbonate values can be interpreted as representing either a continental water source, greatly evaporated prior to dolomitization, a mixed marine-continental water source or a purely marine water source. If the water source was of continental origin then a significant enrichment of δ^{18}O by evaporation must have occurred. The absence of sulphate and halite minerals from the Blue Anchor Formation argues against large scale evaporation so the water source was probably not of purely continental origin. The absence of a marine fauna suggests that the waters were not of purely marine origin. Therefore the available evidence, although scanty, indicates a mixed marine-continental water source for the Blue Anchor Formation. This marks a transition between the predominantly continental brine regime of the Hawton Member and the fully marine conditions of the Penarth Group.

Fig. 5.10. Fields of associated $\delta^{34}S‰$ sulphate and $\delta^{18}O‰$ carbonate analyses from the Upper Triassic of the East Midlands. Also shown is the expected field for ^{34}S sulphate and ^{18}O carbonate precipitated from evaporated marine brines. The origin and evolution of the brines supplying each field is shown. The diagram highlights the fact that minor sulphide input can effectively isotopically mask the major sulphate source. The ^{18}O dolomite analyses fall into distinct marine and continental brine fields. This enables a more accurate model of brine origins to be constructed than would be possible solely from ^{34}S sulphate analyses.
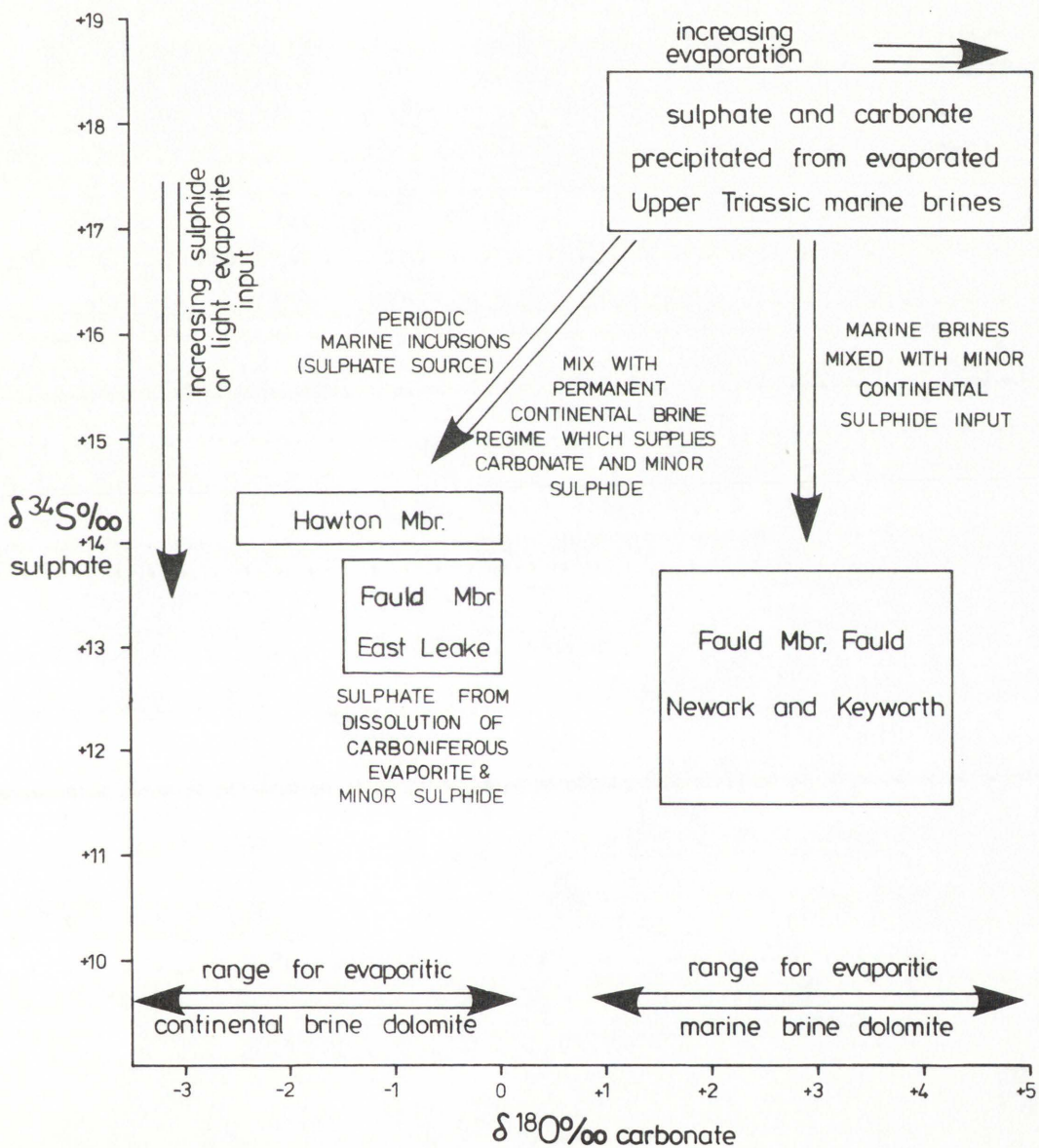
EAST MIDLANDS AREA CARBONATES - CARBON ISOTOPES.

The analyses show a broad range of values from +3‰ to -2‰ with samples from Blyton clustering around -4‰ (Fig. 5.8). The results can be divided into two groups:-

(a)   the red bed lithologies (-0.5‰ to +3‰)

(b)   the grey-green lithologies and Blue Anchor Formation (-2‰ to 0‰)

The red bed lithologies can be subdivided according to whether deposition occurred in a marine or continental brine regime, as previously noted. The strata with a marine brine source (Fauld Member) have a range of +1.0‰ to +3.0‰ with a mean of +1.7‰. The strata with a continental brine source (Hawton Member and some Fauld Member) have a range of -0.4‰ to +1.0‰ with a mean of +0.4‰. This difference could be due to the continental brines deriving increased levels of isotopically light organic carbon from surrounding landmasses.

The lacustrine grey-green lithologies are depleted in ^{13}C. The lakes may have supported the growth of micro-organisms which would have been the source of light carbon. The Blue Anchor Formation is similarly depleted in ^{13}C but to a lesser extent than the lacustrine horizons. A massive dolomite showed minor depletion (δ^{13}C of -0.32‰) whilst dolomitic shaley mudstones infilling desiccation cracks in the dolomite show more organic influence (δ^{13}C of -1.22‰). This indicates that dolomitization of the massive horizon occurred during a period of high salinity when organic productivity was low. There then followed a period of desiccation. The overlying shaley mudstones were deposited on the return of lower salinity waters when organic productivity increased.

OTHER TRIASSIC SULPHATE ANALYSES.

Cheshire Basin.

Sample RT1 supplied by R. Tucker is an anhydrite residue obtained by dissolving massively bedded halite from the Anisian Northwich Halite

Formation, Winsford. It gave a δ^{34}S value of +15.9‰. Claypool et al.(1980)

applied a correction of +0.6‰ to account for ^{34}S fractionation effects when

sulphate crystallizes in the halite facies. If this correction is applied to

RT1 it gives a value of +16.5‰ which is 2‰ to 4‰ lighter than the

estimated Anisian marine sulphate value (Fig. 5.4). However construction of

the Anisian section of the curve may be unduly biased by the 'Röt event'

(Holser, 1977) (Fig. 5.4). If the Röt data are ignored as a localized event

(Pilot et al., 1972; Claypool et al., 1980) then Anisian marine sulphate may

well have had a value between +17‰ and +18‰ (Fig. 5.4). The sulphate

facies is represented by three samples of nodular sulphate from Norian strata

near Whitchurch. These gave δ^{34}S values which cluster around +13.0‰,

markedly lighter than the estimated Norian marine sulphate value of δ^{34}S +

18‰ (Fig. 5.4).

The difference can be interpreted in terms of a variable source for the

brines. The halite and associated sulphate of the halite facies crystallized

from marine brines, whereas the isotopically lighter sulphate facies crystal-

lized from continental brines. This supports the hypotheses that the immense

volume of halite within the Cheshire Basin requires a marine source

(Evans et al., 1968; Arthurton, 1980), and that the halite-free beds which

contain minor sulphate (i.e. the sulphate facies) accumulated in a largely

continental water system (Arthurton, 1980).

Somerset Basin.

The average for the analyses of the three Norian gypsum samples from the

South Welsh coast is +12.6‰. This is 5.4‰ lighter than the average Norian

marine value and as such is strongly indicative of a continental water

source. Tucker (1977, 1978) interpreted this sequence as representing a

shore line or marginal lake facies. The isotopic evidence appears to indicate

that the large temporary lakes were formed by periods of heavy continental

run-off from the adjacent Triassic palaeolandmass.

Sample 62 from the Norian and sample 71 from the lower part of the Blue

Anchor Formation (Rhaetian) (Warrington et al.,1980) in the central part of the Somerset Basin have $\delta^{34}S$ values of +15.0‰. Sample 64 lies 9 m. above sample 71 in the Blue Anchor Formation and is significantly heavier with a $\delta^{34}S$ value of +16.6‰. The nodular gypsums from which samples 64 and 71 were obtained are part of a typical sabkha sequence of dolomites showing inter-tidal and supratidal structures interbedded with dark grey-green shales (subtidal?). The sabkha sequence probably marks the first stages of the late Triassic marine transgression across the underlying sequence of continental mudstones and siltstones. The isotope values indicate mixing of marine and continental brines in the lower part of the sequence and an increased marine influence in the middle of the sequence. Fully marine conditions were established by the Penarth Group (Mayall, 1981).

Yorkshire Basin.

The three $\delta^{34}S$ results, all of probable Norian age, from the Yorkshire Basin show a very similar trend to the Somerset Basin, i.e. two values of +14.7‰ overlain by a single value of +16.6‰ some 20 m. below the Penarth Group. Carbonate samples from 35 to 55 m. below the Penarth Group at Blyton, Yorkshire, gave $\delta^{18}O$ values which indicate a continental brine source. Therefore a history similar to that for the Somerset Basin is proposed. The basin was predominantly influenced by continental input with periodic marine input in the lower part of the analysed section. In the upper section the $\delta^{34}S$ value indicates strong marine influence, perhaps a foretaste of the Penarth Group transgression. The $\delta^{13}C$ isotope values for the Blyton "green-beds" indicate organic input. The source of organic material could have been temporary lakes which allowed algal growth.

CONCLUSIONS.

This study shows that sulphur isotopes on their own do not give a totally reliable indication of brine origin. Continental waters can have

extremely variable $\delta^{34}S$ values dependant on the rock types exposed in the catchment area. In a situation where the $\delta^{34}S$ values obtained are lighter than normal marine values, it is often impossible to estimate whether continental or marine brines dominated the sabkha. Analysis of ^{18}O / ^{16}O in associated carbonate minerals helps to resolve this problem.

MIDLANDS AREA.

During the deposition of the Fauld Member the Needwood Basin and some of the surrounding areas including Newark were invaded by marine brines. It is estimated that 82.5% of the basinal Tutbury Sulphate Bed is of marine origin. The remaining 17.5% is of continental origin derived from the weathering of pyrite-rich horizons exposed around the basin. Using the constraints applied by the sulphate concentration of modern waters in arid environments, calculations showed it is unlikely that the Needwood Basin marine-continental input volume ratio exceeded 4:1.

The marginal part of the horst area (East Leake) had a continental brine regime with the major sulphate source being the Hathern Anhydrite Series which was exposed on the northern margin of the Charnwood Massif. The Charnwood Massif controlled the hydrology with sulphate rich brine flow being confined to the north-east quadrant. The sulphate content of the brines decreased as they moved away from the sulphate source resulting in thick sulphate deposits being confined to a localized area.

During the deposition of the Hawton Member the whole area changed to a predominantly continental brine regime in which the carbonates were formed. The continental brines mixed with marine influxes which periodically occurred across a large part of the area and gave rise to thin widespread sulphate horizons. The marine-continental volume input ratio at times of marine influx probably lay between 4:1 and 5:1. Micronodular sulphate horizons represent minor marine influxes. There is isotopic evidence for the establishment of temporary lakes during this period.

The Blue Anchor Formation had a mixed marine-continental water source which marks the transition between the predominantly continental regime of the Hawton Member and the fully marine conditions of the Penarth Group.

OTHER TRIASSIC BASINS.

Sulphate from the halite horizons of the Cheshire Basin indicates a marine origin for the brines, whereas overlying sulphates in Norian mudstones are possibly of continental origin. In the Somerset Basin the results from the marginal South Wales sequence point towards a strong continental input into the basin. The central part of the Somerset Basin was subject to periodic marine input with a varying amount of continental input. A sabkha sequence in the Blue Anchor Formation marks the first stage of the marine transgression which eventually gave rise to fully marine conditions in the Penarth Group. The Yorkshire Basin had a very similar history. Sabkha cycles are not known in Yorkshire, but isotope values indicate periods of organic input, perhaps as a result of the formation of temporary lakes.

Note added in proof.

Cortecci et al. (1981) published some interesting $\delta^{34}S$ data obtained from Italian Permian and Triassic marine sulphates. Sulphates of probable Upper Triassic age show a range of values from +15 to +17.5‰ (average +16.4‰) in agreement with the results of Claypool et al. (1980) for world wide samples. Whilst these results may lend additional support for a Norian marine sulphate value of about +16.5‰, rather than +18‰ as assumed in this study, the overall concept of a mixed continental – marine brine regime for the East Midlands Triassic remains unchanged.

CHAPTER 6.

STRONTIUM ISOTOPIC COMPOSITION OF TRIASSIC SULPHATE

HORIZONS IN THE ENGLISH MIDLANDS.

INTRODUCTION.

As detailed in Chapter 5, a controversy surrounds the source of the brines from which extensive evaporite deposits formed in the Upper Triassic; previous theories ranged from marine (Evans et al.,1968; Warrington, 1970) through alternating marine-continental (Wills, 1970) to purely continental sources (Bosworth, 1913; Tucker, 1977, 1978). Study of sulphate ^{34}S and dolomite ^{13}C and ^{18}O (Chapter 5) has shown that the brine regime alternated between brines derived from a predominantly marine source and those derived from a continental source. Sulphate deposited from predominantly marine brines contained between 10 and 15% sulphate derived from a continental source (oxidation of pyrite). In the East Leake - Barrow-upon-Soar area the Tutbury Sulphate Bed was deposited in a continental brine regime which derived its high sulphate content from the dissolution of the adjacent Carboniferous Hathern Anhydrite Series (Chapter 5).

East Midlands Triassic sulphate contains between 500 and 1500 ppm strontium (Aljubouri, 1972). In gypsum the strontium forms discrete areas of celestite ($SrSO_4$) whilst it is evenly distributed by isomorphous replacement within the anhydrite lattice (Aljubouri, 1972). In the present study 16 sulphate samples were analysed for their $^{87}Sr/^{86}Sr$ ratio with the aim of providing another method of assessing the relative degree of continental and marine brine input. This study complements the S, C, O study (Chapter 5) and the reader is referred to Chapter 5 for background information on locations, stratigraphy, sedimentology and structural influences on sedimentation.

SAMPLING AND ANALYTICAL PROCEDURE.

Sixteen sulphate samples previously analysed for ^{34}S were selected. To avoid radiogenic ^{87}Sr contamination the samples were devoid of

mudstone. Twelve samples came from the East Midlands Triassic and for comparative purposes four samples came from the Somerset Basin Triassic and the Hathern Anhydrite Series. Approximately 1 gm. of samples was either drilled from the centre of sulphate nodules or, in the case of the Hathern Anhydrite Series, obtained from drill cuttings. A process of cation exchange chromatography is used to extract the strontium prior to analysis on a VG MicroMass 30 mass spectrometer for $^{87}Sr/^{86}Sr$. These were the first samples of calcium sulphate to be analysed at the Institute of Geological Sciences Geochemical Unit; the extraction process is therefore described in some detail.

0.05 to 0.1 gm. of powdered sulphate was dissolved in 5 ml. of 2.5M HCL. The process is aided by gentle heating prior to leaving to stand for 12 hours for equilibration to take place. Insoluble residue is centrifuged down and 1 ml. of the supernatant liquid is loaded onto an ion exchange column filled with Dowex 30 resin previously washed with 2.5M HCL. A small volume of 2.5M HCL is then used to wash in the solution followed by the elution of a further 25 ml. of 2.5M HCL. A further 15 ml. of 2.5M HCL is then added to the column and is collected after draining through. The liquid is evaporated to dryness and the residue dissolved in 1.5 ml. 2.5M HCL. The separation process is now repeated; 1 ml. of the dissolved residue is washed into a primed cation exchange column followed by elution with 28 ml. of 2.5M HCL. 12 ml. of 2.5M HCL is then added, collected and evaporated to dryness. The double separation process is required to completely remove Ca^{2+} from the sample. Mass spectrometer filament beads are prepared by dissolving the final residue in a drop of deionized water and transferring this to a filament loaded with a drop of phosphoric acid. The filament is heated, boiling off the liquid and partially de-gassing the sample, prior to loading in the mass spectrometer.

RESULTS.

The results, corrected to an Eimer and Amend $SrCO_3$ value of 0.70800 standard are shown in Table 6.1. Also given are $\delta^{34}S$ values for each sample (Chapter 5).

Table 6.1   Strontium and sulphur isotope data of gypsum and anhydrite.

Sample No.	Locality and Horizon.	$^{87}Sr/^{86}Sr$	$\delta^{34}S\%_o$
17	Newark,Hawton Member,13.5m below Penarth Group.	0·70870 ± 0·00009	+14·1
11	Newark,Hawton Member,17.8m below Penarth Group.	0·70891 ± 0·00008	+14·2
3	Newark,Hawton Member,23.4m below Penarth Group.	0·70916 ± 0·00010	+14·4
30	N.Leicester,Hawton Mbr,24m below Penarth Group.	0·70890 ± 0·00007	+14·4
19	East Leake, Fauld Member, Tutbury Sulphate.	0·70911 ± 0·00007	+13·8
46	East Leake, Fauld Member, Tutbury Sulphate.	0·70924 ± 0·00007	+12·8
45	East Leake, Fauld Member, Tutbury Sulphate.	0·70907 ± 0·00009	—
29	N.Leicester, Fauld Member, Tutbury Sulphate.	0·70911 ± 0·00007	+13·8
21a	Fauld, Fauld Member, Tutbury Sulphate.	0·70916 ± 0·00007	+13·2
25	Fauld, Fauld Member, Tutbury Sulphate.	0·70920 ± 0·00008	+12·3
73	Fauld, Fauld Member, Tutbury Sulphate.	0·70914 ± 0·00007	+12·2
77	Fauld, Fauld Member, Tutbury Sulphate.	0·70922 ± 0·00007	+12·7
H4	Hathern borehole,depth 1937ft.Hathern Anhydrite	0·70807 ± 0·00007	+19·3
H5	Hathern borehole,depth 2038ft.Hathern Anhydrite	0·70810 ± 0·00007	+19·6
62	Somerset, 58m below Penarth Group.	0·70938 ± 0·00007	+15·0
64	Somerset,Blue Anchor Fm.15m below Penarth Group	0·70906 ± 0·00007	+16·6

DISCUSSION.

STRONTIUM SOURCES AND ISOTOPIC RATIOS.

(a) Marine waters.

Strontium in the oceans originates primarily from the weathering of:-

(1) old igneous and metamorphic rocks and of sedimentary

material derived from them;

(2) young volcanic rocks of basaltic or andesitic composition;

(3) marine carbonate rocks.

Each source contributes strontium of differing isotopic composition

dependent on age and Rb/Sr ratio (Faure and Powell, 1972; Faure, 1977;

Faure et al., 1978). Old sialic rocks are enriched in radiogenic Sr^{87} and

therefore have elevated $^{87}Sr/^{86}Sr$ ratios whereas young volcanic rocks,

such as those extruded at mid-ocean ridges, typically have low $^{87}Sr/^{86}Sr$

ranging from 0·7020 to 0·7060. The $^{87}Sr/^{86}Sr$ ratios of marine limestones

range from about 0·7065 to 0·7090, the latter being the value of modern

seawater (Faure and Powell, 1972; Faure, 1977; Faure et al., 1978). The

$^{87}Sr/^{86}Sr$ ratio of seawater is dependent on the relative proportions of

strontium contributed by the above three sources. Past changes in the

relative contributions resulted in the $^{87}Sr/^{86}Sr$ ratio of marine water

varying through time. As with sulphur, various "age-curves" of the

$^{87}Sr/^{86}Sr$ ratio of seawater through time have been constructed from analyses

of $^{87}Sr/^{86}Sr$ ratios of marine carbonate rocks free from clay minerals

(Peterman et al., 1970; Faure and Powell, 1972; Faure,1977; Faure et al.,

1978). The curve of Faure et al. (1978) (Fig. 6.1) dealing specifically

with the Trias and Lower Jurassic shows that the $^{87}Sr/^{86}Sr$ marine seawater

ratio drops constantly from the middle Norian to Rhaetian (Fig. 6.1).

Faure et al. (1978) attach average values of $^{87}Sr/^{86}Sr$ 0·70791± 0·00014

and 0·70784± 0·00028 to the Rhaetian and Norian stages respectively. For

the purposes of this study a value of 0·70800 appears to be the best estimate

for the marine water $^{87}Sr/^{86}Sr$ ratio in the upper Norian.

Fig. 6.1.   Solid line depicts the estimated $^{87}Sr/^{86}Sr$ rate of marine

waters from the Middle Triassic to the Lower Jurassic. The

stippled box covers the range of values obtained in the

present study ($^{87}Sr/$ ^{86}Sr ratio curve based on Faure et al.,

1978).

Fig. 6.2.   Plot of $\delta^{34}S$ against $^{87}Sr/$ ^{86}Sr for Upper Triassic sulphates

of the East Midlands. Also shown are the estimated

composition ($\delta^{34}S$ vs. $^{87}Sr/$ ^{86}Sr) of Upper Triassic marine

water and analyses of the Hathern Anhydrite Series. The

effects of increasing continental input are to decrease

(in general) $\delta^{34}S$ and increase $^{87}Sr/$ ^{86}Sr ratio of the

brines.

Top figure x-axis labels:

Pl	Si	Ht	Rh&Nr	Ca	Ld	An
Lwr. Jurassic			Upper Triassic		Mid. Triassic	
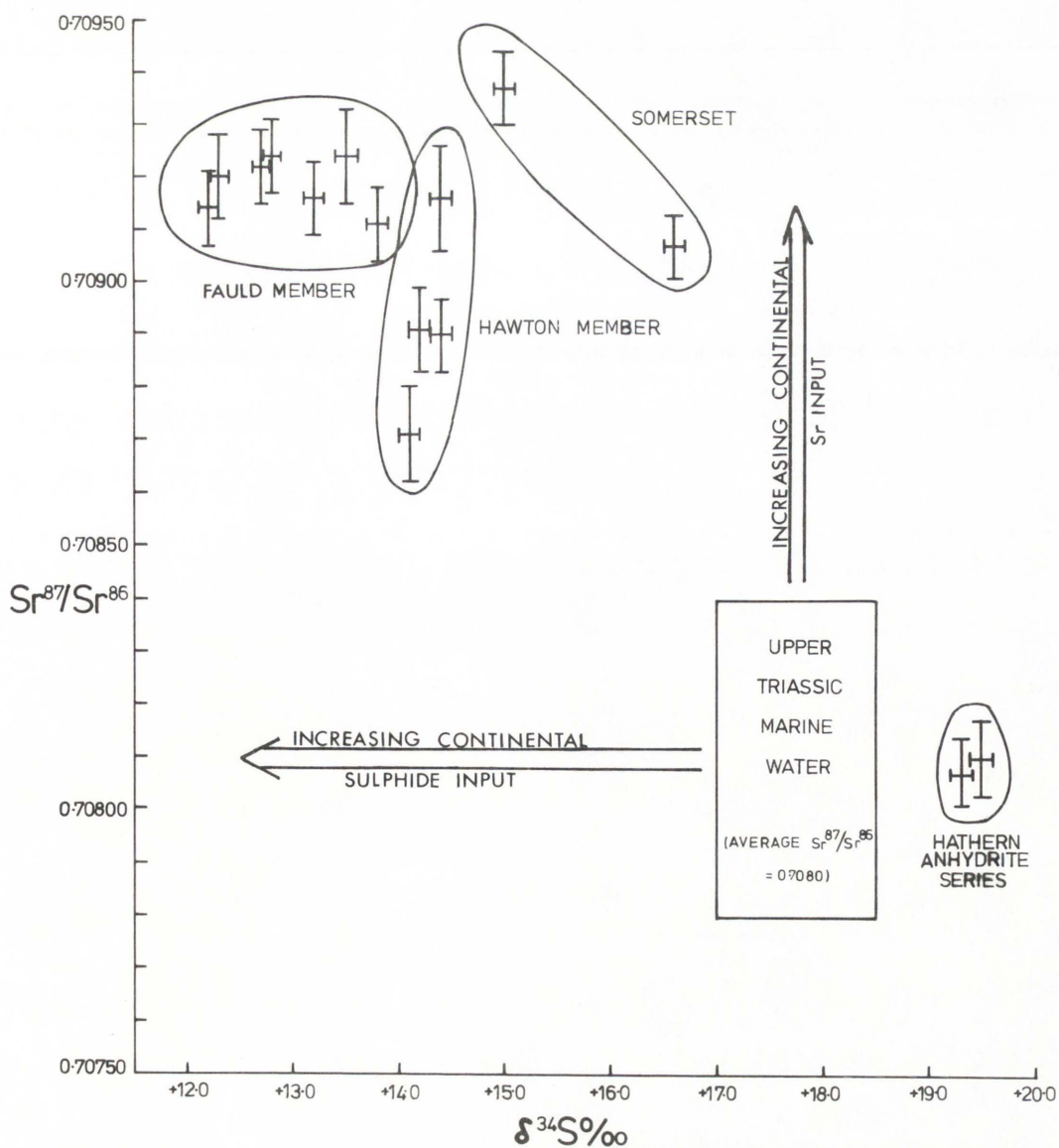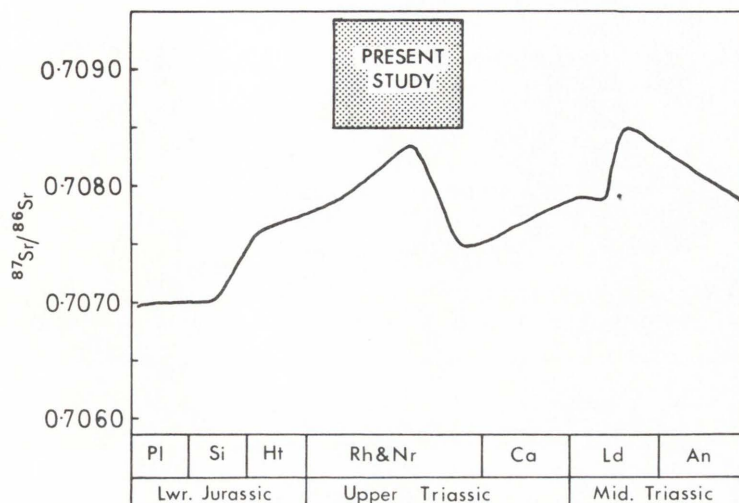
(b) <u>Continental waters</u>.

