THE SEDIMENTARY ENVIRONMENTS AND DIAGENESIS OF NAMURIAN RESERVOIR SANDSTONES: TRUMFLEET FIELD, YORKSHIRE.

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ABSTRACT

Available lithological logs and core samples from the six wells of the Trumfleet Field were studied with a view to defining the diagenetic controls on the reservoir parameters of sandstones occurring between the Gastrioceras subcrenatum and Reticuloceras gracile (Namurian) marine bands.

The study interval was deposited on the northern margin of the Gainsborough Trough, as a cyclical, southwardly prograding fluvio-deltaic complex.

Four major sandstone units up to 25m thick occur within the study interval, each deposited in a period between tectono-eustatic marine transgressions. These coarse grained sandstone units were deposited as multichannel, fining upwards distributary channel fills within a predominantly argillaceous interdistributary and swamp complex. Associated sandstones were deposited by a mixture of crevasse splaying and overbank processes.

Four stages of diagenetic alteration can be identified on the basis of authigenic mineral assemblage and their relative timing as estimated by textural criteria. EARLY DIAGENESIS was controlled by meteoric pore fluids and was manifested by the dissolution of unstable grains and the precipitation of quartz cements. EARLY BURIAL DIAGENESIS was controlled by changes in the pore fluid chemistry as a consequence of the oxidation of organic matter and the infiltration of marine fluids. Siderite and calcite cements are characteristic of this stage. BURIAL DIAGENESIS was controlled by the generation of acidic pore fluids by the thermal maturation of organic matter in adjacent and

downdip shales, which led to the dissolution of feldspar and carbonate and the precipitation of kaolinite. LATE DIAGENESIS was manifested by the precipitation of illite, ankerite and albite cements.

Integrating the diagenetic data to the burial history and modeling the source rock geochemistry of the wells allows the timing of diagenetic events to be predicted within absolute ranges. Early diagenesis took place at the surface and early burial diagenesis occurred during the Carboniferous, from depths of only a few cm. below the sediment/water interface to up to 1500m. Burial diagenesis took place when the rocks passed through the 80^{o} C. isotherm approximately 140 my. Late diagenesis took place during maximum burial prior to Tertiary uplift, when temperatures in excess of 100^{o} C. were reached.

Petrophysical analysis shows that the rocks are poor reservoirs having low porosities and permeabilities (up to 10% and less than 1 mD), and petrographic analysis shows that most of this porosity reduction has been caused by the precipitation of cements, mainly quartz and carbonates, the bulk of the porosity being secondary after carbonate dissolution. The present porosity-permeability distribution in the rock has been overprinted by diagenetic events to such an extent that the primary textural factors have little effect on the present reservoir properties.

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Abbreviations used on plates:

PPL	Plane polarised light
XPL	Crossed polarised light
BEI	Backscattered electron image
SEI	Secondary electron image
XRD	X-Ray diffraction
T1	Trumfleet 1 well
T2	Trumfleet 2 well
A	Askern well
MF	Moss Farm well

Numbers refer to sample depth in feet, except Moss Farm where depths are recorded in meters (drilled).

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CHAPTER ONE

1.INTRODUCTION AND AIMS OF THE STUDY

1.1. INTRODUCTION

Although oil was first discovered in the East Midlands at Hardstoft during the First World War, it was with the discovery of the Eakring Field in 1938 that systematic exploration of the area took place. Despite the recent discoveries at Wytch Farm and Humbly Grove in the South of England, the East Midlands remain the most important onshore petroliferous province in the United Kingdom, the most recent development being the Welton Field.

Carboniferous strata are undoubtedly the most important fossil energy resource of the UK, since not only do they supply virtually all the home produced coal, they also provide the source, reservoirs and seals of the oil and gas in the East Midlands. In addition they source the gas in the southern North Sea Basin and the giant Morecambe gas field. The East Midlands fields are tiny in comparison to their offshore cousins but have proved attractive prospects for the major oil companies. Benificial tax legislation made post war onshore oil exploration profitable, but when these subsidies were removed in 1962 exploration in the East Midlands was halted. The 1973 OPEC oil price rise and the "energy crisis" of the late seventies revived interest in the East Midlands, and onshore exploration in general.

The immediate future of onshore exploration is clouded by the present oil price drop and the glut of oil on world markets, but recent improvements in drilling technology such as microdrilling and improved seismic resolution will ensure that small scale onshore exploration remains attractive.

1.2. AIMS OF THIS STUDY

The study was instigated to improve understanding of the factors controlling reservoir properties of the Carboniferous reservoirs of the East Midlands utilising a sedimentological approach. In particular the diagenetic history of the sandstones was considered worthy of investigation in the light of recent advances in that subject. Remarkably little "modern" diagentic data has been published concerning the East Midlands with the exception of Huggett (1982, 1984) and Hawkins (1972, 1978).

Discussions with Martin Davies of British Petroleum identified the Trumfleet Field as a viable project for study. Situated on the Northern margin of the East Midlands basin (Figure 1:1), on the edge of the Gainsborough Trough, the Trumfleet structure had flowed gas from Namurian reservoir sandstones, although not in commercial quantities. Six wells had been drilled in the structure and core was avialable from four of them, including continuous core coverage of the Upper Namurian reservoir interval from one well.

The aims of the study were initially two fold: to identify the

depositional environments of the Upper Namurian interval of the Trumfleet Field, and to describe the diagenetic history of the sediments with a view to identifying the factors controlling the reservoir properties of the sandstone intervals.

With this data generated it was thought that a diagenetic model could be produced by integration with geochemical maturation data.

The approach called for the following techniques to be applied to the study;

A Sedimentological core logging

B Interpretation of wireline logs

C Facies analysis

D Thin section analysis

E Modal analysis

F Scanning electron microscopy

G Whole rock and clay mineral X-ray dffraction

H Maturation modeling

I Integration of the above data with reference to modern Literature

The study was greatly helped when the Author joined British Gas PLC who kindly allowed access to a geochemical modeling programme, which was used to help produce the diagenetic model outlined in Chapter Nine.

1.3. SUMMARY OF CONCLUSIONS

1: The interval studied was deposited in a fluvial delta top complex as distributary channel, overbank and crevasse sandstones interbedded with siltstones, mudstones, seatearths, coals and shales deposited in a variety of interdistributary and overbank environments.

2: Primary depositional facies does not have a strong influence on reservior properties. Only distributary channel sandstones show reservoir potential, owing to their sheet-like geometries and thicknesses of up to 25m. and coarser grain size The limited areal extent and thickness of the crevasse and overbank facies render them poor potential reserviors. However, the reservoir properties of individual overbank and crevasse core plugs are essentially identical to plugs from the distributary channel facies.

3: Authigenic mineral growth is the single most important diagenetic factor controlling porosity in the sediments. Four stages of diagenesis can be recognised;

A. Early Diagenesis, controlled by meteoric fluids and manifested by the precipitation of minor amounts of quartz and kaolinite.

B. Early Burial Diagenesis. Reducing conditions were established by the oxidation of the abundant organic matter in the sediments, causing the precipitation of siderite and poikilotopic calcite cements.

C. Burial diagenesis was manifested by the precipitation of quartz cements and the generation of acidic pore fluids. Feldspar dissolution was followed immediately by the precipitation of kaolinite cements, negating the potential porosity increase.

D. Late Diagenesis occurred at depth at temperatures in excess of 100^{o} C. Albite, illite and ankerite formed the final assemblage of cements observed in the sediments.
Mechanical and physiochemical compaction played a subordinate role to mineral authigenesis in porosity reduction.

4: The preserved porosity in the sediments is almost entirely secondary feldspar and calcite dissolution porosity with minor relict intergranular primary porosity. Although porosities in excess of 10% are common, permeabilities are low due to the presence of pore blocking kaolinite books in secondary pores. Much porosity is in the form of microporosity between these books.

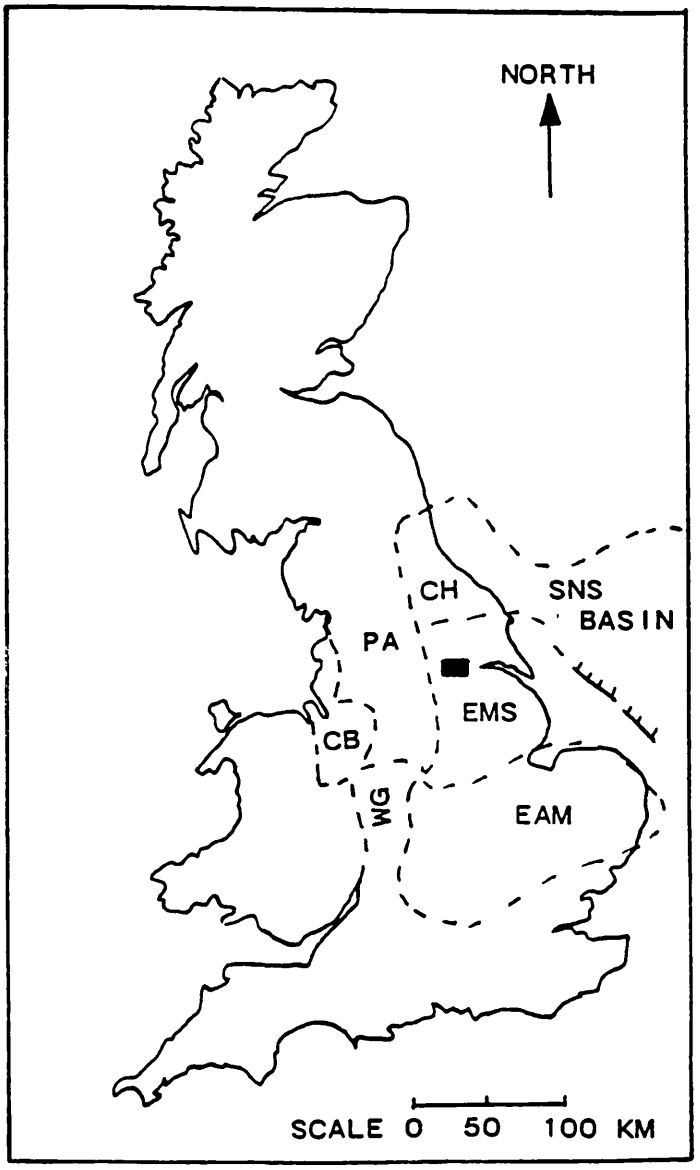
5: Integrating the diagenetic data with palaeogeothermal and burial history models allows the absolute timing of the diagenetic events to be deduced. Early diagenesis occurred at the surface and only operated for a very short period of time. Early burial diagenesis occurred from a few cm below the sediment water interval to burial depths of around 600m during the Carboniferous. Burial diagenesis occurred from Westphalian times, continued during uplift at the end of the Carboniferous and reburial during the Mesozoic. Feldspar dissolution ocurred when the rocks passed the 80 ^{o}C. isotherm (late Jurassic), and Late

diagenesis occurred when the sediments were subjected to temperatures of over 100° C, at the end of the Cretaceous.

FIGURE 1:1

LOCATION MAP

STUDY AREA



STRUCTURAL ELEMENTS

PA	PENNINE ANTICLINE	WG	WOURCESTER GRABEN
EMS	EAST MIDLANDS SHELF	EAM	EAST ANGLIAN MASSIF
CB	CHESHIRE BASIN	CH	CLEVELAND HILLS HIGH

1:4 SCOPE FOR FURTHER WORK

It is not possible to predict if the diagenetic model proposed in this study can be applied to other areas of the East Midlands until a greater published data base exists, however recent work on the Southern North Sea Basin (Cowan 1988, in press) suggests that many similarities exist between diagenetic processes in the the East Midlands and Southern North Sea Basin.

In the form presented in this study the diagenetic model is highly conjectural, temperatures of formation of the various mineral phases have been estimated by reference to published analogues. The most obvious and immediately fruitful area of research would be to estimate the temperatures of formation of the carbonate cements by stable istope analysis and to estimate the age of the illite by K/Ar dating. Fluid inclusion work would provide valuable data on the quartz cements. Unpublished Fission Track analysis has been carried out on the East Midlands sedimnts (Dr Paul Green, pers comm) and release of this data would provide an independent test of the burial model presented here.

The maturation modeling presented in this work is rudimentary, and would ben fit from better control on temperatures and on the mode and timing of basin formation.

It is felt that the approach used here is potentially rewarding, ultimately leading to the development of a predictive diagenetic model if integrated with a basin wide study of diagenesis, facies analysis, organic geochemistry, stable and radioactive isotope studies, as well as detailed structural and basin modeling.

CHAPTER TWO

1.GEOLOGICAL SETTING

2.1.1. LIMITS OF THE CENTRAL PENNINE BASIN

The Central Pennine Basin was only one of a number of basement controlled basins which effected Carboniferous sedimentation throughout Europe. The East Midlands oil fields occur on the eastern margin of the Central Pennine Basin.

During the Upper Carboniferous the Central Pennine Basin was bounded to the south by the Wales Brabant Landmass, and separated from the Northumberland Trough to the north by the emergent Askrigg and Alston blocks (Kent, 1966. Anderton et al, 1979.). The eastern margin is less well defined but is considered to be a northern extension of the Wales-Brabant Landmass, the Linlconshire Shelf (Kent, 1966.). To the West, on the other side of the Pennines there is evidence of thinning of the Carboniferous rocks near Flintshire (Kent, 1968.). Maximum thickness of sediment was deposited along the axis of the Pennine Basin, but similar sediments are known to exist under the Southern North Sea as far as Germany (Ziegler 1982.).

2.1.2. EAST MIDLANDS

The present day East Midlands Basin lies on the eastern margin of the Pennine Hills. Carboniferous rocks outcrop along the axis of the Pennines and dip gently eastwards. Carboniferous

rocks are progressively overlain by Permian, Triassic and Jurassic strata, however, the Sherwood sandstone forms the outcrop over the Trumfleet area.

2.1.3. REGIONAL STRATIGRAPHY

Brunstrom (1963) has described a typical stratigraphical sequence encountered during drilling in the East Midlands;

Lias	0-800	ft	Clays with ironstones. limestone at base
Trias			
Rhaetic	40	ft	Clay with limestone.
Keuper	700	ft	Red beds with gypsum.
Bunter	400	ft	Red Pebbly Sandstone.
Permian			
Upper beds	80	ft	Sandstone and siltstone.
U. Magnesian Lst	100	ft	Grey limestone.
Middle Marls	150	ft	Grey marl.
L. Magnesian Lst	400	ft	Grey dolomitic Limestone
Basal beds	30	ft	Basal conglomerate.
	UNCONFORMITY		
Carboniferous			
Coal Measures	2500	ft	Cyclothemic shales siltstones and coals.
Millstone Grit	100-2000	ft	Sandstones and shales.
Lower Carb.	300	ft	Massive limestones.
	UNCONFORMITY		
Lower Palaeozoic.			

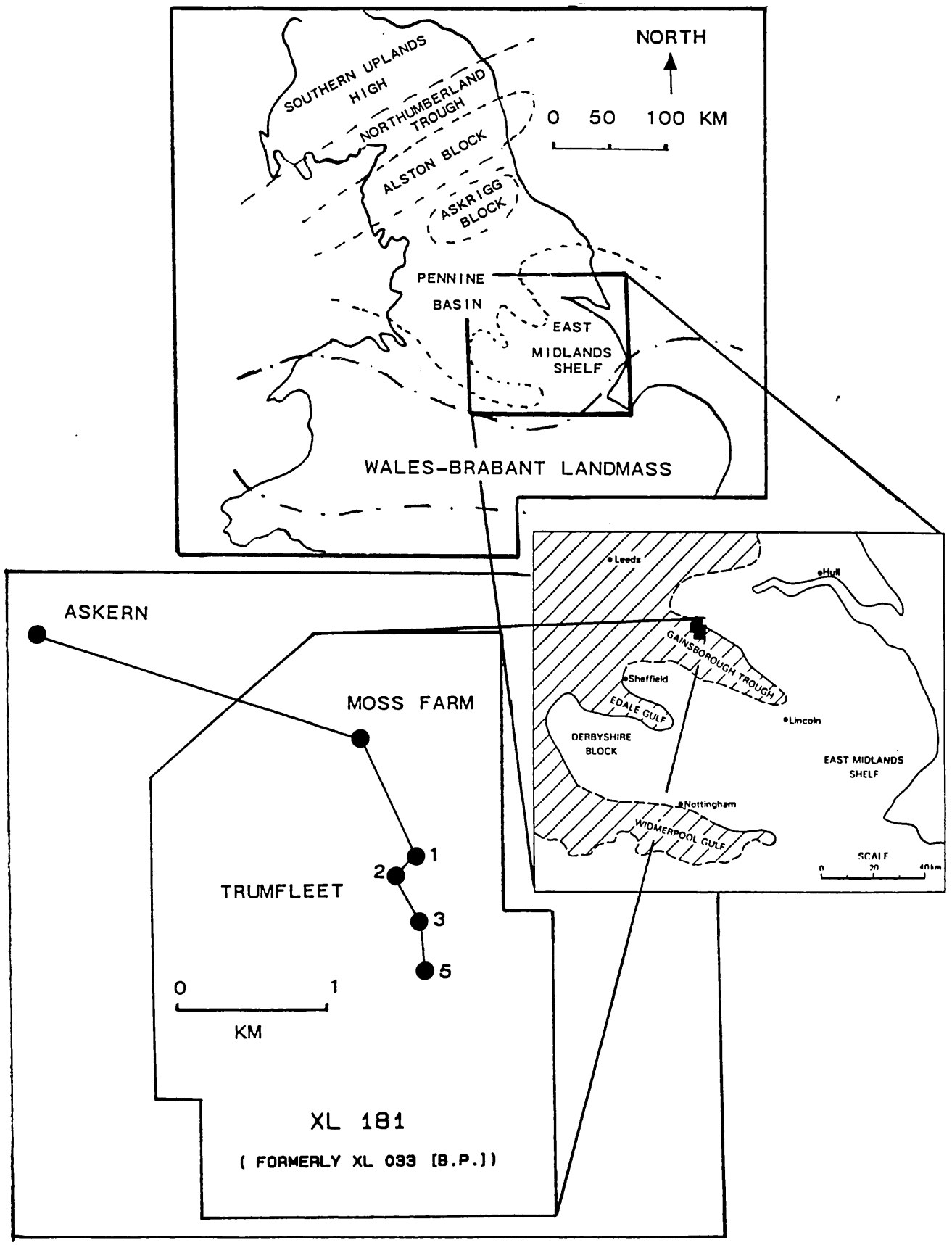
2.1.4. PRE-CARBONIFEROUS ROCKS

Data on pre-Carboniferous rocks are scarce, but Kent (Falcon and Kent, 1960. Kent: 1966.) has shown that the Wales

Brabant Massif is underlain by crystalline basement of possible Precambrian age. The East Midlands are underlain by Lower Palaeozoic quartzites, possibly Cambrian in age.

FIGURE 2:1

LOCATION MAP SHOWING CARBONIFEROUS BASINS



2.1.5. CARBONIFEROUS

I General Stratigraphy

The British Carboniferous System has traditionally been divided into three major lithostratigraphic units:

Coal Measures,

Millstone Grit,

Carboniferous Limestone.

Modern zonation is based on occurrences of goniatites, non marine bivalves and corals (Bisat, 1924, Ramsbottom 1969) but the stages, Westphalian, Namurian, and Dinantian are broadly equivalent to the old lithostratigraphic divisions. Figure 2:2 shows the stratigraphic divisions of the Upper Carboniferous.

FIGURE 2:2 SILESIAN STRATIGRAPHY

SYSTEM	SUB-SYSTEM	SERIES	STAGE	INDEX	Ma
CARBONIFEROUS	SILESIAN	STEPHANIAN	STEPHANIAN C		
			STEPHANIAN B		
			STEPHANIAN A		
			CANTABRIAN		305
		WESTPHALIAN	WESTPHALIAN D		308
			WESTPHALIAN C		
			WESTPHALIAN B	'A'	311
			WESTPHALIAN A	G_2	315
		NAMURIAN	YEADONIAN	G_1	
			MARSDENIAN	R_2	
			KINDERSCOUTIAN	R_1	319
			ALPORTIAN	H_2	
			CHOKIERIAN	H_1	
			ARNSBERGIAN	E_2	
			PENDLEIAN	E_1	326

2.1.6. DINANTIAN

The main outcrop of the Dinantian (Carboniferous Limestone) is in the Derbyshire Dome, and rocks of Dinantian age underlie the Namurian rocks of the East Midlands. Sedimentation was strongly influenced by structurally controlled basement highs, the oldest beds being deposited only in the structural lows (the Troughs and Gulfs of the East Midlands), and younger sediments can be seen to progressively on-lap onto the shelf areas. The oldest Carboniferous rocks are conglomerates which record a major marine transgression. The age of these beds varies (Kent 1966 & 1974); beds of Chadian age (L. Visean) have been found in boreholes in Eakring and Duke's Wood, whilst the earliest transgressive beds on the Alston and Askrigg Blocks are Late Visean (D1-D2) in age.

The major rock types in the basinal areas are impure limestones and shales, whilst the block areas are characterised by the presence of massive limestones and the block margins may have bounding reefs developed (Ramsbottom, 1974). Evaporites indicative of subaerial exposure have been recorded in the Hathern borehole in the Widmerpool Gulf. Ramsbottom (1973) has recognised six major cycles of regression and transgression in the Dinantian which he ascribes to eustatic sea level changes. Possible causes of these sea level changes are reviewed by Johnson (1982), and he postulates that the most likely causes are changes in the volume of the ocean basins due to variations in the rate of sea floor spreading and subduction.

In the East Midlands the Dinantian shelf facies consists mainly of massive crystalline limestones and dolomites. Farther north on the Alston and Askrigg blocks, sandstone and shale cyclothems (the Yoredale Facies) were deposited.

2.1.7. NAMURIAN

The term "Millstone Grit" is applied to beds of Namurian age in England, and it refers to the coarse pebbly feldspathic sandstones which characterise the stage.

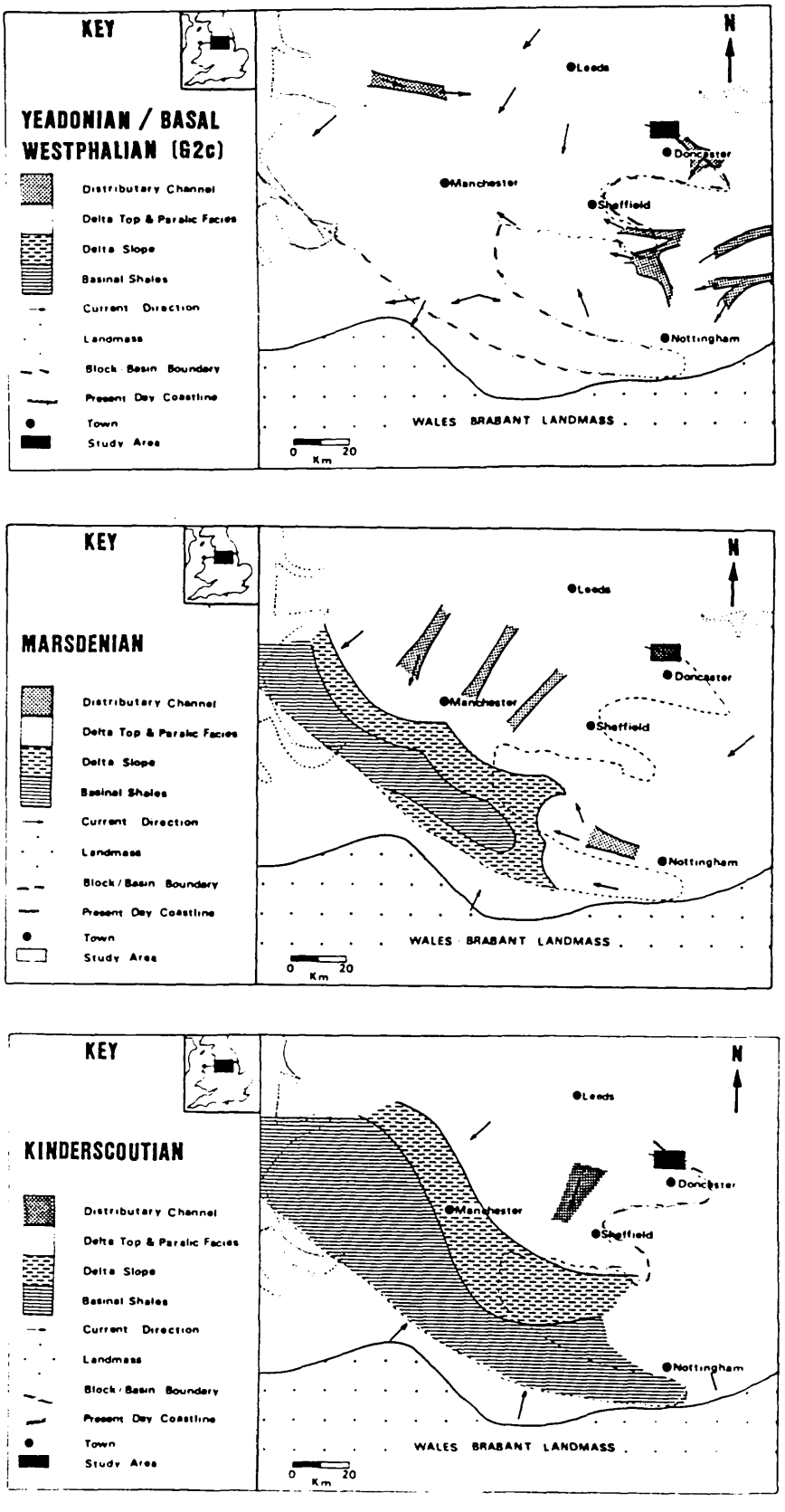
The basement blocks which had affected Dinantian sedimentation continued to exercise their control during the Namurian (Reading, 1964. Kent, 1966. Ramsbottom, 1973.) and the variation in thickness of Namurian strata reflects these relationships (Figures 2:3 and 2:4).

In 1859, Sorby measured cross strata in the Millstone Grit, and postulated that the detritus was derived from the North. In a very detailed paper on the petrography of the Millstone Grit, Gilligan (1920), proposed that the sediments were sourced by a Caledonian granitic and metamorphic landmass and deposited by Southward flowing rivers in a deltaic complex.