There is no evidence of Norian volcanism in the British Isles; the nearest contemporaneous volcanics are in S. Europe and eastern N. America (Hallam and Sellwood, 1976). Strontium would therefore only have been derived from sources (1) and (3), old igneous and metamorphic rocks and marine carbonates. Igneous, metamorphic and clastic rocks form much of the pre-Upper Permian basement of England (Wills, 1971) and they probably continued to dominate Triassic landmasses. Therefore, Triassic continental run-off in the area probably contained strontium with elevated $^{87}Sr/^{86}Sr$ ratios, derived from the weathering of old crustal rocks. Marine carbonates probably only provided a minor amount of strontium, possibly via weathering of Carboniferous limestone. The author assumes a value of 0·7200 for the $^{87}Sr/^{86}Sr$ ratio of continental input; a value previously assumed to be the "average ratio" of old sialic crust (excluding non-clastic material) (Faure, 1977; Faure <u>et al</u>., 1978).

Consideration should be made of the case where evaporite basins derive substantial volumes of sulphate from older sulphates exposed adjacent to the basins. The $^{87}Sr/^{86}Sr$ ratio of the brines and eventual secondary sulphate deposit would be strongly influenced by the initial $^{87}Sr/^{86}Sr$ ratio of the primary sulphate.

(c) <u>Anhydrite - gypsum</u>.

Gypsum generally has lower strontium concentrations than anhydrite, the excess strontium being accommodated as discrete celestite within the gypsum (Aljubouri, 1972). Although dolomite contains significantly less strontium than the precursor calcite, the $^{87}Sr/^{86}Sr$ ratios of dolomite are compatible with those of the calcite precursor (Faure <u>et al</u>., 1978). By analogy to the calcite-dolomite system the author suggests that the sulphate hydration-dehydration process does not affect the $^{87}Sr/^{86}Sr$ ratio of the sulphate; a similar conclusion was reached with regard to fractionation of sulphur isotopes (Holser and Kaplan, 1966; Chapter 5).

DISCUSSION OF RESULTS.

The $^{87}Sr/^{86}Sr$ ratio and $\delta^{34}S$ value of each sample are plotted in
Fig. 6.2. Also shown is the estimated isotopic compositional field of
Norian marine brines. It is evident that all the Triassic samples are
enriched in ^{87}Sr compared to Triassic seawater, but that they all lie
substantially below the 0·7200 ratio of continental crust. The mixing of
the two end-members (continental and marine) provides a ready solution to
the strontium isotope data. However, it is pertinent to recall the proposal
that the Tutbury Sulphate Bed at East Leake is derived from the Hathern
Anhydrite Series (Chapter 5). Samples from the Hathern Anhydrite Series
had an average $^{87}Sr/^{86}Sr$ ratio of 0·70809 in agreement with the marine
water "age-curve" for the Lower Carboniferous (Faure et al.,1978). This
value is substantially below the average $^{87}Sr/^{86}Sr$ ratio of 0·70913 obtained
for the Tutbury Sulphate Bed in the East Leake area. However, continental
brines running off the Charnwood Massif very possibly derived minor amounts
of strontium enriched in ^{87}Sr from the Precambrian and Palaeozoic rocks
exposed in the area. Calculations indicate that a source with a $^{87}Sr/^{86}Sr$
ratio of 0·7200 need contribute less than 10% of the total strontium to
account for the variation between the Hathern Anhydrite Series "source" and
the Tutbury Sulphate Bed "product". Interestingly the associated sulphur
isotope values show that around 15% of the sulphate was derived from
oxidation of pyrite (Chapter 5). This small discrepancy easily lies within
the errors associated with this type of calculation.

As previously mentioned, the mixing of marine and continental waters
provides a ready answer for the remainder of the strontium isotope data.
Considering the data for the Tutbury Sulphate Bed at Fauld Mine,
calculations show that 10% of the strontium was derived from a continental
source, whilst about 20% of the sulphate was derived from continental
oxidized pyrite. Although the "end-member" values were originally
arbitrarily fixed, the overall agreement of the percentages of sulphate and
strontium derived from various sources supports the theories of wholly

continental input in the East Leake area and mixed marine-continental

input in the Fauld Mine area (Chapter 5).

Results from the Hawton Member show a steady decrease in $^{87}Sr/^{86}Sr$

ratios and $\delta^{34}S$ values from the base to the top of the member (Table 6.1,

Fig. 6.2). Analyses from the base of the Hawton Member show close affinity

to Fauld Member analyses. The steady decrease in the $^{87}Sr/^{86}Sr$ ratio may

reflect decreasing continental input, although if this were the case the

$\delta^{34}S$ value should gradually increase. This relationship is well illustrated

by the two samples from the Somerset Basin which indicate increasing marine

influence towards the top of the Mercia Mudstone Group (Table 6.1, Fig.6.2).

Calculations show that continental input provided around 10% of both

strontium and sulphate in the older redbed sample and about 10% of strontium

and 5% of sulphate in the Blue Anchor Formation samples.

CONCLUSIONS.

$^{87}Sr/^{86}Sr$ ratios of East Midlands Norian gypsum and anhydrite indicate

that strontium was not wholly derived from a marine source. The Tutbury

Sulphate Bed at Fauld Mine, the Hawton Member sulphates and the Somerset

Basin sulphates were deposited in a predominantly marine brine regime with

minor input of continental brines containing strontium enriched in ^{87}Sr.

The Tutbury Sulphate Bed at East Leake, although on the basis of strontium

isotopes is indistinguishable from the Tutbury Sulphate Bed at Fauld Mine,

was deposited in a wholly continental brine regime (Chapter 5). At East

Leake much of the sulphate and strontium was derived from the dissolution

of the Hathern Anhydrite Series. The $^{87}Sr/^{86}Sr$ ratio of the Hathern

Anhydrite Series is in agreement with the marine water "age curve" for the

Lower Carboniferous. Minor strontium enriched in ^{87}Sr was derived from

weathering of exposed Precambrian and Palaeozoic rocks. The Somerset Basin

sulphates show increasing marine influence towards the top of the Blue

Anchor Formation.

Calculations of the percentage input of continental and marine waters gave mixing ratios similar to those obtained from sulphur isotope data. The small discrepancies lie within the errors associated with this type of calculation.

CHAPTER 7.

TRIASSIC CLAY ASSEMBLAGES IN THE ENGLISH MIDLANDS

AND THEIR ORIGIN.

INTRODUCTION.

The Mercia Mudstone Group of England consists predominantly of red-bed mudstones with intercalated siltstones and sandstones. In the Midlands area the strata contain abundant deposits of evaporitic calcium sulphate whilst fine-grained carbonate material is frequently dolomitized. Sulphate was derived from both marine waters and the dissolution of older, exposed evaporites by continental groundwaters; deposition occurred in an arid hypersaline environment (Chapter 5). The sediments contain abundant fine-grained silicate minerals which have been the subject of numerous studies (Stephen and MacEwan, 1950, 1951; Honeybourne, 1951; Keeling, 1956; Dumbleton and West, 1966; Davis, 1967; Jeans, 1978). It is well documented that the Mercia Mudstone Group of Britain and equivalent strata in France, Germany and N. Africa contain a suite of magnesium-rich clays which include sepiolite, palygorskite, corrensite, smectite, chlorite and irregular mixed-layer minerals (Jeans, 1978; Lucas, 1962; Lucas and Ataman, 1968; Millot, 1970; Lippmann and Savascin, 1969; Schlenker, 1971). Jeans (1978) proposed that the magnesium-rich clays suite in Britain originated by neoformation, disagreeing with the French (Lucas, 1962; Millot, 1970) and German (Lippmann and Savascin, 1969) schools who proposed that some clays originated by transformation and others by neoformation.

In the present work standard X-ray diffraction techniques were used to study the clay mineralogy of the Trent Formation in the North and East Midlands in an attempt to:

1) provide a comprehensive record of clays present and relate this to previous studies of the area,

2) see how the clay mineralogy of this area compares with that of other areas of the British Triassic,

3) relate the clay mineralogy to environment of deposition,

4) elucidate whether the clays were of a wholly neoformational origin or were of a mixed transformation-neoformational origin.

STRUCTURE AND NOMENCLATURE OF CLAY MINERALS.

NON-SHEET SILICATE MINERALS.

This group has a double chain silicate structure. The linked chains form a 2:1 layer that is continuous in one direction (X) and of limited extent along the perpendicular direction (Y). The end members of this group are palygorskite and sepiolite, the structures of which are shown in Fig. 7.1.

SHEET SILICATE MINERALS.

Clays are predominantly layer silicate minerals which contain both tetrahedral and octahedral sheets. The tetrahedral sheets consist of individual tetrahedra linked with neighbouring tetrahedra by the sharing of three basal oxygens to form a hexagonal mesh pattern (Brindley and Brown,1980) (Fig. 7.2a). The octahedral sheets consist of individual octahedra linked laterally by shared octahedral edges. The octahedral sheet may be either dioctahedral when two in every three octahedra contain a trivalent cation or trioctahedral when all the octahedral cation sites are occupied by divalent cations. The octahedral and immediately adjacent tetrahedral sheets are linked by sharing the fourth tetrahedral oxygen (Brindley and Brown, 1980) (Fig. 7.2b). Also present along this boundary are unshared OH^- groups associated with the tetrahedral layer. When one tetrahedral sheet and one octahedral sheet are linked together the structure is called a 1:1 layer (Fig. 7.2c). A 2:1 layer contains two tetrahedral sheets, the upper of which is inverted, with an intermediate octahedral sheet (Fig. 7.2c). If the 1:1 or 2:1 layers are not electrostatically neutral the excess layer charge is neutralized by various interlayer materials including cations, hydrated cations, hydroxides and octahedral groups and sheets (Brindley and Brown, 1980).

Clays may be effectively classified on the basis of layer type (1:1 or 2:1), layer charge and type of interlayer. Further subdivision is made on the basis of the octahedral sheet type and chemical composition (Brindley and

Fig. 7.1a.   (100) projection of palygorskite structure (Brindley and
              Brown, 1980).

Fig. 7.1b.   (100) projection of sepiolite structure (Brindley and
              Brown, 1980).

| ◯ H₂O | ◯ HYDROXYL | ● Mg or Al |
| ◖ OH₂ | ○ OXYGEN | ▲ SILICON |

| ● Si | ○ O | ⊕ H₂O cryst |
| ⊘ Mg | ⊙ OH | ⊚ H₂O zeol |
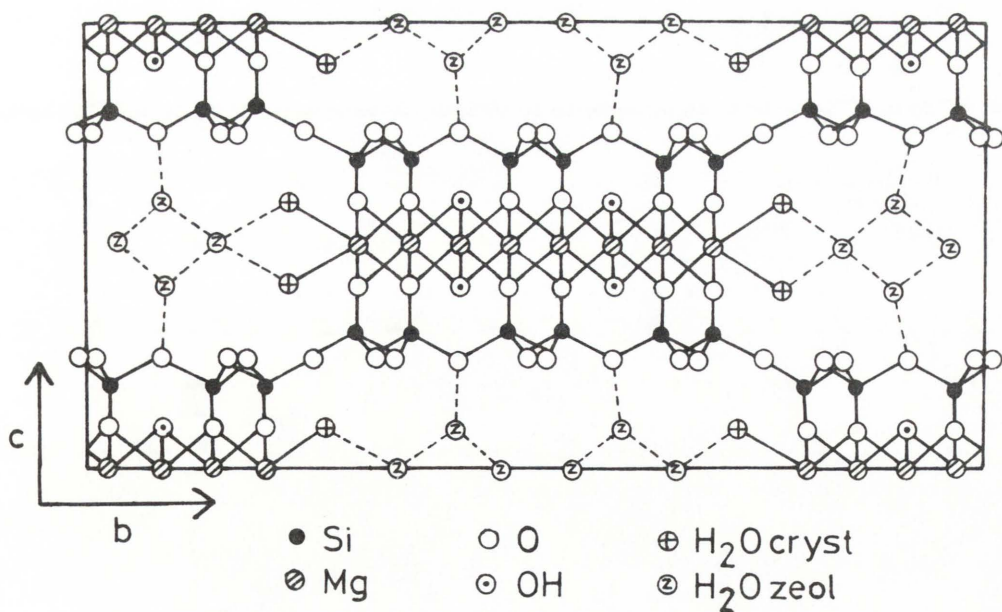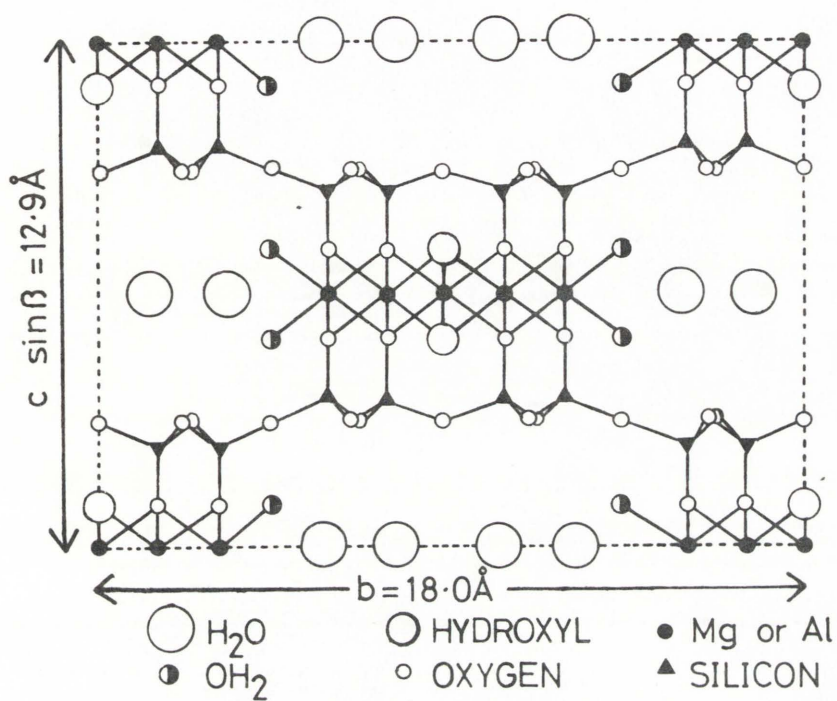
Fig. 7.2.   (a) Plan view of ideally hexagonal tetrahedral sheet

(Brindley and Brown, 1980).   (b) Octahedral sheet,

octahedral cation shaded (Brindley and Brown, 1980).

(c) Junction of sheets to form 1:1 and 2:1 layers

(Brindley and Brown, 1980).

Fig. 7.3.   (010) view of structures of major sheet clay mineral groups.

Layer charge per formula unit = X (Brindley and Brown, 1980).

a)

b)

c)

Tet. Sheet
Oct. Sheet
Tet. Sheet
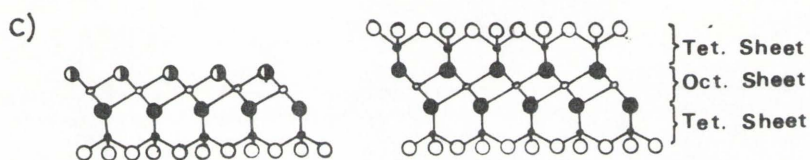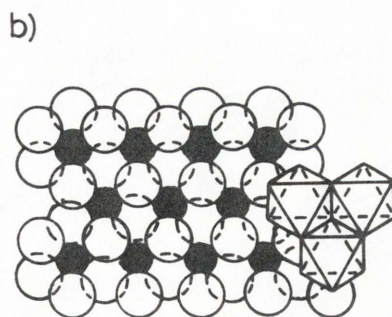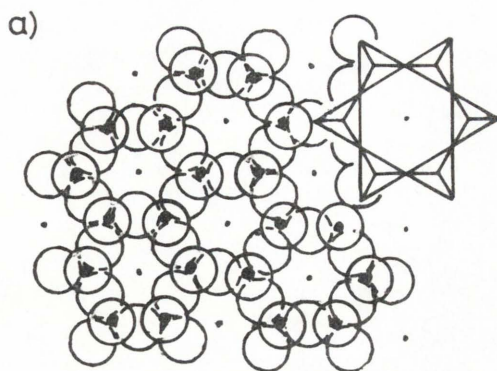
• Tetrahedral cation
○ Octrahedral cation

○ Oxygen
● Oxygen + Hydroxyl (in projection)
◐ Hydroxyl group

c sinß=
7·1–7·3Å

Serpentine–Kaolin
(x~0)

9·1–9·4Å

Talc–
Pyrophyllite
(x~0)

9·6–10·1Å
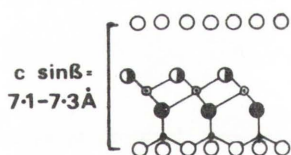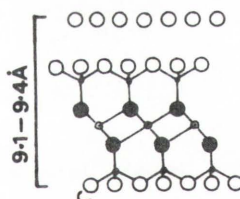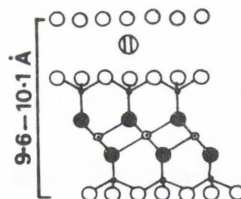
Mica(x~1·0) &
Brittle mica(x~2·0)

14·4–15·6Å
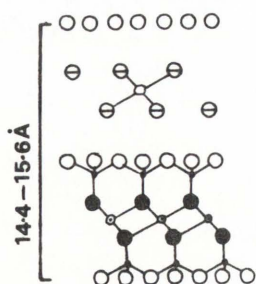
Smectite(x–0·25–0·6)
& Vermiculite(x~0·6–0·9)

14·0–14·4Å
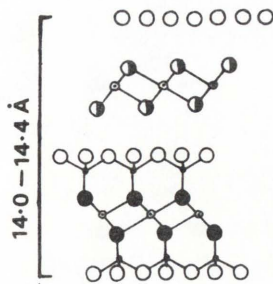
Chlorite
(x is variable)

• Tetrahedral cation
● Octahedral cation
◐ Interlayer cation
o Exchangeable cation

○ Oxygen
◐ Hydroxyl group
⊖ Water molecule
● Oxygen + Hydroxyl
(in projection)

Brown, 1980) (Table 7.1). The structures of the major layered clay mineral groups are shown in Fig. 7.3 (Brindley and Brown, 1980).

TABLE 7.1

Layer type	Group (x = charge per formula unit)	Sub-group
1:1	Serpentine – kaolin (x 0)	Serpentines Kaolins
2:1	Talc – pyrophyllite (x 0)	Talcs Pyrophyllites
	Smectite (x 0.2 - 0.6)	Saponites Montmorillonites
	Vermiculite (x 0.6 - 0.9)	Trioctahedral vermiculites Dioctahedral vermiculites
	Mica (x 1.0)	Trioctahedral micas Dioctahedral micas
	Brittle mica (x 2.0)	Trioctahedral brittle micas Dioctahedral brittle micas
	Chlorite (x variable)	Trioctahedral chlorites Dioctahedral chlorites Di, trioctahedral chlorites
2:1 inverted ribbons	Sepiolite-palygorskite (x variable)	Sepiolites Palygorskites

Classification of phyllosilicates related to clay minerals (from Brindley and Brown, 1980).

MIXED-LAYER CLAY MINERALS.

The sheet phyllosilicate layers have strong internal bonding and weak sheet to sheet bonding (Brindley and Brown, 1980). The basal surfaces of different kinds of layers are geometrically very similar with oxygen or hydroxyl ions in quasi-hexagonal array (Brindley and Brown, 1980). Consequently layers with different internal arrangements can stack together along a line normal to (001) to form a structure termed an interlayered

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or mixed-layer clay. Mixed-layer clays are composed of two, or less frequently, three component clay minerals. Interlayering can be (1) random, for which no discernable pattern exists in the sequence of layer types; (2) ordered, in which layers are stacked in a periodic pattern; or (3) partially ordered, which is intermediate between the two extremes. In this study the distinction between random and partial ordering has not been made, so the term irregular is used for any disordered mixed-layer clay. The most commonly encountered layer types involved in mixed-layer clays are chlorite, illite, smectite and vermiculite.

The polynomial system of nomenclature for random or partially ordered mixed-layer clay minerals (Brown, 1955) has been followed. For example, a smectitic chlorite is an irregular mixed-layer structure consisting of smectite layers and chlorite layers, the latter predominating; a smectite-chlorite is an irregular mixed-layer structure in which smectite and chlorite layers are present in approximately equal proportions. The use of an oblique stroke between two mineral names, e.g. smectite/chlorite, refers to an irregular mixed-layer mineral in which the proportions of the two minerals are not defined.

SAMPLING.

Samples from the following localities were analysed.

1) <u>Newark</u>:- samples taken from a British Gypsum borehole (SK 8276 5285) 3 km. E.S.E. of Newark (Fig. 7.4). The core commenced within the Glen Parva Formation, 4.5 m. below the Blue Anchor Formation and terminated 20.5 m. lower within the Trent Formation (Figs. 7.5 and 7.6) The upper 6 m. of core was not sampled as it showed indications of leaching. Samples were taken at an average interval of 0.5 m. to the base of the core.

2) <u>Keyworth</u>:- samples taken from Institute of Geological Sciences borehole (SK 6235 3175) 0.5 km. N.W. of Keyworth (Fig. 7.4). It penetrated the base of the Glen Parva Formation, the Trent Formation and the uppermost part of the Edwalton Formation (Figs. 7.5 and 7.6). The Glen Parva Formation

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Fig. 7.4.   Outline of the geology of the study area and sampling

localities for the present study and for Jeans (1978).

Geological linework taken from Institute of Geological

Sciences geological maps.

Newark
● BOREHOLE ★

Tutbury
●

★ FAULD MINE

★ BOREHOLE

KEYWORTH
★★ BOREHOLE

East
Leake ★ ★ EAST LEAKE MINE
★ BOREHOLE

CHARNWOOD
MASSIF

Leicester
●

CROFT
▲

KEY

Pre-Mercia
Mudstone Group

Mercia Mudstone Group

Post Mercia Mudstone Group

★ Localities in this study

▲ Localities in Jeans (1978)

10km
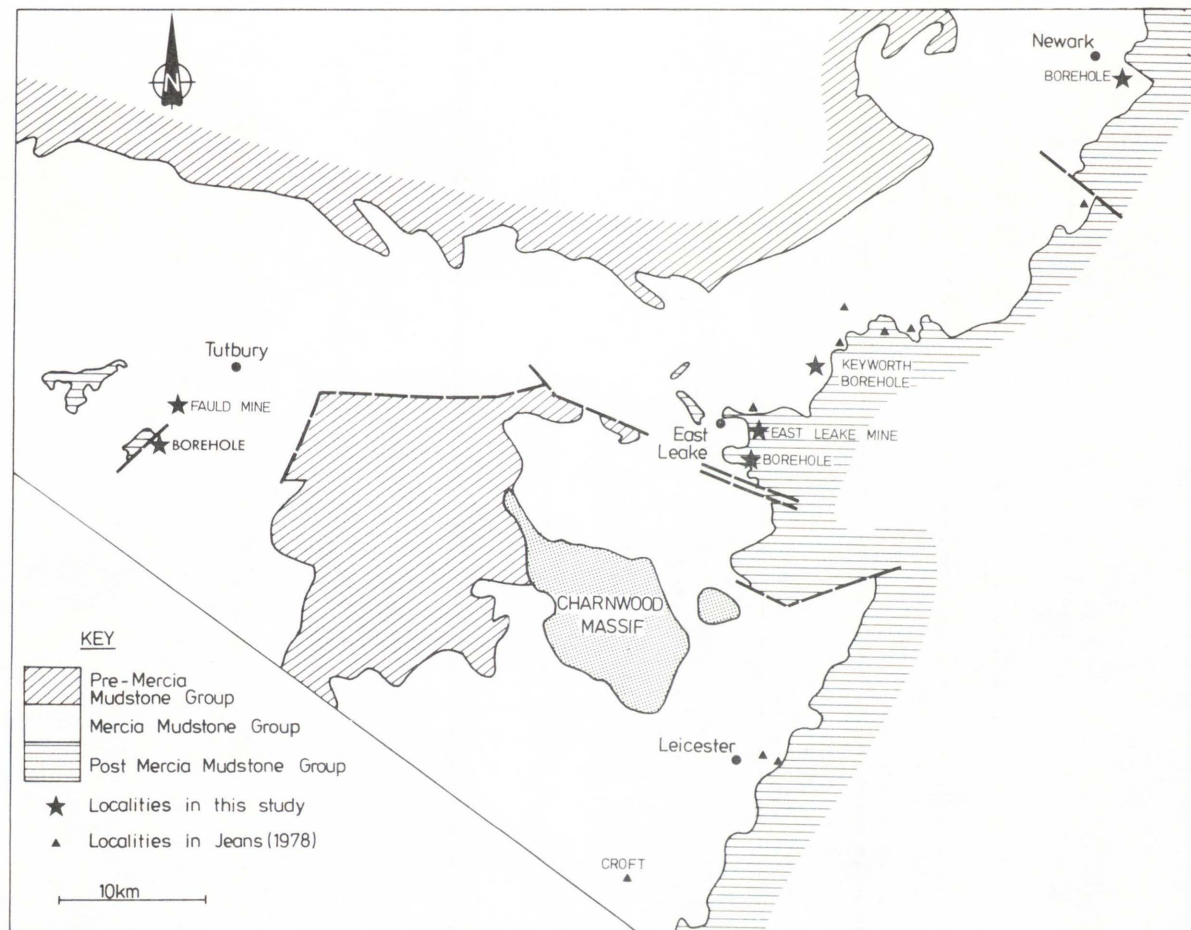
Fig. 7.5.   Stratigraphical units of the East Midlands late Triassic

(after Warrington et al., 1980 and Elliott, 1961).

PENARTH GROUP		
MERCIA MUDSTONE GROUP	BLUE ANCHOR FORMATION	
	GLEN PARVA FORMATION	
	TRENT FORMATION	HAWTON MEMBER
		Tutbury Sulphate Bed
		FAULD MEMBER
	EDWALTON FORMATION	Hollygate Skerries
		Cotgrave Skerries
	HARLEQUIN FORMATION	

Fig. 7.6.   Stratigraphic intervals sampled for localities shown on

Fig. 7.4.