Reading (1964) showed that the basin was filled in two main pulses. The first was a Pedeleo-Arnsbergian period of deposition which filled the Northern basins and led to the first input of coarse clastic sediment into the Central Pennine Basin. The overlying Sabdenian rocks consist predominantly of fine grained clastic sediments, recording a temporary hiatus in the advance of the Millstone Grit delta. The second major phase of basin filling

FIGURE 2:3

NAMURIAN PALAEOGEOGRAPHIES



began during the Kinderscoutian and ended by the Yeadonian (Figure 2:3).

Sandstones derived from a northern source are dominantly first cycle and arkosic, but the sediments derived from the Wales Brabant Massif were derived from lower Palaeozoic sediments and contain less feldspar. These Southerly sourced quartzites are absent in the Kinderscoutian delta although they are present in shallow shelf sediments along the margins of the Wales Brabant Massif and in the North Staffordshire basin (Trewin and Holdsworth, 1973). The Roaches Grit Delta also contains southerly sourced quartzitic sediments (Jones, 1977).

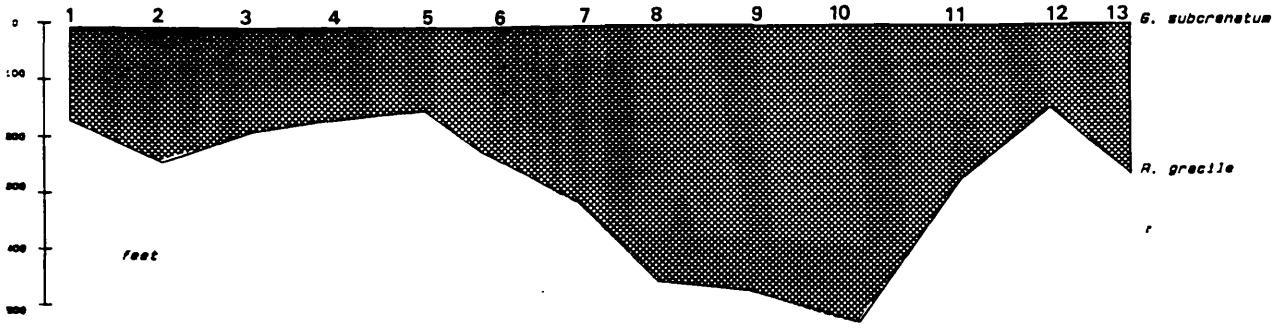
Ramsbottom (1977) has recognised 11 cycles of sedimentation in the Namurian, based upon the study of goniatite biozones, and he ascribed these cycles as being caused by tectonoeuataic sea level changes. Recently some doubt over the nature of the cyclicity of the Namurian has been expressed (Holdsworth and Collinson, 1988). The areal extent of some marine bands is less than would be expected if they were controlled by a truly eustatic mechanism, and Collinson postulates that the distribution of marine horizons may be controlled by salinity variations. During the upper Carboniferous, the basin was separated from marine influence by long seaways to the south west, and south east; subtle variations in the shape of these seaways, runnoff into the basin and tectonic control as well as tectonoeustatic sea level changes probably all contributed to the cyclical nature of Namurian, and Westphalian sedimentation. Steel (1988) described

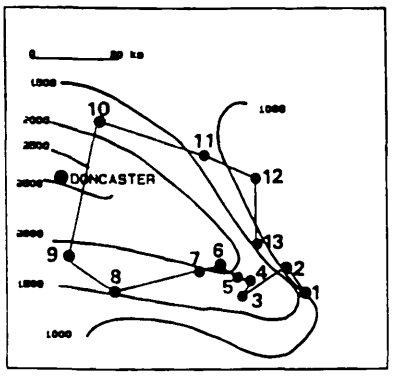
FIGURE 2:4

WELL CORRELATION OVER THE GAINSBOROUGH TROUGH

SHOWING THICKNESS VARIATIONS BETWEEN THE

G. SUBCRENATUM AND R. GRACILE MARINE BANDS.





1 Glentworth
2 Corringham
3 Gainsborough 29
4 Gainsborough 1
5 Gainsborough 2
6 Morton
7 Walkeringham
8 Ranskill
9 Tickhill
10 Trumfleet 1
11 Belton
12 Butterwick
13 Blyton

[From Downing and Howitt, 1968]

Millstone Grit Isopachs

[From Kent, 1968]

the sedimentary history of the Gainsborough area in terms of fluviodeltaic deposition, and showed that no major sandstone deltas of the central Pennine Kinderscoutian type were deposited over the area.

WESTPHALIAN

Up to 2500 metres of Westphalian sediments are known to exist in the East Midlands (Brunstrom, 1963; Kent, 1966). The junction between the Westphalian and Namurian is set at the Pot Clay marine band Reticuloceras subcrenatum, however, the style of sedimentation is a continuation of the pattern set during the late Namurian. Sedimentation during the Westphalian was dominantly cyclic, with shales, sandstones, seat-earths and coals being deposited in lushly vegetated tropical swamps with fresh and brackish water lakes between minor fluvial channels (Goosens et al, 1974).

Sediment input into the lakes was by crevasse splaying, overbank flooding and as minor delta lobes, supplied by low sinuosity streams (Fielding, 1984). Hazeldine + ANDERTON (1980) recognised a braidplain facies in the Westphalian of County Durham which he ascribed to being controlled by hinterland uplift.

In the East Midlands one of the major oil reservoir sandstones is the basal Westphalian Crawshaw/sub-Alton Sandstone. Hawkins (1972) studied this sandstone in the subsurface and was able to show that it was deposited as a number of channel complexes (Fig 2:3). Sandstones are generally thinner

and more argillaceous in the Westphalian than in the Namurian (Brunstrom, 1963), but coals are much thicker and more abundant (Goossens et al, 1974). The effect of block and basin topography is muted in the Westphalian over the Gainsborough Trough (Bedrock, 1984). Guion and Fielding (1988) have described in detail the sedimentary history of the Westphalian A and B in the Pennine Basin.

By the end of the Westphalian increasing stability of the shelf led to the establishment of a continental red bed facies (Johnson, 1982), the uppermost Westphalian and Stephanian was removed throughout most of the area by erosion during late Carboniferous (Hercynian) uplift.

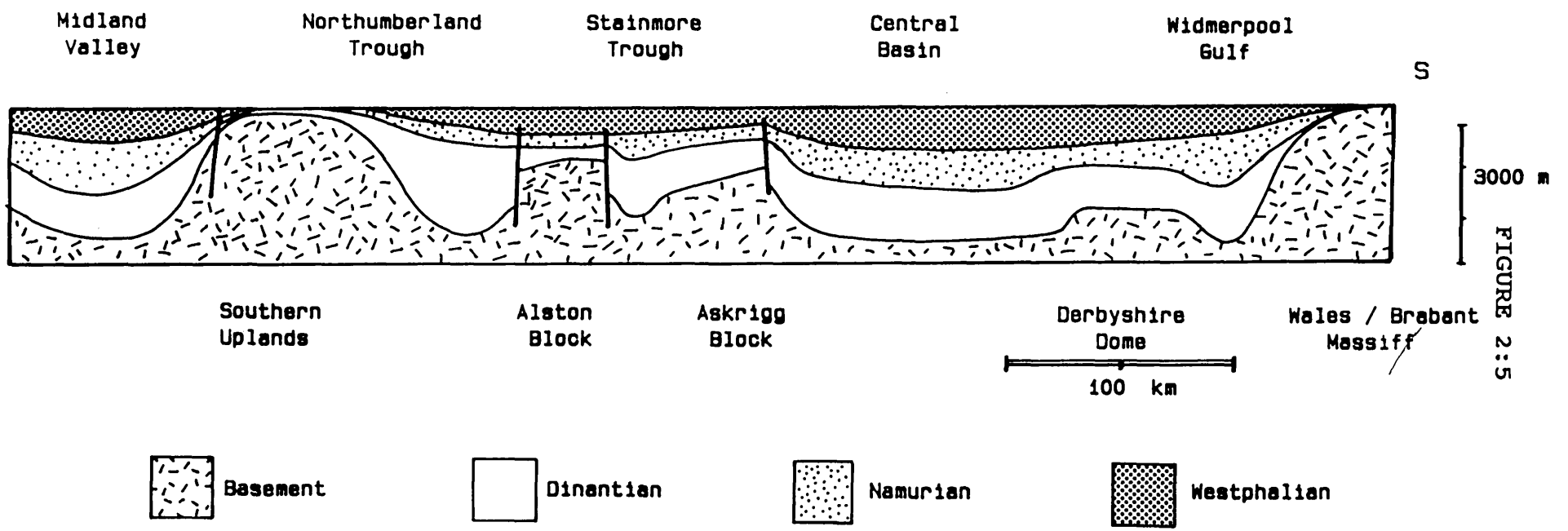
2.1.9. IGNEOUS ROCKS

Up to 600 feet of extrusive and intrusive igneous rocks have been recorded in the East Midlands. Contemporaneous extrusives consist of basalts and tuffs in the Lower Coal Measures. Permo Triassic intrusives were emplaced close to the Namurian/Westphalian boundary (Falcon and Kent, 1960).

2.1.10. HERCYNIAN DEFORMATION

At the end of the Carboniferous period the south and north European plates collided, initiating a southwards dipping subduction zone to the south of England (Shackleton et al, 1982), and the period of extensional tectonics which had led to the

NORTH SOUTH SECTION THROUGH BRITAIN SHOWING CARBONIFEROUS ROCKS



From Johnson 1982

FIGURE 2:5

formation of the Carboniferous Basins of North West Europe came to an end (Leeder, 1982; Johnson, 1982.). South of the Wales Brabant landmass, this event produced tight folding and thrusting of Carboniferous sediments in the south of the country, and the emplacement of granite plutons in Cornwall. Areas to the North of the Wales Brabant Massif were shielded from the major effects of this north south compressional stress regime.

Later earth movements superimposed an east/west compressional stress pattern over the area and this event affected the East Midlands. Structurally positive blocks within the basement were reactivated and the Pennine Basin was uplifted along the locus of maximum deposition (Johnson, 1982). Drape folds of Carboniferous sediments on the eastern margins of the Pennines formed gentle elongate anticlines which comprise the traps for the East Midlands oil fields.

During the Tertiary a second phase of uplift occurred in response to Alpine earth movements and a 1-2 degree easterly regional tilt was imposed over the East Midlands area (Kent, 1954).

2.2. HYDROCARBON HABITAT

2.2.1. EXPLORATION HISTORY

Exploration for hydrocarbons in the East Midlands began during the First World War. Drilling was concentrated on the coal field areas as oil seepages were a commonly observed phenomenon in coal workings.

The first well to produce oil was drilled at Hardstoft in Derbyshire in 1919 (Lees and Tait, 1946). During the late 1930's the Eakring structure, a gentle fold trending roughly North - South, was discovered and produced oil from Millstone Grit sandstones and from the Carboniferous Limestone. Similar structures were drilled and the Dukes' Wood, Kelham Hills and Caunton fields were discovered, all with reservoirs in the Millstone Grit. By the 1960's , Plungar, Egmanton, Bothamsall, Gainsborough, Corringham, Beckingham, South Leverton and Glentworth had been discovered (Falcon and Kent, 1960). Most of the production is from the Upper Millstone Grit sandstones and the basal Coal Measures Crawshaw/Sub-Alton Sandstone (Hawkins, 1972), with minor amounts of production from the Middle Coal Measures, notably at Gainsborough and Glentworth, on the northern margins of the Gainsborough Trough.

Exploration was abandoned in the mid-1960's when tax benefits for U.K. produced oil were removed, making exploration for small fields uneconomic due to the availability of cheap Middle Eastern oil. Interest was revived after the 1973 oil price increases and British Petroleum Development Ltd reopened their exploration

department at Eakring. Exploration is continuing in the 1980's with most of the major oil companies actively exploring in the East Midlands, the Farley's Wood and Welton fields being discovered during this recent period of exploration.

2.2.2. SOURCE ROCKS

Although potential source rocks are abundant in the East Midlands, the precise identification of the actual source rocks has proven problematic. Falcon and Kent (1960) and Hawkins (1972), argued that hydrocarbons migrated into sandstones from adjacent marine and non marine source rocks during the late Carboniferous however, Dobson (1984) has demonstrated that the potential source rocks stratigraphicaly adjacent to the reservoirs are immature and have not generated significant amounts of oil. Dobson showed that only the basal Namurian marine bands have generated significant amounts of hydrocarbons, although most of the oil in the East Midlands is produced from near the Namurian/Westphalian boundary (Dobson, 1984). Potential reservoirs exist in the lower parts of the Namurian sequence so it seems unlikely that oil would by-pass these to collect exclusively near the Westphalian/Namurian boundary. Dobson (1984) proposed that oil may have migrated up along growth faults (Chisholm, 1981) to preferentially concentrate in the more porous channel sands of the Yeadonian and Basal Westphalian. However, Kent (1954) noted that most fault planes in the oil field area are sealed and dry, and he cites the lack of good reservoir facies in the lower Namurian sequence as a possible explanation of the lack of productive lower Namurian reservoirs.

2.2.3. TIMING OF OIL GENERATION

Kent (1954) stated that oil migration occurred in two phases: during the late Carboniferous, and to a lesser extent during the Tertiary, when tilting redistributed the trapped oil. Hawkins (1972,1978) using textural and petrographic criteria, argued that oil migrated into the reservoirs during early (Carboniferous) burial and was concentrated in the anticlines during the Hercynian Deformation. Dobson (1984) calculated burial curves for the East Midlands and using Waples (1980) Time Temperature Index (TTI), proposed that the onset of oil generation occurred during the late Westphalian/Early Permian but that peak oil generation did not occur until the Middle Jurassic, approximately 180 my. Since over 3000m of sediment accumulated in the Central Pennine basin it is highly likely that some hydrocarbons were generated in this area during the Carboniferous. However, the East Midlands Shelf did not suffer such deep burial during the Carboniferous, and probably did not generate significant amounts of hydrocarbons during the Carboniferous. Unpublished fission track data (Dr Paul Green, pers comm) suggests that hydrocarbon generation did not occur until the end Cretaceous.

2.2.4. RESERVOIR ROCKS

All Dinantian reservoirs occur in shelf facies limestones, no reservoirs are present in Basinal (Gulf facies) limestones. Sandstone reservoirs in the Namurian are sheet and shoestring channel sandstones which show a high degree of high lateral variation. Typical porosities are between 10 and 20 per cent

with permeabilities of a few tens of millidarcies. (Downing and Howitt, 1969.) The Middle Grit Group and the Rough Rock form important reservoirs. The most important reservoir in the East Midlands is the Crawshaw/Sub-Alton sandstone (Hawkins, 1972) which lies directly above the Pot Clay Marine Band. The sandstone was deposited as a fluvial channel and barrier bar complex on the delta top, with an associated interdistributary facies (Guion, 1971; Hawkins, 1972). The distributary channel sandstones form the best reservoirs with porosities varying between 6 and 21 per cent and permeabilities of 0.05 to 31 millidarcies. Sandstones higher in the Westphalian generally have lower porosities and permeabilities (5-15% and 20 md) due to a higher clay content and finer grain size (Downing and Howwit, 1969), Huggett (1982).

CHAPTER THREE

PREVIOUS WORK ON DIAGENESIS

31. EARLY WORK

Early work on diagenesis was confined to the study of surface material from the main outcrop area of the Millstone Grit in the Pennines.

One of the earliest workers to study the effects of burial on the rocks was H.C. Sorby (Sorby, 1859). He believed that compaction had occurred by compression of feldspars, squeezing the resultant clay minerals into the pore spaces.

Quartz overgrowths in optical continuity with the host grain, were recorded by Phillips (1881), who also noted the presence of a kaolin mineral.

In a very detailed paper, Gilligan (1920), studied the petrology, and in particular the heavy mineral suites, of many hundreds of samples from the Millstone Grit of the Pennines and Yorkshire. He noted the silica overgrowths and positively identified kaolinite by X.R.D. but did not consider the diagenetic modification of the rocks.

32. 1950 TO 1970

Hemmingway (1957) noted the presence of etch pits on the surfaces of detrital grains and overgrowths, and attributed them to dissolution of quartz at points of contact during early diagenesis. Central to this argument is the implication that silica overgrowth occurred very early during diagenesis.

The first work on subsurface material was carried out by Greensmith (1957), who studied Namurian core material from South Yorkshire. Greensmith argued that silica and carbonate cements were precipitated during early diagenesis, just below the sediment/water interface, silica cements being precipitated in rocks with a fresh pore fluid and carbonate cements being precipitated in rocks with saline pore fluids. He therefore proposed the influence of sedimentary environment on diagenesis, though later studies showed that silica and carbonate cements could be precipitated both during burial and at depth. Greensmith also noted the presence of authigenic kaolinite and identified illite, chlorite and minor amounts of mixed layer clays in the sediments.

Wright (1964) studied the compaction history of Millstone Grit rocks from the Central Pennines, and was able to demonstrate, by quantitatively studying grain contacts, that cementation had occurred during burial. He also noted that the number of grain contacts was greater where clay minerals were absent, and postulated that the clays inhibited quartz cements.

33. 1970 TO 1980

Waugh (1970) used an S.E.M. to study the development of quartz overgrowths from coalescing prisms of authigenic quartz on detrital grains.

The first systematic study of diagenesis was carried out on rocks from the Bothamsall Field by Hawkins (1972, 1978). He postulated that diagenetic modification was controlled mainly by chemical processes, and related diagenesis to sedimentary

environment, concentrating on the reservoir facies (barrier bar and distributary channel sandstones). He showed that the dominant authigenic minerals in the barrier bar sandstones were mixed layer illite smectite, ferroandolomite and ferroancalcite, with minor amounts of quartz. Quartz, ferroandolomite and kaolinite were dominant in the distributary channel sandstones. He found that porosity preservation was greatest in the axis of the distributary channel sandstones and postulated that the coarser grained channel sands would have acted as conduits for fluid flow and would thus have encouraged the preferential migration of hydrocarbons into the channel axes during Carboniferous burial. Authigenic mineral growth would have been inhibited by the presence of hydrocarbons in the pore spaces. He concluded that early oil migration was lithofacies controlled but that secondary migration was structurally controlled as the oil moved into anticlinal traps during the Hercynian deformation.

34. 1980 TO PRESENT

Wilson (1980) studied etch pits on the surface of quartz overgrowths of samples taken from surface exposures. He attributed these pits to leaching by acidic, peat derived surface waters during eodiagenesis.

A geochemical investigation of Mam Tor (Kinderscoutian) turbiditic sandstones and shales was carried out by Spears and Amin (1981). They showed that the major control on the clay mineralogy was the distance of the sediment from the palaeoshoreline and that diagenetic influences were limited. Kaolinite was found to be the dominant clay mineral, but it was more common

in the proximal sandstones than in the more distal shales. They noted that the dominant feldspar was albite, and that most of it was detrital in origin, in contrast to other workers (Gilligan, 1920; Greensmith, 1957; Hawkins, 1972, 1978; Huggett, 1982) who observed that microcline perthite was the dominant feldspar. Spears and Amin concluded that diagenesis had a limited effect on the bulk chemistry of the rocks because the low permeability of the sandstones led to the establishment of a closed system soon after deposition.

Huggett (1982) worked on the clay mineral diagenesis of a Westphalian channel sandstone, the Eagle Sandstone. She noted the following sequence of diagenetic mineral growth: early formation of siderite and pyrite under reducing conditions during early diagenesis, followed by the formation of quartz overgrowths, kaolinitisation and dissolution of feldspars during burial. Though Huggett thought that the channel sandstones would have been in an open system during burial diagenesis, she postulated that the SiO_2 of the quartz overgrowths was supplied by clay mineral reactions with detrital grains and that it was not brought in from the surrounding shales by pore fluid flow during compaction as the shales de-watered. Later diagenetic minerals recorded by Huggett (1982) were illite with minor amounts of dolomite and authigenic albite. Huggett attributes this change in diagenetic assemblage to an increase in pH of the pore fluid caused by H^+ ions being scavenged by clay mineral reactions within the sandstones.

CHAPTER FOUR

EXPERIMENTAL TECHNIQUES

4:1. LOCATION AND SAMPLING

The location of the study area is shown in Fig 2:1. Six wells were studied, but cored samples were available for four wells only. One well, N.C.B. Moss Farm had continuous three inch core coverage of the sandstone intervals of the sequence studied. The remaining wells for which core was available had one and a half inch and three inch core samples at discrete one foot intervals through the sandstone and shale intervals of the sequence studied, though there were frequent gaps. The cores were held by the Institute of Geological Sciences at Keyworth and by British Petroleum Development Ltd. at Eakring.

4:2. INITIAL CORE LOGGING

Core from N.C.B. Moss Farm was stored in boxed 1.5m lengths. As the boxes were stored in no particular order, it was found convenient to log each box on a separate sheet of paper at a scale of 1:10. The core sections were not slabbed. Cores from the Askern and Trumfleet 1 and 2 wells consisted of small sections of core, representative of one foot intervals. These wells were logged on the arbitrary scale of lcm to one foot (approximately 1:30). Information was recorded graphically on the logging sheets using a method adapted from Selley (1968). Vertical grain size profiles were constructed by reference to a

Wentworth comparison chart. The separate logging sheets were combined and re-drawn on a scale of 1:200 for comparison with the 1:200 wireline logs.

4:3. WIRELINE LOGS

Since the wells were drilled over a twenty year period, the quality of the logs varied considerably. The Gamma Ray log was available for all the wells studied, and proved to be the most useful for correlation and environmental interpretation. Marine bands contain condensed sequences which give higher Gamma Ray values than the surrounding rocks (Knowles, 1964, Hawkins, 1972,). The Gamma Ray log was used to identify and correlate these bands. Details of wireline logging theory and practice are set out in Asquith et al (1982).

TABLE 4:1

Wireline Logs Available in this study.

Well	Logs
Moss Farm	GR/SP/DLL/SONIC
Trumfleet 1	GR/SP to 3300 feet
Trumfleet 2	GR/SP/
Trumfleet 3	GR/SP/DLL
Trumfleet 5	GR/SP/DLL
Askern	GR/SP/DLL

GR: Gamma ray, SP: spontaneous potential, DLL: Dual Laterolog.

4:4. SAMPLING

Samples for petrographic and petrophysical examination were taken at approximately one metre intervals for Moss Farm and at

three foot intervals, wherever possible, over the sandstone units from the remaining wells. Selected samples from Moss Farm were slabbed for more detailed examination of sedimentary structures Samples from the other wells were generally too small to yield much useful information on sedimentary structures. Selected shale samples were collected for whole rock and clay mineral X.R.D. analysis.

4:5. OPTICAL PETROGRAPHY

4:5.1. THIN SECTION PREPARATION

Slabs measuring approximately 25 by 15 by 5 mm were cut from each sample and vacuum impregnated with blue stained araldite. To minimise any damage to hydrated clay minerals, the temperature was kept below 40 ^{o}C. during impregnation. Samples were stained with alizarin red S and potassium ferricyanide to differentiate between the various carbonate minerals present (Dickson, 1966). Most authors recommend immersing the section in a .2% solution of HCl for one to two minutes, but immersion for this length of time was found to cause considerable etching to the surface of the thin section. Immersion for twenty to thirty seconds was found to be sufficient. Feldspars were not stained.

4:5.2. PETROGRAPHY

A petrographic description sheet was prepared for each sample. Maximum, minimum and mean grain sizes were recorded by holding a transparent Wentworth size comparison chart against the section. Grain shape and sorting were recorded by reference to visual comparison charts (Scholle 1979).

4:5.3. MODAL ANALYSIS (Point Counting)

A ten memory Swift automatic point counter fitted with a mechanical click stop stage was used. As most sections contained more than ten grain types, some of the less numerous

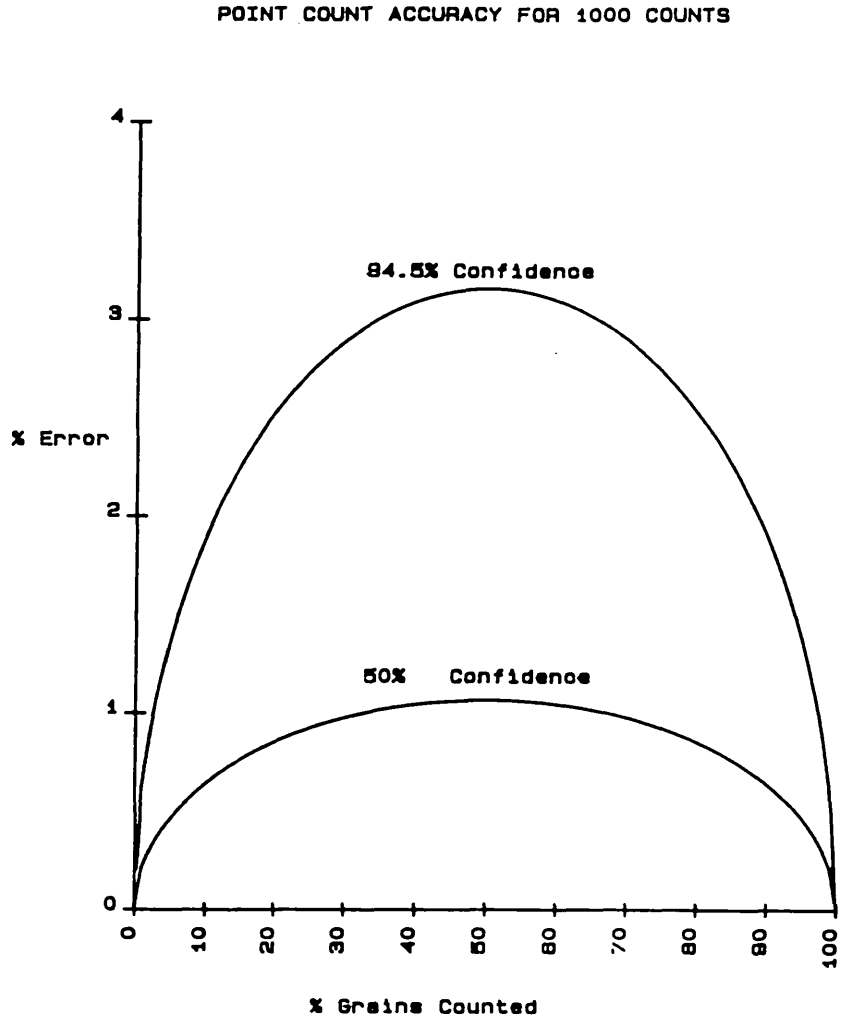
classes were tallied by hand on the petrographic sheets. The maximum possible spacing between counts was used to reduce the bias caused by large grains and to cover as much of the section as possible. Van der Plas (1959) and Griffiths (1967) have fully discussed the principles of modal analysis. Galehouse (1971) discussed the statistical constraints on the accuracy of point count data. Fig 4:1 was constructed using equations given in Galehouse (1971). It was found that reliable results could be obtained by using 1000 counts per section. No large improvement in accuracy was found by making 2000 counts per section (Fig 4:2). Five hundred points were counted twice from each section, and the results compared before summation. If a large discrepancy was found between each set of counts a further 500 points were counted. Wherever possible sections were cut perpendicular to the bedding to reduce bias caused by thin laminae rich in organic matter and by the presence of platy minerals. However because of the small sample size a number of thin sections had to be cut parallel to the bedding. To reduce bias the stage was moved frequently in the "x" direction and if necessary a further five hundred points were counted.

4:5.4. COUNTING METHODOLOGY

Point counting is essentially an exercise in classification, and in order to obtain internally meaningful results, strict rules must be applied when using the technique. The set of rules by which grains are assigned to the various classes will vary depending on the object of the study. The object of point counting in this study was to ascertain the mineralogical

FIGURE 4:1

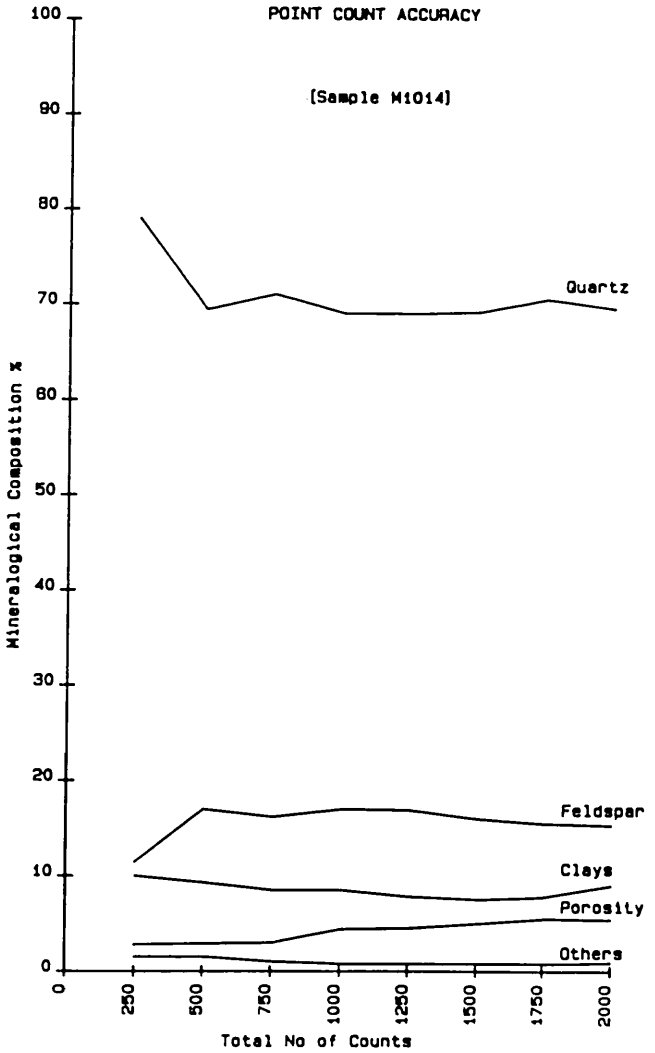
POINT COUNT REPRODUCIBILITY FOR 1000 POINTS COUNTED AT 94.5 AND 50% CONFIDENCE LEVELS



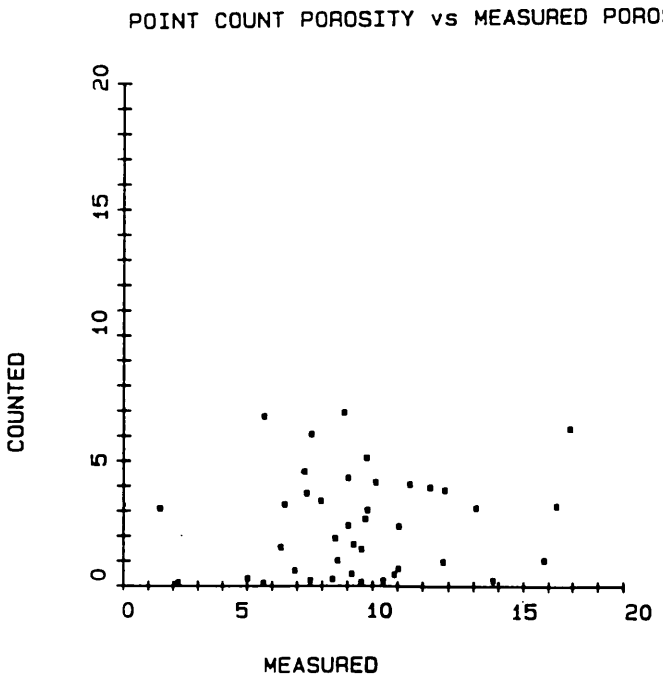
From Galehouse (1971).

FIGURES 4:2 AND 4:3

4:2



4:3



composition of the rock, rather than simply to classify it with respect to one of the many sandstone classifications. With this in mind, wherever possible, grains were assigned into classes on the basis of composition rather than type. For example, chert and quartz were assigned to the same class. Where polymineralic grains, such as rock fragments or grains which had been partially altered, were encountered, the mineral directly under the crosswires was counted, rather than assigning the grain to a generic category. Thus, a rock fragment consisting of quartz, feldspar and mica could be counted as any one of its constituents depending on which one happened to fall under the crosswires. Collecting data of this type gives information on the chemical composition of the rock, which is more useful in considering diagenesis than the data collected by the more traditional methods, which can give a misleading impression of composition. Some classifications, for instance, consider polycrystalline quartz to be a rock fragment. It is quite possible under such schemes to have a sediment which comprises 60% rock fragments but be composed entirely of quartz grains. This sample would behave quite differently during diagenesis than one which consisted of 60% of a more labile "rock fragment". Rock fragments were only counted in this study when they were too fine grained to allow the individual minerals to be discerned.

4:5.5. POROSITY

Point count porosity was found to bear little relation to measured porosity (Fig 4:3), owing to the presence of significant microporosity in clay filled pores and the fact that significant

amounts of the observed porosity was secondary in origin. Porosity was counted as total porosity and the relative amounts of primary and secondary porosity estimated visually.

The criteria for recognition of secondary porosity are set out in Schmidt and McDonald (1979). The main types of secondary porosity encountered were feldspar and calcite dissolution. The amounts of calcite dissolution are difficult to estimate but it appears to be widespread.

4:6. X-RAY DIFFRACTION (X.R.D.)

4:6.1. INTRODUCTION

The principles of X-ray diffractometry are explained in Brindley and Brown (1980). Whole rock and clay mineral separates were run using cobalt K alpha radiation in a Phillips diffractometer goniometer. Output was as a continuous paper chart which was run at a speed of 1 cm. per minute to give a scale of 1 cm = 1 ^{o} 2 ϕ on the chart when the goniometer was run at a speed of one degree per minute.

4:6.2. WHOLE ROCK X.R.D.

Selected whole rock samples were run as a check on the bulk mineralogy as interpreted from thin section analysis. Shales and seat earths, which were too fine grained to be studied optically were studied using X.R.D.

4:6.3. SAMPLE PREPARATION

Samples were dry crushed using an iron pestle and mortar, prior to grinding in an agate Teema mill for about ten minutes, until the sample was smooth when rubbed between the fingers. Any supplementary grinding necessary was carried out in the X.R.D. Lab using a small agate pestle and mortar. The rock powder was pressed into a cavity mount with a glass slide. Samples were run in the diffractometer from 5 to 70 ^{o} 2 ϕ and the resultant charts were indexed and interpreted by reference to tables in

Brindley and Brown (1980), Brown (1961),and to standard diffractograms run in the Geology Department X.R.D. lab.

4:6.4. CLAY MINERAL X.R.D.

The clay size fraction of any rock is defined as the fraction less than 2µm in size. Clay minerals frequently form crystals up to 20µm in diameter. During S.E.M. examination of samples in this study, kaolinite was found to form booklets with individual plates of more than 15µm across. It was, therefore, decided to sample the less than 20µm size fraction to prepare orientated clay discs for X.R.D. analysis, even though this size fraction was contaminated by non clay minerals.

4:6.5. SAMPLE PREPARATION

Selected samples were gently disaggregated in an iron pestle and mortar before being placed in screw topped jars which contained de-ionised water. A few drops of ammonia were added to act as a deflocculant. Where flocculation was severe the samples were washed with de-ionised water to remove any soluble salts present. The samples were further disaggregated by placing them in an ultrasonic bath for approximately one hour. The jars were then shaken vigorously by hand and left to stand for twenty minutes before the top 5 to 7 cm. was poured off. No significant differences in the resultant diffractograms were found using this method than by separating the clay fraction by centrifuge. Centrifuged samples were spun for three minutes at 1000 RPM. Orientated clay discs were prepared by filtering the supernatant liquid through a porous ceramic tile under vacuum, using the

technique described by Shaw (1970). The process was repeated, if necessary, until the entire surface of the ceramic tile was covered by an opaque film of clay.

TABLE 4:2

Time taken for particles of 2.65 g/cm^3, to settle 5 cm. in water at 20 degrees Centigrade.

Size µm	Time
50	22 secs
20	2 min 20 secs
5	37 mins 30 secs
2	3 hours 50 mins

(From Brindley & Brown, 1980)

To investigate the effect of grain size on the resultant diffractograms, a series of discs from the same sample were prepared with settling times of 10, 20, 30 and 40 minutes. A slight increase in the height of the kaolinite peak and a decrease in the height of the illite peak was noted in the coarsest sample (10 minutes), the 20, 30 and 40 minutes samples showed no significant differences in peak height or position. A disc was observed in the S.E.M. to check the size of particles present (Plate 4:1). The maximum size of particles on the surface of the disc was seen to be approximately 20µm, but some scattered flakes of mica over 50µm across were observed. This mica contamination contributes to the very sharp "illite" peak seen on some diffractograms.

4:6.6. INTERPRETATION OF ORIENTATED DISCS

Each disc was run in an air dried, ethylene glycol solvated and heated state, and interpretation was based on integrating information from the three diffractograms. Each separate run was

carried out using the same sensitivity and time constant. Basal reflections were used for identification of mineral species with reference to Carroll (1970) and Brindley & Brown (1980). No attempt at quantifying the relative amounts of the clay mineral species present was made using diffractograms.

4:7. SCANNING ELECTRON MICROSCOPY

4:7.1. INTRODUCTION

The S.E.M. consists of an electron optical system, which produces a finely focused beam of electrons, and a signal processing unit which detects, amplifies and displays a signal produced in the specimen by the electron beam's interaction with it. By raster scanning the specimen in synchronisation with a cathode ray tube, a two dimensional image of the signal's interaction with the specimen can be built up. The various modes of operation of the S.E.M. depend on detecting the differing ways in which the electron beam interacts with the specimen. Three modes of operation were used in this study; the emissive mode (both backscattered and secondary), the luminescent mode, and the X-ray mode. Details of the modes of operation of the S.E.M. can be found in White et al, (1980) and Goldstein et al, (1981). Two S.E.M.s were used in the study. A JEOL J.S.M. 35 was used for secondary electron imaging and cathodoluminescence, and a JEOL J.S.M. 733 for secondary electron imaging, backscattered electron imaging and for quantitative X-ray analysis.

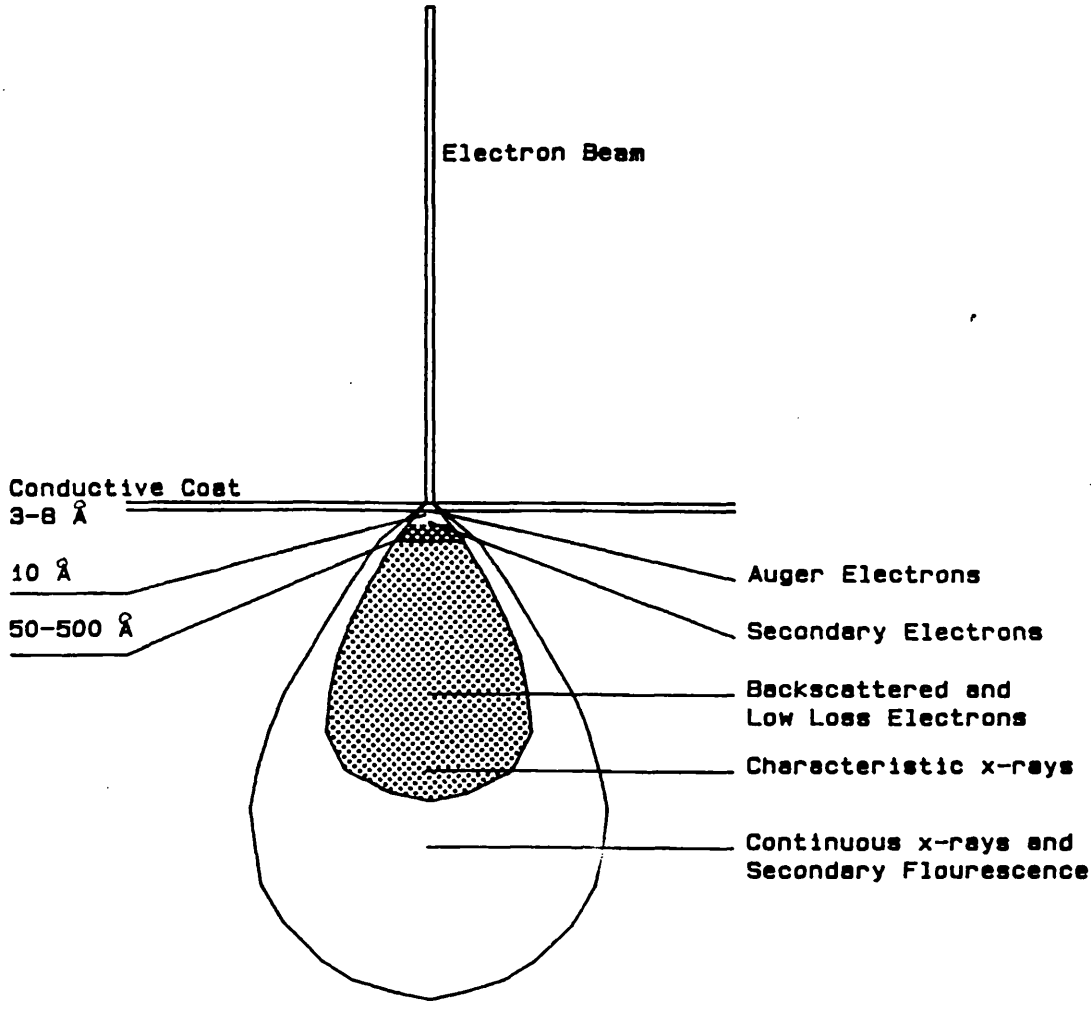
4:7.2. EMISSIVE MODE

4:7.3. SECONDARY ELECTRON IMAGING (S.E.I)

Secondary electrons are low energy electrons (<50ev) which are produced by the electron beam's interaction with the surface of the specimen to a depth of 50 - 500 Angstroms (Fig 4:4).

FIGURE 4:4

ENERGY DISSIPATION VOLUME



Adapted from Goldstien et al, 1981.

These electrons provide information about the surface topography of the specimen and are the electrons used to produce the familiar scanning electron photomicrographs. The photomicrographs resemble normal light images but have much greater resolution and depth of focus. It must be remembered that electron photomicrographs contain only two dimensional information despite the fact that they appear three dimensional. For this reason stereo pairs of photomicrographs were taken using the tilting method described in White et al, (1980). When operating the S.E.M., three dimensional relationships can be observed by rocking the specimen stage back and forth in the specimen chamber.

4:7.4. SAMPLE PREPARATION FOR S.E.I.

Freshly fractured surfaces were prepared prior to mounting on aluminium stubs as the well cemented nature of the rock made it difficult to fracture the surfaces after they were mounted. The sides of the specimen were painted with conductive paint then were placed in a vacuum chamber for 24 hours to remove any volatile hydrocarbons in the rock which could destroy the coating when placed under the electron beam. A 300 Angstrom thick gold palladium coat was applied in a sputter coater, and if the sample exhibited charging under the beam, a further 2-300 Angstrom thick coat was applied.

4:7.5. BACKSCATTERED ELECTRON IMAGING (B.E.I.)

Backscattered and low loss electrons are high energy electrons which are involved in elastic scattering events within

the specimen.(Fig 4:4) Backscattered electrons have energies approaching that of the beam current and the intensity of the backscattered signal is related to the mean atomic number (Z) of the specimen. High Z phases appear bright on the C.R.T., screen whilst low Z phases appear darker. Under ideal conditions an atomic number contrast of 1 between phases can be detected. The backscattering coefficient is the ratio of the backscattered current to the total beam current. For compounds the backscattering coefficient is given by the following equation.

EQ 4:1

$$N = \sum_{i-1}^{n} C(i)n(i)$$

N	Weighted mean B.S.E. coefficient.
C(i)	Concentration by weight of each element in the compound.
n(i)	Elemental B.S.E. coefficient. (Tabulated in Reed, 1975)
n	Number of elements.
	From Hall & Lloyd, (1982).

Using the above equation it is possible to estimate the backscattering coefficient for geological materials. Table 4:3 gives the backscattering coefficient for some common minerals encountered in this study.

Fig 4:4 shows that the excitation volume for backscattered electrons is larger than that for secondary electrons and the resolution is correspondingly less than for secondary electron imaging. Goldstein et al, (1981) give a detailed treatment of beam-specimen interactions.

TABLE 4:3

MINERAL	B.S.E. COEFFICIENT
ALBITE	.1307
ANORTHITE	.1451
APATITE	.1759
BOITITE	.1658
CHLORITE(FeO)	.1859
CHLORITE(Mg)	.1226
DOLOMITE	.1259
ILLITE	.1407
KAOLINITE	.1264
MAGNETITE	.2427
MUSCOVITE	.1389
ORTHOCLASE	.1446
PYRITE	.4262
SIDERITE	.1899
SILICA	.1317

4:7.6. SAMPLE PREPARATION FOR B.E.I.

Polished thin sections and blocks were used for backscattered imaging, since topographic effects must be minimised to enhance the atomic number contrast. Thicker than usual sections were prepared, as for normal optical use, then polished to 30 micron thickness. A conductive carbon coat was applied to the surface immediately prior to examination.

4:8. ENERGY DISPERSIVE X-RAY ANALYSIS (EDAX)

4:8.1. INTRODUCTION

The principles of quantum mechanics are utilised in energy dispersive analysis. Bombarding electrons produce X-rays in the specimen in two ways. Firstly, continuous X-rays (Bremsstrahlung) are produced when the bombarding electrons are decelerated by the electrical fields within the specimen. This energy, (and therefore wavelength of the radiation) is proportional to the energy lost by the bombarding electron and a continuous spectrum of radiation is produced. Secondly, characteristic X-rays are produced when a bound electron is ejected from a shell within an atom of the specimen, leaving the atom in an excited state, with a vacancy in its electron shell. This vacancy is quickly filled by an electron from an outer shell which gives off energy as it does so. This energy can be released as x-rays or by the ejection of an Auger electron without any radiation produced. The energy of this radiation is equal to the energy difference between the vacant and donor shell and is characteristic of the element. Only radiation produced by electrons from the the K, L and M shells is used in analysis since the N, O, P and Q shells of heavy elements produce radiation in the optical spectrum. It is not possible to detect elements lighter than Na^+ using EDAX. For elements with atomic weights between those of Na^+ and Zn^+ the K radiation is used, for elements between Zn^+ and Ag^+ the L series, and for elements heavier than Ag^+ the M series lines are measured.

4:8.2. QUALITATIVE ANALYSIS

Qualitative EDAX analysis was carried out routinely on the J.S.M. 35 and the J.S.M. 733. Spot analyses or raster analyses can be made on any part of the specimen viewed in the CRT screen, although care must be taken when using rough specimens that the X-ray detector has a clear "line of sight" to to analysed area. Frequently, when studying clay minerals in pores, it is not possible to analyse the phase under study as the pore walls stop the X-rays produced from reaching the detector. (Fig 4:5).

4:8.3. QUANTITATIVE ANALYSIS

Quantitative energy dispersive analysis was carried out using a Link Systems 860 analyser, with ZAF4 software. Cobalt was used as a standard. Sructural formulae were estimated using the "last element by difference" technique which assumes that the number of 0 atoms is known. Details of quantitative energy dispersive x-ray analysis are given in Goldstein et al (1981). The basic principles of analysis outlined above apply, however quantitative analysis incorporates corrections for geometrical and atomic weight factors.

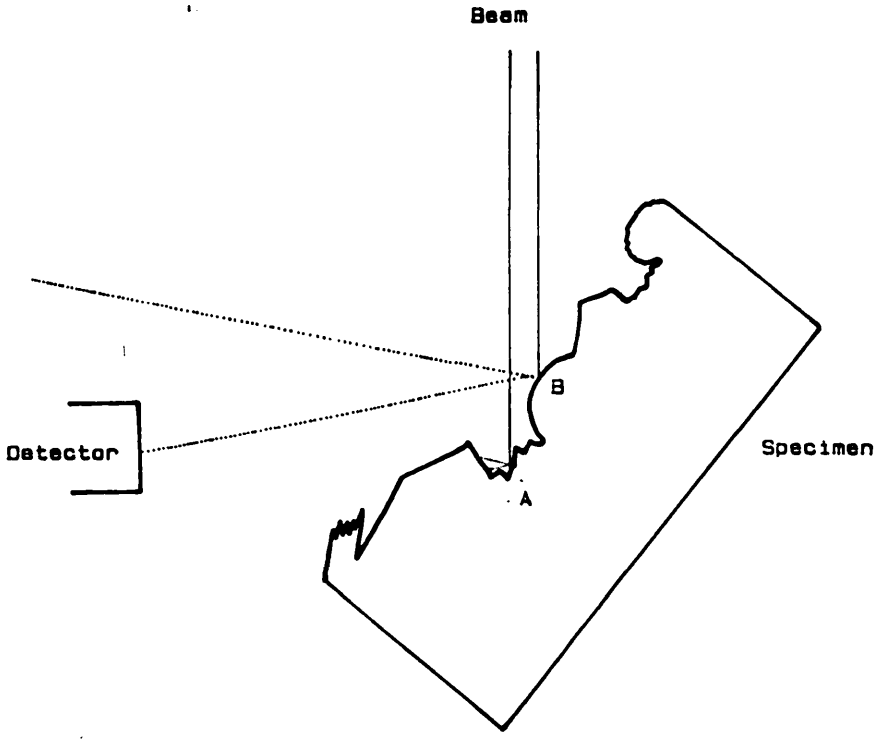
4:8.4. CATHODOLUMINESCENCE (C.L.)

Cathodoluminescence is light in the visible spectrum emitted when a semiconductor or insulator is bombarded with electrons. Details of the mechanism of production are given in Goldstien et al, (1981). The spectrum of light emitted is sensitive to

impurities in the substance bombarded; therefore, viewing a specimen in C.L. mode can reveal details which were invisible using other techniques. Authigenic overgrowths are often very pure and exhibit little or no C.L.. This property was used to define the margins of detrital cores of overgrown quartz grains and to detect any pressure dissolution present. Polished thin sections were examined in C.L. mode using the J.S.M. 35 fitted with a C.L. detector.

FIGURE 4:5

EFFECT OF SURFACE TOPOGRAPHY ON ANALYSIS



x-rays generated at A will not be detected

x-rays generated at B will be detected

4:9. VITRINITE REFLECTANCE

Vitrinite reflectance measurements (Ro %) of polished blocks of shale samples from the Trumfleet 1 well were carried out by the B.P. Research Centre at Sunbury. The reflectivity (in oil immersion) of vitrinite macerals increases with the temperature to which the macerals are subjected. Measurement of vitrinite reflectance, therefore, enables an estimate of the maximum temperature to which a sample has been subjected since the reflectivity value is not decreased by subjecting the samples to lower temperatures (Waples, 1980. Hunt, 1979).

Vitrinite reflectance values can be used to "window" the maturity of source rocks with respect to hydrocarbon generation (Hunt, 1979: Table 4:3), however the relationship between Ro and maturity can vary from basin to basin and care must be taken in interpreting vitrinite reflectance data.

4:10. LOPATIN TTI MATURITY MODELING

The maturity of a particular source rock can be modelled by considering it's burial history and the temeperatures to which it has been subjected as a consequence (Waples, 1980). The accuracy of such studies is dependant on being able to accurately define the burial history and the evolution of the heatflow (and therefore geothermal gradients) of a particular basin. The Lopatin time temperature index (TTI) can be defined by the following equation:

EQ 4:2

$$TTI = \sum_{i=1}^{n} (ti)(yi)$$

where ti is the length of time in millions of years spent by the rock at the ith temperature interval, and yi is the factor by which the rate of oil generation changes with increase in temperature (Waples, 1984)

TTI modeling was carried out for the Trumfleet 1 well using a geochemical modeling programme "MATURE" (Copyright IGI 1896), The programme and the computing facilities were supplied by the courtesy of British Gas plc.

TABLE 4:3

MATURITY	VITRINITE REFLECTANCE Ro (%)	TTI INDEX
ONSET OF OIL GENERATION	0.55	10
PEAK OIL GENERATION	0.65	15
WET GAS GENERATION	1.00	70

Comparison of TTI index with vitrinite reflectance.

From Waples, 1980.

The relationship between Ro and TTI is that given in Waples (1980), since the data required to estimate a basin specific Ro/TTI equation was not available.