FAULD
BOREHOLE

FAULD MINE
AND MINE
BOREHOLE

EAST LEAKE
MINE AND
BOREHOLE

KEYWORTH

NEWARK

PENARTH
GROUP

BLUE ANCHOR
FORMATION

GLEN PARVA
FORMATION

HAWTON
MEMBER

TRENT FORMATION

FAULD
MEMBER

EDWALTON
FORMATION

MINE

MINE

BOREHOLE

KEY

Tutbury Sulphate
Bed

Sampled strata

m ⌈ 0
  ⊢ 5
  ⌊ 10
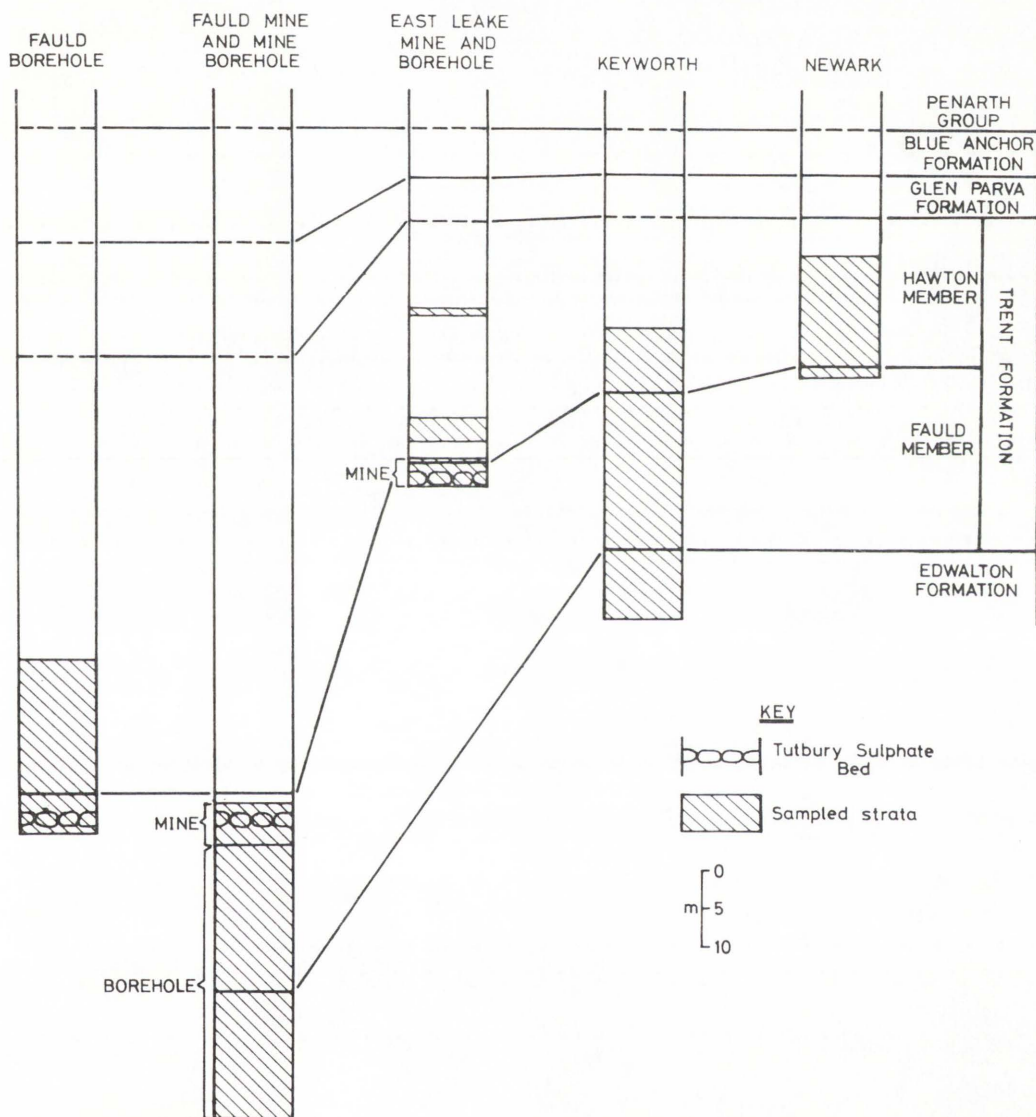
and uppermost 14 m. of the Trent Formation were very badly leached. Sampling started 1.3 m. below the leached strata and covered 38 m. of strata, terminating below the Hollygate Sandstones in the top of the Edwalton Formation (Figs. 7.5 and 7.6).

3) <u>East Leake:</u> - British Gypsum borehole (SK 5796 2485), 3 km. S.E. of East Leake (Fig. 7.4). The borehole passed through the Blue Anchor Formation, ? Glen Parva Formation and the upper 43 m. of the Trent Formation. Samples were taken at depths of 17.73 m., 17.98 m., 31.52 m. and 34.64 m. below the base of the Blue Anchor Formation (Fig. 7.6). The Tutbury Sulphate Bed was encountered 38.57 m. below the Blue Anchor Formation (Figs. 7.5 and 7.6).

East Leake Mine, East Leake. Two sample sets were taken at (SK 5808 2718) and SK 5808 2665). Each set sampled a mine pillar from the roof to the floor and included the Tutbury Sulphate Bed (Fig. 7.6).

4) <u>Fauld</u>:- British Gypsum borehole (SK 1631 2472), 6 km. S.W. of Tutbury (Fig. 7.4). The borehole penetrated the Trent Formation, terminating just below the Tutbury Sulphate Bed. Samples were taken from 20 m. above the Tutbury Sulphate Bed to the base of the hole (Fig. 7.6).

Fauld Mine (SK 181 283) 3 km. W.S.W. of Tutbury (Fig. 7.4). Four sample sets were taken, each of which sampled 2.70 m. of strata between the mine roof and floor. In addition one set sampled 2.70 m. of strata exposed in a drivage below the normal floor level (Fig. 7.6).

A borehole in the drivage floor (SK 163 256) penetrated 36 m. of strata, the lower 17 m. of which lie in the Edwalton Formation (Fig. 7.5). The borehole was sampled from 2 m. below the mine floor to its base (Fig. 7.6).

Jeans (1978) analysed samples from the following localities within the study area (Fig. 7.4). Brackets contain analysed horizons:

Croft Quarry (SP 512 965), Gypsy Lane brick pit, Leicester, SK 616 070 (Hawton Member, Glen Parva Formation); New Star brick pit, Leicester SK 616 075 (Hawton Member); Bunny brick pit, near East Leake, SK 581 286 (Hawton Member); Cropwell Bishop gypsum pit, SK 675 358 (Hawton Member); Staunton-in-the-Vale gypsum pit SK 804 445 (Hawton Member, Glen

Parva Formation); Windmill Hill marl pit, Cotgrave, SK 644 358 (Fauld Member); Owthorpe borehole SK 666 325 (Blue Anchor, Glen Parva Formations); Clipston borehole, SK 638 333 (Trent, Edwalton Formations).

METHOD

Sample Preparation.

Whenever possible, samples were taken from sulphate-free mudstones and siltstones. Seventy-five to one hundred grams of each sample were crushed fine enough to pass through a 60 mesh sieve. The sample was then shaken with 1% sodium hexa-metaphosphate on a mechanical shaker for two hours. The sample was then allowed to stand for six hours at room temperature, after which only particles of less than 2 μm. effective settling diameter remained in suspension. A 40 ml. sample of this suspension was pipetted off and flocculated with a few drops of concentrated calcium chloride solution (480 g^{-1}l). After centrifuging for ten minutes the resulting clear super-natant liquid was poured off and two smears of the compacted clay were made on glass slides using a piece of flexible perspex. The smears were dried in a desiccator and the resulting orientated aggregates used for X-ray analysis. Any sulphate rapidly broke down to a fine powder when the samples were crushed and remained in the clay fraction. The presence of sulphate did not seriously affect the clay diffraction patterns.

It should be mentioned that the first thirty samples were prepared by a different method. Initially samples were treated with cold 2N. acetic acid to remove any carbonate material. As soon as the carbonate had dissolved the samples were washed repeatedly with distilled water prior to treatment with sodium hexa-metaphosphate. However, dolomite took two weeks to dissolve and so the acid treatment was abandoned as the presence of dolomite in the clay fraction did not affect the diffraction patterns obtained for the clay minerals.

After dispersion with sodium hexa-metaphosphate a few millilitres of the

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suspended clay fraction were drawn off and dropped on to two round glass slides to form large globules. These were allowed to dry, resulting in a thin film of clay being deposited on the glass surface. However, the clay films frequently cracked and curled rendering them useless for X-ray diffraction. Various methods, including grinding the glass surface prior to loading with clay suspension, were used to try and overcome this problem, without any great success. The sample holder in the diffracto-meter was then changed to accommodate glass slides. Very few of the clay smears cracked and none curled, so this latter method proved highly satisfactory.

X-Ray Analysis.

Analysis by X-ray diffraction of the orientated aggregates of the clay fraction was carried out on a Phillips PW 1050/25 diffractometer and X-ray generator in the Department of Geology, Leicester University.

The following machine settings were used: generator, 20mA, 40kV, to produce Cu Kα radiation which was passed through a nickel filter. Diffractometer, 1° divergence slit, and 0.2 mm. receiving slit; rate of rotation $\frac{1}{2}^{\circ}$ 2 /min; chart rate 1 cm/min; time constant 4 s., counting rate 100, 200 or 400 counts s^{-1} so as to obtain maximum peak height.

Three runs were made for each sample before and after various chemical and thermal treatments. One untreated smear slide was run from 2° to 35° 2θ and then again after being heated over pre-heated glycerol in an enclosed vessel for 2 to 2$\frac{1}{4}$ hours at 110° C. This treatment causes minerals containing swelling interlayers to expand. The other smear slide was heated for 2 hours in a muffle furnace at 440° C., cooled and then X-rayed from 2° to 15° 2θ Selected samples then had this second smear slide reheated to 550° C. for 2 hours, cooled and X-rayed.

Analysis of the three or four diffractometer traces obtained for each sample enabled in particular the basal 001 spacings of the different clay mineral groups to be identified using mineralogical data given by Brown (1961),

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Jeans (1978) and Brindley and Brown (1980). Identification was undertaken to group level only; further differentiation would require analysis of randomly orientated clay powders over a large 2θ range of 2° to 65° (Brindley and Brown, 1980).

CLAYS RECOGNIZED IN THIS STUDY.

The criteria used to identify each clay mineral group recorded in this study are considered below.

Illite: - is the group name for clay-size micaceous components in argillaceous sediments (Brindley and Brown, 1980; Grim, Bray and Bradley, 1937). Illite is identified by the 001 peak between 9.8Å and 10.0Å in the untreated sample which is unaffected by glycerolation or heat treatment (Fig. 7.7). Illite 002 and 003 peaks occur at 5.0Å and 3.33Å respectively.

Chlorite:- is identified by the 001 peak at 14Å and the 002 peak at 7Å in the untreated sample (Fig. 7.7). The latter peak is of more use as the 14Å peak is frequently masked by peaks of other minerals. The peaks are unaffected by glycerolation or heating to 440° C. On heating to 550° C. the 002 peak often disappears whilst the 001 peak increases in intensity (Fig.7.7). This is due to a structural rearrangement of macrocrystalline chlorite (Brindley and Ali, 1950; Brindley and Brown, 1980). Heating also enables one to differentiate chlorite from kaolinite which has its 001 peak at 7Å. Kaolinite breaks down to a poorly ordered meta kaolin when heated at 450° to 550° C. (Brindley and Brown, 1980).

Sepiolite:- is identified by the 001 peak at 12.2Å which is unaffected by glycerolation or 440° C. heating (Figs. 7.8 and 7.9). On 550° C. heating this peak often migrates to 10.4Å where it can be masked by the illite 001 peak.

Palygorskite:- is identified by the 110 peak at 10.5Å which is unaffected by glycerolation but disappears with 440° C. heating. The 110 peak is frequently partially masked by the 001 illite peak (Fig. 7.9).

Fig. 7.7.   Characteristic X-ray diffraction peaks of illite and chlorite

and their behaviour under various chemical and thermal treat-

ments. Note collapse of chlorite 7Å peak on 550°C heating

and relative increase of chlorite 14Å peak. Sample EL 7389.
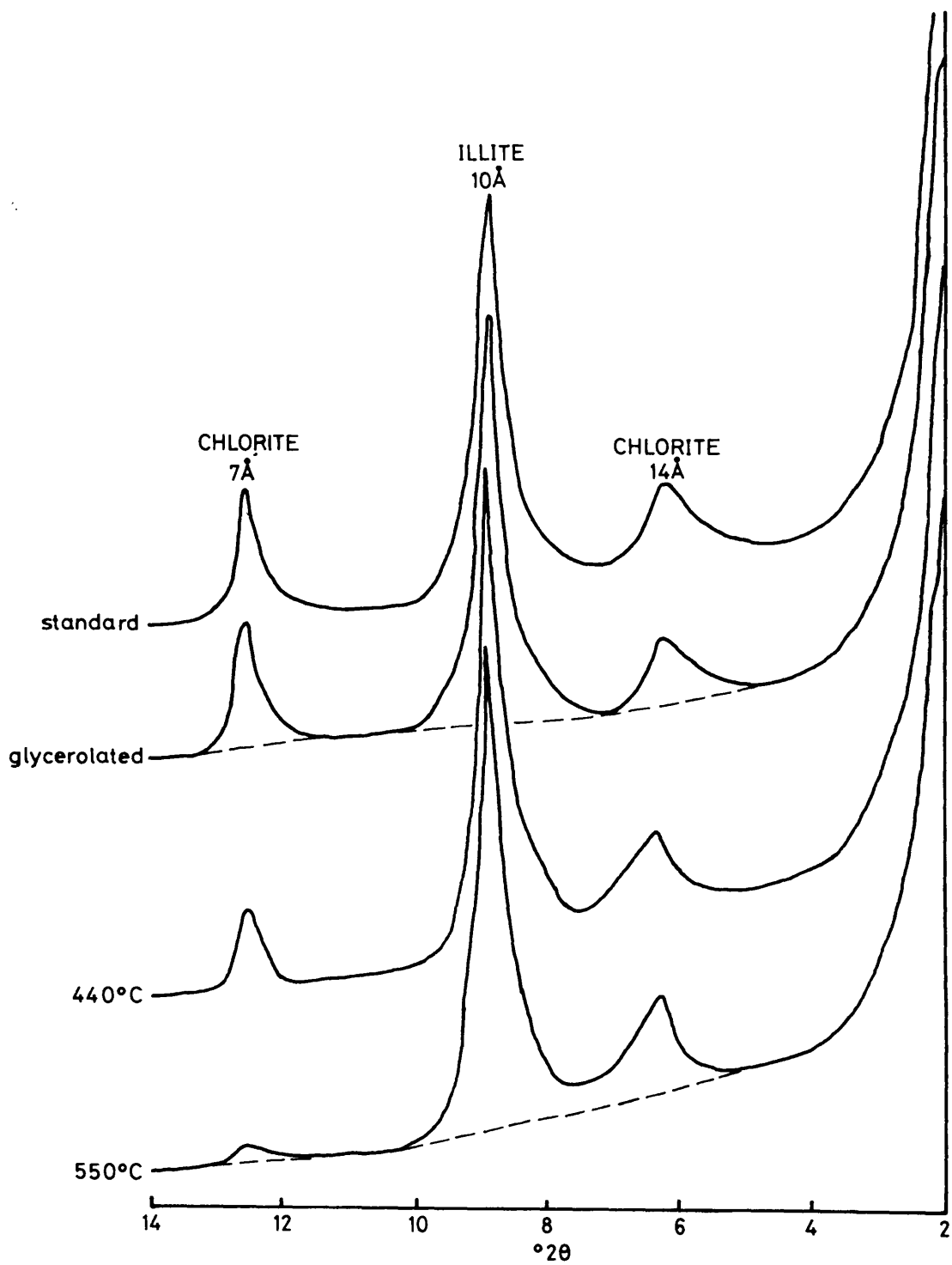
Fig. 7.8. Characteristic X-ray diffraction peak of sepiolite at

12.2Å is unaffected by either glycerolation or 440° C

heating but it collapses on 550° C heating. Other minerals

present are illite, chlorite and irregular mixed-layer

smectite/chlorite. The migration of the smectite/chlorite

peak on glycerolation and its marked collapse on heating

indicates that the mixed-layer mineral contains a high

proportion of smectite. Sample K 4135.
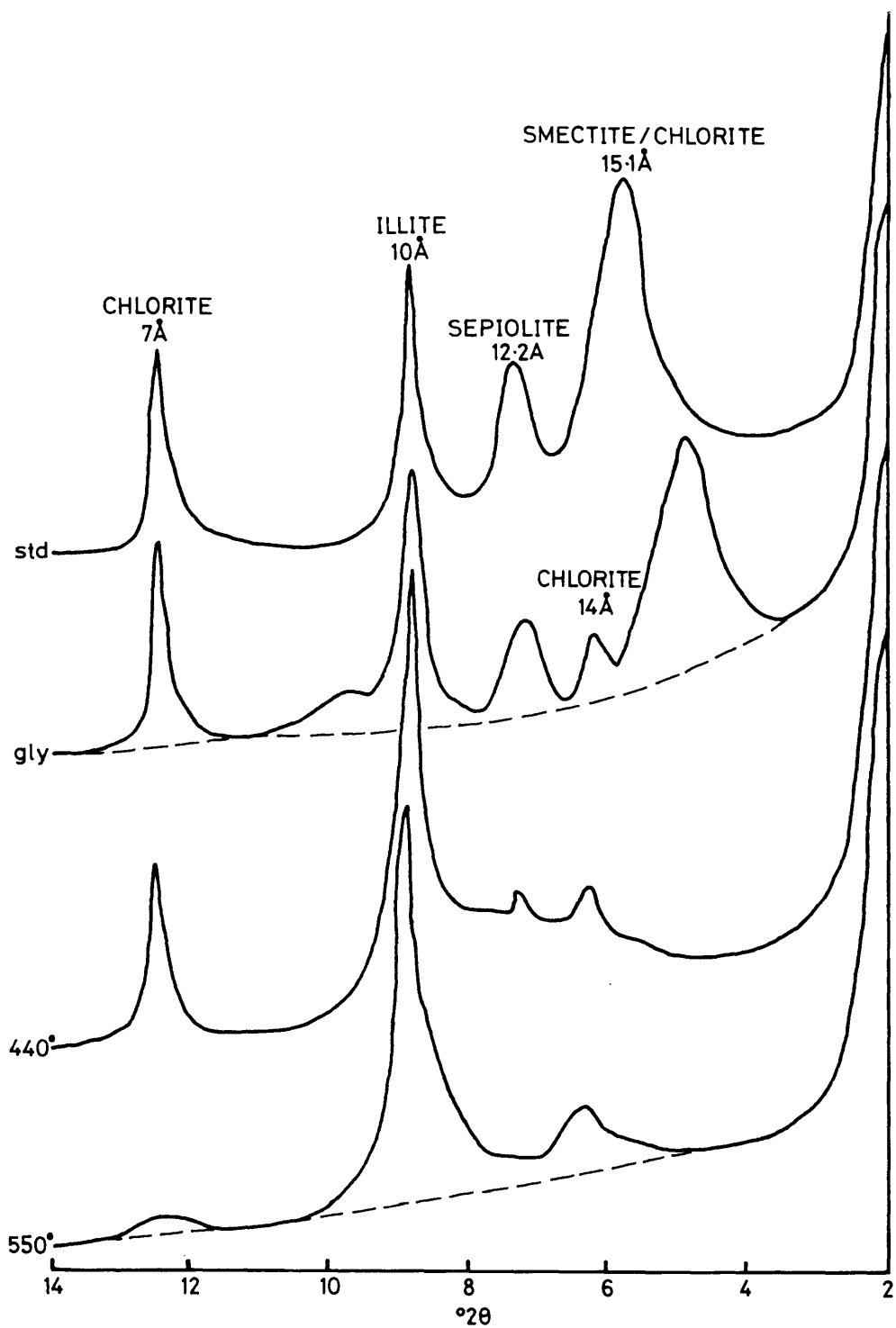
Fig. 7.9. Characteristic X-ray diffraction peak of palygorskite at

10.5Å is unaffected by glycerolation but collapses at

440° C heating. Also present are illite, chlorite, sepiolite

and mixed-layer smectite/chlorite. Note the 12.7Å position

of the Na^+ saturated smectite/chlorite peak compared to the

15.1Å Ca^{2+} saturated smectite/chlorite peak in Fig. 7.8.

Sample EL3.

SMECTITE / CHLORITE
(Na⁺ saturated)
12·7Å

PALYGORSKITE

ILLITE
10Å
10·5Å

CHLORITE
7Å

SEPIOLITE
12·2Å

SM./
CHL.

std

gly

440°

550°

14   12   10   8   6   4   2
°2θ
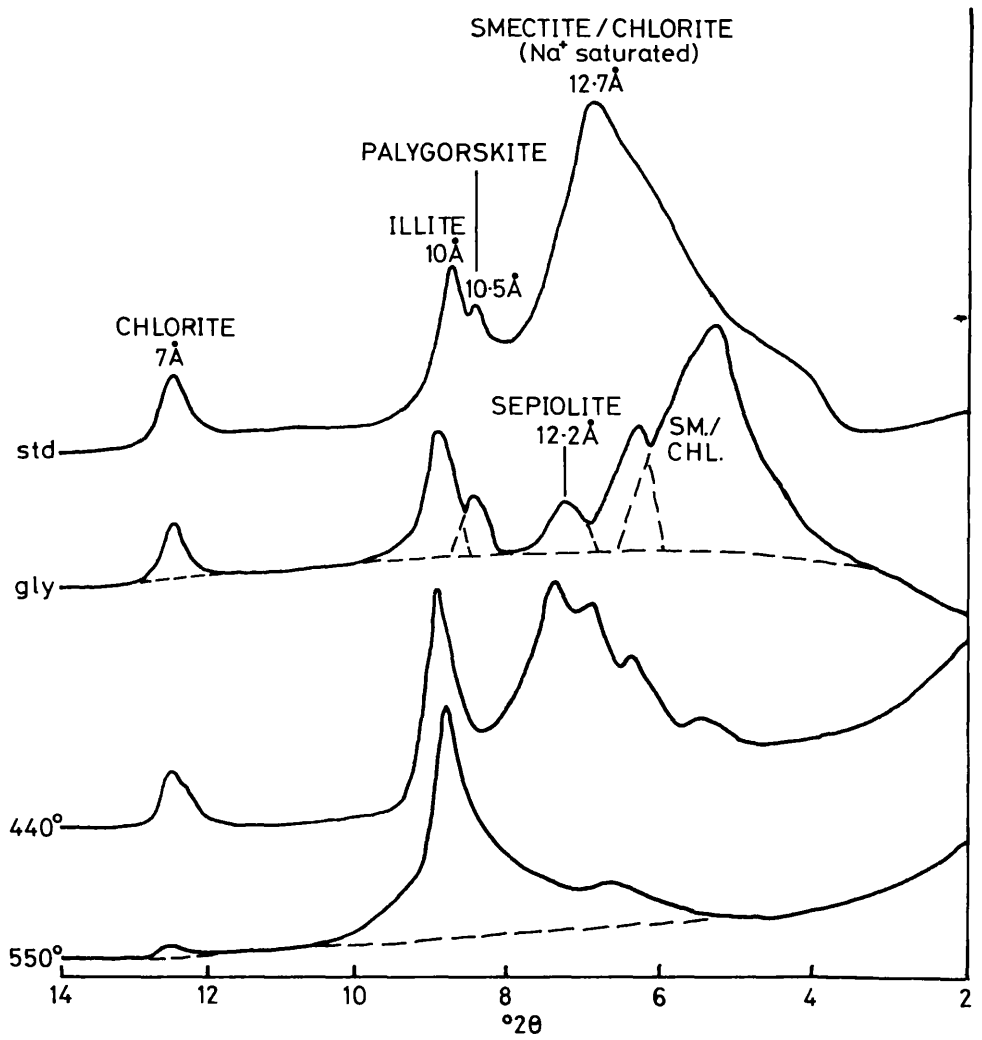
Mixed-layer minerals:- have diffraction characteristics intermediate between those of the constituent layer types. This results from the recorded diffractions for each crystal being the average of diffractions from each layer type within that crystal. Identification of a mixed-layer mineral is dependant on recognition of the constituent layer types. An irregular mixed-layer mineral will be represented by its 001 peak; secondary peaks are absent. A regular or ordered mixed-layer mineral will have both 001 and secondary peaks. The 001 peak is frequently a "super-lattice" peak resulting from the regular ordering of the constituent layer types. The "super-lattice" peak position equals the sum of the basal spacings of the constituent layer types. For example corrensite, a regular smectite-chlorite has an 001 peak at 29Å which equals 15Å (smectite) + 14Å (chlorite) (Lippmann,1954). This mineral was not recorded in the present study.

The only irregular mixed-layer mineral identified in the present study is smectite/chlorite which has properties characteristic of both smectite and chlorite. Smectite in an untreated sample is identified by an 001 peak between 15.0 and 15.5Å when saturated with calcium or magnesium cations. If smectite is saturated with sodium cations the 001 peak lies between 12.2 and 12.7Å (Brindley and Brown, 1980). On glycerolation both peaks expand to around 17.8Å whilst on heating to 440° C. the untreated peaks collapse to between 9.6 and 10.0Å . This collapse reflects the dehydration of interlayer areas (Brindley and Brown, 1980). (Fig. 7.3). The relevant properties of chlorite have already been stated. When saturated with calcium cations in an untreated sample smectite/chlorite has a broad peak between 14.5 and 15.0Å (Fig. 7.10), whilst the same peak occurs between 12.5 and 14.0Å (Fig. 7.9) when the sample is saturated with sodium cations. On glycerolation the shift in peak position is dependant upon the proportion of smectite within the smectite/chlorite; the greater the smectite content the further the peak migrates towards the expanded smectite position of 17.8Å (Figs. 7.10a, b). If chlorite predominates in the mixed-layer mineral the expansion on glycerolation will be slight (Fig. 7.10a). When heated to 440° C. smectite/

Fig. 7.10a.   Illite, chlorite and irregular mixed-layer smectite/
chlorite which may contain up to 50% smectite.   On heating,
the peak partially collapses to between 10 and 14Å -
compare this to the total collapse of the smectite/
chlorite peaks in Figs. 7.8 and 7.9.   In this case the
greater resistance to heat may reflect a higher chlorite
content.   The (001) chlorite peak is properly revealed on
migration of smectite/chlorite peak after glycerolation.
Sample F193/3.

ILLITE
10Å

SMECTITE / CHLORITE
14·5Å

CHLORITE
7Å

Sm./Chl.
15·9Å

std

CHLORITE
14Å

gly

Sm./Chl.

440°

Sm./Chl.