PLATE 4:1

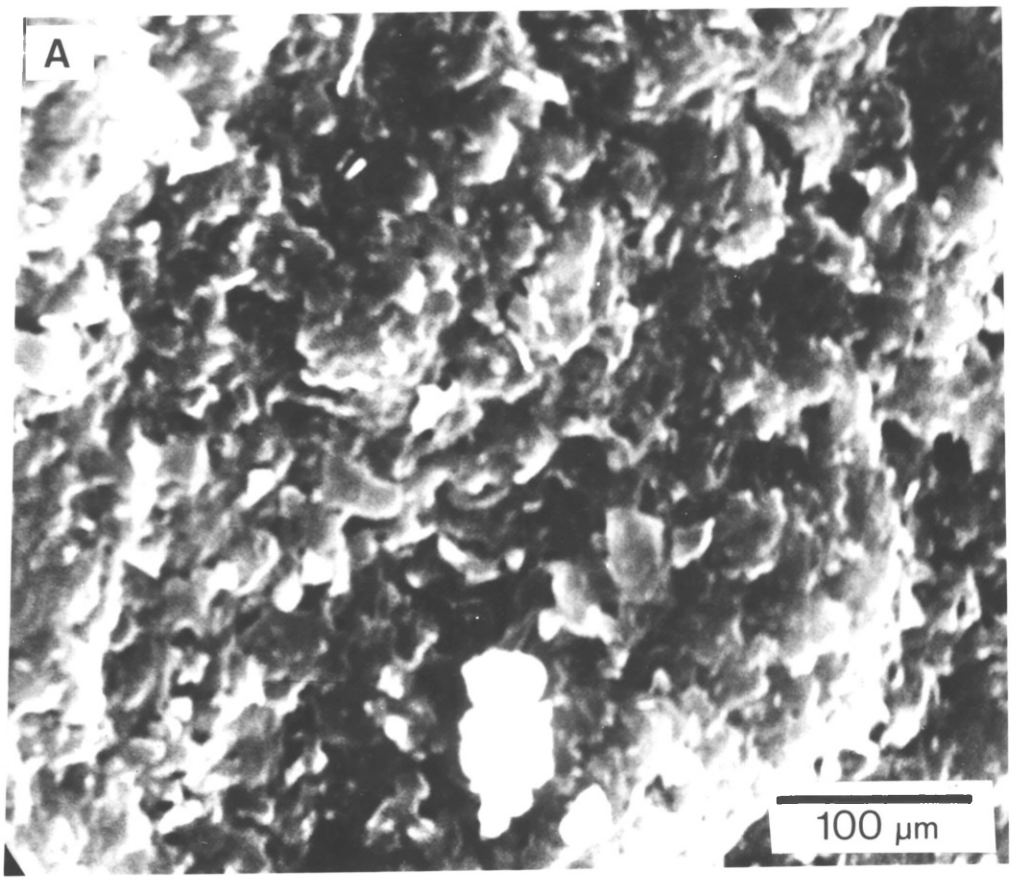
A

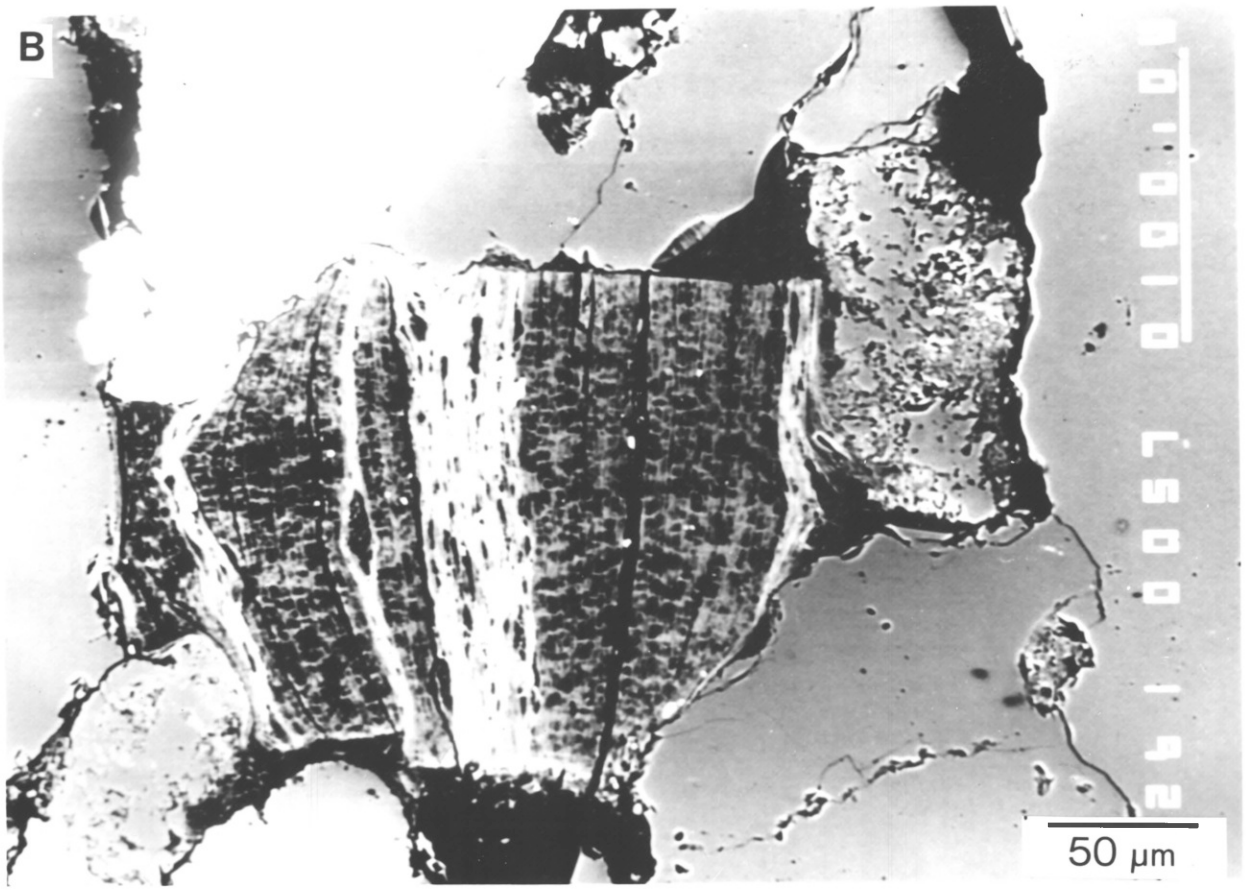
SEI image of the surface of a disc used for clay mineral XRD. The surface is not completely flat, and can be seen to consist of flocculated aggregates of clay minerals.

B

Backscattered electron image of a sandstone sample. The shades of grey on the photomicrograph reflect the atomic number contrast between the phases on the specimen. The darkest patches correspond to pore spaces, whilst the mid grey phases are quartz and feldspar. Ferromagensian minerals appear light owing to the higher atomic number of the atoms within the crystal lattice (Especially Iron).

Plate 4:1





NB: The term "shelf" used in this thesis refers to a structurally positive, slowly subsiding block (Kent. 1966) and not the continental shelf.

CHAPTER FIVE

SEDIMENTARY ENVIRONMENTS

5:1. THE TRUMFLEET FIELD

The Trumfleet structure is an irregularly shaped faulted dome situated on the Northern margin of the Gainsborough Trough at the intersection of the Askern-Spital High and the Don fault system (Fig 5:1, Fig 5.2)

Seven wells were drilled over an 22 year period from 1957 to 1979, and the licence (XL 033) was abandoned by B.P. in 1980. The field produced virtually no oil on drill stem testing but did produce uneconomic amounts of gas from Namurian sandstones. R.T.Z. Oil and Gas have been awarded an exploration licence over licence XL 033 in which the Moss Farm and Trumfleet wells were drilled, XL 181. The Askern Well was drilled in licence PL 161, which is now operated by Candecca.

Figure 5:2 shows a conjectural cross section through the field, and Figure 5:3 shows the stratigraphy of the studied interval.

5:2. SEDIMENTARY CYCLES (Mesothems)

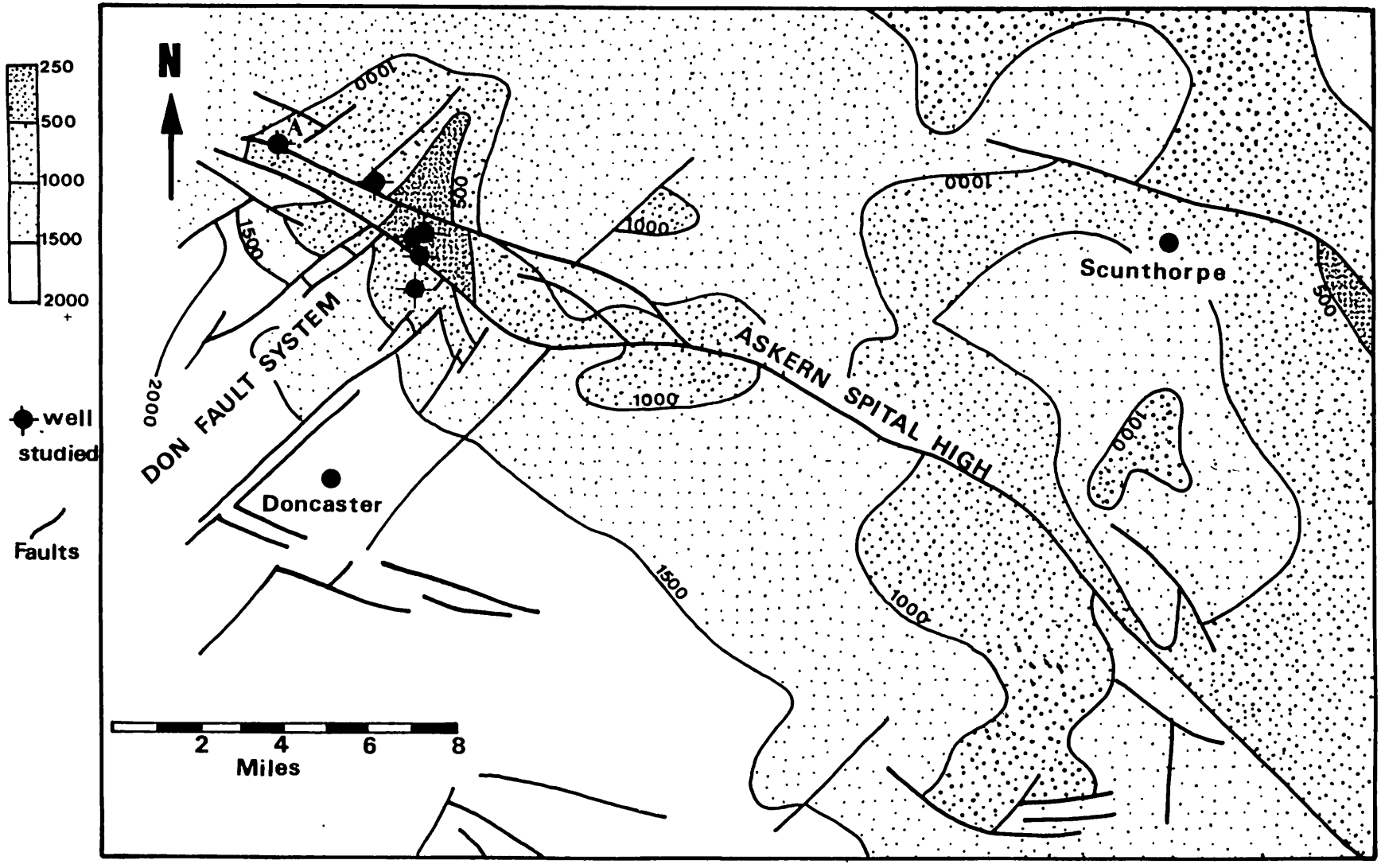
Ramsbottom (1977) proposed a sequence of 11 mesothems in the Namurian of Britain. Each mesothem commences with a transgressive marine phase at the base and is characterised by a shallowing upward sequence often with a seatearth and coal at the top.

Of the 11 cycles proposed by Ramsbottom, three (N9 to N11)

are recognised over the interval studied, although the base of N9 is not identified as no characteristic goniatites were identified below the R. gracile marine band (fig 5.3)

Recent work (Holdsworth and Collinson 1988) has cast some doubt over the tectonoeustatic nature of the origin of Ramsbottom's mesothems, but in no way diminishes their usefulness as correlatable marker horizons.

Structural Contour Map (Depth to Top Hard Coal)



Contours in feet

From Kent (1966)

FIGURE 5:1

CONJECTURAL SECTION THROUGH THE TRUMFLEET FIELD

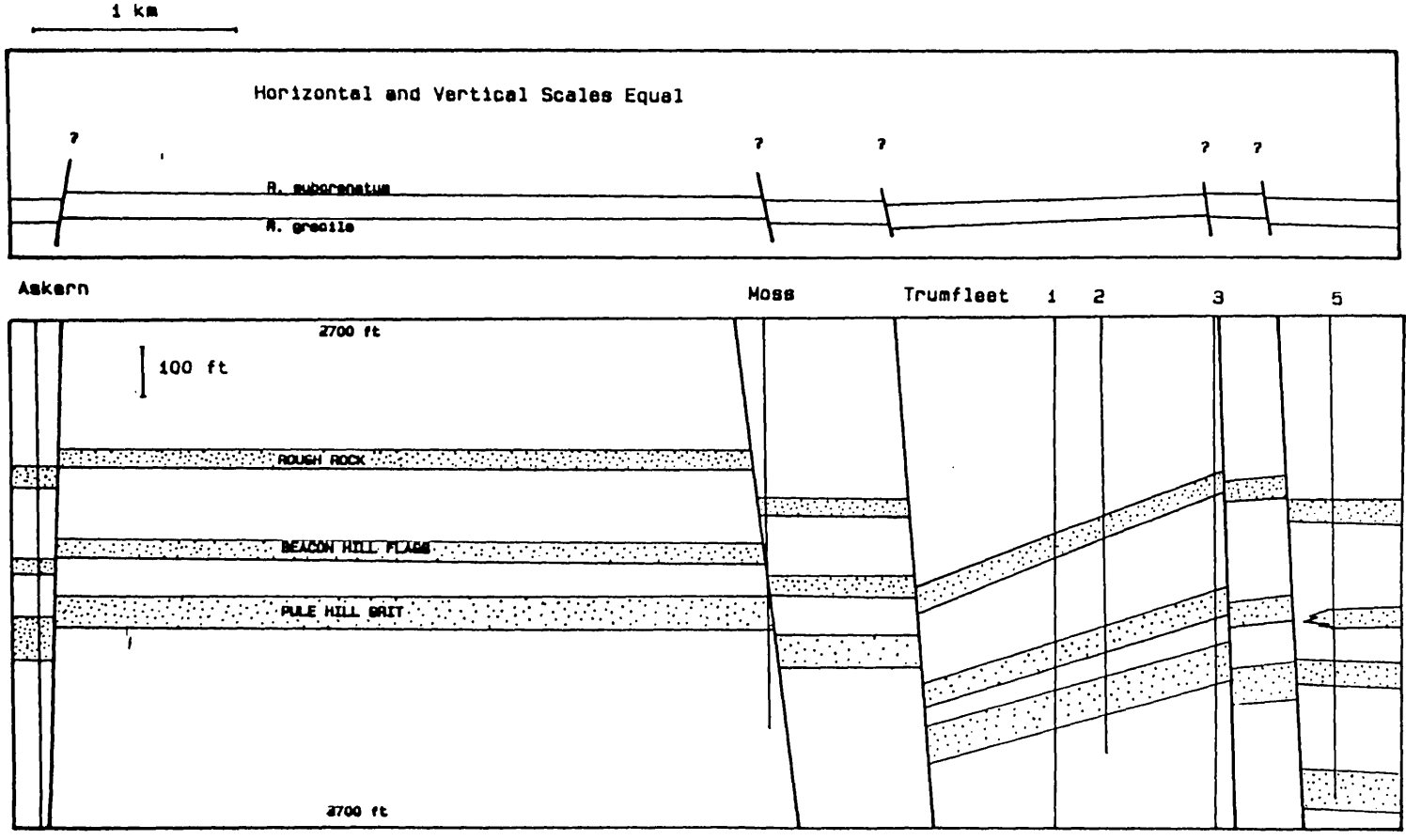
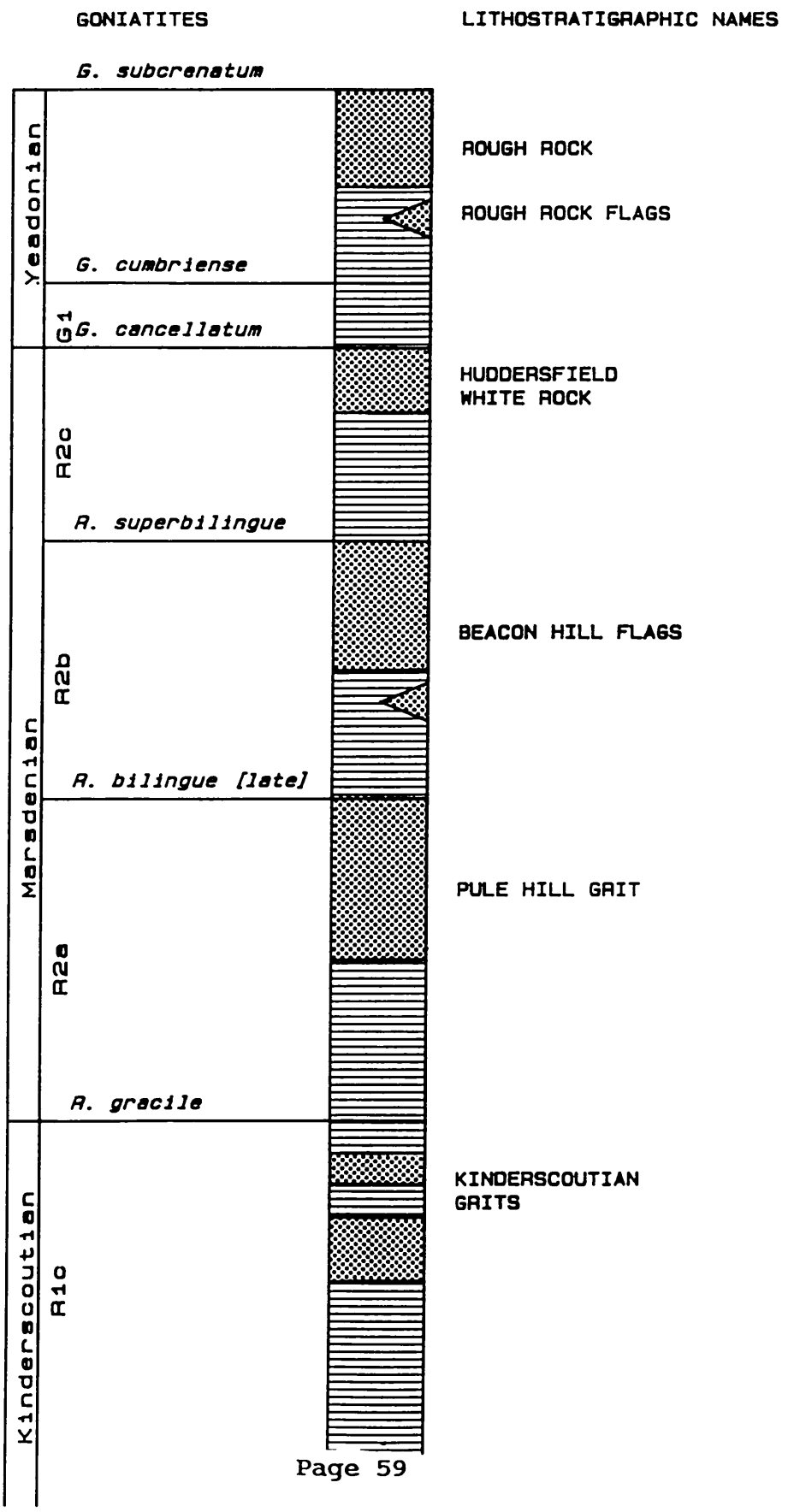


FIGURE 5:2

FIGURE 5:3

STRATIGRAPHY OF THE TRUMFLEET FIELD



5:3. CORRELATION

5:3.1. INTRODUCTION

Lithostratigraphic correlation was carried out using B.P. composite logs which show lithology, gamma ray, S.P. and resistivity. The gamma ray log was used preferentially as it was the only wireline log available for all the wells studied. All core depths were corrected to agree with the gamma ray logs.

The marine bands encountered in the Moss Farm well were identified by Dr. W.H.C. Ramsbottom of the B.G.S. in Leeds (Ramsbottom 1980) and where possible these were identified on the wireline logs and extrapolated to the other wells.

5:3.2. MARINE BANDS

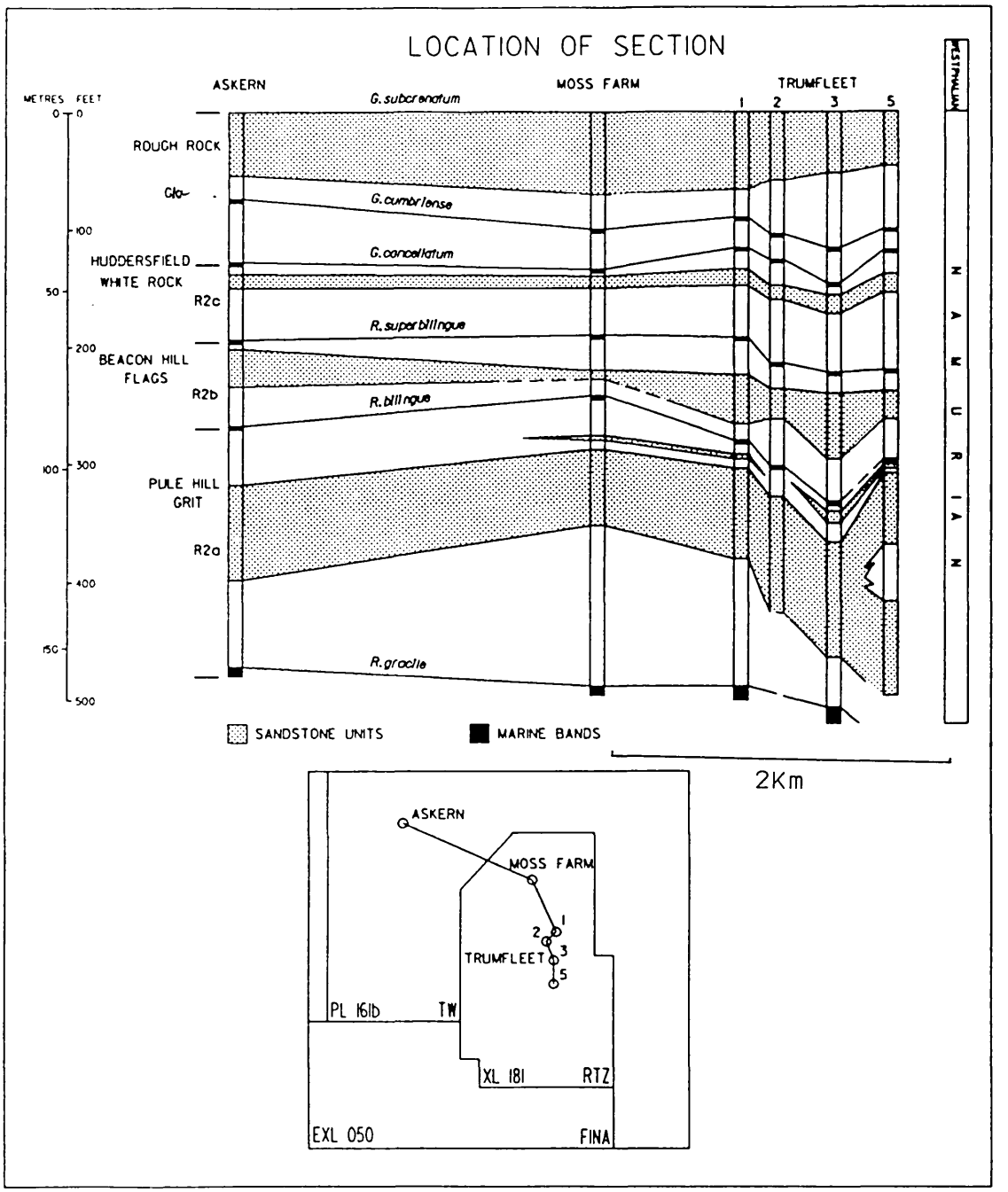
The Reticuloceras gracile marine band is particularly thickly developed and produces a marked gamma ray peak which can be traced from Moss Farm into Trumfleet 1, 3 and Askern.

The Gastrioceras subcrenatum marine band does not produce a marked increase in gamma ray activity but a gamma ray peak was observed 2 m below a recorded occurrence of marine goniatites, at the top of the Rough Rock (when the depth was corrected to agree with the log depth (Fig 5:10)) This peak was observed in all the wells studied and the G. subcrenatum marine band was picked 2 m above this peak.

The G. cumbriense and G. cancellatum marine bands produce well defined gamma ray deflections and are easily identifiable in the wells studied. The two marine bands are separated by a zone

FIGURE 5:4

WELL CORRELATION ACROSS THE TRUMFLEET FIELD



containing a non marine bivalve fauna which also produces an increase in gamma ray activity (Fig. 5.4). The G. cumbriense band is over 2m. thick in Moss Farm, whilst it is rarely more than 30cm. thick in the Gainsborough Trough (Ramsbottom 1980). Thick development of marine bands is characteristic of shelf or near coastal environments, however they may not contain their full suite of fossils on the shelf areas.