550°

14   12   10   8   6   4   2

°2θ
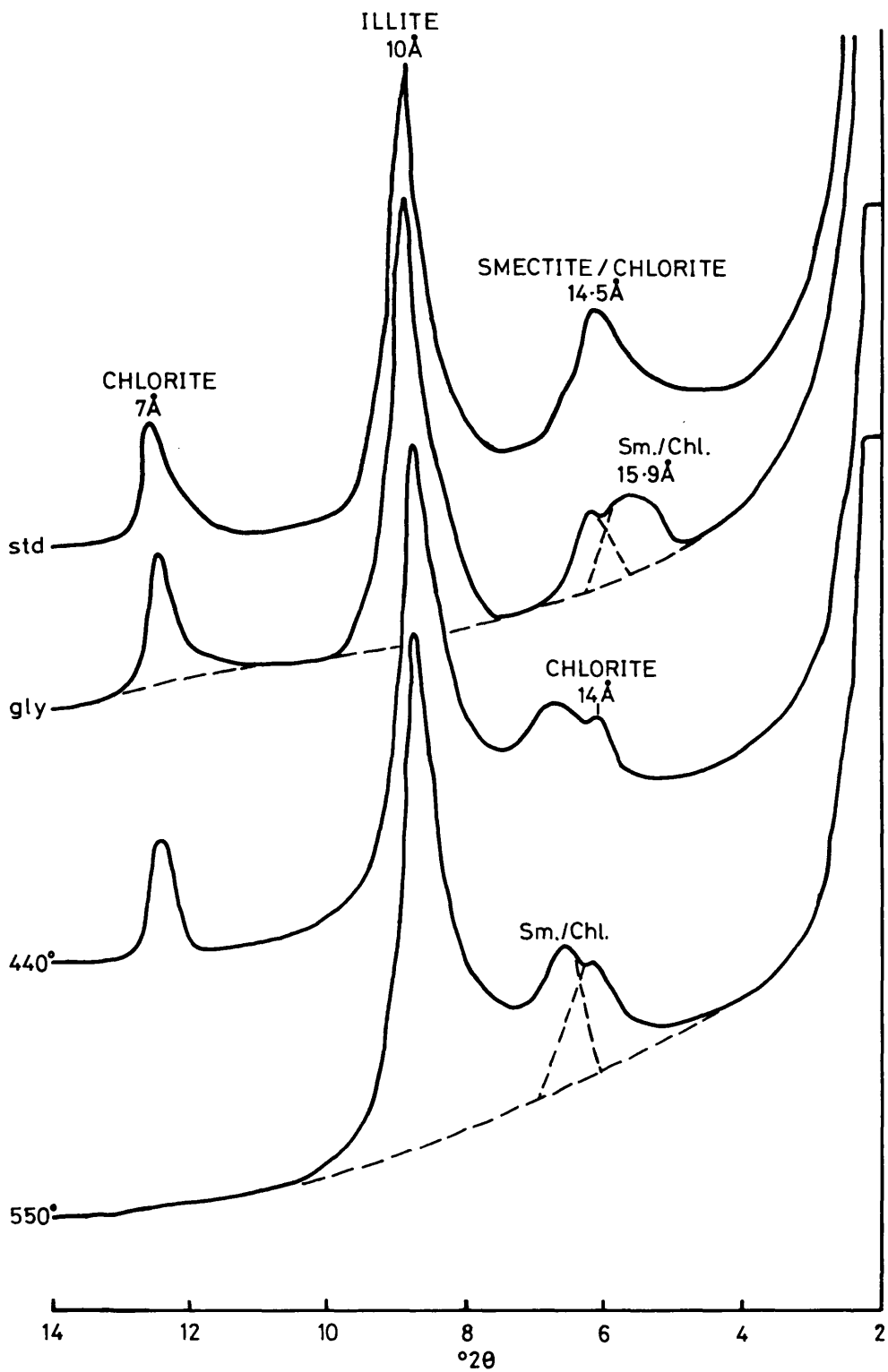
Fig. 7.10b.   Illite, chlorite and smectite/chlorite containing a high

proportion of smectite. On glycerolation the smectite/

chlorite peak migrates towards the expanded smectite

position (17.8Å) and it readily collapses to a broad peak

between 10 and 14Å on heating. Sample F500/D.

SMECTITE / CHLORITE
15·0Å

16·8Å

ILLITE
10Å

CHLORITE
7Å

std

gly

Sm./Chl.

440°

550°

14   12   10   8   6   4   2

°2θ
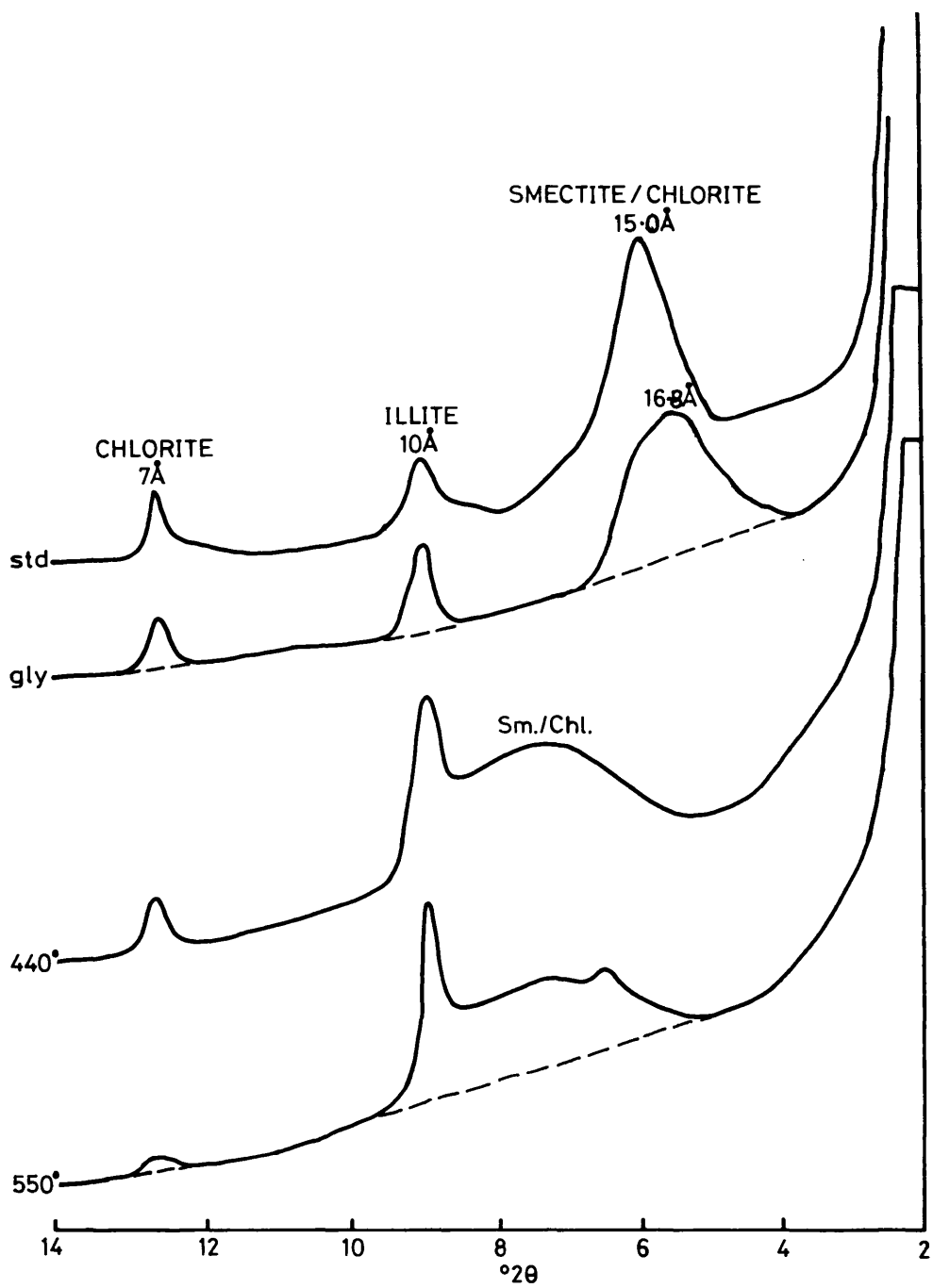
chlorite collapses to a broad peak between 10.0 and 14.0Å, which as previously noted are the positions of the two constituent layer types after heating. The position of the smectite/chlorite peak after heating reflects the proportions of smectite and chlorite within the mineral (Fig. 7.10).

Another common mixed-layer mineral which could be confused with smectite/chlorite is smectite/mica. The 001 peak positions of smectite/ chlorite and smectite/mica are very similar in both the standard and glycerolated traces. On heating to 440°C. smectite/mica collapses to between 10 and 9.8Å, a property which distinguishes it from smectite/chlorite.

SEMI-QUANTITATIVE ANALYSIS OF DIFFRACTOMETER TRACES.

The following method was used to obtain a semi-quantitative assessment of the clay mineralogy in each sample. Base lines, representing background scattering, were drawn on the diffractometer trace for each glycerolated sample between 2° and 14° 2θ (Figs. 7.7 - 7.10). The area of a diagnostic peak for each mineral present was then measured. The peaks used were: illite, 10Å; chlorite 7Å; sepiolite, 12.2Å; palygorskite, 10.5Å; and smectite/chlorite, 14 to 17.8Å. When measuring smectite/chlorite and palygorskite peak areas compensation was made for possible interference from the respective adjacent peaks of 14Å chlorite and 10Å illite. For any sample the peak areas give the relative intensities of the basal reflections from the orientated smear slide. The relative intensities (= peak areas) are then converted to relative proportions by the following procedure which is based on that used by Biscaye (1965) and Bradshaw (1976). The relative intensities of illite and chlorite are multiplied by four and two respectively to account for variations in the degree of X-ray scattering by each mineral (Brindley and Brown, 1980). Similarly sepiolite and palygorskite relative intensities are multiplied by eight and six respectively. A multiplication factor for smectite/chlorite has not been published. Biscaye (1965) multiplied the relative intensity of smectite by one and that of chlorite by two. It was

decided to multiply the smectite/chlorite relative intensity by either 1.25, 1.50 or 1.75 dependant on the estimated proportion of Ca^{2+} saturated smectite within the mineral. The smectite content is estimated by measuring the expansion of the smectite/chlorite on glycerolation. If the mineral expands by 1.8 to 2.7$\mathring{A}$ (i.e. it contains 66 to 99% smectite) then the relative intensity is multiplied by 1.25; peaks expanding between 0.9 to 1.8$\mathring{A}$ (33 to 66% smectite) and 0 to 0.9$\mathring{A}$ (0 to 33% smectite) are multiplied by 1.50 and 1.75 respectively. The relative proportions of minerals in each sample are then expressed as percentages.

RESULTS.

The semi-quantitative results for each locality are shown in Figs. 7.11 – 7.17. Estimated smectite contents based on the expansion of smectite/chlorite on glycerolation are plotted as frequency graphs in Fig. 7.18.

DISTRIBUTION OF CLAY MINERALS.

Illite occurs throughout the sequence at each sampling area. The Hawton Member contains 70 to 90% illite whilst the Fauld Member has an overall lower illite content of between 15 and 75% (Figs. 7.11 – 7.17). The Edwalton Formation contains 15 to 70% illite (Figs. 7.12 and 7.17).

Chlorite occurs throughout the sequence at each sampling area. The Hawton and Fauld Members both contain 2 to 20% chlorite (Figs.7.11 – 7.17). The Edwalton Formation contains 5 to 20% chlorite (Figs. 7.12 and 7.17).

Smectite/chlorite: irregular mixed-layer smectite/chlorite is recorded in most samples. The relative proportions of this mineral have been used to sub-divide the Trent Formation into the Fauld Member and overlying Hawton Member (Chapter 2). The Hawton Member clay fraction generally contains between 0 and 15% smectite/chlorite, whilst the Fauld Member generally contains more than 15% smectite/chlorite (Figs.7.11 – 7.17). For example, the average smectite/chlorite content of the Fauld and Hawton Members at Newark is 23.2 and 8.7% respectively (Fig.7.11). The maximum smectite/chlorite content recorded in this study is 80% from Fauld Mine (Fig.7.16). The proportion of smectite

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Fig. 7.11.   Clay mineralogy of the Hawton Member and top of the Fauld
             Member near Newark (Fig. 7.6). Samples taken from British
             Gypsum core material. Depths from surface.

KEY
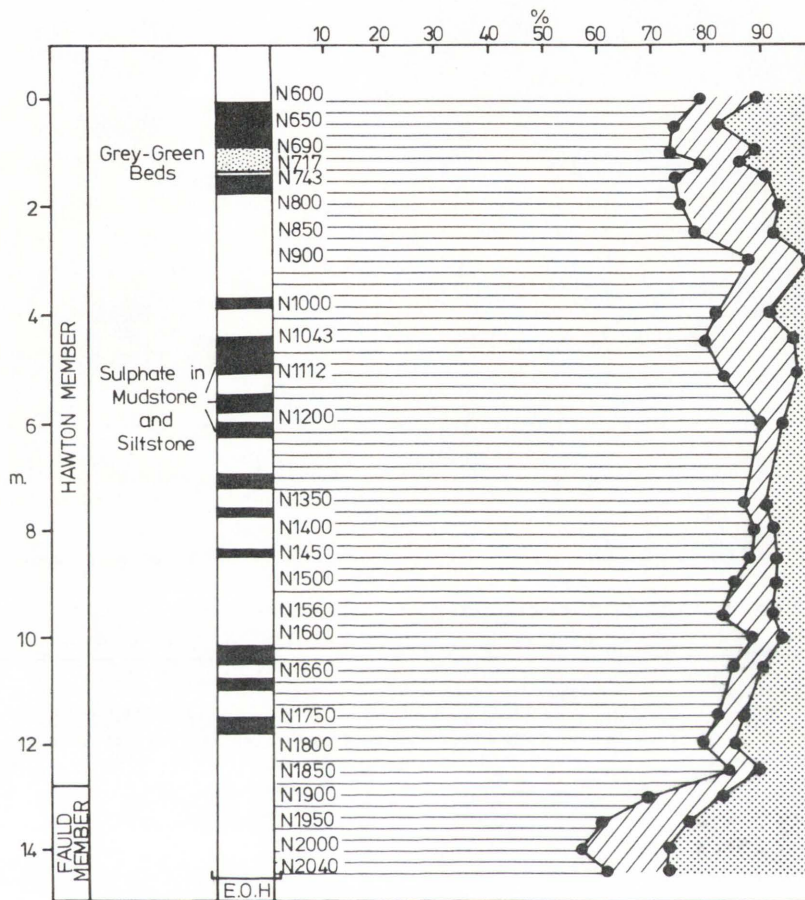
Illite   Sepiolite

Chlorite   Palygorskite

Smectite / Chlorite

Fig. 7.12.   Clay mineralogy of the I.G.S. borehole, Keyworth (Fig.7.6).

Depths from surface. Key as on Fig. 7.11.
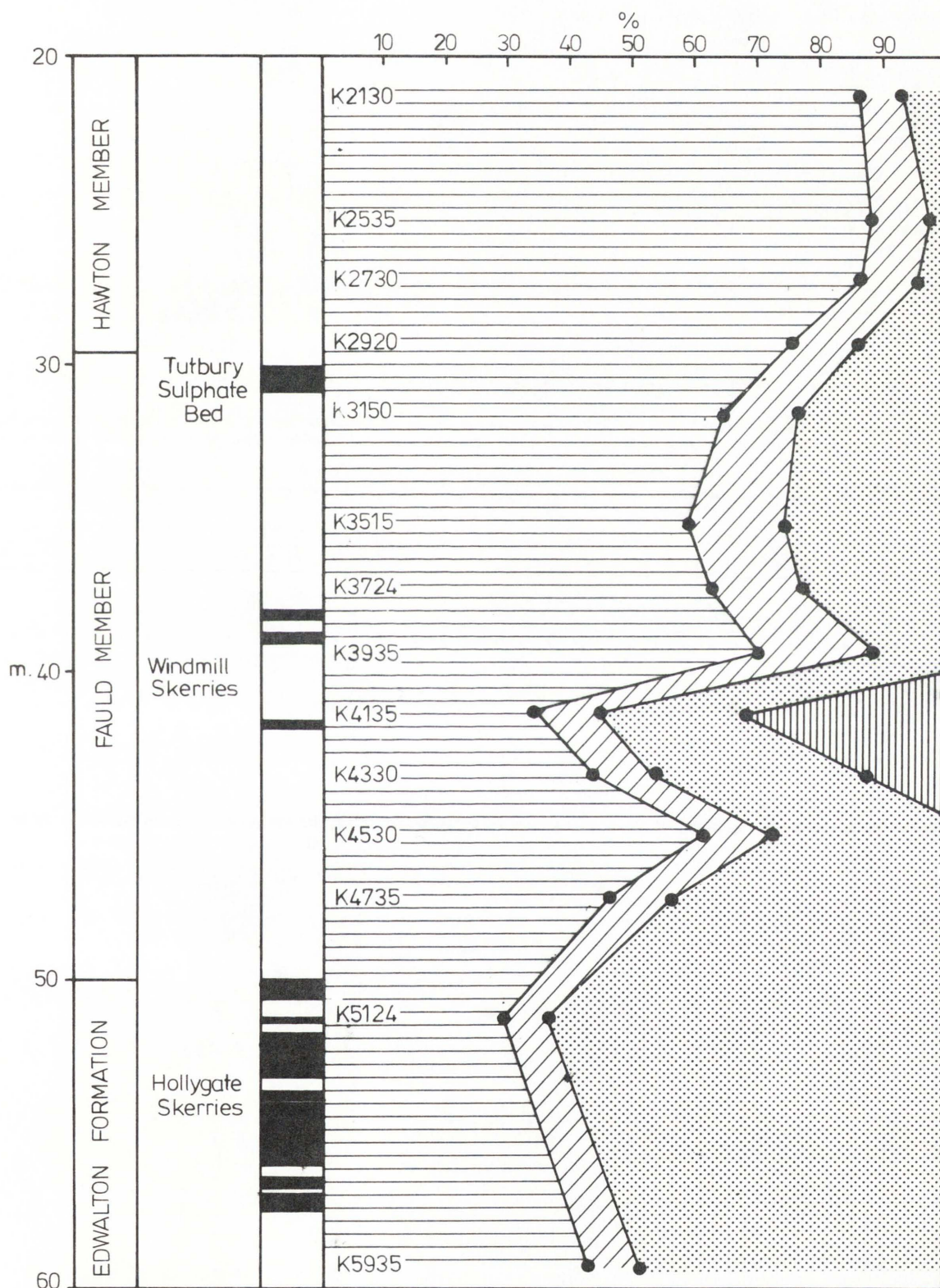
Fig. 7.13.   Clay mineralogy of the Hawton Member near East Leake

(Fig. 7.6).   Samples taken from British Gypsum core

material (note large open hole interval between two cores).

Depths taken from top of upper core. Key as on Fig. 7.11.
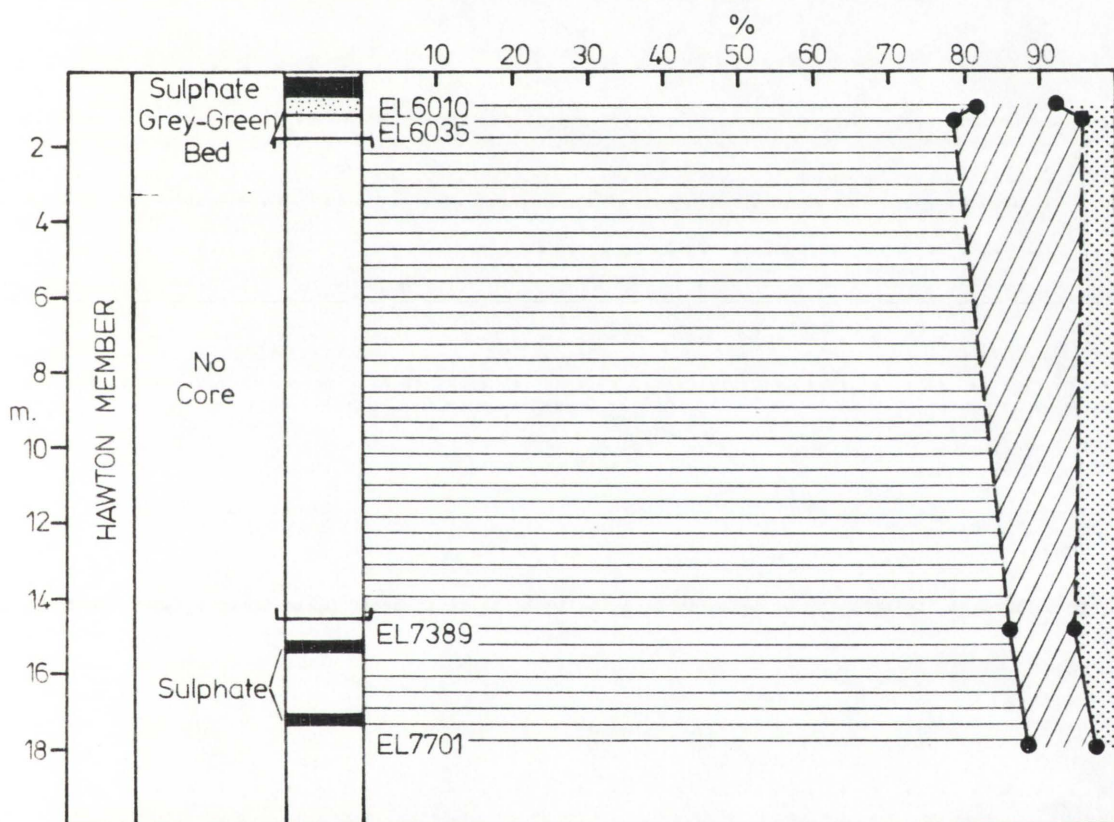
Fig. 7.14.   Clay mineralogy of the strata exposed in East Leake Mine

(Fig. 7.6).   The two sample localities are 530 m apart

but show a very similar distribution of clay minerals.

Key as on Fig. 7.11.
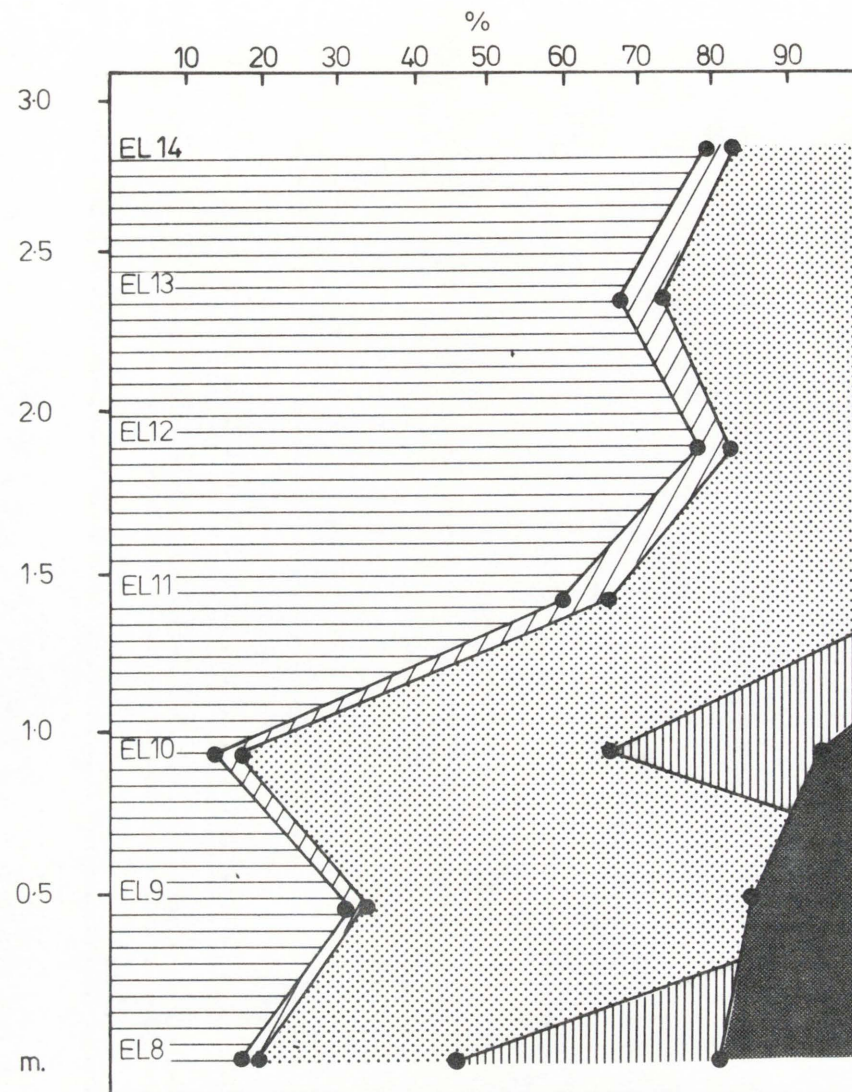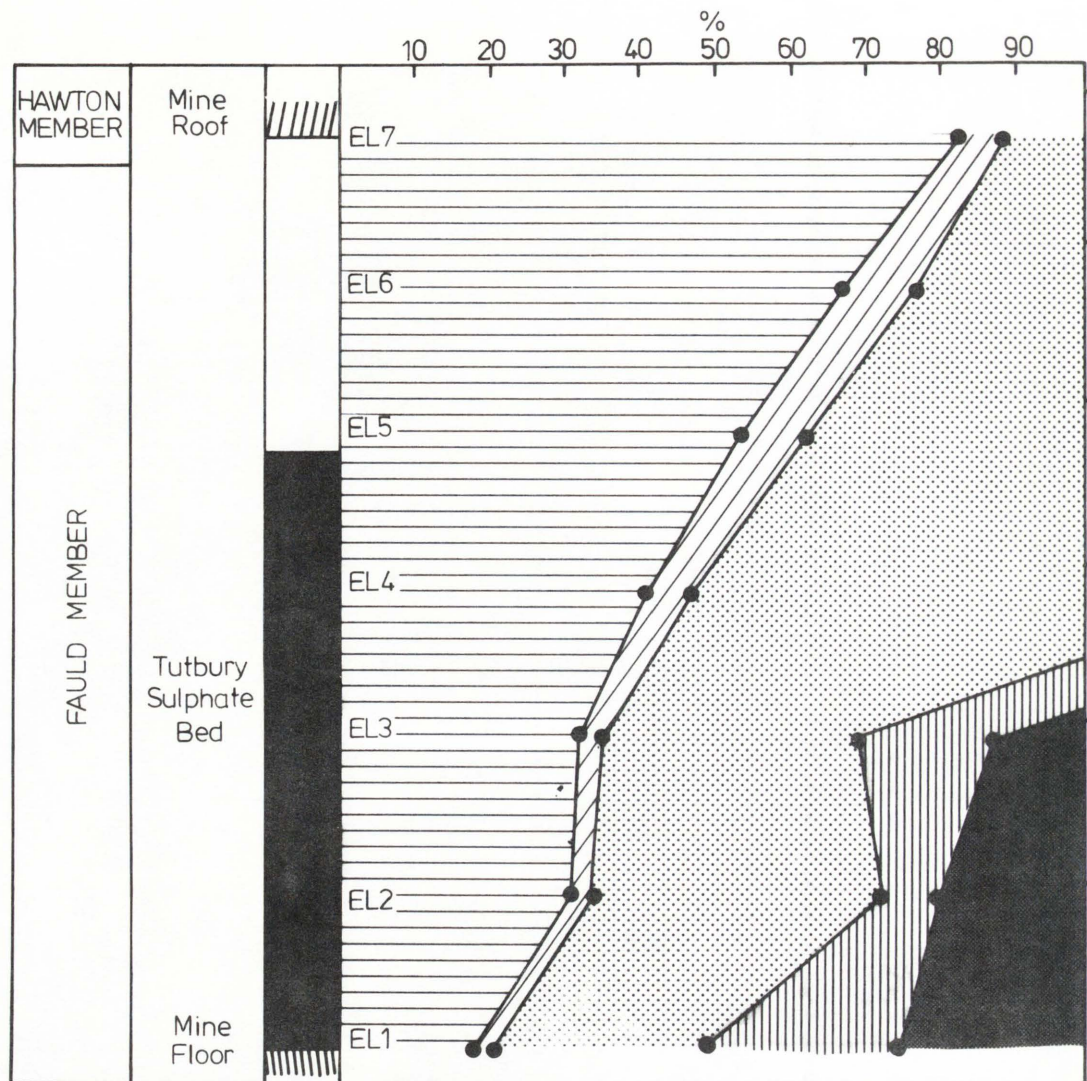
Fig. 7.15.   Clay mineralogy of the Hawton Member and upper prt of the

Fauld Member cored in a British Gypsum borehole near Fauld

Mine (Fig. 7.6). Depths given from top of core.

Key as on Fig. 7.11.
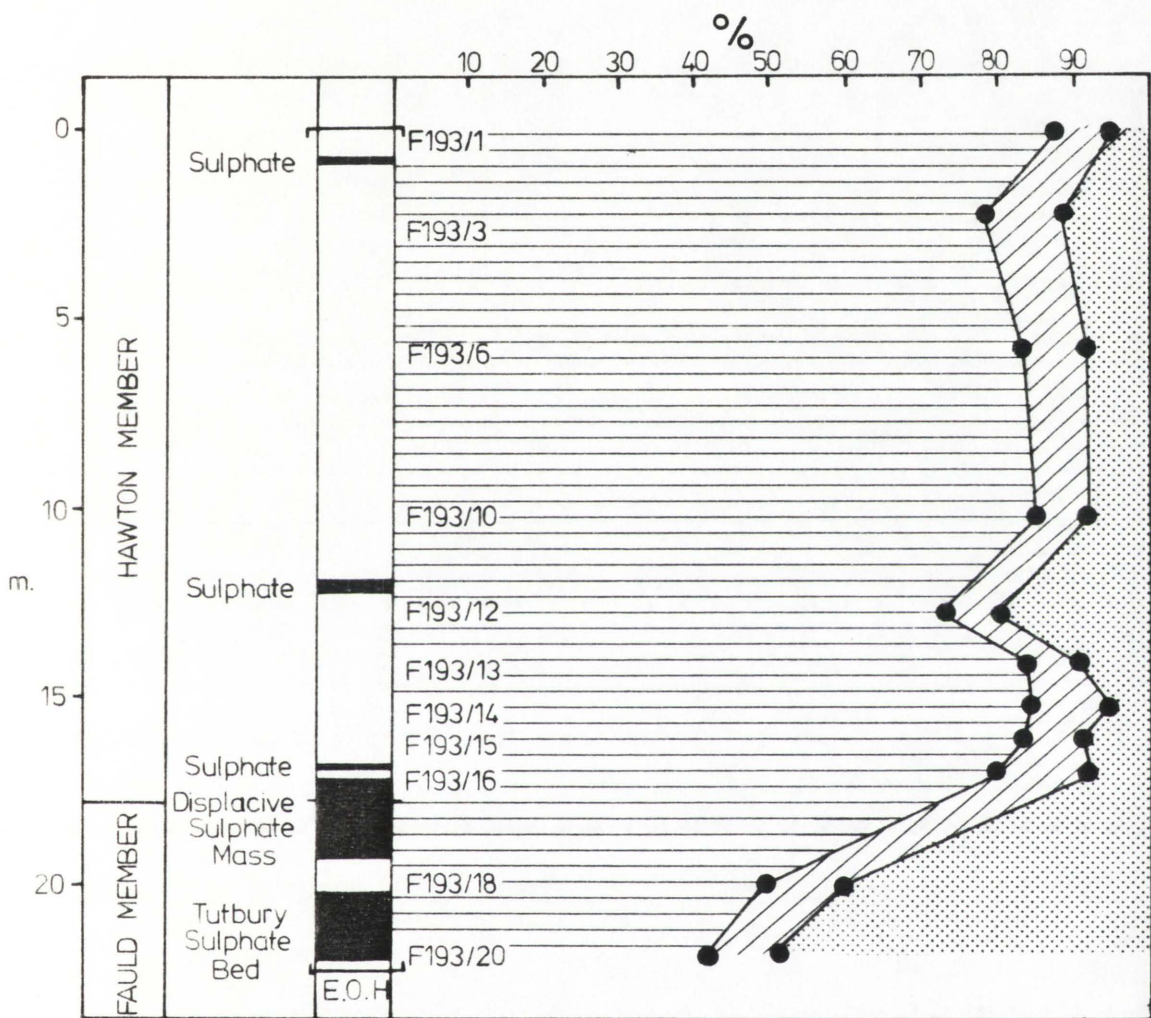
Fig. 7.16a.   Clay mineralogy of strata exposed in F$_a$uld Mine (Fig.7.6).
Both sample sets are of the typical mine face.   Depths
from mine roof.   Key as on Fig. 7.11.
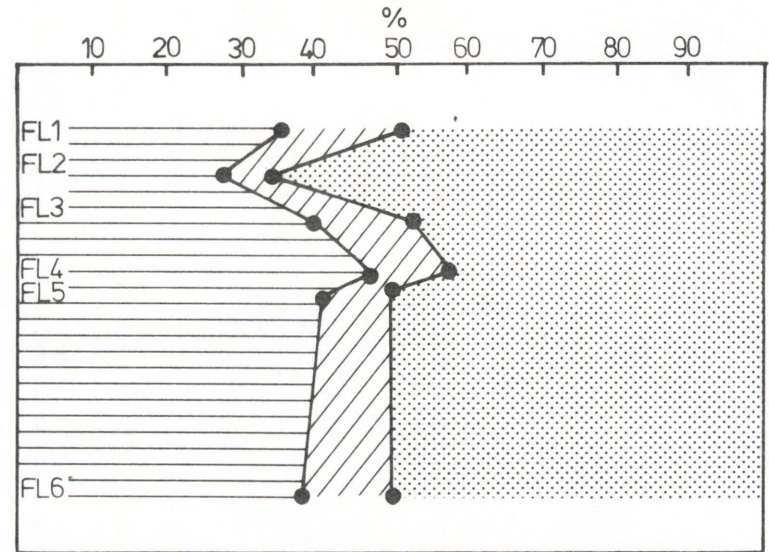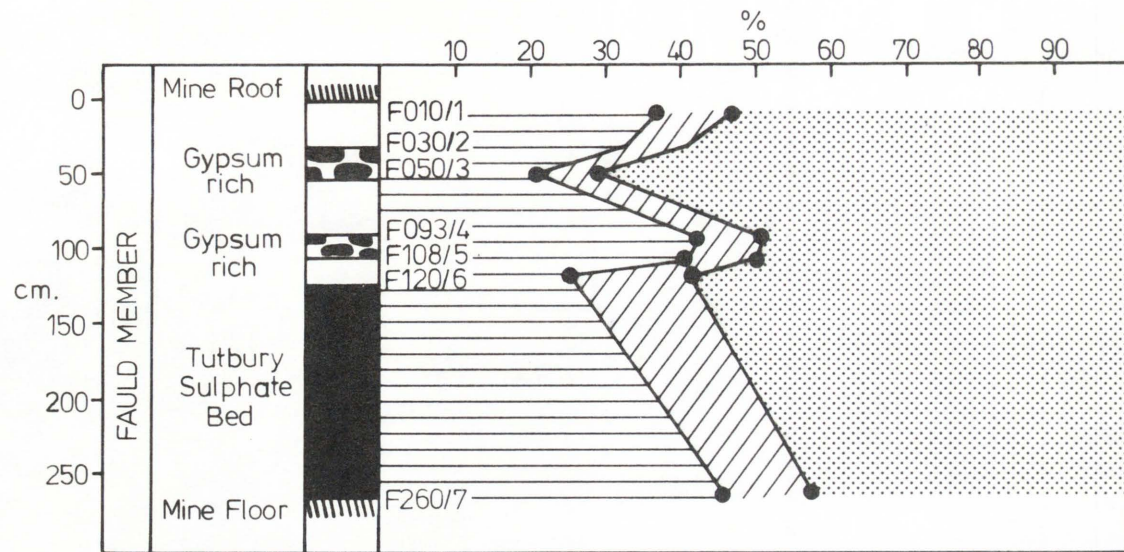
Fig. 7.16b.   Clay mineralogy of strata exposed in Fauld Mine (Fig.7.6).

Sample set F010/D to F500/D samples a section on a drivage

(or heading) which passes underneath the Tutbury Sulphate

Bed.   Depths from mine roof.   Key as on Fig. 7.11.
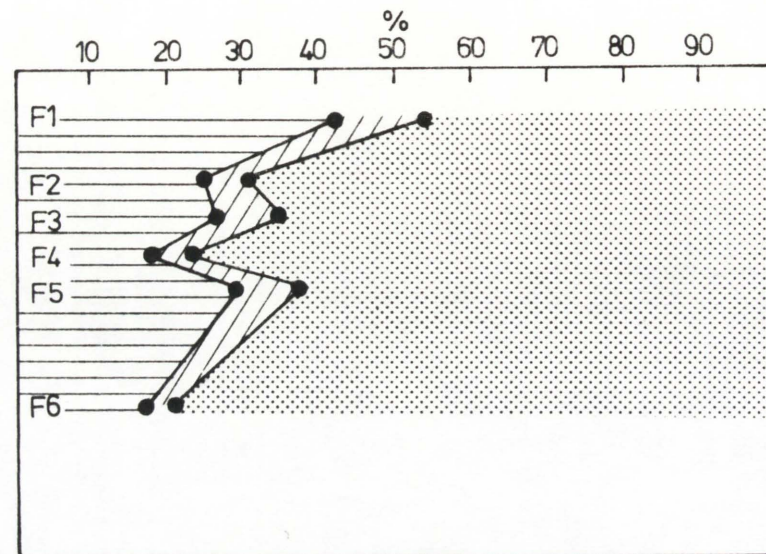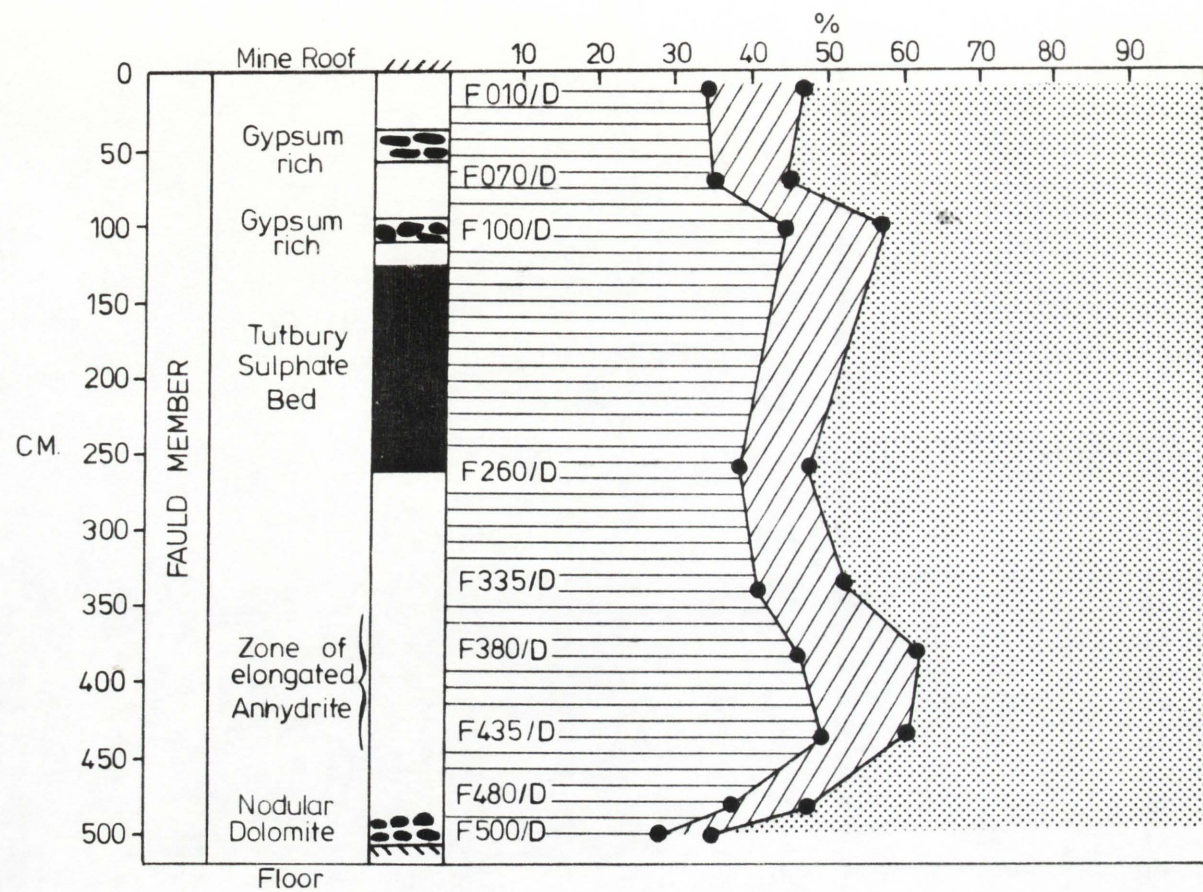
Fig. 7.17.   Clay mineralogy of the borehole drilled in the floor of
             Fauld Mine (Fig. 7.6).   The borehole cored parts of the
             Fauld Member and Edwalton Formation.   Depths given from
             mine floor.   Key as on Fig. 7.11.
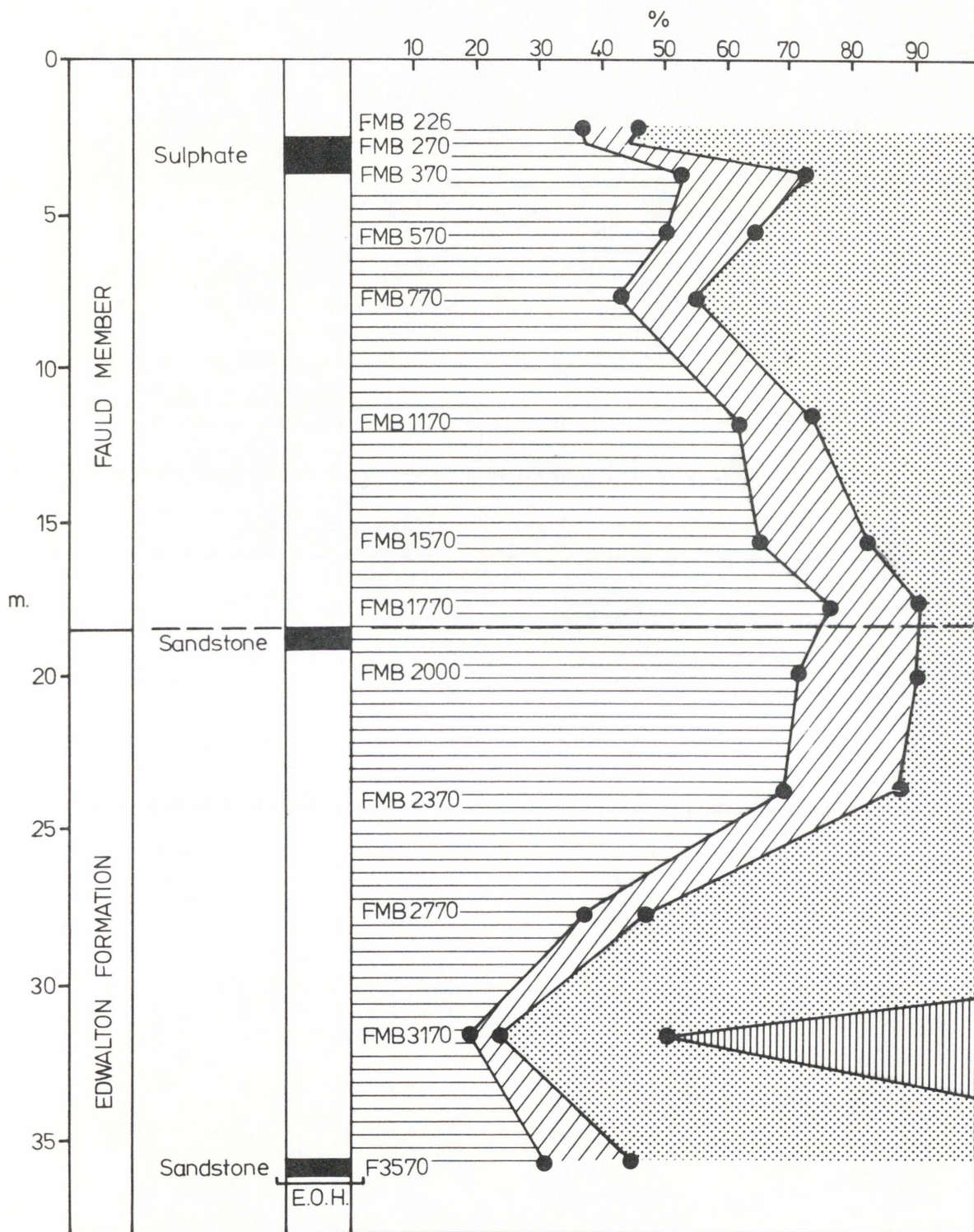
within the smectite/chlorite varies from 0 to 99% (Table 7.2, Fig. 7.18).

TABLE 7.2

		FAULD	EAST LEAKE	KEYWORTH	NEWARK
HAWTON MEMBER	Range	0 - 1.61Å 0 - 60%	0 - 1.25Å 0 - 45%	0 - 1.48Å 0 - 50%	0.30 - 1.45Å 10 - 50%
	Average	0.95Å 35%	0.52Å 20%	0.67Å 25%	0.89Å 30%
FAULD MEMBER	Range	0.30-2.62Å 10 - 90%	0.41-3.43Å 15 - 99%	1.10-2.90Å 40 - 95%	0.30 - 0.65 10 - 25%
	Average	1.54Å 55%	1.58Å 55%	1.98Å 70%	0.48Å 20%
EDWALTON FORMATION	Range	0.83-2.49Å 30 - 85%		1.56-2.53Å 55 - 90%	
	Average	1.72Å 60%		2.05Å 70%	

Estimated smectite content within irregular mixed-layer
smectite/chlorite in the Hawton and Fauld Members and Edwalton Formation.

The Hawton Member mixed-layer mineral is generally a smectitic-chlorite,
i.e. it contains less than 50% smectite (Table 7.2), whilst the Fauld
Member mixed-layer mineral is generally a chloritic-smectite, i.e. it
contains greater than 50% smectite (Table 7.2). The few analyses for the
uppermost part of the Edwalton Formation indicate that the mixed-layer
mineral is a chloritic-smectite containing an average of 65% smectite
(Table 7.2).

The ubiquitous occurrence of illite and chlorite confirms the results
of earlier studies (Dumbleton and West, 1966; Jeans, 1978). The correlation
based upon smectite/chlorite content is applicable to sections analysed by

Fig. 7.18.   Estimated smectite contents in irregular mixed-layer

smectite/chlorite. The smectite content is estimated from

the degree of expansion shown by the mineral on glycero-

lation; pure smectite expands by 2.8$\overset{\circ}{A}$.

HAWTON (NEWARK) BOREHOLE

⊠ Fauld member sample

Å-Expansion

% Smectite

KEYWORTH BOREHOLE

⊠ Hawton member sample

Å-Expansion

% Smectite

EAST LEAKE MINE

⊠ Hawton member sample

Å-Expansion

% Smectite

FAULD MINE & BOREHOLES

⊠ Hawton member sample

Å Expansion

% Smectite
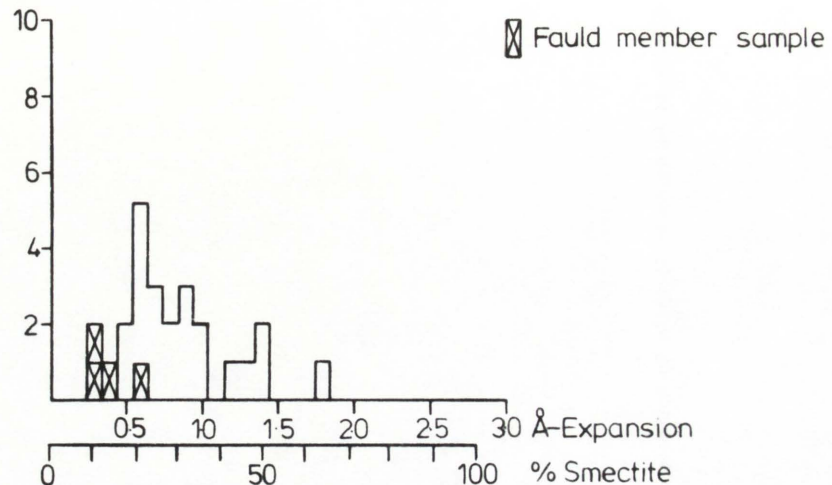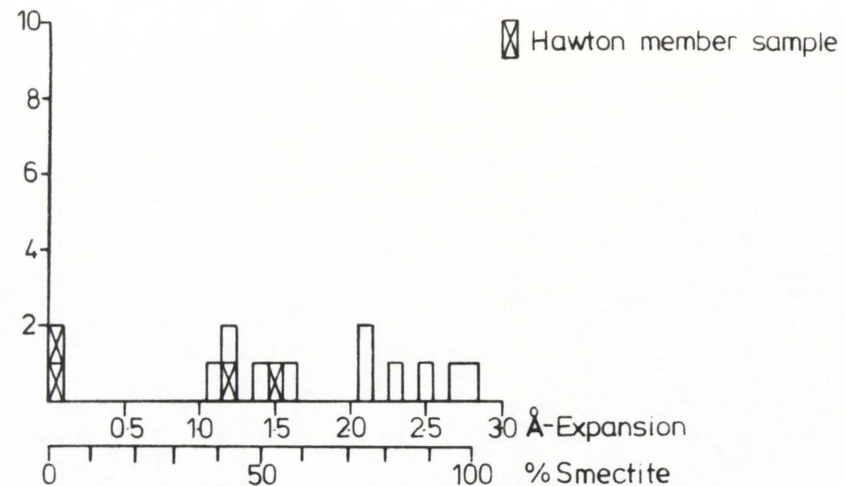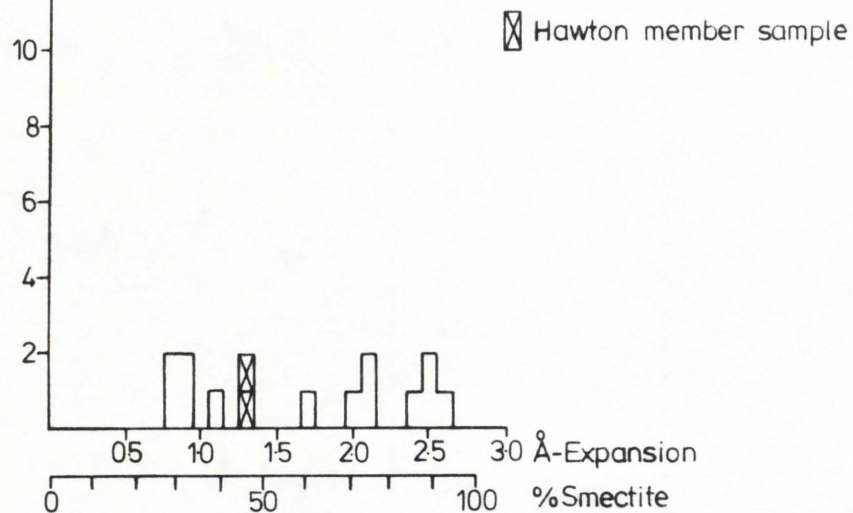
Jeans (1978) (Chapter 2), in the East Midlands as well as other areas of Britain. For example Bunny brick pit, Cropwell Bishop and Staunton gypsum pits, Gypsy Lane brick pit, all expose part of the Hawton Member and all have low smectite/chlorite (Jeans, 1978). The Hawton – Fauld Member boundary is marked in the Clipston borehole by an increase in smectite/chlorite (Jeans, 1978).

Sepiolite is recorded in the Fauld Member at Keyworth (Fig. 7.12) and East Leake (Fig. 7.14) and the Edwalton Formation at Fauld (Fig. 7.17). At Keyworth sepiolite occurs in 2 m. of strata immediately underlying the sandy horizons previously called the Windmill Skerries by Elliott (1961). Jeans (1978) recorded sepiolite from just below the Windmill Skerries at Windmill Hill (SK 644 358) near Cotgrave and from two horizons in the National Coal Board Clipston borehole, the upper horizon of which correlates with the sepiolite-bearing horizon in the Keyworth borehole and Windmill Hill exposure. This is a total distance of 5 km. The lower horizon lies 1 m. above the base of the Trent Formation (Jeans, 1978) and has not been detected in the present study.

At East Leake sepiolite is recorded from the middle and basal sections of the Tutbury Sulphate Bed (Fig. 7.14).

Palygorskite is recorded only from the Tutbury Sulphate Bed at East Leake and is associated with sepiolite (Fig. 7.14).

S.E.M. STUDIES OF CLAY MINERAL PETROLOGY.

Numerous mudstone samples were examined under the scanning electron microscope in an attempt to relate textural features to specific clay minerals previously identified on the X.R.D. However illite, chlorite and smectite/chlorite all have a plate-like morphology and proved to be indistinguishable. The qualitative E.D.A.X. (energy dispersive analytical system) attachment on the S.E.M. was also unable to differentiate between these three minerals. The distinctive fibrous habit of sepiolite and palygorskite (Jeans, 1978; Hassouba and Shaw, 1980) could not be identified

in the present study.

DISTRIBUTION OF SODIUM IN MUDSTONES ASSOCIATED WITH THE TUTBURY SULPHATE BED, EAST LEAKE MINE.

Fourteen samples from East Leake Mine, the clay mineralogy of which is shown in Fig. 7.14, were also analysed for their soluble salt content. Of particular interest is the sodium content which was determined on a flame photometer by British Gypsum Ltd., at East Leake. Assuming that Na^+ is contained only within the non-evaporitic fraction of the sample, the analyses are corrected to take account of the sulphate content in each sample, as high sulphate contents would result in the apparent lowering of Na^+ content. Both uncorrected and corrected Na^+ contents are shown in Fig. 7.19.

The smectite content of the smectite/chlorite mixed-layer minerals was estimated from the degree of expansion shown on glycerolation (Fig. 7.20). It is readily apparent that Na^+ and smectite contents increase sympathetically. Smectite has a high cation exchange capacity (100 to 160 meq : 100g) and readily absorbs cations such as sodium into the interlayer space (Brindley and Brown, 1980). It is therefore proposed that sodium in the analysed samples is present in the interlayer space associated with the smectite component of the irregular mixed-layer smectite/chlorite mineral.