A small deflection produced in the R. superbilingue marine band was correlatable with other wells but characteristic goniatites of the R. bilingue zone were absent from Moss Farm. A thin Lingula band is taken to represent the R. bilingue marine band in Moss Farm (Ramsbottom, 1980), which produces no gamma ray deflection. The situation is complicated by the presence of a fault which intersects Moss Farm between R. bilingue and R. superbilingue, removing the Beacon Hill Flags section from the well (Figs 5.2 & 5:4) . The R. bilingue marine band was therefore taken to be coincident with the Lingula band at 989 m in Moss Farm, but was picked in the other wells as being coincident with a gamma ray peak approximately 8 m above the top of the Pule Hill Grit (Fig 5:4, Fig 5:9). The position of the top of the R2a zone is therefore somewhat arbitrarily chosen in the wells studied as it is likely that the R. bilingue marine band was not developed over the area. The distribution and thickness of marine bands is highly variable, Kent (1966) states that marine bands are commonly absent or do not contain their full suite of fossils over the shelf areas, even though they may be thickly developed. and Ramsbottom (1980) notes that the absence of characteristic goniatites in the R. bilingue marine band is characteristic

of deposition in the shelf environment. No goniatites characteristic of the R. bilingue zone were identified in the other wells so it seems likely that owing to the position of the field on the edge of the Gainsborough Trough, there was not a sufficiently developed marine transgression during R. bilingue times to allow goniatites to colonise the Trumfleet area.

5:4. DESCRIPTIONS AND ENVIRONMENTAL INTERPRETATION

5:4.1. PRE-*R. GRACILE* (Kinderscoutian and undifferentiated)

INTRODUCTION

Four wells penetrated strata below the *R. gracile* marine band: Trumfleet 1 and 3, Moss Farm and Askern: however, no core was available from Trumfleet 3. Trumfleet 1 and Askern reached total depth in the Dinantian at 4780 ft (1457 m) and at 4706 ft (1453 m) respectively. Detailed environmental interpretation was carried out only on the uppermost sandstone unit, and no petrographic interpretation was carried out.

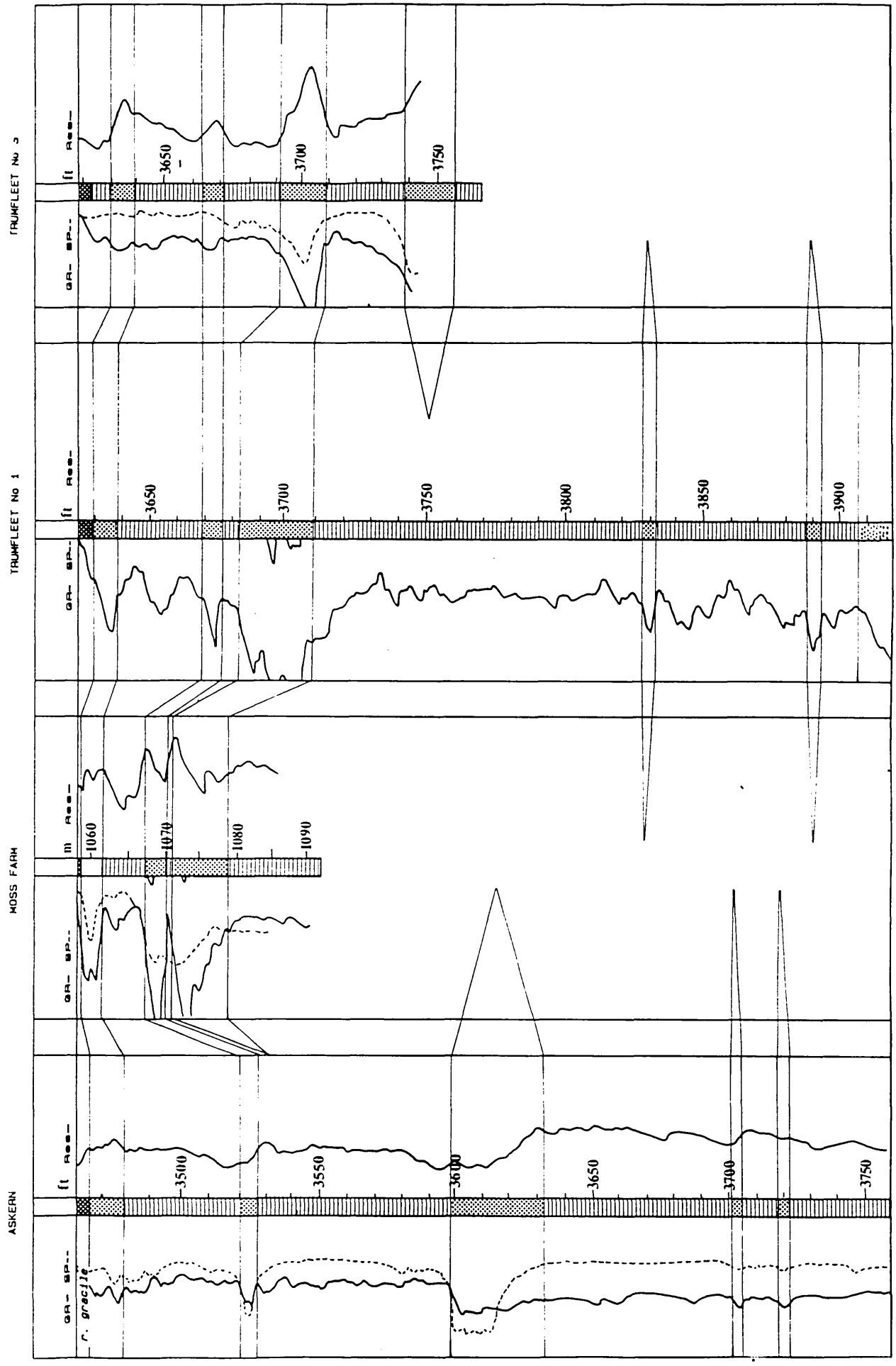
5:4.2. GENERAL DESCRIPTION

The Dinantian rocks from Trumfleet 1 consist of tight coarsely crystalline limestones which are overlain by dominantly argillaceous Namurian strata, consisting of dark grey and black shales and mudstones, interlaminated with siltstones and thin sandstones, and occasional thin limestones. Seatearths and thin coals are present throughout the sequence, and the sandstones often contain siderite and chamosite. No diagnostic goniatite zone can be identified below *R. gracile*, but non marine bivalves and *Lingula* are present in the siltstones and shales.

The thickest sandstones (up to 15 m thick) are developed at the top of the 1400 ft (427 m) sequence (Fig 5:5). The gamma ray log shows a gradual upward decrease in activity over a 150 ft interval from 3850 to 3600 ft (1174 to 1100 m) and the S.P. shows a corresponding "coarsening upwards" sequence over the same

FIGURE 5:5

GAMMA RAY CORRELATION OF KINDERSCOUTIAN SANDSTONES



interval (Fig 5:5), however no thick sand units are developed.

5:4.3. GENERAL INTERPRETATION

Kent (1966) stated that the Dinantian rocks deposited on the shelf or block areas were commonly coarsely crystalline pure limestones, similar to those recorded in Trumfleet 1, whereas rocks deposited in basinal areas were limestones interbedded with argillaceous deposits. The absence of any marine fauna and the presence of non marine bivalves and seatearths and coals indicate deposition in shallow fresh or brackish water and swamp conditions. The presence of early siderite and chamosite indicate a reducing geochemical environment (See chapter 7).

5:4.4. DISCUSSION

Fig 2.3 shows that by the end of the Kinderscoutian, about the time of the deposition of the R. gracile marine band, the Kinderscoutian delta had advanced some 100 km South of the Trumfleet area, filling the Central Pennine Basin with over 2000m of clastic sediments of which the majority were argillaceous (Anderton et al, 1979). No thick sandstone units representing major deltaic advance as described by Walker (1966), Collinson (1968), Reading (1964), McCabe (1977) are present in the Trumfleet area. Steele (1988) has interpreted the apparent coarsening upwards sequence described above as the lateral equivalent of the Kinderscoutian Delta, however, the Trumfleet sequence was produced by deposition on a stable shelf area on the eastern margin of the Kinderscoutian delta and this marginal position probably explains the lack of thick coarse clastics developed in the Trumfleet area.

5:5. Pre-R2 SANDSTONES (Kinderscoutian Grits)

5:5.1. DISTRIBUTION

The sandstones occurring up to 30 m below the R. gracile marine band are termed "Kinderscoutian Grits" by B.P. geologists and the name is used informally in this thesis. Fig. 5:5 shows the distribution of these sandstones in the wells studied and Fig. 5:6 shows the detailed logs for the sandstones for which core was available. Two sandstone units separated by a dark fossiliferous shale containing plant fragments, Lingula and non marine bivalves. No petrographic analysis was carried on these sandstones.

5:5.2. UPPER SANDSTONE UNIT

The upper sandstone unit is approximately 2 m thick and consists of dark grey/brown, medium to fine grained sandstones which are poorly sorted. Interbeds of shale rich in organic matter are common. The sandstones are weakly carbonaceous and siderite is concentrated into bands which are rich in organic matter and mica (mainly muscovite). A thin (3 cm) coal seam is present at the top of the unit and is underlain by 20 cm of silty shale containing rootlets and organic matter.

5:5.3. LOWER SANDSTONE UNIT

The lower Kinderscoutian Sandstone unit is approximately 18 m thick, the basal contact is difficult to identify as the base of the unit is gradational with the underlying siltstones and shales. It shows a composite structure, comprising two

Fold Out

coarsening upwards units approximately 4 m. thick, separated by a 1-5 m thick silty shale (Fig 5:6). The sandstones are fine to medium grained and are dark grey in colour. Sorting is variable but the sands are generally argillaceous and micaceous and often contain bands rich in organic matter. Interbeds of siltstone and shale are common throughout the sandstone unit. Thin coal seams and rooted horizons are present at the top of each of the coarsening upward units (Fig 5:6). The coals are 1 to 2 cm thick and are sulphurous.

5:5.4. SEDIMENTARY STRUCTURES

The base of the upper unit is erosive on the underlying shale, with ripple cross laminated sands lying immediately above the erosion surface. The rest of the unit shows indistinct parallel lamination picked out by mica rich laminae or discontinuous carbonaceous laminae. The parallel lamination is often disturbed by bioturbation, load casting and water escape structures, giving a mottled appearance to the sandstone.

5:5.5. ASSOCIATED SEDIMENTS

5:5.5.1. SILTY SHALES

The argillaceous unit separating the upper and lower sandstone units varies from 7 m thick in Moss Farm to 13 m in Trumfleet 1. The unit consists of black silty shales and dark grey siltstones rich in plant fragments and containing a Lingula and non marine bivalve macrofauna. Bioturbation, and water

escape structures are present. The more silty horizons show micro current ripple lamination with an amplitude of less than 1 cm. Siderite nodules up to 2 mm in diameter occur in discrete horizons.

5:5.5.2. MARINE BANDS

No characteristic goniatites were recorded below the R. gracile marine band in the wells studied since very little core was cut in any of them below this zone. Lingula, Posidonia and fish fragments were identified at 3944 ft (1202 m) in the Trumfleet No 1 well, and Lingula and fish remains were present from 3515 to 3520 ft (1071-1073 m) in the Askern well.

5:5.5.3. SEDIMENTARY STRUCTURES

The lower unit sandstones show ripple cross lamination throughout the unit as well as rare trough cross lamination. Soft sediment deformation structures; load casting, water escape structures, slumping and faulting, are common where sandstones and siltstones are interbedded (Plate 5:1). Bioturbation, possibly Pelecypodichnus? was recorded near the base of the lower sandstone unit in the Moss Farm well.

5:6. INTERPRETATION

5:6.1. INTRODUCTION

The sediments deposited in the uppermost pre R. gracile sequence shown in Figure 5:6 were deposited in two distinct sedimentary environments; as siltstones and shales of the interdistributary facies, and as sandstones and siltstones

deposited by overbank flooding and crevasse splaying, building minor delta lobes. The sequence found in the Trumfleet Field is similar to that described by Elliot (1974, 1978) and to that recorded in the Durham Coalfield by Fielding (1984).

5:6.2. THE INTERDISTRIBUTARY FACIES

Rocks of the interdistributary facies are predominantly siltstones and silty shales, rich in organic matter, which were laid down in shallow bays and lakes between distributary channels on the delta plain (Coleman and Gagliano, 1965, Hawkins, 1972, Elliot, 1974, 1978, Fielding, 1984).

The rocks are dark in colour, show parallel lamination and contain abundant organic matter and plant fragments. They are micaceous and often contain pyrite and siderite. Occasionally, the bedding is disrupted by bioturbation and the rocks contain non-marine bivalves in discrete horizons. The dark grey to black colour, the predominance of parallel lamination and the presence of alternations of silt and shale indicate deposition from suspension; and the lack of a true marine fauna indicate a fresh to brackish water environment (Coleman and Gagliano, 1965, Elliot, 1974, 1978). Fielding (1984) describes anoxic lake floor sediments which are similar to those at Trumfleet, although they contain pyrite. Both pyrite and siderite are present in rocks of the interdistributary facies at Trumfleet indicating either temporal or/and spatial variations of the geochemical environment. The "anoxic lake floor" facies of Fielding was not differentiated from oxygenated substrata in this study.

The term "interdistributary facies" as used in this thesis is

used as a general term for dark, parallel laminated, argillaceous rocks which were deposited in a variety of sub environments (fresh water lake bottom, interdistributary bay) on the delta plain. This facies most closely corresponds to the "interdistributary facies" as described by Hawkins (1972).

5:6.3. CREVASSE SPLAY FACIES

The crevasse splay facies consists of the argillaceous sandstones of the upper and lower Kinderscoutian sandstone units shown on Figure 5:6; interbedded with fine silt and shale laminae which show a coarsening upwards grain size profiles and log motif (Figs 5:5 & 5:6). Individual crevasse lobe deposits are 2-6 m thick and the sandstones commonly exhibit ripple cross lamination and parallel lamination, as well as soft sediment deformation structures. It is probable that these rocks were deposited by a mixture of true crevasse splaying and overbank flooding, but it was not thought viable to attempt to differentiate between the two processes owing to the paucity of samples and the lack of continuous exposure.

Also included in this facies are thin coals and seatearths which were deposited in a swamp sub-environment often on top of the crevasse deposit (Fig 5:6). This facies corresponds to the "medial crevasse splay facies" of Fielding (1984). The upper Kinderscoutian unit described previously was deposited as a crevasse splay lobe over which a swamp conditions developed prior to the R. gracile transgression. The lower Kinderscoutian unit was deposited as two stacked crevasse splay lobes separated by a crevasse channel sandstone and a fine grained channel fill

deposit (Fig 5:5).

5:6.4. CREVASSE CHANNEL FACIES

A 3 m thick, fining upwards sandstone body in the middle of the lower Kinderscoutian unit has been interpreted as being deposited as a crevasse channel sandstone on the basis of its erosive base and fining upwards nature, as well as the presence of trough cross lamination passing upwards into ripple and parallel lamination. Similar crevasse channel sandstones were described by Elliot, (1974, 1978), Guion, (1971) and Fielding, (1984). The erosive base and the presence of trough cross bedding indicate deposition from a higher energy environment than the crevasse splay sandstones. Fielding (1984) describes crevasse sandstones as being deposited in confined and partially confined channels which may quickly become blocked with sediment.

5:6.5. CONCLUSIONS

The pre R. gracile deposits of the cored sections of the Trumfleet field were deposited as a series of crevasse splay lobe and overbank sandstones prograding over interdistributary area of fresh and brackish water lakes and bays on the delta plain of the eastern margin of the Kinderscoutian delta.

5:7. R2a Zone (Marsdenian)

5:7.1. INTRODUCTION

The thickness of the R2a zone varies between 68m in Moss Farm and 102m in Trumfleet 3. The base of the zone was not drilled in Trumfleet 5. The entire zone thickens towards the South, the direction of the Gainsborough Trough (Fig 5:4) The zone corresponds to Ramsbottom's N 10 cycle (Ramsbottom, 1977) and is predominantly argillaceous although one major 25 to 35 m thick sandstone unit, the Pule Hill Grit (Also named as Trumfleet Grit by B.P.) occurs towards the top of this zone (Fig 5:4, Fig 5:7).This sandstone unit splits into two separate units in the Trumfleet 5 well. The base of the zone is marked by the R. gracile marine band.

5:7.2. R2a SANDSTONES (Pule Hill Grit)

The Pule Hill Grit consists of light grey and brown poorly sorted sandstones, which are predominantly coarse grained but contain occasional conglomeratic horizons with clasts up to 5 cm in diameter: the most common large clasts are polycrystalline quartz grains, feldspars and occasional granitic rock fragments. The sandstones contain 10 to 15% feldspar and up to 2% mica and white kaolinite is conspicuous in hand specimen. The top 5m of the unit consist of interbedded fine to coarse sandstones interbedded with siltstones and dark grey shales which pass upwards into a 1.5 m thick sandstone containing roolets and carbonaceous laminae.

The gamma ray log shows a box like motif with a sharp base

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and top although the S.P. log shows a more gradational top to the unit The base of the sand becomes more gradational in Trumfleet 3, and the Trumfleet 5 logs show two separate sandstone units (probably faulted repeat sections), although the base of the lower one was not penetrated (Fig 5:7). Grain size profiles show that the main part of the sandstone consists of stacked fining upwards units with an erosive base. Pebbly lag conglomerates at the base of these units pass upwards into massive coarse grained sandstones.

5:7.3. ASSOCIATED SEDIMENTS

5:7.3.1. SILTSTONES AND SHALES

The base of the Pule Hill Grit overlies interlaminated siltstones and shales which show very regular lamination, with individual beds up to 1cm thick. These rocks are similar to the "striped siltstones" described by Collinson (1968). The striped siltstones are underlain by dark grey silty flat bedded shales (Fig 5:7). Lingula was identified in samples from a depth of 1041m in the Moss Farm well. The shales also contain Palaeonielo and fish fragments.

5:7.3.2. MARINE BANDS

The Reticuloceras gracile marine band is 9m thick in the Moss Farm well and as well as R. gracile it contains Lingula mytiloides, Dunbarella, Palaeoneilo, Anthracoceratites and mollusc spat. Ramsbottom (1980) comments that the thick development of this marine band and the presence of benthonic

species is indicative of a shelf or near coastal environment.

5:7.4. SEDIMENTARY STRUCTURES

Due to the coarse grained nature of the Pule Hill Grit, bedding is often toodifficult to discern. Where it is observed in the massive part of the sandstone, large scale trough cross bedding is the most common bedding type. The bedding appears to be planar on the scale of the core but the dip angle varies in relation to the core axis so it is probable that it was produced by trough cross lamination (Plate 5:2a). Intraformational shale rip-up clasts are present in the microconglomeratic lag deposits; these have often been flattened by mechanical compaction and now appear as lenses of shale.

Ripple cross lamination is preserved more commonly towards the top of the Pule Hill Grit and is the most common bedding type in the medium and fine grained sandstones of this unit.

Core from the Moss Farm well shows a variety of deformational features, mainly soft sediment slumping and faulting. At a depth of 1007m in Moss Farm the core has a brecciated appearance with slumped and faulted beds juxtaposed and rotated with respect to each other. (Plate 5:3). Carbonaceous tubes, filled with sand are common. These form by the infilling of the hollowed out stems of plants (Plate 5:3). Wavy bedding and discontinuous carbonaceous laminae are conspicuous in the medium to fine grained parts of the Pule Hill Grit.

Parallel lamination is the dominant bedding type in the argillaceous sediments.

5:7.5. INTERPRETATION

5:7.5.1. INTRODUCTION

Five distinct facies can be recognised from the R2a zone. A distributary channel, overbank and crevasse splay facies, corresponding to the arenaceous members of the zone with an interdistributary siltstone and shale facies and a marine shale facies corresponding to the argillaceous members of the zone.

5:7.5.2. MARINE SHALE FACIES

The marine part of the R2a zone is represented by the R. gracile marine band and is identified by the presence of a marine fauna and its characteristic gamma ray log motif (Fig 5:7).

5:7.5.3. THE INTERDISTRIBUTARY FACIES

The interdistributary facies comprises the striped siltstone lithofacies and the underlying shales. This facies has variable thickness and is essentially identical to the pre-R. gracile interdistributary facies in lithology and environment of deposition.

5:7.5.4. THE DISTRIBUTARY CHANNEL FACIES

The distributary channel facies comprises that part of the Pule Hill Grit which displays a box like log motif and consists of coarse grained, pebbly, trough and ripple cross-laminated sandstones. In detail the distributary channel sandstones are seen to comprise a number of stacked fining upwards units. An argillaceous channel fill deposit is present in the middle of the

Pule Hill Grit in the Askern and Trumfleet 2 and 3 wells. English Carboniferous distributary channel sandstones have been described by Collinson (1968), Hawkins (1972), McCabe (1977), Jones (1977), Guion (1971), Huggett (1982) and Oklo (1983) and Steel (1988). Fisher (1969) astates that box like log motifs are characteristic of distributary channel sandstones in the Mississippi Gulf Coast.

5:7.5.5. OVERBANK FACIES

The overbank facies is composed of the interlaminated sandstones, shales and siltstones which make up the upper part of the Pule Hill Grit (Fig 5:7). The sandstones show ripple and wavy lamination and are occasionally erosive on the underlying shale beds. The shale is silty and normally shows parallel lamination. Sphaerosiderite nodules and thin seatearths with rooted horizons are also common. The bed boundaries often show load casting and bioturbation is common. Coleman and Gagliano (1965) noted that load casting was rare in subaerial levee deposits but was common in subaqueous levees.

The overbank facies was deposited as a series of flood pulses over the interdistributary area. The facies corresponds to the "subaqueous levee" deposits of Coleman and Gagliano (1965) and the "levee" deposits of Hawkins (1972) and the "overbank" facies of Fielding (1984).

5:7.5.6. CREVASSE SPLAY FACIES (Minor mouth bar)

The crevasse splay facies is represented by a coarsening upwards unit at the base of the Pule Hill Grit in the Trumfleet 3

well, identified solely on the basis of its log motif. It probably represents minor mouth bar type deposition in deeper water, at the downstream end of the Pule Hill Grit channel, prior to the deposition of the erosively based channel sandstone, since the Trumfleet 3 well is closer to the depositional low of the Gainsborough Trough. Minor mouth-bar deposition is gradational to crevasse splay deposition (Fielding, 1984) and the unit is included within the crevasse splay facies.

5:7.5.7. CONCLUSIONS

Rocks of the R2a zone were deposited on the delta plain in interdistributary bays and lakes with occasional incursions of sand sized material as crevasse and overbank deposits. A major distributary channel, represented by the coarse channel fill deposits of the Pule Hill Grit, was active in the Trumfleet area during R2a times.

5:8. R2b ZONE (Marsdenian)

5:8.1. INTRODUCTION

The R2b zone contains one major sandstone body, the Beacon Hill Flags (Fig 5:4). The entire zone reaches a maximum thickness of 30 m but only 8 m is present in Moss Farm as a the bore intersects a fault plane (Fig 5:2). No continuous core from this interval was available from Moss Farm so study of the R2b zone was confined to log interpretation and the study of representative one foot core samples of the sandstone unit. Sample coverage over the sandstone unit was patchy in all the wells studied (Fig 5:8).

5:8.2. R2b SANDSTONES (Beacon Hill Flags)

The lithology of the Beacon Hill Flags is highly variable both in grain size and sandstone type. The coarser parts of the unit are quartzites which contain up to 95% quartz. They are brown in colour and consist of poorly sorted angular and subangular grains (Fig 5:8). Two types of siderite are present; microcrystalline siderite is common throughout and nodular sphaerosiderite occurs in the finer grained rocks at the top of the unit where it comprises up to 80% of the rock. Carbonaceous material is common throughout and is often concentrated in mica rich laminae.

A thin coal and siliceous seatearth are present at the top of the unit. The grain size profiles are highly variable and interpretation is hindered by lack of data; generally a fining

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upward sequence is observed. The thickness of the sandstone unit is also highly variable ranging from 6 to 20 m.

In the Trumfleet 1 well the lithological data show a fining upwards unit with a sharp base and gradational top although the gamma ray log suggests a gradational base and top. The S.P. log more closely relates to the grain size profile.

In the Trumfleet 2 well, the gamma and S.P. logs show a bell shaped motif whilst in Trumfleet 3 the unit has a multichannel appearance.

5:8.3. ASSOCIATED SEDIMENTS

5:8.3.1. SILTSTONES AND SHALES

The siltstones and shales underlying the Beacon Hill Flags are dark grey in colour, rich in mica and carbonaceous material and contain common interbeds of parallel laminated and massive siliceous seatearths.

5:8.3.2. MARINE BANDS

The Reticuloceras bilingue marine band is the lower bounding marine band in the R2b zone. No goniatites characteristic of this zone were found in any of the wells studied, so the base of the R2b zone was taken to be coincident with a Lingula band at 989m in the Moss Farm well.Ramsbottom (1980) noted that the R. bilingue marine band commonly deteriorates in the shelf areas.

5:8.4. SEDIMENTARY STRUCTURES

Small scale ripple cross lamination was the most common structure observed and cross bedding can be inferred by the variable bedding angle to the core axis. Pelecypodichnus? and a flat burrow were observed in the Askern well at 3701 ft (1128 m). Rootlets were observed in siliceous the seatearths.

The most commonly observed structure in the siltstones and shales was parallel lamination.

5:8.5. INTERPRETATION

5:8.5.1. MARINE SHALE FACIES

No true marine transgressive facies was recognised at the top of the R2b zone, the R. bilingue marine band is represented by a thin Lingula band.

5:8.5.2. LEVEE AND OVERBANK FACIES

The levee and overbank facies exhibits a fining upwards log motif and occur in the upper part of the Beacon Hill Flags. The thin development of the Beacon Hill Flags in Trumfleet 2 and Askern, the variable lithology, and the abundance of sphaerosiderite nodules indicate that the Beacon Hill Flags in the Trumfleet 2, 5 and Askern wells were probably deposited entirely as overbank sandstones, with no channel development. These sandstones show bioturbation and load casting.

A 2m thick siliceous seatearth and a thin coal are present at the above this facies.

5:8.5.3. DISTRIBUTARY CHANNEL FACIES

Distributary channel sandstones occur in the Trumfleet 1 and Trumfleet 3 wells only, and are characterised by a massive log motif. These sands contain less siderite than the overbank facies, and where siderite is present it is of the microcrystalline variety. The channel sandstones contain up to 95% quartz and are often cemented by ankerite. They show visual porosities of 2-13% as compared to 0-4% for the overbank and levee sandstones. Unfortunately no continuous core was available from this facies and 1ft core samples were available only from Trumfleet 1.