MECHANISMS OF CLAY FORMATION.

In view of the controversy regarding the origin of clay minerals in Triassic red beds it is worthwhile considering the possible modes of origin of different clay minerals documented in this study.

There are three major processes by which clays may originate:-

1) inheritance.

2) transformation.

3) neoformation.

Fig. 7.19.   Sodium content of mudstones in East Leake Mine.

Uncorrected values are results received from laboratory

whilst corrected results take account of the presence of

sulphate in the sample. The two sample sets were also

analysed for clay mineralogy (Fig. 7.14).

Fig. 7.20.   Estimated smectite content in smectite/chlorite of above

samples;   the sample symbols correspond to those in

Fig. 7.19. Note the similarity of the Na^+ content curves

and associated smectite content curves indicating that the

sodium is present in the interlayer space of the smectite

component of the mixed layer smectite/chlorite (see text).

'UNCORRECTED' Na$^+$ ————
CORRECTED Na$^+$ ————

%Na$^+$

ESTIMATED% SMECTITE IN SMECTITE / CHLORITE
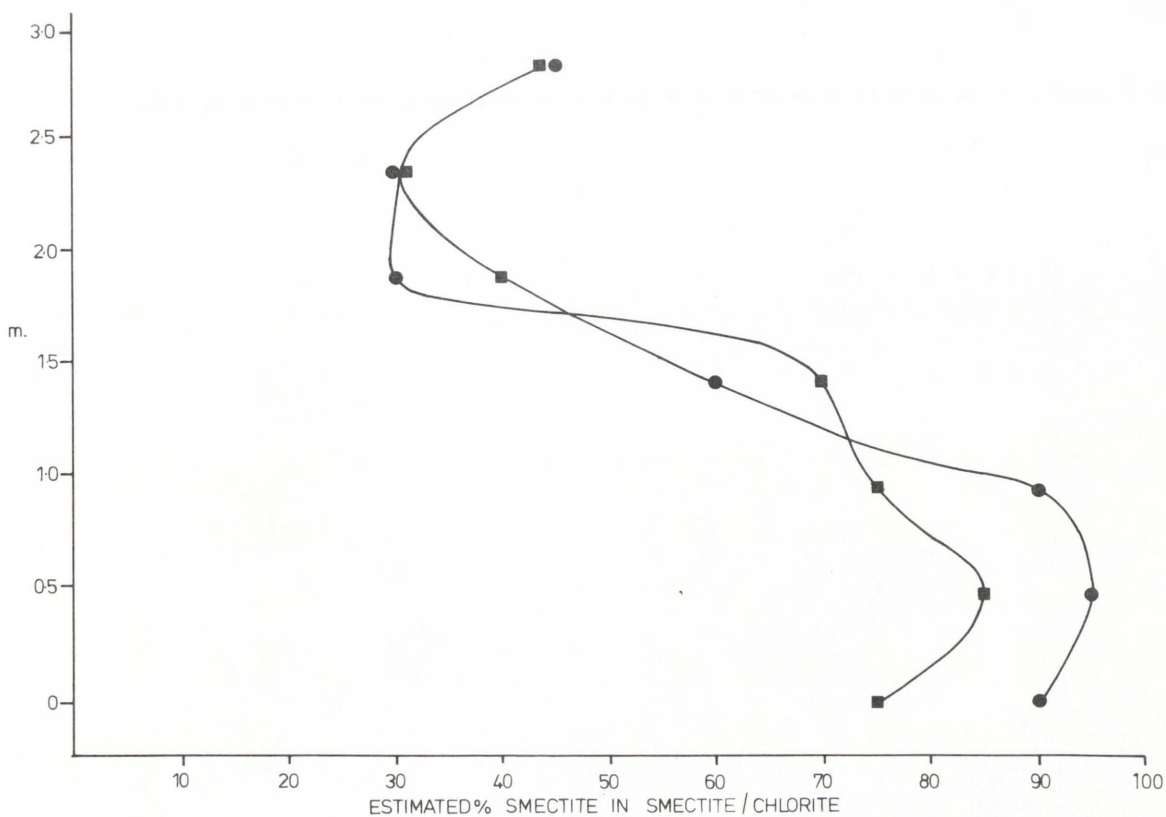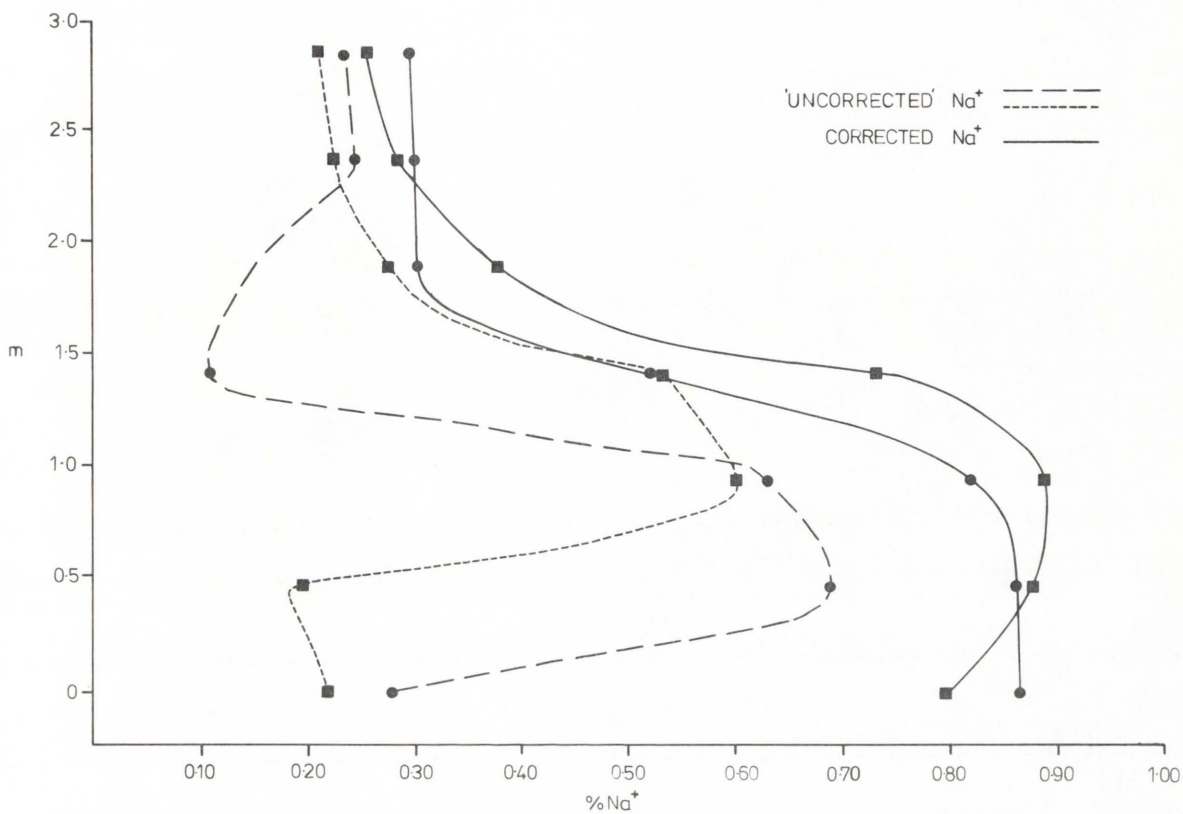
INHERITANCE.

As the term suggests this process covers the weathering of parent rock
to form clay particles which are transported and deposited in a sedimentary
basin. Following Millot (1970) the term is restricted to particulate
material which remains inert during sedimentation and diagenesis.

TRANSFORMATION.

This applies to changes that modify a clay mineral without altering
its 1:1 or 2:1 structural type. Transformational processes are subdivided
into degradation and aggradation.

Degradational transformation is characteristic of continental weathering
and soil formation when cation deficient solutions principally leach cations
from minerals. Addition of water molecules in the interlayer space also
occurs (Millot, 1970), giving rise to clays with variable basal spacing
such as smectite and vermiculite (Millot, 1970).

Aggradational transformation is the opposite of degradational transformation
as it involves the reconstruction of clay minerals from the degraded minerals
provided by weathering. Aggradation proceeds by the adsorption of cations
into the unstable degraded lattice resulting in the formation of the original
mineral or, if modification of the lattice occurs, a new mineral.
Aggradation only occurs in cation rich environments such as during transport
within the sedimentary basin and during diagenesis (Millot, 1970; Dunoyer
de Segonzac, 1970). There is a whole gamut of aggradations that corresponds
to the range of degraded minerals, involving simple fixation of cations to
the complete reorganization of clay minerals. Consequently discussion is
limited to aggradation reactions directly applicable to minerals documented
in the present study.

Degraded illite and chlorite are re-aggraded during early diagenesis by
adsorption of potassium and magnesium ions respectively (Dunoyer de Segonzac,
1970; Millot, 1970). Aggradation of chlorite can occur during trans-
portation as magnesium is adsorbed in preference to potassium at high Mg:K

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concentration ratios (Powers, 1959; Millot, 1970). At lower Mg:K ratios potassium adsorption occurs, giving rise to illite 1Md (disordered stacking sequence, Brindley and Brown, 1980), which is the common illite polytype resulting from aggradational processes. Later burial diagenesis enhances the crystallinity of both illite and chlorite (Shaw, 1980; Dunoyer de Segonzac, 1970).

Smectite reacts with Mg^{2+}-rich solutions to form mixed-layer minerals such as smectite-chlorite, which in turn can eventually form chlorite (Fig. 7.21), whilst reaction with K^+-rich solutions results in the formation of mixed-layer smectite-illite and eventually illite (Fig. 7.21) (Dunoyer de Segonzac, 1970; Millot, 1970; Shaw, 1980). The latter reaction series was recorded in the upper 50 m. of East African lake sediments with smectite reacting with brines of unusually high K:Na ratio (Singer and Stoffers, 1980).

Of particular importance to this study is the work by Lucas (1962) on the hypersaline Triassic sediments of France and Morocco. Lucas (1962) concluded that two mineral sequences could be followed from the edge of the Jura (France) depositional basin into its centre. The sequence commenced at the basin margin with degraded illite, minor chlorite and kaolinite, all supplied by erosion of adjacent landmasses. Continental run-off diluted near-shore brines but the salinity increased towards the centre of the basin due to evaporation and decreasing continental dilution effects. As the original poorly crystalline illite moved further towards the basin centre it reacted with brines of increasing salinity resulting in the following mineral sequences:-

    1) degraded illite→open illite→irregular mixed-layer smectite-mica →irregular mixed-layer smectite-chlorite→corrensite→irregular mixed-layer chlorite-swelling chlorite→well crystallized chlorite;

    2) degraded illite→well crystallized illite.

These aggradational transformation reactions may have occurred during early diagenesis when the detritus equilibrated with the sediment's pore waters, rather than during the period of transportation (Krumm, 1969; Donoyer

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Fig. 7.21.   Transformation reactions involving smectite in magnesium

and potassium rich environments (taken from Dunoyer de

Segonzac, 1970).
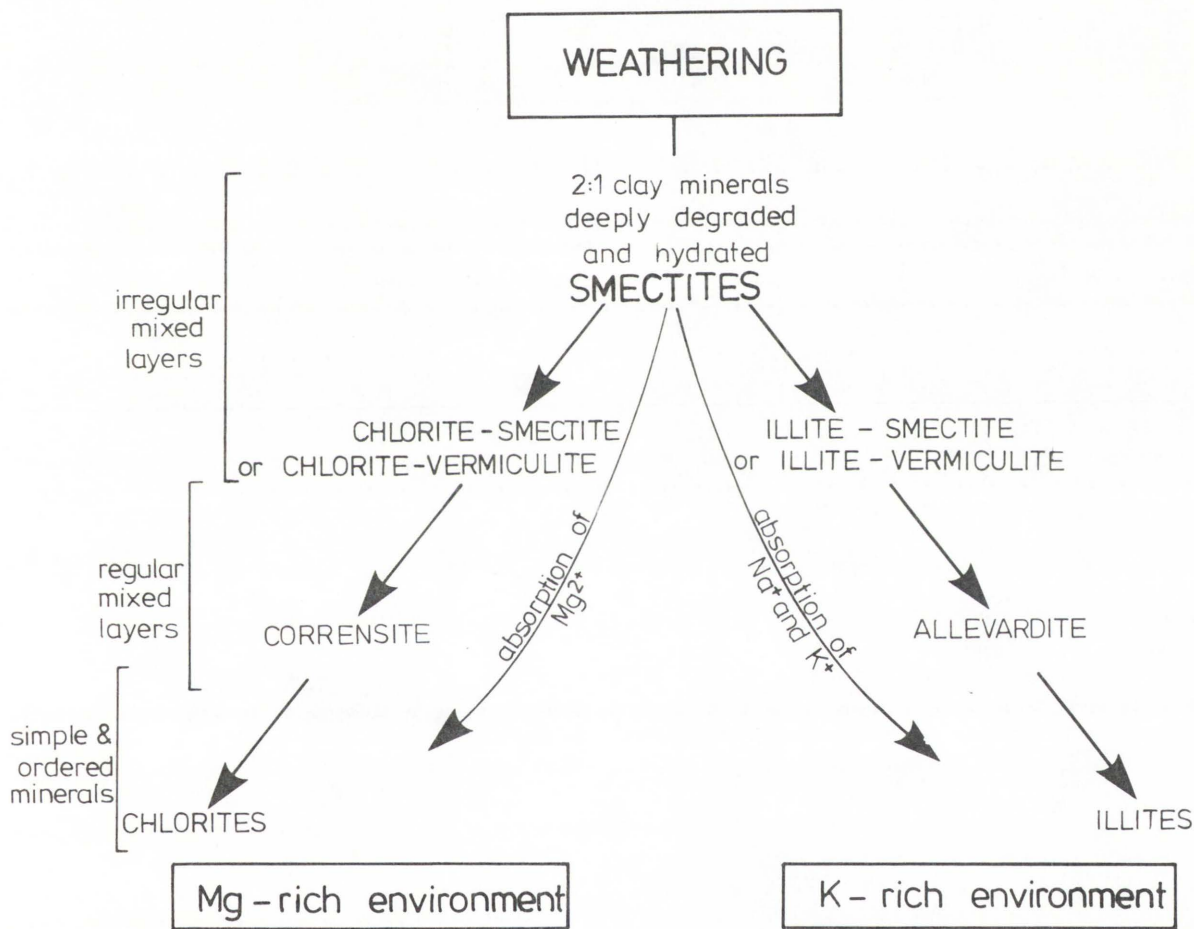
de Segonzac, 1969). However, Millot (1970) drew attention to the fact that the transformed chlorites occur as well formed euhedral crystals in the order of 0.5 to 1.0 μm. whilst the original degraded illite particles were around 0.1 μm. diameter. The growth of the chlorite crystals could therefore be interpreted as a neoformational process; however, the classification of this series as transformational reactions is justified because the original material was pre-existing crystal lattices. Neoformation, as described later, is the formation of crystals from material in solution.

A similar post-depositional transformation reaction suggested by Lippmann and Savaşçin (1969) involves the alteration of detrital mica by magnesium-rich pore water to form corrensite:

detrital mica + Mg^{2+} -rich pore solutions $\rightarrow$ corrensite + K^+
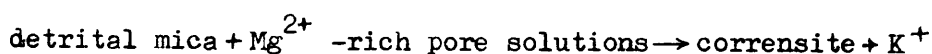
Schlenker (1971) postulated that the potassium released by the corrensite forming reaction reconstituted originally open mica containing swelling interlayers to give smectite-mica.

Aggradational reactions are important during burial diagenesis when increased pressures and temperatures aid the dehydration of smectite (Perry and Hower, 1972; Shaw, 1980), adsorption of Mg^{2+} or K^+ results in formation of chlorite or illite type minerals respectively (Fig.7.21) (Aoyagi and Kazama, 1980; Shaw, 1980).

NEOFORMATION.

This involves clays being formed from solutions that may or may not be supplied with alumino-silicate debris such as amorphous gels (Millot, 1970; Grim, 1968). Pre-existing silicate sheet lattices are not utilized during neo-formation, hence differentiating it from transformation. Neoformation occurs in weathering, sedimentary and diagenetic environments. As with transform-ational clays, discussion is restricted to minerals of interest to this study.

a) Neoformation in the weathering environment:

Smectite is neoformed in neutral to alkaline, medium intensity weathering conditions, such as those found in temperate climates where drainage is moderately good (Jackson, 1959; Millot, 1970; Velde, 1977). Kaolinite is formed in more intense weathering conditions which result in the leaching and

removal of alkali cations (Jackson, 1959; Millot, 1970; Velde, 1977).
Kaolinite is unstable in alkaline conditions when the presence of alkali
cations, especially calcium, will inhibit its formation (Millot,1970), or
result in the breakdown of pre-existing kaolinite (Velde,1977). The absence
of kaolinite in the Trent Formation mudstones is attributed to alkaline
conditions of weathering and deposition combined with restricted rainfall.

b) Neoformation in sediments:

The most commonly documented neoformational minerals found in evaporitic
sedimentary environments are smectite, palygorskite and sepiolite. Millot
(1970) described a lateral sequence of smectite-palygorskite-sepiolite in
peripheral Tertiary sedimentary basins of Western Africa (Fig.7.22). The
sequence reflects a gradual decrease in aluminium and concomitant increase in
magnesium content of the waters. Aluminium is the least soluble element in
solution, hence it is depleted preferentially (Millot,1970). Fontes et al.,
(1967) documented a similar, but vertical, neoformational sequence in the
wholly continental evaporite bearing (Fontes and Letolle,1976) Tertiary beds
of the Paris Basin. Like the African Tertiary Basins run-off from surrounding
landmasses washed amorphous silica and magnesium into the basin where pre-
cipitation of sulphate raised the Mg^{2+} : Ca^{2+} ratio resulting in the
neoformation of sepiolite.

Neoformational palygorskite has been reported associated with gypsum and
caliche cappings on the Egyptian coastal plain (Hassouba and Shaw, 1980).
McLean et al., (1972) reported neoformational palygorskite and sepiolite from
saline continental lakes, whilst Parry and Reeves (1968) documented a similar
occurrence of neoformational sepiolite.

Sepiolite and palygorskite have been previously recorded in the Mercia
Mudstones (Keeling, 1956; Freeman, 1964; Dumbleton and West, 1966;
Davis, 1967; Jeans, 1978) as well as from the Triassic red beds of France
(Millot, 1970; Lucas, 1962) and Morocco (Lucas, 1962).

c) Neoformation during diagenesis:

The formation of smectite, sepiolite and palygorskite very probably
continues during early diagenesis (Millot, 1970). Other diagenetic

Fig. 7.22.   Neoformation of smectite (Sm), palygorskite (Pa) and

sepiolite (Sep) in peripheral Tertiary basins in West

Africa. Detrital input is restricted to the basin margins

and is composed of kaolinite (K), illite (I) and

chlorite (Ch). The neoformational sequence represents a

gradual loss of Al^{3+} and increase in availability of Mg^{2+}

as one moves from the shore line into the basin (taken from

Millot, 1970).
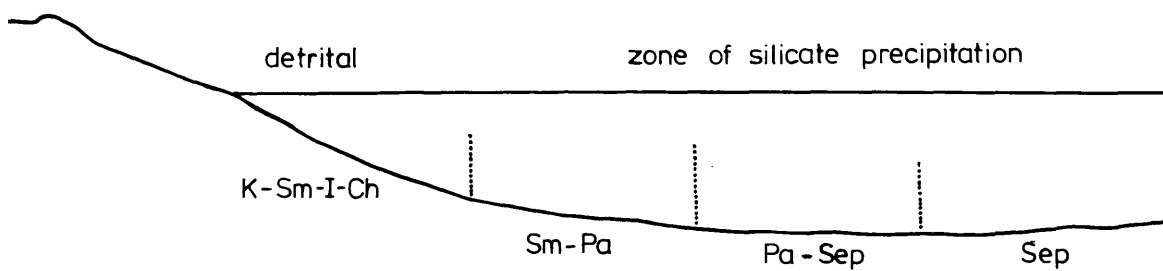
neoformational processes are of little interest to this study.

ORIGINS OF TRENT FORMATION CLAYS.

Each mineral will be discussed separately, but obviously inter-relationships between the different minerals do occur and these help in building up an overall model.

ILLITE.

Illite in the Triassic red-bed mudstone facies has been assigned a detrital origin by Lucas and Bronner (1961); Lucas (1962); Dumbleton and West (1966); Lucas and Ataman (1968); Millot (1970) and Jeans (1978). All these authors, with the exception of Jeans (1978), proposed that illite was initially degraded and that aggradation occurred on contact with hyper-saline brines. Jeans (1978) proposed that both illite and chlorite were of purely detrital origin and had not undergone any aggradational reactions.

However, the present study has revealed evidence of variations in the crystallinity of illite, suggesting that either the original detrital illite varied in crystallinity, or that aggradational reactions have occurred. The crystallinity of illite within a group of samples can be compared by measuring the width of the $10\overset{\circ}{A}$ peak at half its height (Esquevin, 1969; Dunoyer de Segonzac, 1970). As the crystallinity improves, with illite tending towards well ordered mica, the peak shape sharpens hence narrowing the half peak width. Crystallinity is plotted against the intensity ratio of the $5\overset{\circ}{A}$ and $10\overset{\circ}{A}$ peaks (I (002) / I (001)). This qualitative ratio gives an indication of the chemical composition of the octahedral layer. Average $10\overset{\circ}{A}$ illite has one-eighth of its octahedral cations in the form of Mg^{2+} or Fe^{2+} whereas ideal muscovite contains only aluminium in its octahedral sites (Weaver and Pollard, 1973). The (I (002) / I (001)) ratio varies as an inverse function of the Fe^{2+} and Mg^{2+} content of the octahedral layer (Esquevin, 1969).

Illite crystallinities and intensity ratios from Newark, Keyworth,

East Leake and Fauld are plotted in Fig. 7.23. Illite crystallinities fall into two groups with only slight overlap:

    1) Width of 001 peak at half height is greater than 9 mm.

    2) Width of 001 peak at half height is less than 9 mm.

Surprisingly the groups show a distinct stratigraphic zonation with group 1 being composed solely of analyses from the Hawton Member at Newark, Keyworth and Fauld, whilst group 2 is composed solely of analyses from the Fauld Member. The majority of samples have an (I (002) / I (001)) ratio of between 0.4 and 0.7 with no significant trend towards enrichment of octahedral aluminium. The spread in intensity ratios may reflect variations in the composition of the original detrital material with four samples with high value intensity ratios representing highly aluminous detrital material.

The variations in crystallinity could be due to either:-

    a) variations in degree of weathering of detrital illite prior to deposition; the greater the weathering then the poorer the detrital illite crystallinity, which, unless aggradation processes sympathetically increase will be reflected in the crystallinity measurements. Weathering is strongly controlled by climate, hence the decrease in illite crystallinity could reflect a change in palaeoclimate, possibly as increased rainfall during deposition of the Hawton Member.

    b) variations in particle size; finer grained particles are more sensitive to degradation and hydration during weathering and transport than larger detrital particles (Dunoyer de Segonzac, 1970). Alterations in particle size could result from a change in weathering processes, particle source or mode of particle transport, such as aeolian to aquatic.

    c) variations in degree of aggradation, of which a major control would be the availability of K^+ compared to the volume of degraded illite. Solutions rich in potassium could produce more highly crystallized illite from degraded detrital illite. Hence the Fauld Member may have had K^+ rich pore fluids with a high K^+ to volume of degraded illite ratio, whilst the Hawton Member had low availability of potassium in relation to the volume of

Fig. 7.23.   Plot of illite crystallinity (width of 001 peak at $\frac{1}{2}$ height)
and the ratio of intensities of the 002 and 001 peaks. The
illites of the Fauld Member are more crystalline than those
of the Hawton Member. The majority of the illites have
very similar intensity ratios indicating that the Fe^{2+} and
Mg^{2+} content of the octahedral layer does not significantly
change (see text).
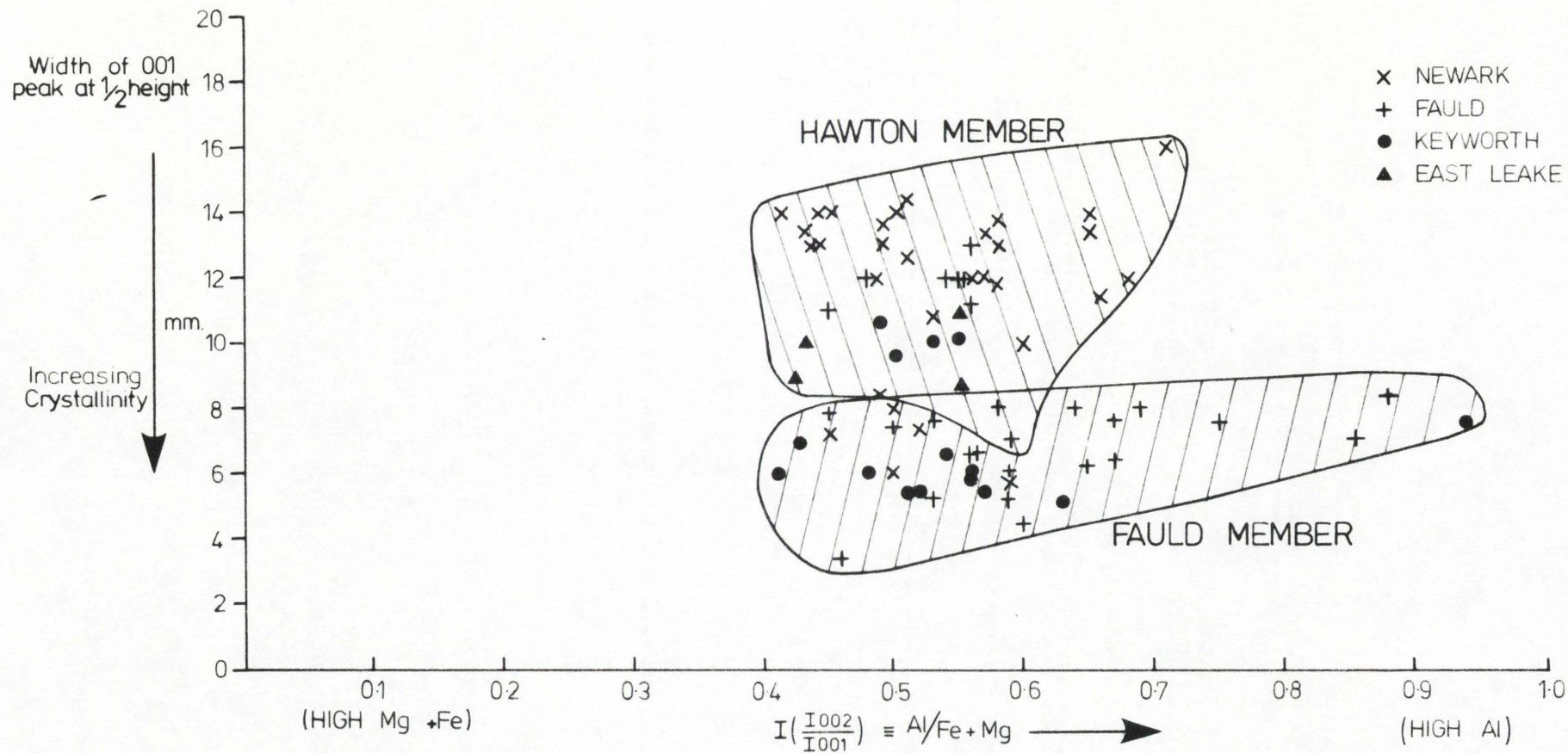
degraded illite.