The basal part of the Beacon Hill Flags in Trumfleet 1 has been assigned to the distributary channel facies on the basis of its grain size lithology, and log motif.

5:8.5.4. CREVASSE SPLAY FACIES

The crevasse splay facies consists predominantly of siltstones and shales interbedded with thin sandstones and seatearths. The facies shows coarsening upwards log motifs and underlies the Beacon Hill Flags.

No core material was available for rocks of this facies in any of the wells studied and its identification is based entirely upon the interpretation of B.P. composite logs.

5:8.5.5. INTERDISTRIBUTARY FACIES

The interdistributary facies consists of dark siltstones and shales and is essentially identical to the interdistributary facies previously described in the pre-R. gracile sequence. It is

identified on the basis of its argillaceous lithology (as noted on B.P. logs) and log response, since no samples were available for study from this interval.

5:8.5.6. CONCLUSIONS

The rocks of the R2b zone were deposited on the delta plain by overbank deposition and crevasse splaying. The overbank deposits are characterised by fining upward log motifs and by sandstones rich in sphaerosiderite nodules which are interbedded with siltstones and shales. The crevasse splay deposits are lithologically similar but are characterised by coarsening upwards log motifs which reflect the progradational nature of their deposition (Elliot, 1974, 1978, Fielding, 1984).

Distributary channel sandstones are not laterally extensive and are characterised by a quartz-rich lithology, coarser grain size, and by a fining upwards grain size profile.

5:9. R2c ZONE (Marsdenian)

5:9.1. INTRODUCTION

The R2c zone varies in thickness from 22m in Trumfleet 1 to 32m in Trumfleet 5. One sandstone unit, the Huddersfield White Rock occurs near the top, and is immediately overlain by the Gastrioceras subcrenatum marine band.

Continuous core was available from Moss Farm and 1ft core samples were available for Trumfleet 1.

5:9.2. R2b SANDSTONES (Huddersfield White Rock)

The Huddersfield White Rock (Fig. 5:9) consists of light brown, medium grained, poorly sorted sandstones which grade downwards into siltstones. The unit is capped by a siliceous seatearth and a 6 inch thick (15 cm) coal seam. The sandstone is the thinnest of those studied and the maximum thickness reaches 4m. It is argillaceous and micaceous and contains abundant sphaerosiderite nodules. A shale 1-2m thick splits the sandstone into two units. The uppermost unit contains abundant carbonaceous material and rooted horizons; Carbonicola was recorded from the top, immediately below the seatearth. The lower unit, which grades down into the underlying siltstones, is interlaminated with thin shale interbeds up to 1cm thick. The Gastrioceras cancellatum marine band rests directly on top of the coal at the top of the upper unit (Fig. 5:9).

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5:9.3. ASSOCIATED SEDIMENTS

5:9.3.1. SEATEARTHS AND COALS

No core samples of the thick seatearth below the Huddersfield White Rock were available for any of the wells studied, although core of the siliceous seatearth deposited at the top of the Huddersfield White Rock was. This seatearth is light to dark grey in colour and appears to be structureless, apart from carbonaceous veining caused by the presence of rootlets.

The lower seatearth is described on the B.P. logs as being dark grey and brown, siliceous and containing ironstone. A 15cm thick coal seam is present at 3273 ft (997 m) in Trumfleet 1 and is underlain by a 6m seatearth which can be traced into Trumfleet 2, 3 and 5 but is absent from Moss Farm and Askern.

A number of coal seams up to 15 cm thick are present within the seatearths. They show little lateral continuity and cannot be traced from well to well.

5:9.3.2. SILTSTONES AND SHALES

The underlying argillaceous sediments contain Lingula and siderite nodules are common. The shales are silty and exhibit parallel lamination. Fish scales and Carbonicola are present in the shales overlying the Huddersfield White Rock, immediately below the G. cancellatum marine band.

5:9.3.3. MARINE BANDS

The Reticuloceras superbilingue marine band is 6m thick in Moss Farm and exhibits an increase in gamma ray activity at the

top and bottom of the band (Fig 5:9). Fossils present are R. superbilingue, Orbiculoidea, Palaeoniscid scales, Productid fragments and Lingula. Fish scales are present in the overlying shales.

Ramsbottom (1980) noted that the band shows thick development which, along with the presence of Lingula and Orbiculiodea is indicative of deposition in a shelf type environment.

5:9.3.4. SEDIMENTARY STRUCTURES

Current produced structures in the upper part of the Huddersfield White Rock consist of vague wavy lamination picked out by argillaceous bands. The most common structure in this part of the unit is parallel lamination of sandstone and shale interbeds. The base of the upper part of the Huddersfield White Rock shows load casting into the underlying shale. Ripple cross lamination is present in the lower part of the which also contains a high proportion of sphaerosiderite nodules. The nodules are associated with carbonaceous laminae which are often concentrated on the lee side of ripples. This lower unit appears mottled due to extensive bioturbation, and bivalve escape structures are present at the top of the upper sandstone unit, below a rooted horizon.

5:9.4. INTERPRETATION

Four distinct sedimentary facies can be recognised in the R2c zone; crevasse splay sandstones, interdistributary siltstones and shales, with a swamp subfacies represented by the presence of

a thick seatearth which is not associated with any sandstone deposition, and a marine shale facies.

5:9.4.1. MARINE SHALE FACIES

The marine band at the base of the zone is the R. superbilingue band and is identified by the presence of the characteristic R. superbilingue goniatite and its attendant marine fauna. The presence of a benthonic fauna suggests shallow water conditions during the deposition of this marine band (Ramsbottom, 1980).

5:9.4.2. INTERDISTRIBUTARY FACIES

The interdistributary facies consists of interlaminated siltstones and shales with occasional thin sandstones and seatearths, corresponding to a swamp subfacies where they are thickly developed. The environments were identified on the basis of their argillaceous lithology as identified from wireline logs since no core samples of this facies were available for study.

The swamp subfacies corresponds to the seatearths which occur below the Huddersfield White Rock, which were identified on the B.P. lithological logs.

5:9.4.3. CREVASSE SPLAY FACIES

The Huddersfield White Rock was deposited as a crevasse splay lobe sand body. The coarsening upward nature of the unit and the presence of bioturbation, and extensive load casting as well as the seatearth and coal seam present at the top of the sandstone unit are indicative of deposition as a prograding

crevasse lobe sand body (Elliot 1974, 1978. Fielding, 1984).

5:9.4.4. CONCLUSIONS

The rocks of the R2c zone were deposited under very shallow water and emergent conditions as indicated by the presence of seatearths and coals throughout the sequence. The Huddersfield White Rock was deposited as as couplet of crevasse splay lobes, prograding over interdistributary swamp and lake deposits (Fig 5:9).

5:10. Gla ZONE (Yeadonian)

5:10.1. INTRODUCTION

The Gla zone is the uppermost zone studied and the upper bounding marine band, Gastrioceras subcrenatum, marks the division between the Namurian and Westphalian. The Gla zone varies in thickness from 43m in Moss Farm to 75m in Trumfleet 1, and contains one sandstone unit, the Rough Rock. The Rough Rock is approximately 25m thick in all the wells studied, the difference in thickness of the zone from well to well is made up by the underlying argillaceous section of the zone (Fig. 5:10).

5:10.2. Gla SANDSTONES (Rough Rock)

Sandstones of the Rough Rock are medium to coarse grained, grey in colour, and are interbedded with thin, 1-2 cm thick siltstone and shale beds, especially in the upper part of the sandstone body. Where the rocks are carbonate cemented, the appear light grey in colour and large poikilotopic calcite crystals can be seen on freshly fractured surfaces.

The gamma ray and S.P. logs show a bell shaped, 20 m thick fining upwards motif leading up to a 1 m thick shale bed which is overlain by a further 3m of sandstone, interbedded with siltstone and shale (Fig 5:10). The lowermost unit consists of massive medium grained sandstones which contain discrete carbonate (ankerite and ferroancalcite) cemented zones. In Trumfleet 1 and 2, a 5 m thick coarsening upwards subunit can be seen on the gamma and S.P. logs at the top of a lower sandstone unit (Fig. 5:10).

Fold Out

No samples of this part of the sandstone were available for study.

5:10.3. ASSOCIATED SEDIMENTS

5:10.3.1. MARINE BANDS

Three gamma ray peaks corresponding to fossiliferous horizons are observed in the Gla zone (Fig 5:10)). Two contain characteristic goniatites (Gastrioceras cancellatum and Gastrioceras cumbriense) and are true marine bands. They are separated by a band containing a non marine bivalve and Lingula fauna.

The higher marine band, the G. cumbriense band, is 2.2. m thick and contains Lingula, Dunbarella, Anthracoceratites, G. crenulatum, G. cumbriense and fish fragments. Ramsbottom (1980) notes that the 2 m thickness of this band is normal for a near coastal or shelf succession.

The non marine bivalve band contains Carbonicola, Niadites and Giesina.

The G. cancellatum band is 3.3 m thick and contains Lingula, Dunbarella, Agastrioceras, Anthracoeratites and G. cancellatum. This suite of fossils represents the uppermost of three faunas characteristic of the G. cancellatum and is indicative of deposition on a shelf or near coastal environment (Ramsbottom, 1980).

5:10.3.2. SILTSTONES AND SHALES

The siltstones and shales underlying the Rough Rock are

dark grey to black in colour, micaceous and contain carbonaceous material. They are intimately interbedded and commonly show parallel lamination of the interbeds.

5:10.3.3. SEDIMENTARY STRUCTURES

The most common structures exhibited by the medium grained sandstones of the Rough Rock are planar and trough cross lamination and ripple cross lamination. Bedding is delineated out by siderite and carbonaceous rich interlaminae. Where shale interbeds are present in the sandstones, load casting is common Mottling due to bioturbation, is common in the upper fine grained part of the Rough Rock, as are wavy bedding and discontinuous carbonaceous laminae. Unlike the Pule Hill Grit no internal erosion surfaces or intraformational lag conglomerates were observed in the main part of the sandbody, although planar and trough cross bedded units were observed. The underlying shales and siltstones show parallel lamination of siltstone-shale interbeds and bioturbation.

5:10.4. INTERPRETATION

5:10.4.1. MARINE SHALE FACIES

The marine shale facies is represented by two horizons below the Rough Rock, and they are identified by an increase in gamma ray activity and by the presence of a marine fauna (Fig 5:10)). The Lower marine band is the G. cancellatum marine band, and it marks the base of the Gla zone.

5:10.4.2. INTERDISTRIBUTARY FACIES

The interdistributary facies comprises the siltstones and shales underlying the Rough Rock and between the G. cumbriense and G. cancellatum marine bands. The facies is identified on the basis of lithology and log response and is essentially similar to rocks of this facies described in the pre-R. gracile sequence

5:10.4.3. DISTRIBUTARY CHANNEL FACIES

The distributary channel facies comprises the massive and cross bedded parts of the Rough Rock and is identified by the presence of medium grained, trough and ripple cross-laminated sandstones which exhibit a bell shaped gamma ray and S.P. log motif (Fig 5:10)). Fisher (1969) states that bell shaped log motifs are characteristic of channel sands. Walker (1978) has shown that trough cross bedded sandstones which pass upwards into ripple laminated sandstones are common features of channel sandstones. Similar motifs were recorded by Steel (1988).

The upper part of the channel sandstone shows an upward decrease in gamma ray activity and S.P. response. As no samples of this subunit were available for study, precise identification of the environment of deposition is uncertain, however although it has been assigned to the channel facies.

A shale bed, approximately 1 m thick, sits on top of the main stacked channel sequence of the Rough Rock and was probably deposited as an abandoned channel fill. This subunit is

developed in all the wells studied, and probably represents a temporary abandonment of the channel system.

5:10.4.4. OVERBANK AND CREVASSE SPLAY FACIES

The overbank and crevasse facies comprises the upper part of the Rough Rock and consists of sandstone interbedded with siltstones and shales. A seatearth and thin coal are present at the top of the unit. The facies is identified on the basis of the interbedded nature of the sandstones and argillaceous deposits and the presence of the seatearth and coal, indicating a shallow water environment (Elliot, 1974. Fielding, 1984).

5:10.4.5. CONCLUSIONS

The rocks of the Gla zone were deposited on the delta plain in predominantly non marine interdistributary bays and lakes. Three marine incursions introduced marine conditions for short periods of time. The Rough Rock was deposited as a distributary channel and overbank complex.

PLATE 5:1

A

Moss Farm, Pule Hill Grit

Slabbed core sample of distributary channel facies sandstone. Note parallel lamination caused by grain size variations and sealed vertical fracture on right hand side of core. Fractures break cemented grains and are often filled with calcite or kaolinite. No hydrocarbon stains were observed on these surfaces.

B

Moss Farm, Pule Hill Grit

Interlaminated siltstone and mudstone deposited in the overbank environment, showing bioturbation mottling.

Plate 5:1

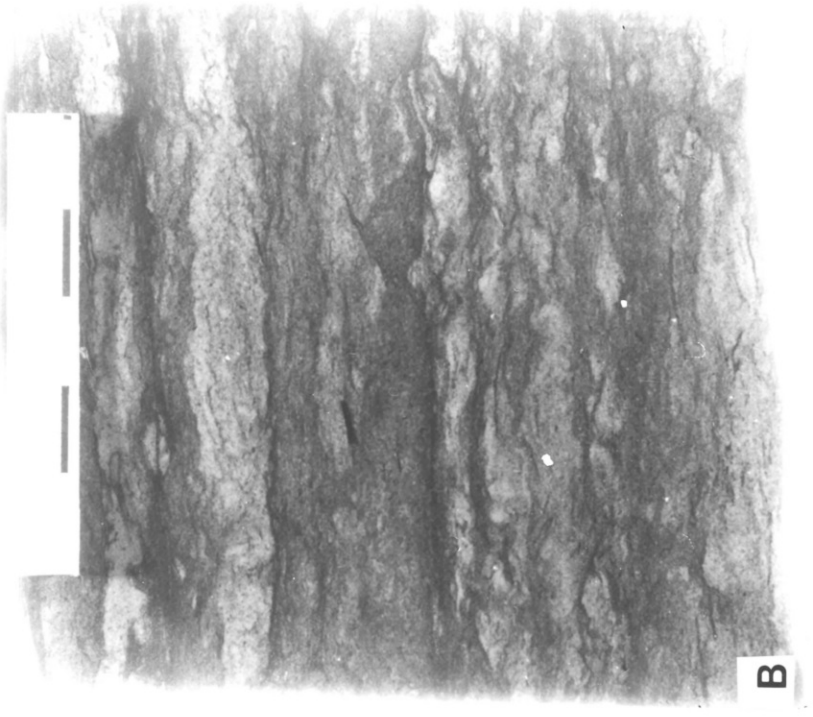




PLATE 5:2

A

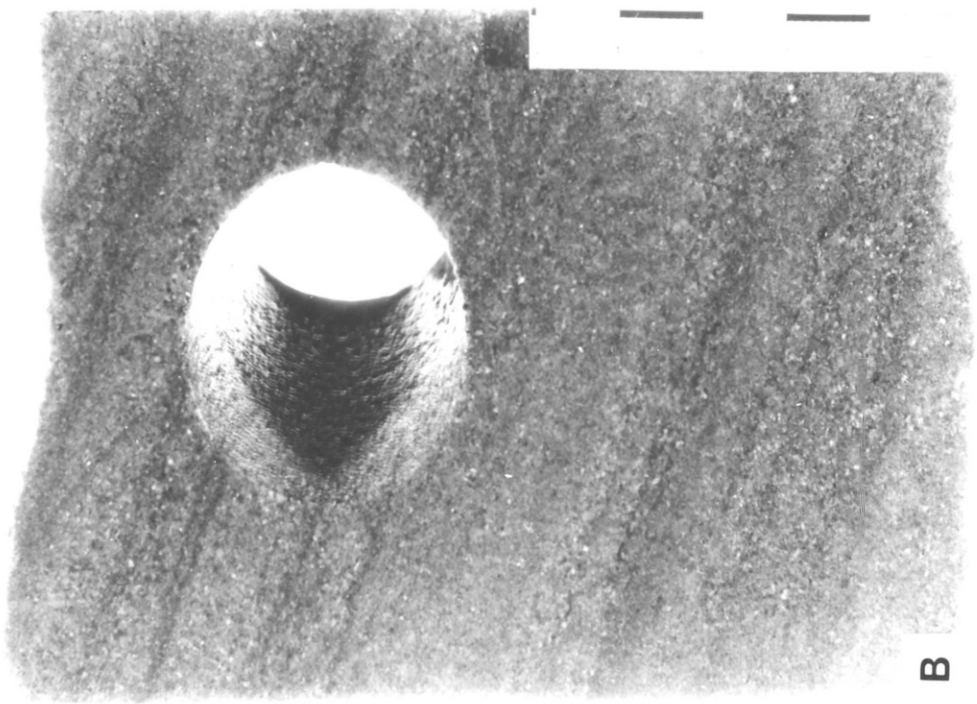
Moss Farm Rough Rock

Overbank facies sandstone showing ripple lamination and abundant argillaceous interlaminae. speckled appearance of the core is caused by the presence of sphaerosiderite nodules.

B

Moss Farm Rough Rock

Well developed planar cross-lamination in distributary channel sandstones. Lamination is picked out by grain size variations. More commonly: the distributary channel facies displayed massive bedding. Large scale trough cross-lamination may be present but undetectable on the scale of the core.


B

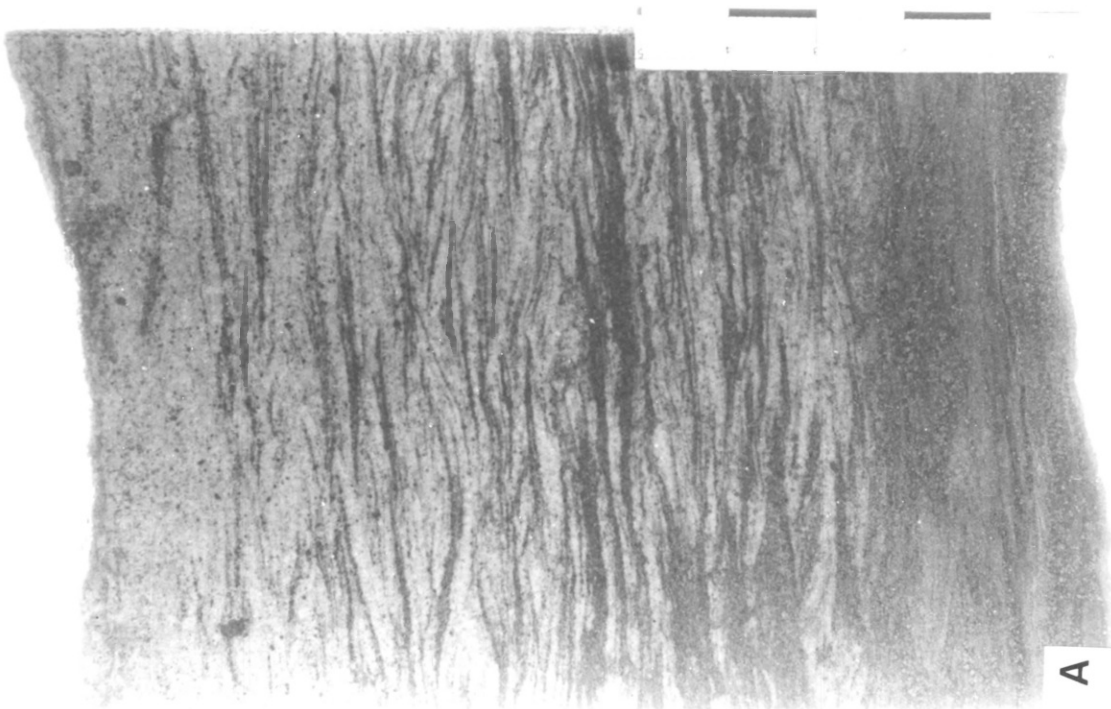

A

PLATE 5:3

A

MOSS FARM PULE HILL GRIT

Interdistributary channel sandstone showing disturbed channel slump deposit with abundant organic matter. Note sand infilled plant stems. Disturbed zone passes upwards into trough cross bedded coarse grained sandstones.

B

MOSS FARM ROUGH ROCK

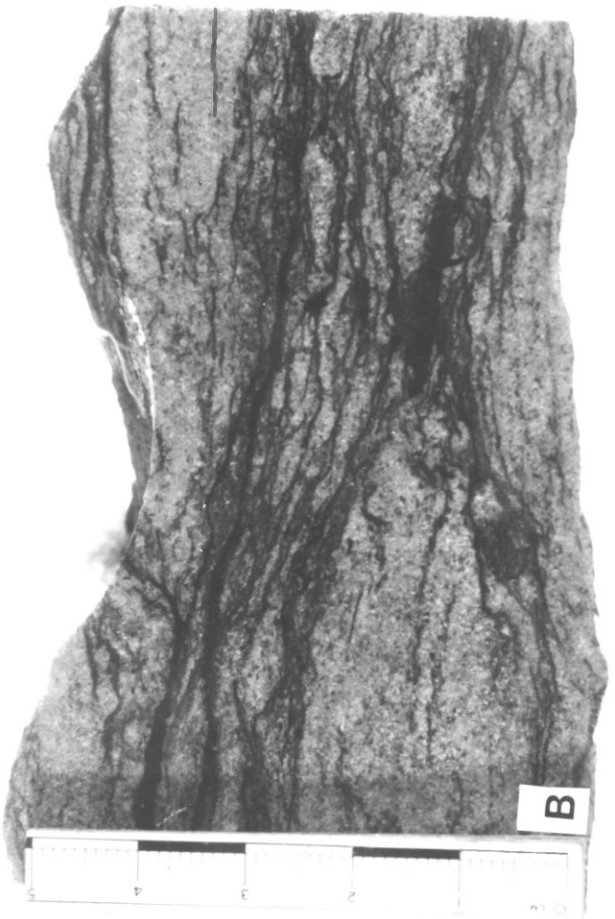
Organic matter rich sandstone from near a distributary channel base. Wavy type lamination is caused by differential compaction.

C

MOSS FARM Kinderscoutian

Interdistributary facies sandy siltstone. Note sandy interbeds and complex nature of disturbed lamination caused by a mixture of load casting, water escape, and microfracturing.

Plate 5:3







CHAPTER SIX

6 PETROGRAPHY: DETRITAL GRAINS AND ASSOCIATED SEDIMENTS

6:1. INTRODUCTION

The detrital mineralogy is one of the principal factors which influence fluid-rock interactions, and is therefore of vital importance in any diagenetic study. It is, however, especially in the case of clay minerals, often difficult to determine whether specific minerals are detrital in origin, or if they have formed authigenically during diagenesis. It is not possible to consider the detrital mineralogy in exclusion, therefore, diagenetic effects will be alluded to from time to time in this section. For a detailed discussion of diagenesis, the reader is referred to Chapters Seven and Eight.

6:2. DEFINITIONS

The terms and concepts used regarding the diagenesis of the sediments studied will crop up from time to time in this Chapter so it is convenient to give some definitions here.

Diagenesis is defined as all changes taking place in the rock after deposition, prior to the onset of metamorphism and equilibrium mineral reactions (Hutcheon, 1982).

Early diagenesis refers to diagenetic reactions controlled by the initial pore fluid chemistry, reflecting the sedimentary environment (the eodiagenetic regime of Choquette and Pray 1970).

Burial diagenesis refers to changes taking place as a result

of burial and increasing temperature and overburden pressure.

Detrital minerals are allochthonous minerals which were introduced into the rock during sedimentary deposition.

Authigenic minerals are defined as those minerals which have formed within the sediment, or at the sediment surface after deposition.

Neoformed minerals are authigenic minerals in which the crystal structure is not inherited from a detrital precursor, and have formed by direct precipitation from the pore fluid.

6:3. DETRITAL MINERALS

6:3.1. QUARTZ

Detrital quartz grains vary from 36 to 93% of the rock, although the mean content is around 70%. Quartz percentages as low as 45% are recorded locally where early carbonate cements are present. The quartz consists predominantly of unstrained monocrystalline varieties, though polycrystalline and strained grains can locally constitute up to 30% of total quartz content. Boehm lamellae, indicating intense strain, are common in the detrital quartz grains. Chert grains are rare.

The quartz grains showed varying degrees of overgrowth in all the samples studied, with up to 23% of the total quartz content being authigenic. Estimation of the relative amounts of authigenic quartz was hindered by the lack of complete "dust rims" in most samples (Plate 6:1A). The point count data, therefore, overestimates the amounts of detrital quartz present by up to 20% of the total mode.

Inclusions in the quartz grains consist mainly of small needles, probably of rutile, and some isolated crystals of zircon, and siderite.

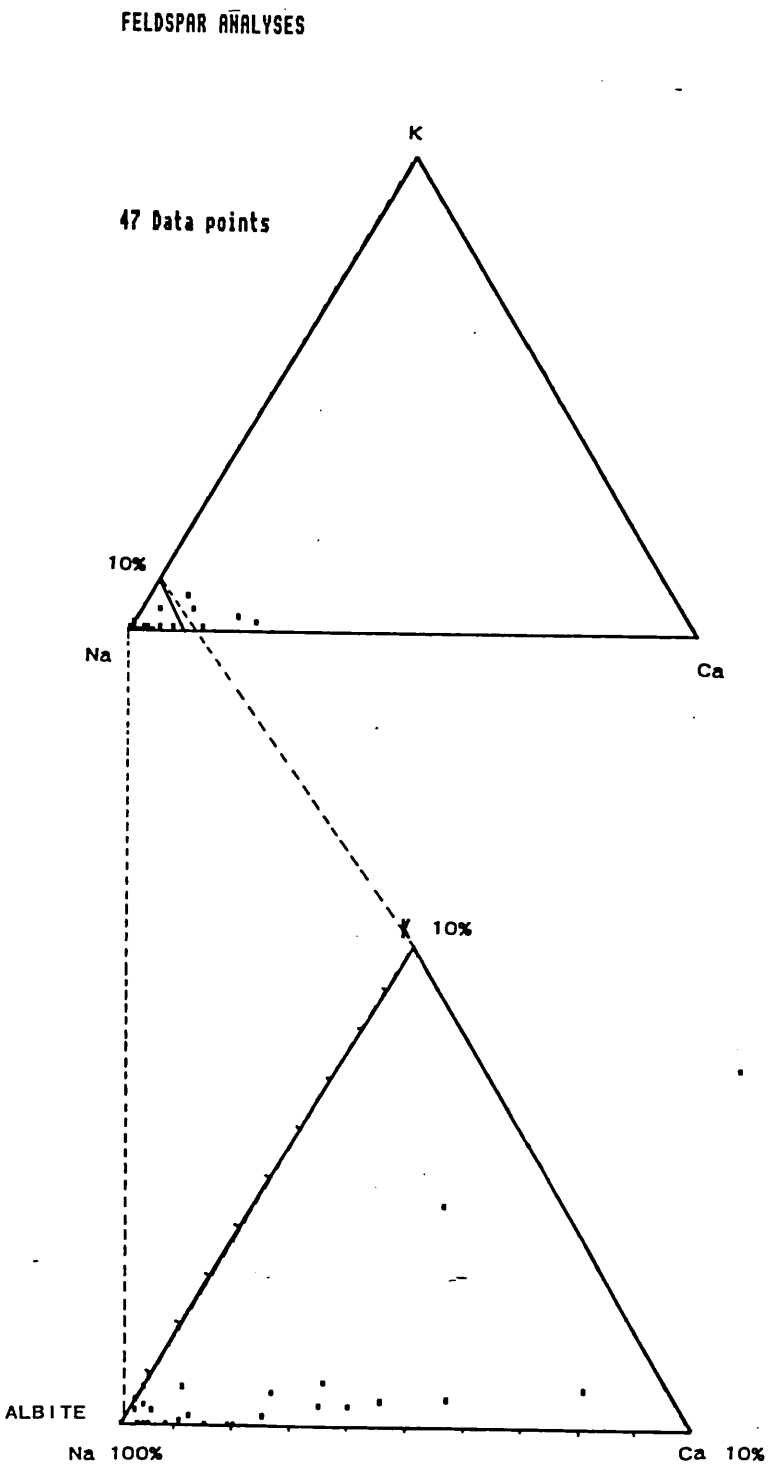
6:3.2. FELDSPAR

Plagioclase and alkali feldspars constitute from 0.5 to 17 % of the rock volume with mean values of around 8%. Untwinned feldspars constitute 90% of the total feldspar content and microprobe analyses show these to be almost pure albite (Fig 6:1, Plate 6:3 A).

6:3.3. Untwinned "Alkali" Feldspars

Untwinned feldspars accounted for 90% of the feldspar in the rocks studied. Hawkins (1972) noted that the most common type in Namurian and Westphalian sandstones of the Bothamsall field was untwinned alkali feldspar, but other workers, notably Huggett (1982) and Gilligan (1919), found that microcline and microcline microperthite were the dominant feldspar types in the Namurian and Westphalian rocks of the East Midlands and Central Pennines, However, it has recently been shown by Walker (1984) that perthite type textures can form by the authigenic albitisation of alkali feldspar. The lack of microcline microperthite in the rocks of the Trumfleet Field, therefore, either reflects localised variations in provenance or is a consequence of complete albitisation of alkali feldspar grains. Untwinned feldspars show varying degrees of sericitisation, and in some cases the orientation of the sericitic micaceous clays provide the only indication that a particular grain was originally a feldspar (Plates 6:2A and &:2B). Hawkins (1972) and Huggett (1982) considered this alteration of feldspar to be diagenetic but no textural evidence of a diagenetic origin of sericite found in this study. In addition, the presence of highly sericitised and fresh feldspars together in the same rock suggests that the sericitisation was pre-depositional. Diagenetic alteration of untwinned feldspar is important and will be considered in detail in Chapter 8.

FIGURE 6:1


FELDSPAR ANALYSES
47 Data points
K
10%
Na
Ca
10%
ALBITE
Na 100%
Ca 10%

6:3.4. Twinned Plagioclase Feldspars

Twinned plagioclase grains mainly show albite type twinning and have oligoclase and bytownite compositions (probe data). They account for approximately 10% of the feldspar population.

Plagioclase grains, especially microcline, appear less altered than the untwinned alkali feldspars, although they are usually more highly fractured (Plate 6:3B). The untwinned grains show less fracturing and the fractures have been healed by authigenic feldspar overgrowths.

6:3.5. MICA

Mica grains range from trace amounts to a maximum of 14% of the rock, but generally comprise less than 2% of the volume of the sediment. They consist dominantly of muscovite with lesser amounts of biotite. A very rare blue pleochroic mica, possibly Mn rich muscovite (Deer, Howie and Zussman, 1966.) was identified from two sections from the Beacon Hill Flags.

Muscovite appears relatively fresh and is often concentrated into discrete horizons rich in organic matter and detrital clay minerals (Plate 6:4A). Individual grains are often bent and fractured but some long grains remain straight and there is evidence of pressure solution along their length (Plate 6:4B). Huggett (1982) demonstrated that similar grains from the Westphalian of the East Midlands are K^+ deficient and are in the process of altering to illite, even although, optically, the grains appear fresh.

All biotite grains show varying stages of alteration to chlorite and kaolinite minerals as well as siderite. In some

cases all the biotite has altered to chlorite and the grains show typical "Berlin Blue" birefringence (Plate 6:5B). It is not clear whether this chloritisation in pre or post depositional.

6:3.6. ROCK FRAGMENTS

As stated in Chapter 3, rock fragments were not considered as such in the modal analyses and the constituent minerals of the rock fragments were counted. However a number of types of rockfragment were recognised.

Tuffaceous fragments, which appear as green chlorite rich masses are present in trace (less than 1%) amounts. Sedimentary rock fragments consist mainly of claystone fragments, which often contain authigenic microcrystalline siderite. These grains are often squeezed and bent between framework grains, and can constitute up to 3% of the volume of the rock in the overbank and crevasse splay sandstones. They also occur as rip up clasts in the channel sandstones.

Metamorphic rock fragments consist of mica schist comprising silt sized quartz and muscovite grains as well as quartz schists.

Igneous granitic rock fragments commonly recorded by other workers were not as abundant in this study as had been reported for rocks of similar ages, but the bulk of the rock appears to have been derived from a granitic/metamorphic source. It is probable that granitic clasts were disaggregated in transport.

6:3.7. ORGANIC MATTER

Organic matter is common in the sandstones where it can constitute up 10% of the rock, although it usually varies from

traces to 3%. It varies in colour from reddish brown and partially transparent to black and opaque, and tends to be concentrated into discrete bands with mica (Plates 6:6A 6:6B, 5:3B). The presence of organic matter in the channel sandstones as well as the overbank sandstones has significant implications for the early diagenesis of the rocks. If oxidising acidic meteorically derived fluids were present throughout early diagenesis it would be expected that the organic matter would be oxidised and would not be preserved in the sandstones (Chapters Seven and Eight).

Vitrinite reflectance measurements carried out on organic matter from the interdistributary shales vary from 0.73 to 1.02% Ro indicating that the rocks were subjected to a maximum burial temperature of approximately 115^{o} C. (Waples, 1980).

6:4. DETRITAL CLAY MINERALS

Whole rock and clay fraction XRD showed that kaolinite, illite and chlorite were present in varying abundances in all the samples studied, and in almost all cases, minor amounts of mixed layer clay minerals were present. Kaolinite is dominant. Detrital clay minerals are difficult to identify if they have undergone diagenetic modification and have the morphology of true authigenic phases. It is often possible, however, to use textural criteria to distinguish between detrital and authigenic phases (Shaw: In Hobson, 1980), but where small amounts only are concerned these criteria may be of little use. Highly birefringent clay minerals were observed in thin section between grain contacts (Plates 6:7A and 6:7B), but location alone is not conclusive proof of detrital origin, since authigenic (mixed

layer?) clays were observed during SEM examination coating pressure solution surfaces. Illite was observed within intraformational mudstone fragments and in clay and mica rich seams, so a detrital origin for some at least some illite seems likely.

Detrital kaolinite was probably deposited in the sandstones, since studies of modern deltas (Elliot, 1978) have shown that kaolinite is a commonly deposited in deltaic sands of tropical regions. It is probable that detrital feldspars had suffered some hydrothermal kaolinitisation prior to deposition, as this is a common feature of acid igneous rocks. However, all the kaolinite observed was in the form of books and vermiculae showing fresh crystal faces, therefore, it was postulated that all kaolinite is authigenic in origin.

6:5. ACCESSORY MINERALS

The most common heavy minerals observed were zircon, sphene and monazite, with lesser amounts of rutile, tourmaline, and apatite. Gilligan (1919) noted that the most common heavy mineral observed in heavy mineral separates of Millstone Grit sandstones was garnet, and he found no apatite in these sandstones, except as acicular grains in quartz from pebbles from the Millstone Grits. The apatite observed in this study exists as large clear angular grains of the same size as the framework grains.

The most common detrital opaque mineral observed is magnetite, Pyrite is common but it is of authigenic origin.

6:6. ASSOCIATED SEDIMENTS

6:6.1. SHALES

Whole rock XRD showed that the shales above and below the sandstones contain a mineral assemblage similar to that of the sandstones. XRD analyses of orientated clay fractions give stronger illite and chlorite peaks than those of the sandstones. Only traces of expandable clay minerals were recorded.

The minerals in the shales identified from XRD are: quartz, feldspar (mainly plagioclase), mica/illite, mixed layer illite smectite (possibly chlorite/smectite), kaolinite, chlorite, siderite, calcite, pyrite. No attempt was made to differentiate between detrital and authigenic phases in the shales.

It is likely that similar mudrocks were the source of the fluids migrating through the sandstones during burial diagenesis. The fact that only minor amounts of smectite or illite/smectite were detected indicates that either little smectite was deposited in the mudrocks, or that the expandble fractions have been altered to illite during burial diagenesis. The latter explanation is more likely since it is probable that smectite and illite/smectite was present at deposition, as these minerals are commonly derived during weathering. In addition the vitrinite reflectance data indicates that sufficient temperatures have been reached in the sediments to allow alteration of mixed layer clay minerals to illite to take place (Boles and Franks, 1979.).

6:6.2. SEATEARTHS

Two samples of seatearth facies were studied by XRD. A sample from the top of the Rough Rock, from 917m depth, within a 0.5m thick claystone from the Moss Farm well, was found to consist entirely of kaolinite and quartz in roughly equal proportions. Traces of carbonaceous material were present in rootlets within the sediment. This restricted mineralogy reflects the results of subaerial leaching as a consequence of intense tropical weathering in an oxidising environment. Oxidising, meteoricaly derived fluids leach ferromagnesian minerals and alter feldspars to kaolinite, since the pH is lowered by the presence of humic acids derived from the biodegradation of organic matter at the surface. Ultimately this process leads to the development of bauxitic soils.

Analysis of an iron stained seatearth from the top of the Beacon Hill Flags in Trumfleet 2, is shown in Table 6:1.

TABLE 6:1 Modal Analysis of Fe Stained Seatearth

MINERAL	%
QUARTZ	6.0
FELDSPAR	2.0
MICA	1.6
KAOLINITE	12.4
ILLITE	0.4
SIDERITE	79.6
Total	100.0

(SOURCE: Point count analysis and XRD)

Over 90% of the volume of the rock is authigenic (Plate 6:8A and B), the siderite and kaolinite having formed during early diagenesis. The siderite formed via subaqueous early diagenesis

in a reducing environment. A more detailed discussion of the role of siderite in diagenesis is given in Chapter 8.

It would seem on the basis of these limited data that seatearth formation occurred in differing environments; in a subaqueous swamp forming siderite rich sediments. Where the availibilty of iron is limited, kaolinite rich sediments form. It is probable that this latter type form subaerially due to the action of meteoric waters free of iron.

6:7. CLASSIFICATION

The purpose of sandstone classification in this study is simply to compare the compositions of the sandstones with respect to each other, and not to provide direct comparisons with rocks studied by other workers. It was felt that assigning grains to various rock type classes would diminish the amount of compositional information obtained, and that it would be more helpful to have an understanding of the mineralogical composition of the rocks, since this has a more direct effect on diagenesis and wireline log response.

Traditional compositional sandstone classifications are plotted on a ternary diagram with quartz, feldspar, and lithic grains as the end members. Reviews of the various schemes of sandstone classification can be found in Hawkins (1972), and Pettijohn et al (1972). The advantages and disadvantages of the various classifications are explained in these publications.

As rock fragments were not counted in this study a classification which does not use lithic grains as an end member was sought, and a classification similar to that of Krynine's (1948) was used (Fig 6:2). The Krynine scheme plots quartz and chert, mica and chlorite, and feldspar together with kaolinite as the end members of a ternary diagram. This classification was amended by the addition of illite in the mica/chlorite corner of the diagram (Fig 6:5). The plotting of feldspar together with kaolin and mica, illite and chlorite together partially

negates the effects of diagenetic alteration of feldspar to kaolin, and muscovite and biotite to illite and chlorite respectively To assume, however that all the kaolin, illite and chlorite in any particular sample are simply the products of alteration of feldspar and mica is a gross oversimplification.

Figures 6:4 to 6:6 indicate that most of the samples studied show little variation in composition and plot as orthoquartzites and arkoses. The Beacon Hill Flags show greater variation in composition than the Rough Rock and Pule Hill Grits, two samples plotting as low grade greywackes on account of their high clay content. There is no simple relationship between sedimentary environment and composition but it was generally observed that the channel sandstones contain more kaolinite than the overbank or crevasse sandstones, therefore channel sands tend to plot as arkoses whilst the overbank sandstones plot towards the greywacke end on the diagram. .The use of the term greywacke in the classification is unfortunate since it has textural implications. A better term would be "clayey sandstone".

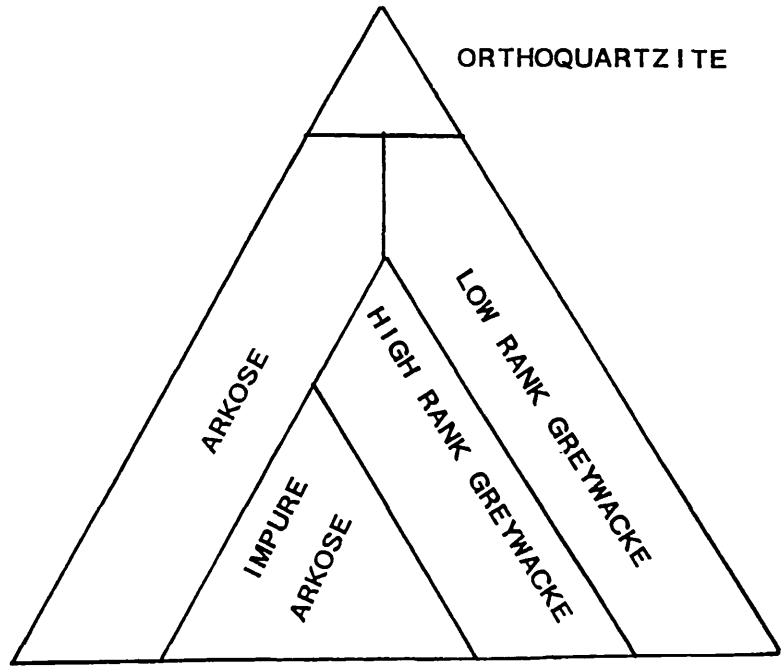
6:8. PROVENANCE

It has long been proposed that the source of the sediment for the Namurian delta in the Central Pennines was a metamorphic and granitic landmass situated to the north (Sorby, 1859; Gilligan, 1919.). The presence of alkali feldspar and apatite from the Trumfleet area suggest that a granitic source the provided sediment for the sandstones. The large apatite grains probably came from a pegmatitic source, associated with the granite, or from the Lower Carboniferous limestones. The lack of microcline

KRYNINE 1948

QUARTZ/CHERT

FIGURE 6:2



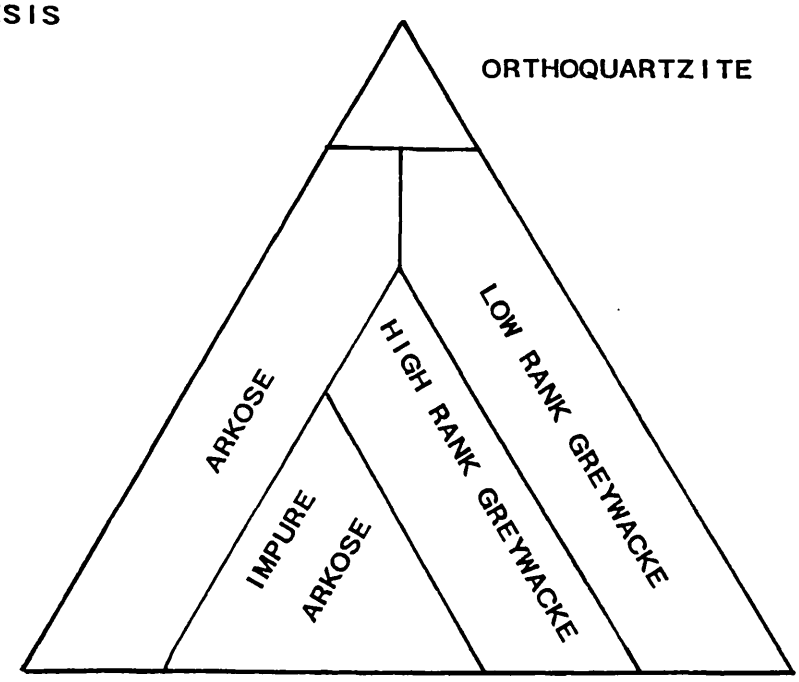
FELDSPAR/KAOLIN

MICA/CHLORITE

QUARTZ/CHERT

FIGURE 6:3

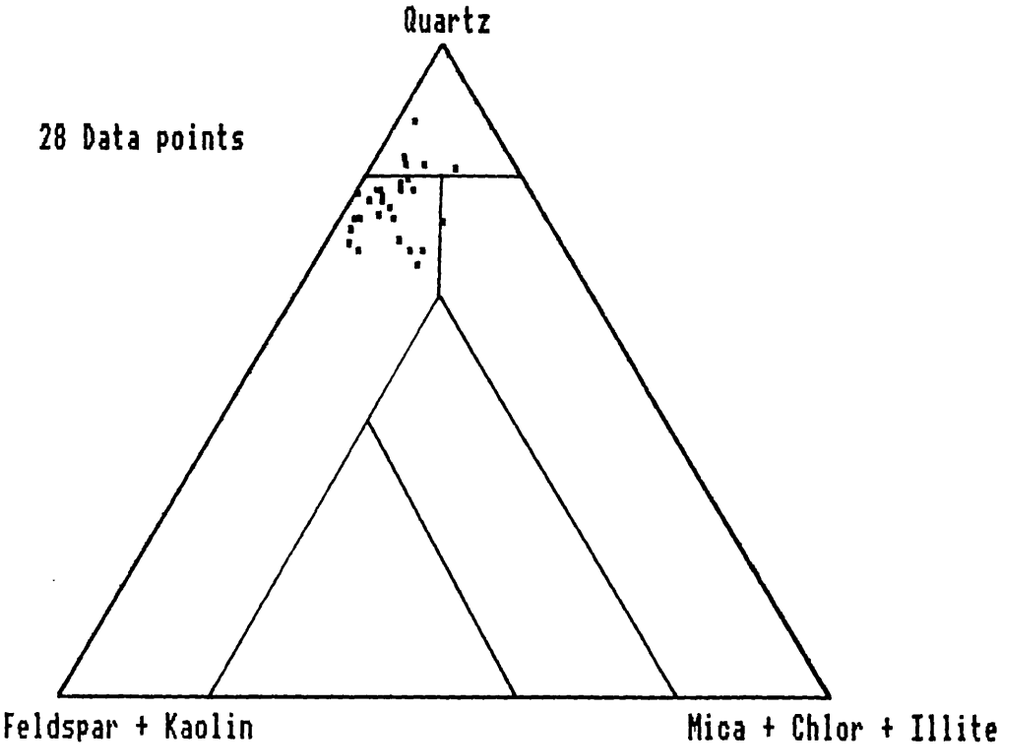
THIS THESIS



FELDSPAR/KAOLIN

MICA/CHLORITE/ILLITE

ROUGH ROCK FIGURE 6:4



BEACON HILL FLAGS FIGURE 6:5

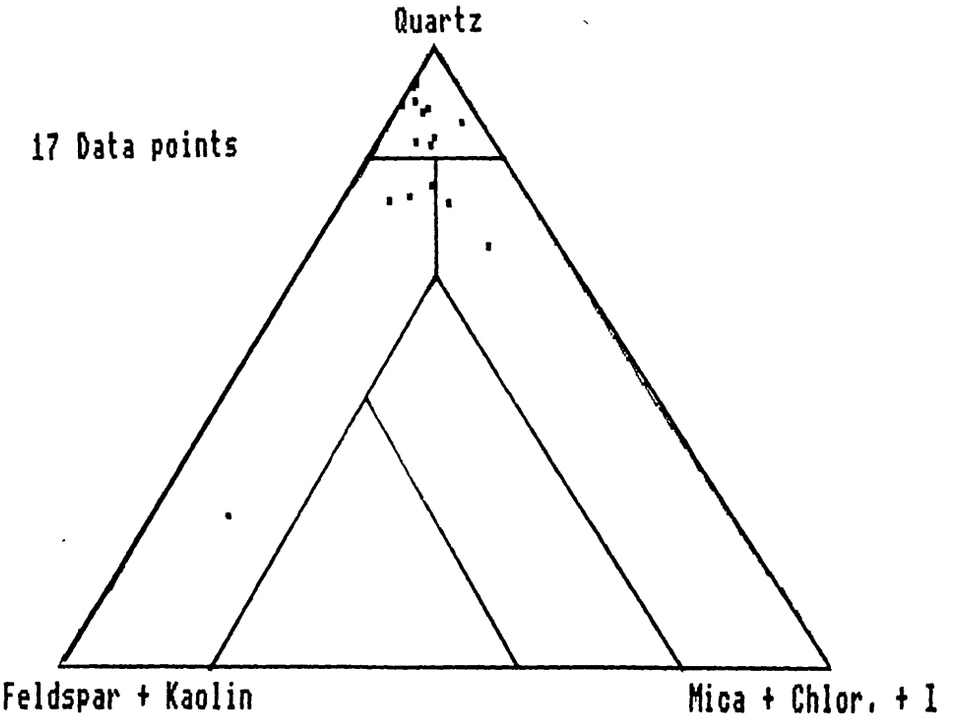


FIGURE 6:6

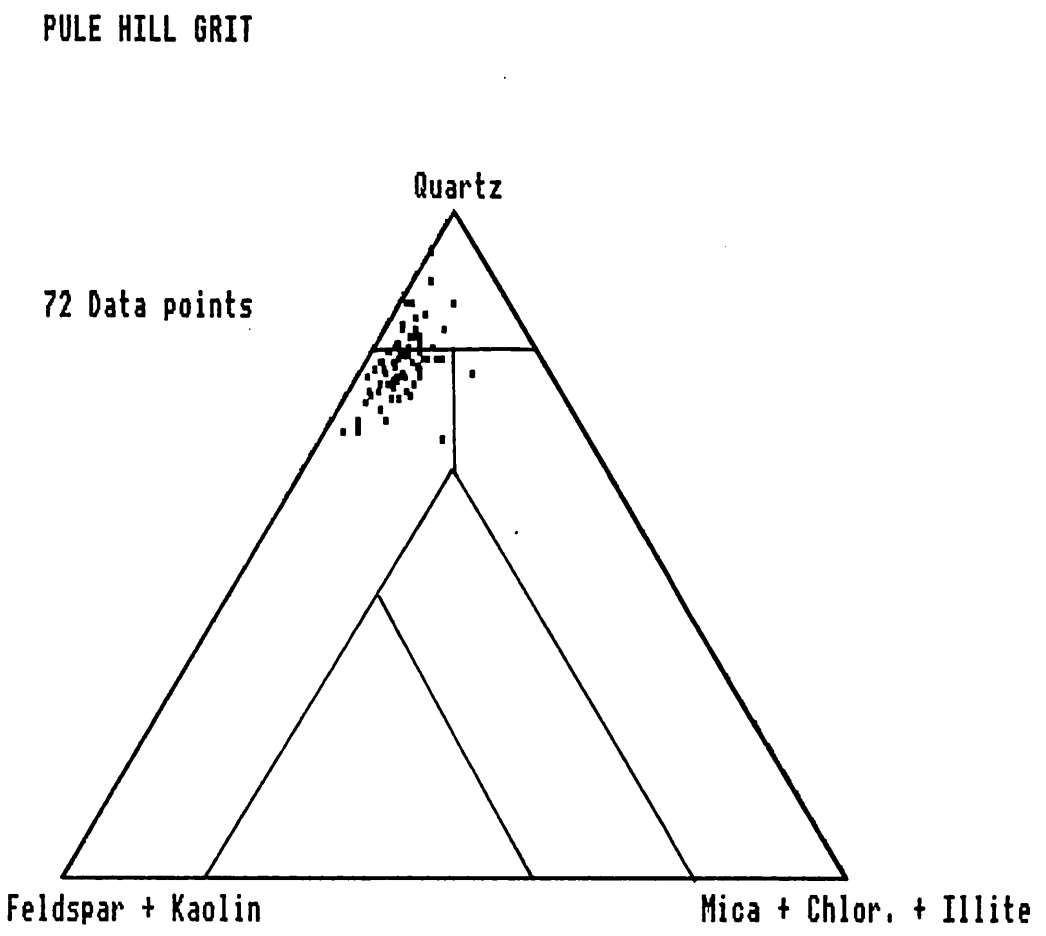

PULE HILL GRIT
Quartz
72 Data points
Feldspar + Kaolin
Mica + Chlor. + Illite

PLATE 6:1

A

Trumfleet 1 Pule Hill Grit 4386 ft PPL

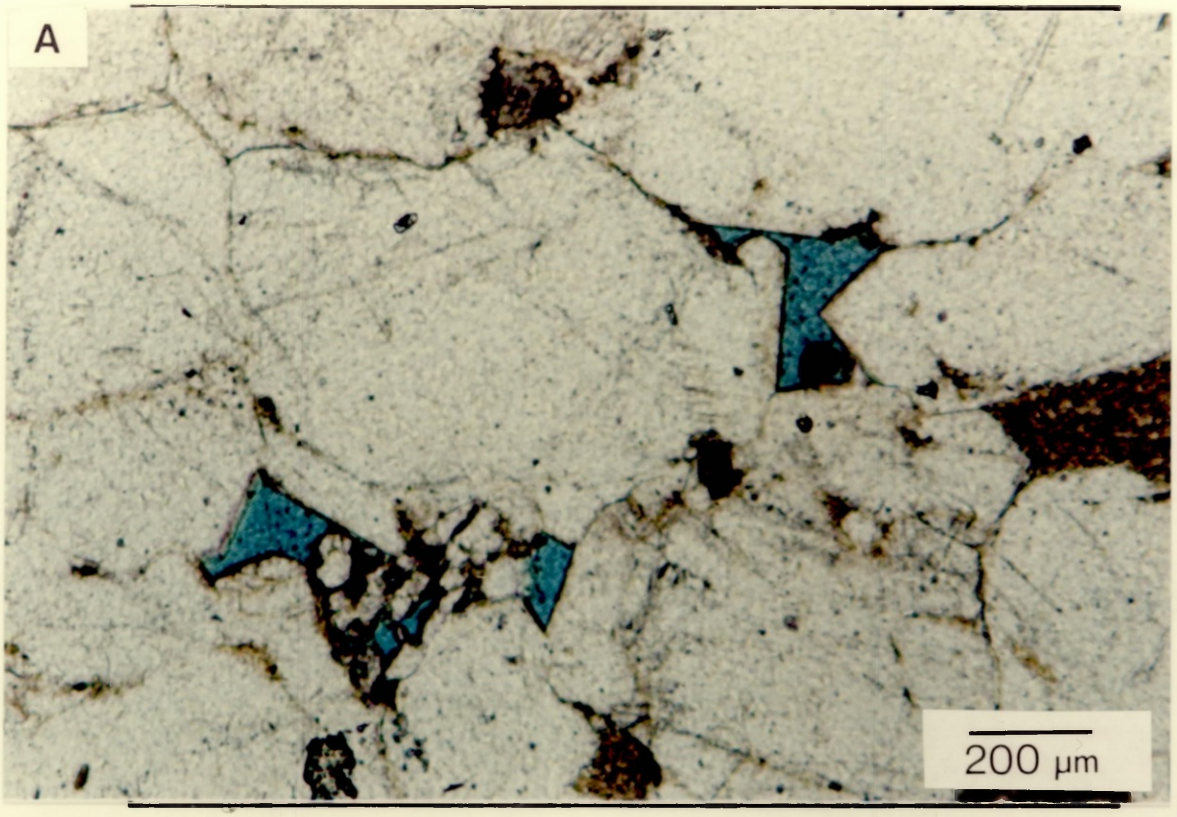
Detrital quartz grains. Note poorly developed dust rims and well developed overgrowths.