It is unlikely that the variation in crystallinity is due to late burial diagenesis as this would enhance the crystallization of both Hawton Member and Fauld Member illites in a similar manner. Further discussion of reasons for the variations in illite crystallinity is left until after a discussion on the other clay minerals present.

CHLORITE.

Chlorite may be of detrital origin (Lucas, 1962; Lucas and Ataman,1968; Jeans, 1978), formed by dehydration of smectite in a magnesium-rich environment (Shaw, 1980; Dunoyer de Segonzac, 1970) (Fig. 7.21), or formed by magnesium rich solutions transforming degraded illite (Lucas, 1962). In the latter two formation processes chlorite is the end-member of a transformation reaction series, which as previously discussed have irregular and regular mixed-layer stages. Whilst irregular mixed-layer smectite/chlorite is associated with illite and chlorite in the Trent Formation, the intermediate corrensite phase has not been recorded in the present study. The absence of corrensite and the lack of an inverse relationship between irregular mixed-layer content and chlorite content argues against chlorite having a transformational origin. Therefore, chlorite in the Trent Formation was probably derived from weathering of surrounding landmasses and, like illite initially degraded chloritic material underwent an aggradation process on contact with magnesium-rich solutions present in the depositional basin.

SMECTITE/CHLORITE.

Many authors favour an aggradational origin for mixed-layer minerals associated with evaporites. Füchtbauer and Goldschmidt (1956, 1959) recorded transformational corrensite from the Zechstein, Germany; Grim, Droste and Bradley (1960) and Bodine (1978) recorded irregular smectite-chlorite, corrensite and irregular vermiculite-chlorite from the Permian, New Mexico; Lucas (1962), Lucas and Ataman (1968), Lippmann and Savasçin (1969) and

Millot (1970) record a variety of transformational mixed-layer minerals from the European Triassic.

An alternative neoformational origin for the mixed-layer minerals in the British Triassic was proposed by Jeans (1978). To support his proposal Jeans (1978) cited the following as evidence precluding all origins except neoformation:-

1) petrographic studies showed diagenetic growth of "exotic" mixed-layer minerals;

2) the absence of clay mineral phases representing the mineralogical intermediate stages between the detrital assemblage of mica and chlorite and the associated mixed-layer clay minerals;

3) the absence of chlorite-dominated assemblages;

4) detrital mica and chlorite occur invariably as distinct phases in samples containing the exotic assemblage;

5) a conspicuous correlation between clay mineral assemblages and lithologies of the Mercia Mudstone Group of the South Devon coast;

6) the absence of any relationship between illite content and corrensite content.

Whilst each of these points appears to support Jeans' hypothesis, critical examination of each proves otherwise.

1) Jeans recorded small irregular plates encrusting both euhedral carbonate, generally dolomite, rhombs and large detrital mica and chlorite plates. Jeans identified these plates and other small, thin, ragged crystals as being smectite/mica. The mixed-layer mineral is associated with well formed fibrous sepiolite and palygorskite. Jeans concluded that these minerals were formed by post-depositional early diagenetic crystallization, a conclusion with which the author agrees. However, this statement does not differentiate whether smectite/mica had a neoformational or aggradational origin, as both processes occur during early diagenesis. One might expect neoformational minerals to exhibit a higher degree of crystallinity than minerals formed by aggradation. For example, neoformational palygorskite and

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sepiolite typically occur as delicate, well formed fibres (Jeans, 1978; Hassouba and Shaw, 1980). The "irregular plates" and "small, thin ragged crystals" of smectite/mica (Jeans, 1978) are perhaps more typical of aggradation reactions, especially as smectite/mica is one of the early stages in the transition from degraded illite to chlorite (Lucas, 1962). Lucas (1962) records beautifully euhedral transformational chlorite crystals from the French Triassic indicating that mineral crystallinities improve as aggradation proceeds along the transformational series. Jeans (1978) failed to find petrographic evidence for neoformational growth of smectite/chlorite, corrensite or chlorite, all of which he postulated to have a neoformational origin.

The author feels that the smectite/mica textures described by Jeans (1978) could equally well result from early diagenetic aggradation of degraded detrital illitic lattices as proposed by Lucas (1962).

2) Jeans (1978) stated that the absence of intermediate clay mineral phases between the detrital illite and chlorite phase and the mixed-layer clays is evidence that the mixed-layer clays did not form by transformation of detrital illite and chlorite. The transformational reaction series of Lucas (1962) and Lucas and Ataman (1968) commences with degraded illite transforming to open illite. A parallel reaction series involves degraded illite aggrading to well crystallized illite (Lucas, 1962). Unfortunately Jeans (1978) did not undertake any measurements of illite crystallinity, but the present author has shown that significant variations in illite crystallinity exist within the Mercia Mudstone Group (Fig. 7.23). Such variations could represent the parallel reaction series, although it admittedly occurs in a vertical rather than horizontal direction. The illite in the Hawton Member may correspond to Lucas's open-illite phase. In order to observe any possible change from degraded to open illite one would need to sample close to the sediment source (? London Landmass).

Smectite/mica, the next mineral of Lucas's series, is recorded from the Mercia Mudstone Group on the South Devon coast (Jeans, 1978) (Fig. 7.24).

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Fig. 7.24.   Map of distribution of clay mineral assemblages in the late Triassic sediments of England and Wales. Smectite/ mica is confined to the South Devon coast and the relationship of this assemblage to the corrensite and smectite/ chlorite assemblages to the north is uncertain. Note that corrensite and chlorite confined to graben sequences (Stowell Park and Wilkesley boreholes) whilst smectite/ chlorite is common on the horst areas (Newark and Upton borehole). Clay mineral assemblages outside the study area taken from Jeans (1978).

CORR
+
CHL
3 •

+ NEWARK

FAULD
+ MINE
SM/CHL

?

?

?

?

CORR
+
SM/CHL
1 • • 2
SM/CHL

?

?

?

?

SM/MICA

S. DEVON
COAST

N

0   km.   60
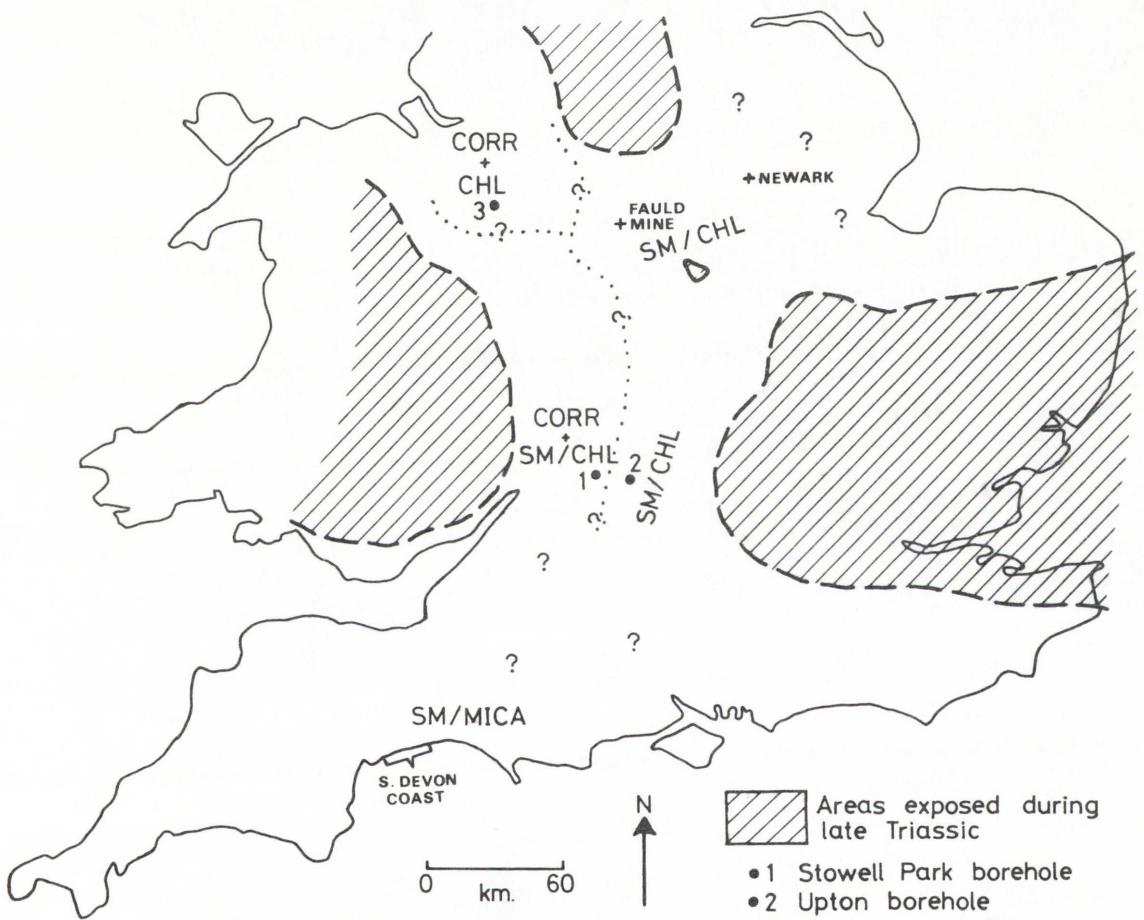
Areas exposed during
late Triassic

• 1 Stowell Park borehole
• 2 Upton borehole
• 3 Wilkesley borehole

CHL = Chlorite
CORR = Corrensite
SM/CHL = Smectite/Chlorite
SM/MICA = Smectite/Mica

As one moves northwards to the Gloucester and Oxford area smectite/mica is replaced by irregular smectite/chlorite and corrensite (Jeans, 1978) (Fig. 7.24) which are the subsequent two phases of Lucas's series. The intermediate stage between smectite/chlorite and smectite/mica is represented by a broad diffuse reflection between 10$\overset{\circ}{A}$ and 14$\overset{\circ}{A}$ (Lucas, 1962). Jeans did not record this type of reflection which, he argues, indicates that trans-formational reactions have not occurred. However, when one considers that Jeans did not sample between the smectite/mica-rich South Devon area and the smectite/chlorite-rich Gloucestershire area, his conclusion becomes somewhat tenuous in nature. Lucas' mineral transformation series is a lateral phenomena and so consequently one should sample between South Devon and Gloucestershire to elucidate the relationship between smectite/mica and smectite/chlorite. An interesting point is the distribution of irregular smectite/chlorite and corrensite. Corrensite is restricted to a graben sequence cored in the Stowell Park borehole (Audley-Charles, 1970b; Jeans, 1978) (Fig. 7.24), whilst irregular smectite/chlorite occurs in the Stowell Park borehole and the Upton borehole (Fig. 7.24) which cored a thin Triassic sequence lying close to the London Landmass (Audley-Charles,1970b; Jeans, 1978). A similar distribution of these two minerals occurs between the present study area which is characterized by the predominance of irregular smectite/chlorite and the corrensite-rich graben sequence of the Cheshire Basin (Jeans, 1978) (Fig. 7.24).

Lucas (1962) postulated that with increasing brine salinity corrensite would progressively replace irregular mixed-layer smectite/chlorite. The Triassic grabens of Britain were, in general, areas of increased salinity as evidenced by extensive halite deposits (Audley-Charles, 1970b). Surrounding areas, for example Upton, Oxfordshire and the present study area, were influenced by brines of an overall lower salinity, indicated by sulphate deposits and/or scanty carbonate. Therefore, the distribution of corrensite and irregular smectite/chlorite in the Mercia Mudstone Group correlates with palaeo-salinity patterns, in agreement with the transformational mineral

series model of Lucas (1962).

The final two phases of Lucas' transformational series are irregular mixed-layer chlorite/swelling chlorite and well crystallized chlorite. The term swelling chlorite was restricted by Lucas (1962) to a chlorite type mineral in which the octahedral layer is incomplete and allows a certain degree of expansion whilst maintaining the heat stability of chlorite. Swelling chlorite was originally identified in the Mercia Mudstone Group by Stephen and MacEwen (1950, 1951) and subsequently by Dumbleton and West (1966) and Davis (1967). Honeybourne (1951) identified a similar mineral as a mixed-layer vermiculite-chlorite. Swelling chlorite as defined by Lucas (1962) was not identified by Jeans (1978) or the present author. Interestingly, one of Stephen and MacEwen's (1951) swelling chlorite samples came from Abbots Bromley, Staffordshire, 9 km. west of Fauld Mine. They did not record the presence of mixed-layer smectite/chlorite, which is extemely common in Fauld Mine. The X.R.D. traces of the two minerals are very similar in untreated and glycerolated samples. If swelling chlorite is a mixed-layer smectite/chlorite as suggested by Jeans (1978), then the characteristic swelling chlorite 14Å line present after heating at 400° C. may be the 14Å chlorite line, with powder camera techniques failing to reveal the broad 10 to 14Å peak of collapsed irregular smectite/chlorite (Fig. 7.10).

If swelling chlorite in the Mercia Mudstone Group is in fact mis-identified mixed-layer smectite/chlorite, then the chlorite/swelling chlorite phase (Lucas, 1962) has yet to be positively identified. Its apparent absence may reflect a real mineralogical difference from Lucas' transformational series or be simply due to differing methods of sample preparation and interpretation. At present it is not possible to be any more precise.

Abnormally high concentrations of well crystallized non-detrital chlorite are present in the Mercia Mudstone Group of the Cheshire Basin (Jeans, 1978) (Fig. 7.24). Jeans (1978) favoured a neoformational origin for the chlorite but it could equally well represent the final phase of Lucas' mineral series. The restriction of well crystallized chlorite to the

Cheshire Basin, which as previously stated was an area of increased palaeo-salinity, agrees with the clay mineral distribution predicted by the transformational model (Lucas, 1962).

To summarize, the author believes that the Mercia Mudstone Group contains most phases of the degraded illite to chlorite aggradational series (Lucas, 1962). Unidentified phases are (1) degraded illite, (2) a poorly crystallized intermediate phase between smectite/mica and smectite/chlorite and (3) irregular chlorite/swelling chlorite. The absence of the former two may well result from very widely spaced sample points, whilst the absence of the latter may be due to either mineralogical differences or differing methods of sample preparation and interpretation. The degree of aggradation reflects the availability of Mg^{2+} with the chlorite "end-member" being confined to strata deposited in hypersaline conditions.

3) Non-detrital chlorite has been identified in the Mercia Mudstone Group of the Cheshire Basin (Jeans, 1978) but as stated by Jeans (1978) it does not dominate the total clay mineral assemblage. In the Northwich Halite Formation (Warrington et al., 1980) chlorite is the only non-detrital clay mineral present in 80% of the samples, whereas the Wilkesley Halite Formation (Warrington et al., 1980) contains both chlorite and corrensite (Jeans, 1978). Therefore, chlorite dominates the non-detrital clay assemblage in the Northwich Halite Formation, whilst corrensite dominates the Wilkesley Halite Formation non-detrital clay fraction. This difference may reflect variations in the degree of transformation of corrensite to chlorite; in the Northwich Halite Formation transformation has proceeded to completion whilst in the Wilkesley Halite Formation transformation ceased prior to the total conversion of corrensite to chlorite.

The absence of a chlorite dominated total clay assemblage does not provide evidence against the transformational origin of chlorite. Millot (1970) when referring to the Triassic of the Jura Basin stated "chlorite is never abundant and is found in thin mudstone horizons between halite beds. Corrensite occurs within the halite".

4) The presence of detrital mica and chlorite as distinct phases in samples containing the exotic assemblage was stated (Jeans, 1978) as evidence that the exotic assemblage did not form by transformation from the detrital assemblage. This would be correct if Lucas (1962) and Lucas and Ataman (1968) had originally stated that detrital illite and chlorite were transformed to form exotic mixed-layer clays. Instead Lucas (1962) and Lucas and Ataman (1968) proposed that the exotic clays formed by aggradation of more highly degraded illite in an Mg^{2+} -rich environment, whilst the least degraded illites adsorped potassium to form more highly crystalline illite. Therefore mixed-layer clays reflect both input of highly degraded detrital illite and the presence of Mg^{2+} -rich brines.

5) Jeans claimed to find "a conspicuous correlation between mineral assemblages and lithologies of the South Devon coast". Throughout much of the sequence massive mudstones contain smectite/mica whilst carbonate-rich units and immediately underlying mudstones contain sepiolite and minor smectite/mica. Sandstone units only contained illite and chlorite. Whilst sepiolite is of restricted distribution, smectite/mica is not related to any particular lithology. Its absence from the sandstone units probably reflects an alteration of depositional environment, either in terms of a restricted supply of highly degraded illitic detritus or an Mg^{2+} -deficient brine regime which prevented aggradation of degraded illite. The formation of sepiolite is discussed later.

6) The absence of any relationship between illite and corrensite argues against the corrensite-forming reaction of Lippmann and Savaşcin (1969). However, this statement does not preclude a transformational origin for corrensite as discussed in section 2.

The author proposes that the arguments put forward by Jeans (1978) do not disprove a transformational origin for the mixed-layer clays and secondly do not provide any proof of a neoformational origin for these clays.

A TRANSFORMATIONAL ORIGIN FOR SMECTITE/CHLORITE.

A plot of illite-chlorite-mixed-layer content (Fig. 7.25) illustrates that the illite and mixed-layer contents are the major variables. The relationship of detrital illite and chlorite to mixed-layer mineral is shown in Fig. 7.26. The illite:chlorite ratios for Fauld Member samples lie between 2 and 5 with an average of 3.95 (Fig. 7.27a) whilst the Hawton Member illite:chlorite ratios lie between 4 and 25 with an average of 11 (Fig. 7.27b). The remarkable grouping of the Fauld Member illite:chlorite ratios and their independance from smectite/chlorite content (Fig. 7.26) could be interpreted as evidence of smectite/chlorite having a neoformational origin with the relative input of detrital illite to detrital chlorite being constant throughout the deposition of the Fauld Member.

However, the highly variable illite:chlorite ratios in the Hawton Member indicate marked changes in the composition of the detrital input. This could be due to a change of source or a change in weathering patterns. The major sediment sources in the Upper Triassic are the London Landmass, Pennines and Welsh Landmass (Audley-Charles, 1970b) (Fig. 7.24). These were exposed throughout deposition of the Trent Formation and no other indications of sudden variations in sediment source and supply are known. Similarly weathering patterns must have drastically altered from the "steady-state" conditions of the Fauld Member to produce such variable input during the Hawton Member. Stable isotope studies have shown that an alteration from marine to continental brine regime occurred at the base of the Hawton Member (Chapter 5). Although this change may have been related to an alteration in climate, no other evidence has been found to support the variations in weathering needed to produce the widely spread illite:chlorite ratios in the Hawton Member. Therefore the sediment supply very probably had broadly similar characteristics throughout deposition of the Trent Formation. This implies that the detrital material must have been modified subsequent to deposition, very probably by transformational reactions so producing variable illite:chlorite ratios. If this was the case then smectite/chlorite may have

Fig. 7.25.   Plot of illite, chlorite and smectite/chlorite contents in

the Hawton and Fauld Members and Edwalton Formation in the

East Midlands.   The Fauld Member and Edwalton Formation

contains generally higher smectite/chlorite contents than

the Hawton Member.   Chlorite is nearly constant throughout.

SMECTITE/CHLORITE

+ FAULD
● NEWARK
△ KEYWORTH
□ EAST LEAKE

50

Fauld Member &
Edwalton Formation

50

Hawton Member

ILLITE

50
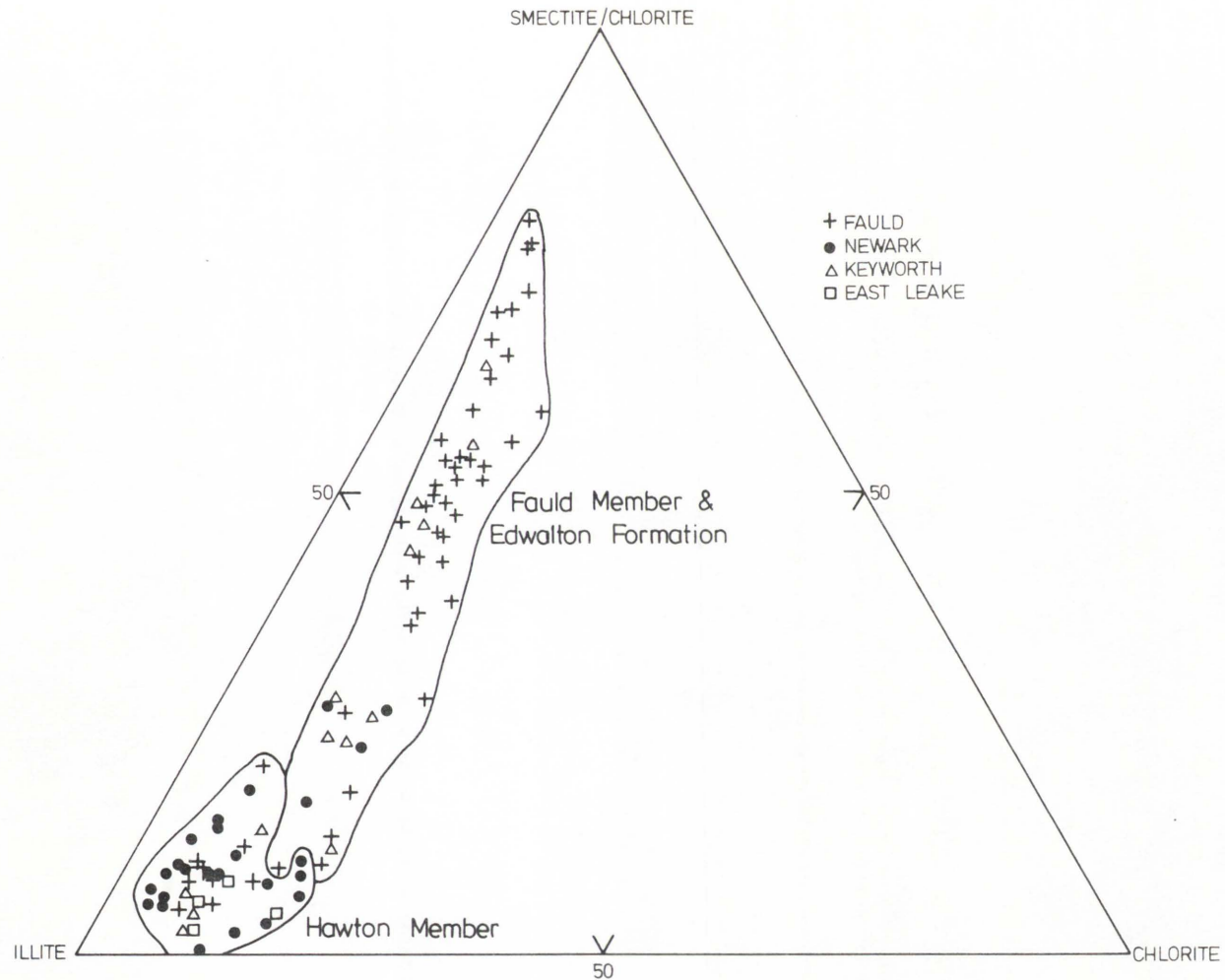
CHLORITE

Fig. 7.26. Plots of illite: chlorite ratio and smectite/chlorite content for samples from Newark borehole, Keyworth borehole and Fauld. Note the remarkable grouping of samples from the Hawton Member and the Fauld Member and Edwalton Formation. This may reflect changes in the Mg : K concentration ratio of the brine regime with smectite/chlorite forming in magnesium-rich brines and illite in potassium-rich brines (see text).

NEWARK BOREHOLE

ILLITE : CHLORITE RATIO

HAWTON MEMBER

FAULD MEMBER

% SMECTITE/CHLORITE CONTENT

KEYWORTH BOREHOLE

ILLITE:CHLORITE RATIO

HAWTON MEMBER

FAULD MEMBER & EDWALTON FORMATION

% SMECTITE / CHLORITE CONTENT

INCREASING AVAILABILITY OF POTASSIUM
Low Mg:K conc^n ratio

ILLITE: CHLORITE RATIO

HAWTON

FAULD MEMBER & EDWALTON FORMATION

FAULD

▲ 193 BOREHOLE
+ FAULD MINE
● FAULD MINE BOREHOLE

% SMECTITE / CHLORITE CONTENT

INCREASING AVAILABILITY OF MAGNESIUM
(High Mg:K conc^n ratio)

Fig. 7.27a.   Illite : chlorite ratios in the Hawton Member.   These

show a widespread distribution with a mean of 10.97.

Fig. 7.27b.   Illite : chlorite ratios in the Fauld Member.   These

show a very well defined gaussian distribution with a

mean of 3.95.

HAWTON MEMBER
Mean 10·97

FAULD MEMBER
Mean 3·95
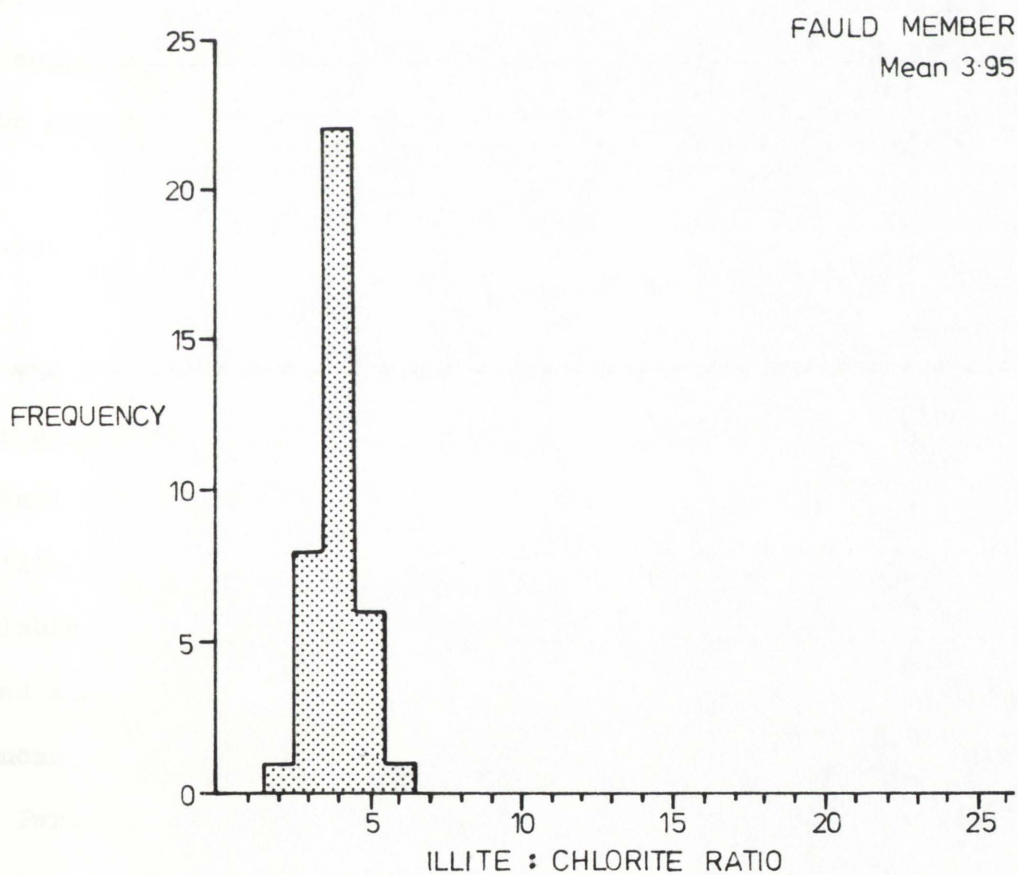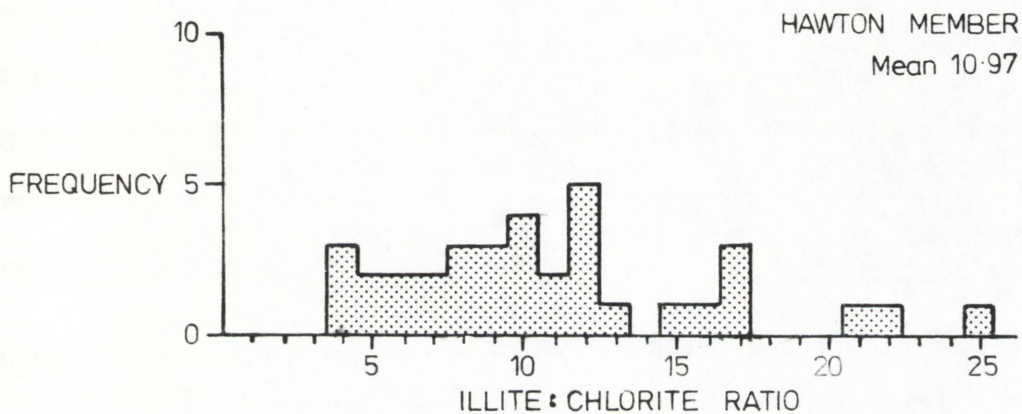
formed by transformation of highly degraded illite (Lucas, 1962; Lucas and Ataman, 1968) instead of by neoformation.

In this model the generation of smectite/chlorite is dependent on the presence of Mg-rich brines, as in the neoformational model, and on a supply of degraded illite. The detrital sediment input can now be divided into three major phases; highly degraded illite, moderately degraded illite and chlorite. In Mg-rich environments highly degraded illites form mixed-layer clays, whilst moderately degraded illites adsorb potassium to form well crystallized illites (Lucas, 1962).

FAULD MEMBER.

The Fauld Member has a constant illite:chlorite ratio (Fig. 7.27a) which, as previously stated, indicates that moderately degraded illite and chlorite were supplied at a constant ratio. From this statement one can make either of two assumptions:

a) the highly degraded illite to moderately degraded illite ratio was also approximately constant, (better expressed as HDI:MDI ratio);

b) the HDI:MDI ratio varied.

The latter is unlikely as it is improbable that the HDI:MDI ratio should alter when the moderately degraded illite:chlorite ratio stays approximately constant (Fig. 7.26). If the HDI:MDI ratio was constant then the increasing smectite/chlorite content must result from the initial aggradation of available highly degraded illite and subsequent enlargement of the newly formed smectite/chlorite crystallites. This model is similar to that proposed by Lucas (1962).

Periods of calcium sulphate precipitation lower the Ca:Mg ratio thereby enhancing the availability of magnesium for formation of mixed-layer minerals. This is illustrated by the high (50 to 70%) smectite/chlorite contents associated with the Tutbury Sulphate Bed (Figs. 7.14, 7.15 and 7.16). Increases in smectite/chlorite content therefore probably reflect rises in the availability of magnesium.

The Fauld Member was deposited in a continental brine regime at East Leake and a marine brine regime at Fauld, Keyworth and Newark (Chapter 5). Smectite/chlorite is not restricted to one brine type, perhaps indicating that both continental and marine brines had high magnesium concentrations.

HAWTON MEMBER.

The Hawton Member has a variable illite:chlorite ratio (Fig.7.27b) which as previously discussed is unlikely to represent original variations in the mildly degraded illite:chlorite ratio. It is proposed that the relative increases in illite content have resulted from highly degraded illites adsorbing potassium instead of magnesium and forming illite as opposed to mixed-layer clays. Magnesium is adsorbed in preference to potassium when the Mg:K concentration ratio is high; at lower Mg:K concentration ratios potassium is preferentially adsorbed (Powers, 1959). The presence of minor smectite/chlorite indicates that the Mg:K concentration ratio was initially high enough for magnesium adsorption to occur. If adsorption took place after burial, when the magnesium reservoir could not be replenished, the Mg:K concentration ratio would fall until a level was reached at which potassium would be preferentially adsorbed. The low smectite/chlorite content suggests that the Mg:K concentration ratio did not remain at a level sufficient for large scale adsorption of magnesium. This may be a reflection of the continental origin of the Hawton Member brines (Chapter 5) as average continental water has a Mg:K concentration ratio approximately half that of marine water (Fyfe, Price and Thompson, 1978). The relatively poor crystalline nature of the Hawton Member illites may be due to the brine potassium concentration being insufficient to aggrade all the degraded illite lattices to highly crystalline illite.

COMPOSITION OF ORIGINAL DETRITUS.

The initial composition of the degraded detritus is problematical. It is likely that the mildly degraded illite:chlorite ratio lay between 3:1 and

6:1 as the only ways of lowering an initially high illite:chlorite ratio are by transformation of a mineral into chlorite, a process not documented in the present study, or by illite remaining in such a highly degraded state that the X-rays were not diffracted, which is rather unlikely. Similarly the highly degraded illite input must have lain between 10 and 20%, equivalent to the minimum content of mixed-layer mineral present in the Fauld Member. If the input of highly degraded illite had been higher than 10 to 20% the excess degraded illite not transformed to mixed-layer mineral either remained in its original state or was transformed to illite. The former is unlikely, whilst the latter would raise the illite:chlorite ratio which, as previously stated, is nearly constant in the Fauld Member.

The plot of illite:chlorite ratio against mixed-layer content can be thought of as a measure of the effect of varying brine composition on the degraded detrital material (Fig. 7.26). Increased levels of smectite/ chlorite are indicative of raised magnesium concentrations (high Mg:K ratio) whilst higher illite:chlorite ratios indicate the increased adsorption of potassium when the brines had low Mg:K ratios.

SEPIOLITE AND PALYGORSKITE.

The commonly recorded delicate nature of palygorskite and sepiolite fibres in evaporitic sediments argues that these are neoformational minerals as it is difficult to envisage how such delicate fabrics could have been preserved during transport and deposition (Hassouba and Shaw, 1980; Jeans, 1978). A neoformational origin has been invoked for many other occurrences of these minerals (Millot, 1970; McLean et al., 1972; Fontes et al., 1967; Lucas, 1962). Therefore, although no evidence was found in this study to argue for or against neoformation, it is very likely that sepiolite and palygorskite in the East Midlands Triassic are of neo- formational origin, in agreement with Jeans' conclusions (1978).

Experimental work provides some interesting data on the chemical conditions required to precipitate sepiolite at 25° C. 1 atm. (Siffert,1962;

Siffert and Wey, 1962; Wollast et al., 1968). The major requirements are:

    a) the pH of the solution from which sepiolite precipitates lies between 8.5 and 9. At higher pH values magnesian smectite and then talc are precipitated. This corresponds to saline basins which are known to have sufficient concentrations of magnesium and silica, but where sepiolite is absent because the solution pH is too high (Velde, 1977);

    b) the silica concentration in aqueous solution lies between 20 and 150 ppm. The silica will be in an amorphous form as solutions are saturated with respect to quartz when they contain 10 to 15 ppm SiO_2 at 25° C. (Velde, 1977; Siever, 1962).

It was shown that sepiolite precipitates with varying Mg:Si molecular ratios; Siffert (1962) used solutions with Mg:Si ratios of 1.43 and 0.70 whilst Wollast et al., (1968) used solutions with an Mg:Si ratio of 140.

In the present study sepiolite is recorded at three horizons; the Tutbury Sulphate Bed (Fauld Member) at East Leake Mine (Fig. 7.14), immediately below the Windmill Skerries (Fauld Member) at Keyworth (Fig.7.12), within the Edwalton Formation underlying Fauld Mine (Fig. 7.17). By comparison with the experimental work the brines associated with the deposition and early diagenesis of these horizons probably had a pH between 8.5 and 9, a high amorphous silica and possibly a high Mg^{2+} content. It is interesting to note the restricted occurrence of sepiolite within the strata studies; for example sepiolite is present within the Tutbury Sulphate Bed at East Leake and not within the laterally equivalent strata at Fauld, Keyworth and Newark. During deposition of the Tutbury Sulphate Bed the latter localities were dominated by marine brines, whereas East Leake had a continental brine regime (Chapter 5). The marine and continental brines may very well have had differing chemistries with amorphous silica content and pH being of particular interest. The weathering of silicate minerals is a major source of amorphous silica in continental groundwaters (Brownlow, 1979). The dissolution of quartz is only of importance at pH > 9 as evidenced by corrosion of detrital quartz in South Australian alkaline lakes (Peterson and von der Borch, 1965).

As sepiolite is not precipitated at such high pH values (Velde, 1977) the latter process was probably unimportant during deposition of the Trent Formation. The distribution of sepiolite may therefore be a reflection of the availability of silica derived from the weathering of adjacent landmasses. The areas with a marine brine regime were not enriched in silica by continental run-off, hence sepiolite was not precipitated.

The second major variable affecting sepiolite formation is pH, which as previously stated, must lie between 8.5 and 9 for sepiolite to precipitate. It is unlikely that the pH was higher than 9 when smectite and talc are precipitated (Velde, 1977); neither mineral is recorded in the present study. It is possible that the pH of the marine brine regime was insufficient to promote sepiolite neoformation.

Neoformation of palygorskite probably occurs under similar conditions to that of sepiolite. The more aluminous nature of palygorskite indicates that the brines had higher Al^{3+} concentrations (Velde, 1977); palygorskite forms when the Al^{3+} : Mg^{2+} ratio is about 1, whereas sepiolite forms when the Al^{3+} : Mg^{2+} ratio is <1 (Velde, 1977; Weaver and Beck, 1977). Aluminium is poorly soluble, hence a solution containing Mg^{2+} and Al^{3+} will initially precipitate aluminous palygorskite in preference to magnesia-rich sepiolite (Millot, 1970). The development of palygorskite in the East Leake area may therefore be a reflection of the proximity of the source of the Si and Al which, in all probability was the Charnwood Massif (Fig. 7.4). Its association with sepiolite (Fig. 7.14) indicates variable Al^{3+} concentrations in the brines. As the formation of palygorskite is related to Al^{3+} availability so sepiolite formation in the strata studied may have been controlled by silica availability rather than brine pH. The absence of these minerals in the Hawton Member may reflect an overall decrease in Al^{3+}, amorphous silica and Mg^{2+} concentrations within the brines or a decrease in brine pH.

Jeans (1978) proposed that clay mineral neoformations in the Mercia Mudstones resulted from chemical reactions occurring between two water masses,

one of which was hypersaline and overlain by a mass with a salinity near to that of unevaporated sea water. The hypersaline mass was enriched in silica whilst the overlying mass provided the required Mg^{2+}. Mixing of the two masses resulted in the precipitation of minerals such as sepiolite; according to Jeans (1978), a reaction comparable with the experiments of Wollast et al.,(1968). Lateral variations in water chemistry and salinity resulted in the neoformation of a series of mixed-layer clay minerals (Jeans, 1978).

However, it has already been shown that the mixed-layer minerals may have formed by transformation. The author believes that such a complex brine mixing model is unnecessary to account for the neoformation of sepiolite and palygorskite. Millot (1970) summarized the experimental work of Siffert and Wey (1962) as "the synthesis of sepiolite is achieved at ordinary temperatures by means of an increase in the Mg^{2+} concentration in siliceous solutions." The continental brines running off the Charnwood Massif contained both amorphous silica and Mg^{2+} with Al^{3+}. Subsequent evaporation raised silica, Mg^{2+} and Al^{3+} concentrations and when calcium sulphate precipitation occurred the Mg^{2+} concentration was raised to a level sufficient to initiate precipitation of sepiolite and palygorskite as proposed for other occurrences of these minerals (Fontes et al., 1967; Hassouba and Shaw, 1980). This reaction is particularly well demonstrated by the association of calcium sulphate, palygorskite and sepiolite within the Tutbury Sulphate Bed at East Leake.

CONCLUSIONS.

The Edwalton and Trent Formations of the Midlands contain ubiquitous illite and chlorite with variable proportions of irregular mixed-layer smectite/chlorite, sepiolite and palygorskite. The distribution of smectite/ chlorite is used to sub-divide the Trent Formation into the smectite/ chlorite-rich Fauld Member and the overlying smectite/chlorite-poor Hawton Member (Chapter 2).

Illite and chlorite were derived from the weathering of adjacent landmasses and were deposited as a detrital assemblage in varying states of degradation. After deposition the degraded illite and chlorite were in contact with predominantly hypersaline brines and underwent aggradation. Degraded chlorite was aggraded by adsorption of Mg^{2+}. The aggradation of illite was more complex and depended on the degree of degradation and the brine composition.

During deposition of the Fauld Member predominantly marine brines contained an abundance of Mg^{2+} with a high Mg:K concentration ratio. The more highly degraded open-lattice illite preferentially adsorbed Mg^{2+} and underwent a transformational reaction forming irregular mixed-layer smectite/chlorite. When the brines had a very high Mg^{2+} content, for example during the precipitation of thick sulphate horizons, the initially small smectite/chlorite crystallites increased in size; a process which borders on neoformation. This resulted in high levels of smectite-chlorite associated with major sulphate development. More moderately degraded illite adsorbed potassium to form well crystallized illite.

During deposition of the Hawton Member brines of continental origin had a restricted supply of Mg^{2+} and a low Mg:K concentration ratio. Highly degraded illite initially adsorbed Mg^{2+} to form mixed-layer smectite/chlorite. This lowered the Mg:K concentration ratio still further and resulted in the preferential adsorption of potassium and the formation of illite. The moderately degraded illite also adsorbed potassium, improving its crystallinity. However, the overall crystallinity of these illites is lower than that of illites associated with Mg^{2+}-rich brines. This reflects K^+ supply and demand; in the Hawton Member brine regime a higher proportion of degraded illite adsorbed K^+ than in the Fauld Member brine regime. The Hawton Member regime may have had a lower K^+ concentration despite its lower Mg:K concentration ratio.

Sepiolite and palygorskite formed by direct precipitation from brines with high Mg^{2+} and amorphous silica contents and pH values between 8.5 and 9.

Palygorskite formed when the Al^{3+}:Mg^{2+} ratio~1 whereas sepiolite formed when Al^{3+}:Mg^{2+}< 1. Within the Fauld Member the distributions of palygorskite and sepiolite were controlled by the concentration of aluminium and amorphous silica respectively, both of which were derived from the weathering of adjacent landmasses such as the Charnwood Massif. The poor solubility of Al^{3+} very possibly restricted palygorskite formation to areas more closely adjacent to the landmasses. The absence of these minerals in the Hawton Member may reflect a decrease in brine pH and/or a decrease in the concentrations of Mg^{2+}, Al^{3+} and amorphous silica.

CHAPTER 8.

CONCLUSIONS.

CONCLUSIONS.

Each chapter has its own set of conclusions to which the reader is referred. The following is an amalgamation and summary of those conclusions.

A new approach to the problem of correlation of fine-grained unfossiliferous red-beds is illustrated in Chapter 2. The lithologically monotonous Trent Formation was subdivided on the basis of coincident variations in dolomite oxygen isotope geochemistry and clay mineralogy into the Fauld Member and overlying Hawton Member. The Fauld Member is characterized by the presence of a Mg-rich clay suite ($>15\%$) and minor illite and chlorite. The Hawton Member contains illite, chlorite and minor Mg-rich clays ($<15\%$). The $\delta^{18}O$ dolomite values indicated that the brine regime of the Fauld Member was predominantly marine in grabens and horst margin areas, whilst the centre of the horst block was continental. The Hawton Member had a predominantly continental brine regime. The correlation was tentatively extended into Cheshire, Gloucestershire and Oxfordshire.

The Upper Triassic of the East Midlands was deposited on an extensive peneplain which had areas of upstanding minor relief (Charnwood Massif, Pennines, London Landmass). Horst and graben structures formed during the early and middle Triassic influenced deposition of late Triassic sediments. Sedimentological studies showed that four major facies can be recognized within the East Midlands Upper Triassic;

   (a) marginal facies of breccias, sandstones and siltstones associated with the gradual burial of wind-eroded stoney desert surfaces which had considerable topographic relief;

   (b) red-bed mudstone and siltstone facies, interpreted as saline mudflat deposits. It is uncertain whether deposition occurred subaqueously or subaerially, as bedding was probably destroyed by aeolian reworking and massive intrasediment salt growth;

(c) skerry and minor sandstone facies, deposited from sheet floods running off surrounding landmasses;

(d) grey-green facies containing both massive and laminated siltstones and mudstones with a few thin sandstones. This facies is interpreted as lacustrine deposits, whilst the presence of mudcracks and reworked carbonate crust material indicate the temporary nature of the lakes.

Dolomite, present throughout much of the sequence, formed by the early diagenetic replacement of $CaCO_3$ in an Mg^{2+}-rich brine regime. Halite and gypsum or anhydrite precipitated interstitially from hypersaline brines. In some cases dissolution of halite cubes and their subsequent infilling by adjacent sands gave rise to pseudomorphs on both upper and lower surfaces of sandstones. Both lateral and vertical development of the Tutbury Sulphate Bed in the East Leake - Barrow-upon-Soar area was controlled by brine sulphate concentration. Evidence that the Tutbury Sulphate Bed was exposed and eroded, either concomitant with or shortly after its formation, is provided by channels cut into the top of the bed in Fauld Mine. Polygonal networks of vertically orientated sulphate veins are interpreted as mudcracks infilled by sulphate during early diagenesis.

The sediments record variations in the degree of aridity of the climate. In the Edwalton Formation and Fauld Member an arid saline mudflat environment prevailed although minor sheet flooding occurred. The mudflats were the site of gypsum and halite growth. Saline mudflats dominate the Hawton Member, but the establishment of temporary lakes signals a decrease in aridity. In the Glen Parva and Blue Anchor Formations further decreases in aridity are indicated by the presence of fish-bearing lakes. The Penarth Group marine transgression signals the ending of a substantial period of continental sedimentation.

In Chapter 4 large sulphate masses which truncate the Tutbury Sulphate Bed in Fauld Mine are recognized as diapiric structures. The displacive nature of the diapirs is illustrated by the partial or complete over-

turning or truncation of adjacent strata. The diapirs were emplaced as gypsum which was subsequently dehydrated resulting in voids in the cores of now anhydritic diapirs. Dehydration waters leached halite from mudstones trapped within the diapirs. Later evaporation of these waters which had been trapped in the voids gave rise to halite cubes. The source bed of the diapir sulphate is now marked by lenses of elongated anhydrite nodules in dolomitized mudstones containing minute sulphate fragments. The diapirs originated when movement along an adjacent fault line triggered flowage of the source bed gypsum which was saturated with gypsum-rich waters.

In Chapter 5 it is shown that analyses of $\delta^{34}S$ sulphate, when taken in isolation, do not give a totally reliable indication of brine origin. In situations where $\delta^{34}S$ values obtained are lighter than marine $\delta^{34}S$ values for that period, it is difficult to estimate whether continental or marine brines dominated the evaporite basin. Analysis of $\delta^{18}O$ in associated carbonates helps to resolve this problem. During deposition of the Fauld Member marine brines invaded grabens (Needwood Basin) and adjacent horst margin areas (Newark). However, about 20% of the sulphate was derived from a continental source and the marine – continental input ratio exceeded 4:1. The central part of the horst area (East Leake) had a continental brine regime with sulphate being derived from the Carboniferous Hathern Anhydrite Series exposed on the northern margin of the Charnwood Massif. The Charnwood Massif controlled the palaeohydrology with sulphate-rich brine flow being restricted to the N.E. quadrant.

In the Hawton Member, carbonates formed in a predominantly continental brine regime, whilst marine influxes gave rise to thin widespread sulphate horizons. During such influxes the marine – continental input ratio lay between 4:1 and 5:1. Carbon isotope results support the sedimentological interpretation of the grey-green facies as being lacustrine. During the Blue Anchor Formation a mixed marine – continental brine regime prevailed, marking the transition to the fully marine Penarth Group. Other Triassic basins (Cheshire, Somerset, Yorkshire) show similar

variations between marine and continental brine sources.

Analyses of the isotopic ratio of Sr in sulphates confirm the results of the sulphur isotope study. The strontium is not purely of marine origin; calculations indicate that around 10% of strontium was derived from weathering of continental material.

Two discrete and stratigraphically distinct clay suites are recognized in Chapter 7. One suite contains predominantly illite and chlorite with minor (<15%) irregular mixed-layer smectite/chlorite (characterizes the Hawton Member); the other suite contains increased irregular mixed-layer smectite/chlorite (>15%) with minor illite and chlorite and occasional sepiolite and palygorskite (characterizes the Fauld Member).

Illite and chlorite were supplied to the basin as detritus in varying degrees of degradation. In the basin, chlorite was reaggraded by adsorption of Mg^{2+} from hypersaline brines. Illite aggradation was more complex, depending on degree of degradation and brine composition. In the Fauld Member highly degraded illite preferentially adsorbed Mg^{2+} from alkaline Mg^{2+}-rich brines with a high Mg/K concentration ratio. This resulted in the transformation of the illite to irregular mixed-layer smectite/chlorite. At very high Mg^{2+} levels such as periods of sulphate precipitation, the smectite/chlorite crystallites increased in size; a process bordering on neoformation. More moderately degraded illite adsorbed K^+ to form well crystallized illite. The same brine regime provided the conditions necessary for the neoformation of sepiolite and palygorskite; their distribution was controlled by the concentration of amorphous silica and aluminium respectively. Both elements were derived from adjacent land-masses and the poor solubility of aluminium very possibly restricted palygorskite formation to areas more closely adjacent to landmasses, e.g. East Leake.

In the Hawton Member the brines had a low Mg:K concentration ratio. Hence only a limited amount of smectite/chlorite was formed before the degraded illite adsorbed K^+ instead of Mg^{2+} and formed more crystalline

illite. However, the lower crystallinity of these illites compared to those associated with Mg^{2+}-rich brines, may be due to an overall lower K^+ concentration in the brines and an increased proportion of illite adsorbing available K^+. Similarly, the absence of sepiolite and palygorskite in the Hawton Member may be due to a decrease in the concentration of Mg^{2+}, Al^{3+} and amorphous silica, or a decrease in pH. The clay mineralogy therefore provides evidence of a marked alteration in brine chemistry. In the grabens (Needwood Basin) and adjacent areas (Newark) this change correlates with a change in brine source, i.e. from marine to continental. However, the centre of the horst block maintains a continental regime throughout, hence in this area the chemical variation reflects a change in brine evolution patterns (a climatic control?).

SUGGESTIONS FOR FURTHER RESEARCH.

(1) A geochemical study of Triassic evaporites in N.W. England (Cheshire Basin, Walney Island) and Northern Ireland. Analysis of $\delta^{34}S$ sulphate combined with K/Br profiles of associated halite and $\delta^{18}O$ carbonate would provide data on brine sources for this area from the upper Scythian (Rot) to Carnian. If K/Br and $\delta^{18}O$ values of halite and dolomite respectively are indicative of marine brines, then the $\delta^{34}S$ values could be used to construct a local $\delta^{34}S$ age curve and to improve the resolution of the "worldwide sulphate age curve" (Claypool et al., 1980).

(2) Further detailed analysis of clay mineral assemblages both in pre-Norian East Midlands Triassic and other areas of England. Clay mineral studies should be complemented by analysis of $\delta^{18}O$ carbonate. The correlation method outlined in Chapter 2 could then be used over a wide area of England.

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<u>The Trent, Glen Parva and Blue Anchor Formations (Upper Triassic)</u>

<u>of the East Midlands and their Sulphate Deposits.</u>

S. R. TAYLOR

<u>ABSTRACT</u>

Norian and Rhaetian (Upper Triassic) sediments from the
Mercia Mudstone Group of the East Midlands are described. Four
facies are recognized; (1) sheet flood, (2) red-bed saline
mudflat, (3) lacustrine and (4) marginal. Sedimentation occurred
on an extensive peneplain which had areas of minor relief.
Interstitial sulphate precipitation from hypersaline brines formed
widespread evaporitic gypsum and anhydrite horizons. Dolomitization
of carbonate material and possibly extensive interstitial halite
growth also occurred. Displacive sulphate masses in Fauld Mine are
recognized to be diapiric structures associated with an identifiable
source bed. Fault movement triggered flowage of water-saturated,
source bed gypsum.

A new approach to correlating fine-grained, unfossiliferous
red-beds utilizes variations in clay mineralogy and $\delta^{18}O$ dolomite.
The Trent Formation is subdivided by this method into the Fauld
Member and overlying Hawton Member. An isotopic study of the Trent
Formation sulphates (S, Sr) and dolomites (C, O) indicates that
during deposition of the Fauld Member, grabens and adjacent areas
had a marine brine regime with minor continental input (up to 20%
sulphate derived from continental sulphides). The horst area was
strongly influenced by continental brines which derived sulphate
from the exposed Carboniferous Hathern Anhydrite Series. The
Hawton Member brine regime was predominantly continental with
periodic marine influxes.

Two distinct clay suites are recognized. The first is
characterized by a predominance of detrital illite and chlorite
(= Hawton Member); the second is characterized by the presence of
either Mg-rich smectite/chlorite and/or palygorskite and/or
sepiolite with minor illite and chlorite (= Fauld Member).
Sepiolite and palygorskite were neoformed, whilst smectite/chlorite
was transformed from degraded detrital illite. Both processes
occurred in an alkaline brine regime rich in Mg^{2+} and amorphous
silica. The distribution of Mg-rich clays reflects availability
of Mg^{2+}, Al^{3+} and Si within the brines.