B

Askern Pule Hill Grit 3324 ft XPL

Poly and monocrystalline varieties of quartz.

Plate 6:1



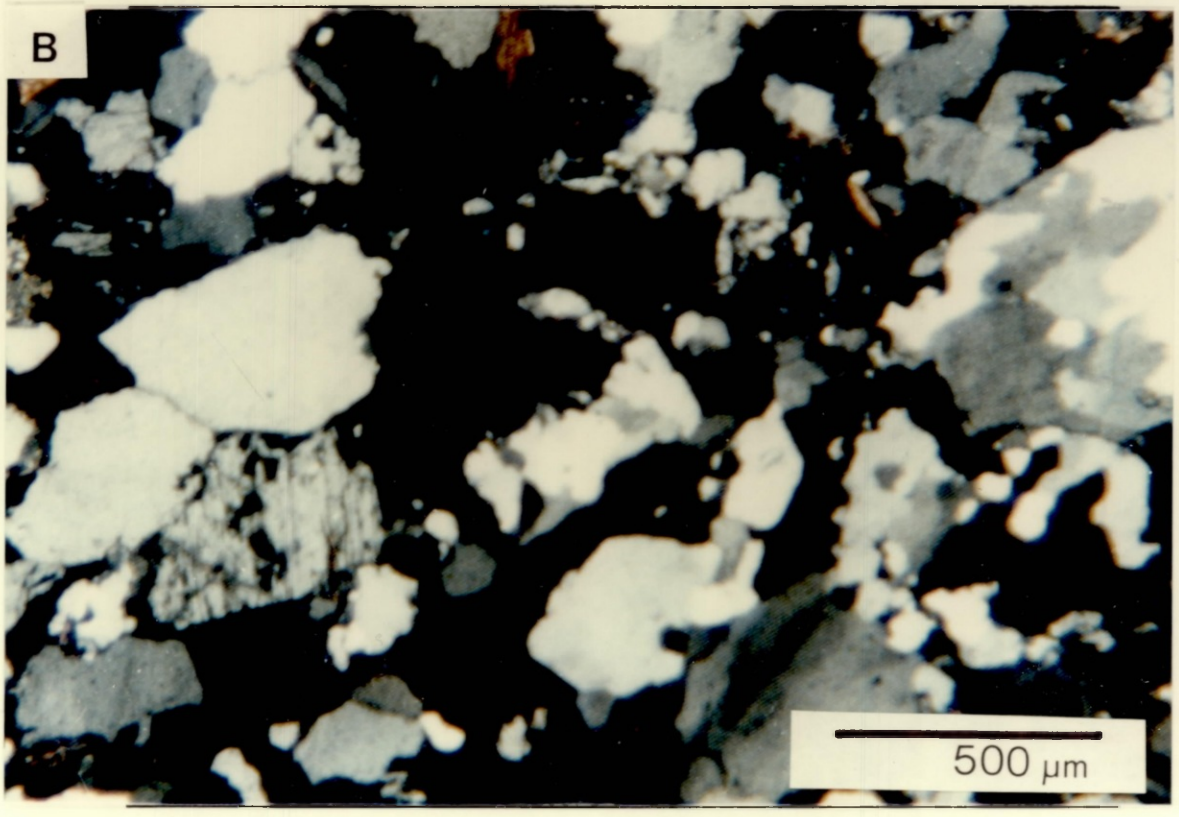


PLATE 6:2

A

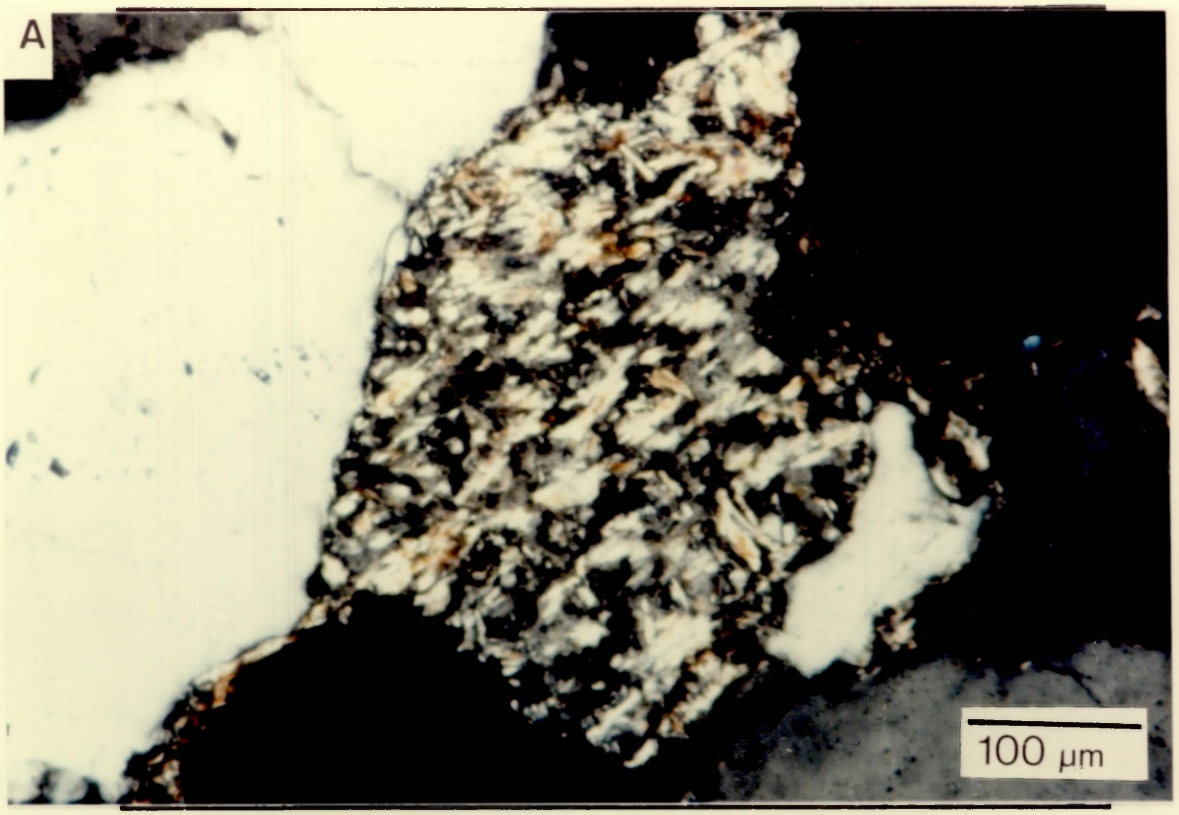
Trumfleet 1, Pule Hill Grit 3486 ft. XPL

Altered feldspar grain showing alteration to highly birefringent sericite and low birefringence kaolinite. The observation that highly sericitised grains appear in the same rocks as fresh grains of the same composition, suggests that sericitisation is predepositional.

B

Trumfleet 2, Pule Hill Grit 3396 ft. XPL

Partially sericitised feldspar grain.


A
100 µm

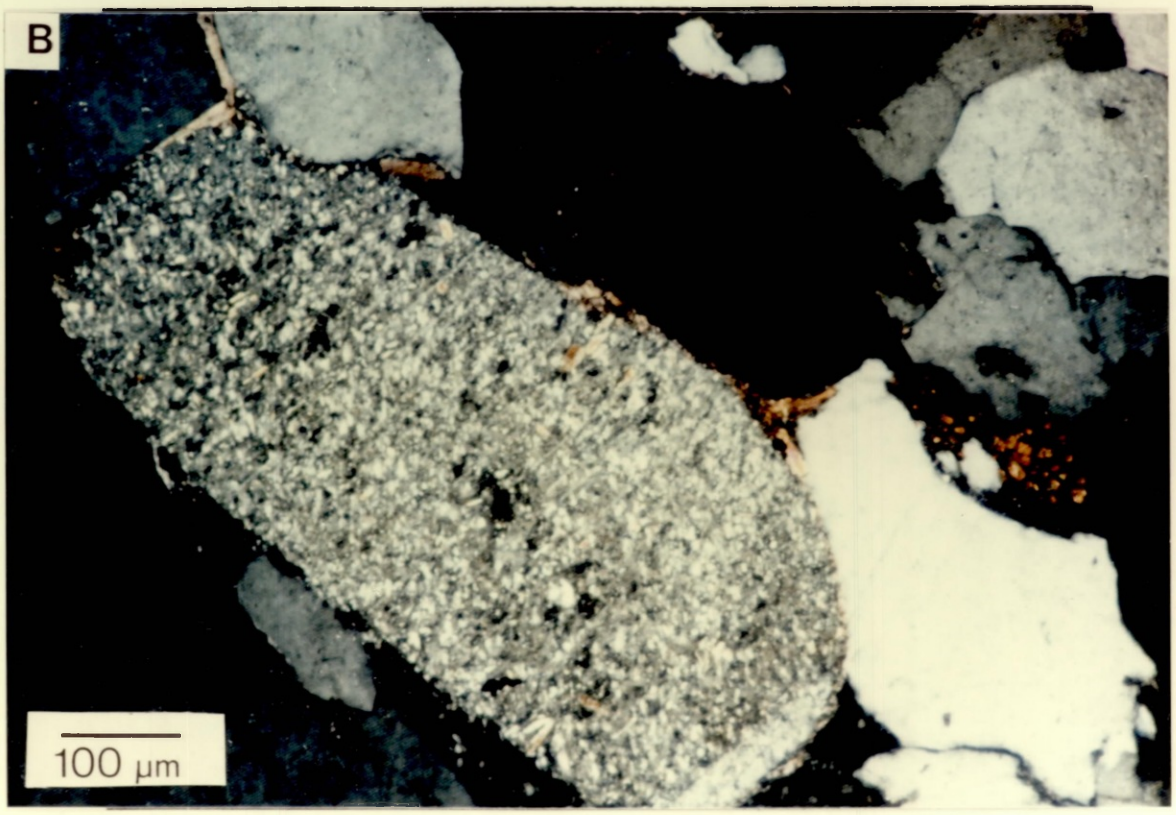

B
100 µm

PLATE 6:3

A

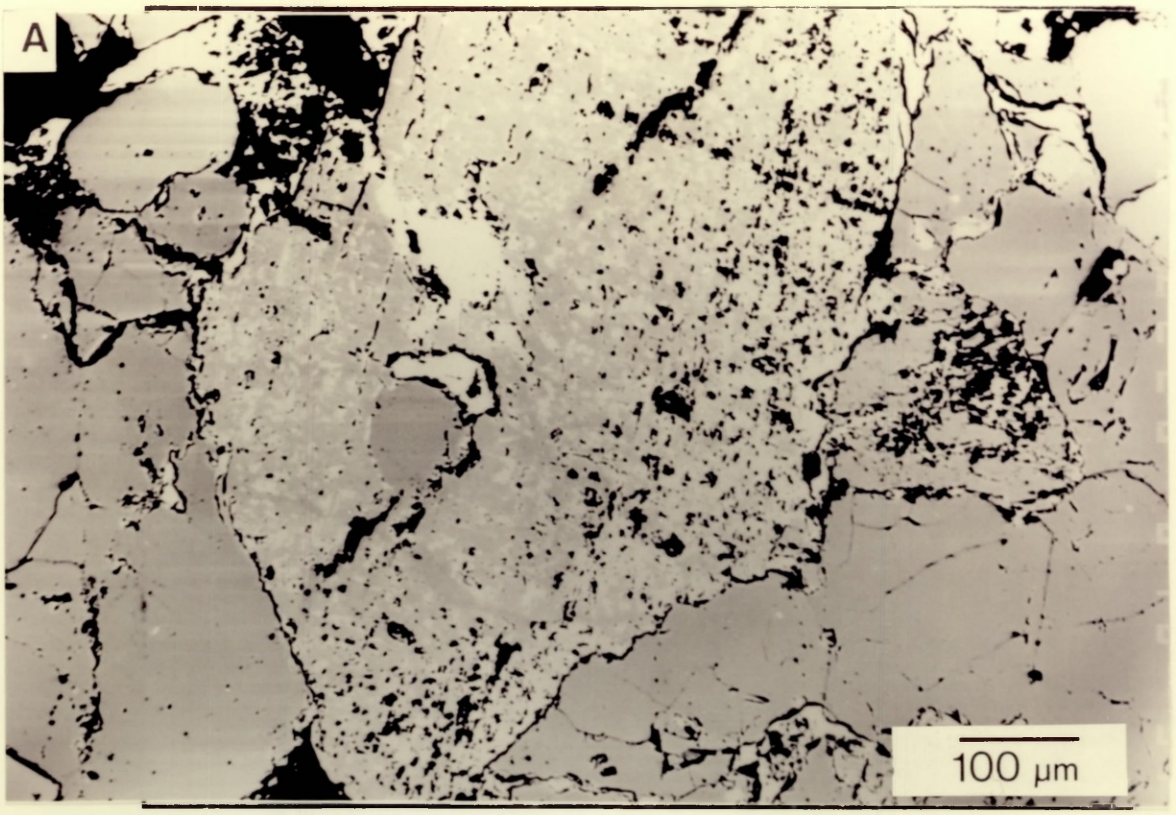
Moss Farm, Pule Hill Grit 1017 m. BSE

Large partially albitised feldspar grain. Lighter patches contain K^+ background (same shade as surrounding quartz) is albite.

B

Askern Pule, Hill Grit 3348 ft. XPL

Fractured plagioclase grain. Note fresh appearance despite fractures. The presence of fractured plagioclase grains in the same rock as highly dissolved yet unfractured alkali feldspar grains implies that dissolution occurred after compaction


A
100 µm

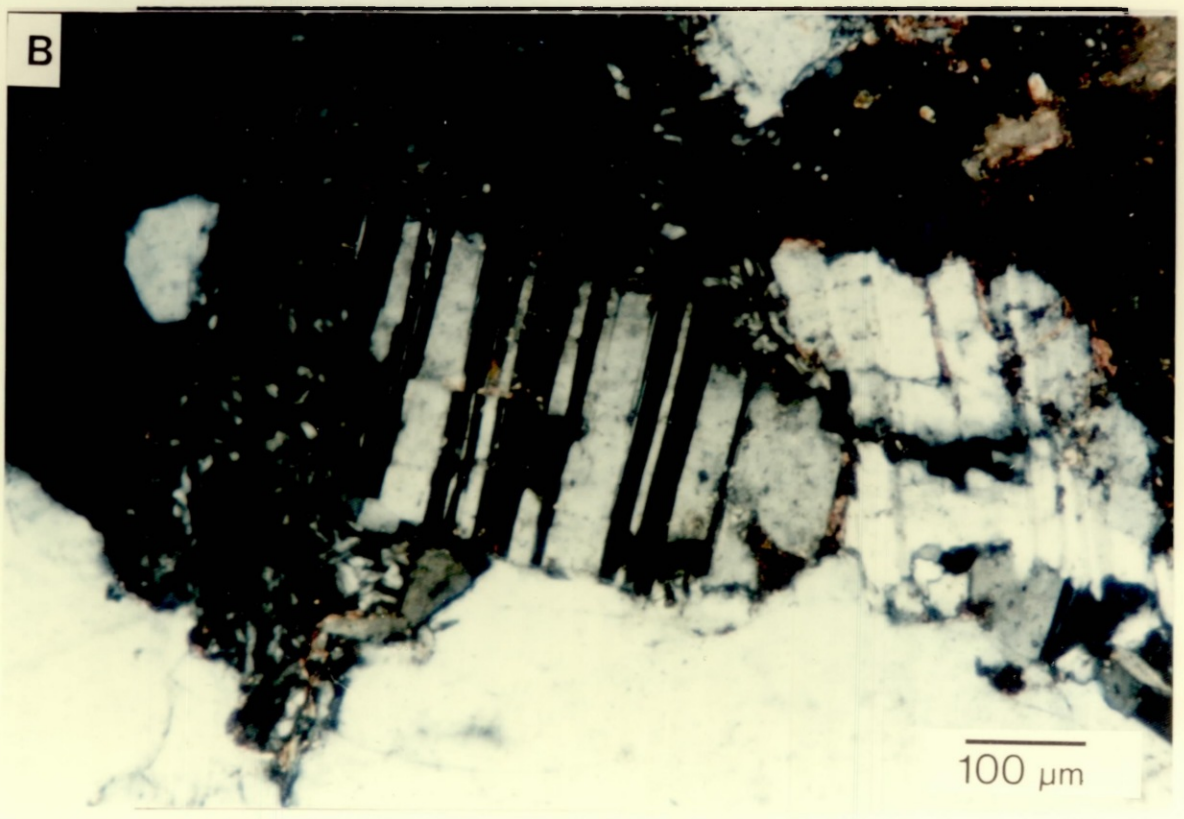

B
100 µm

PLATE 6:4

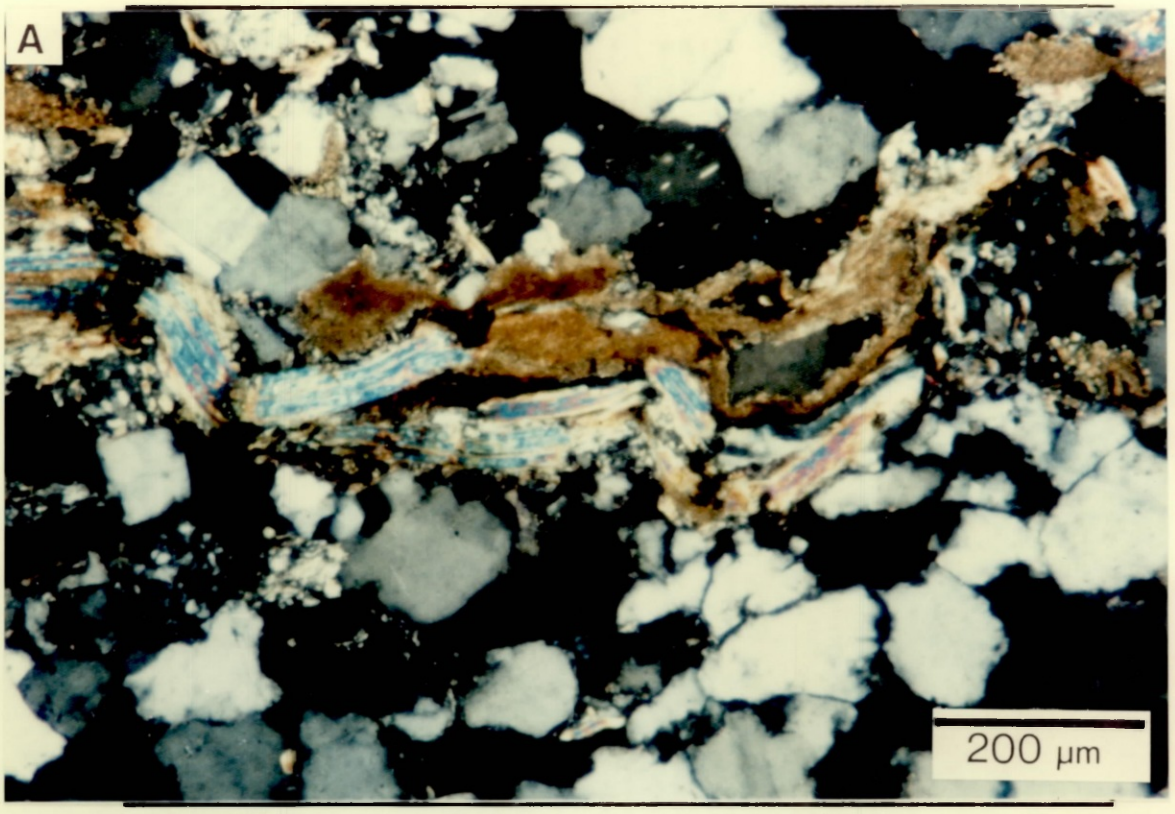
A

Moss Farm, Rough Rock 924.57m. XPL

Mica/argillaceous lamina in sandstone. Note bending and splitting of mica between framework grains as a response to stress.

B Moss Farm, Pule Hill Grit 1017m. XPL

Large muscovite grain which has survived compaction fresh and intact, by responding to stress by pressure dissolution rather than fracturing. Note straight contacts between quartz grains; it appears that up to 30% of the quartz has been removed. Section angled at 45% degrees to bedding to show maximum birefringence.


A
200 µm

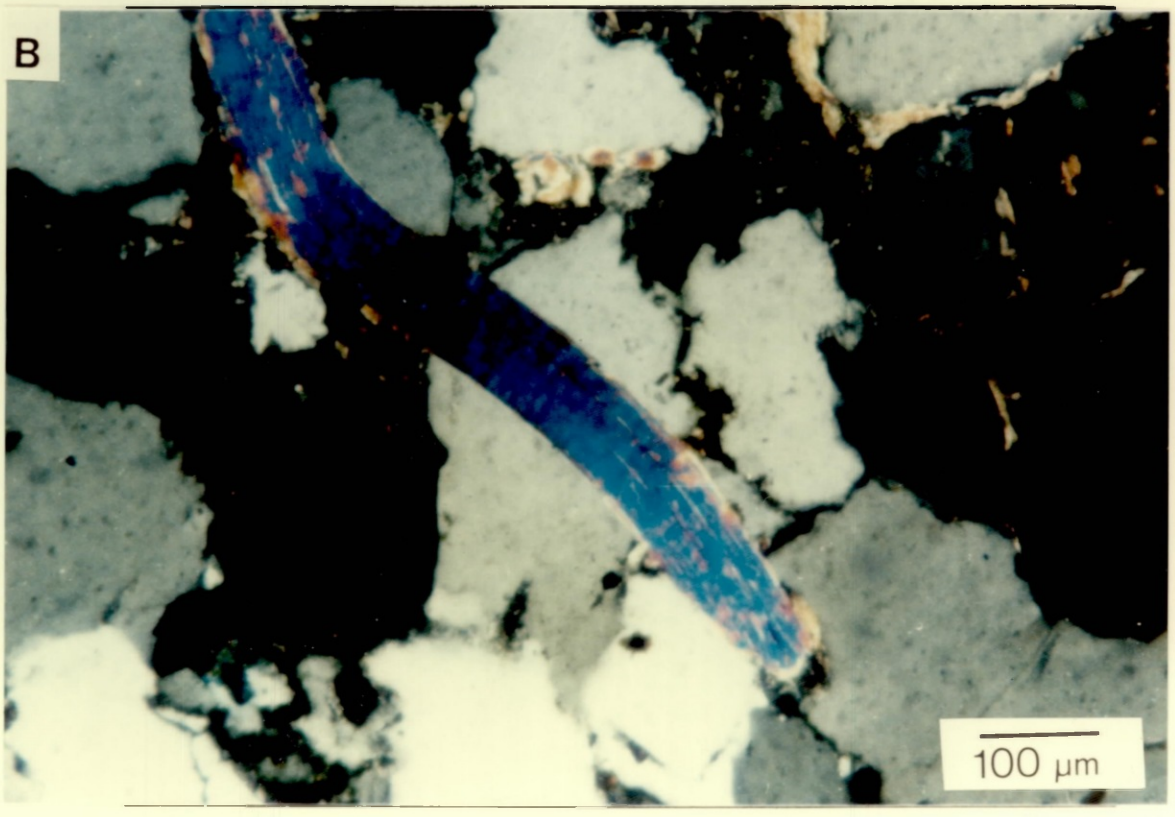

B
100 µm

PLATE 6:5

A

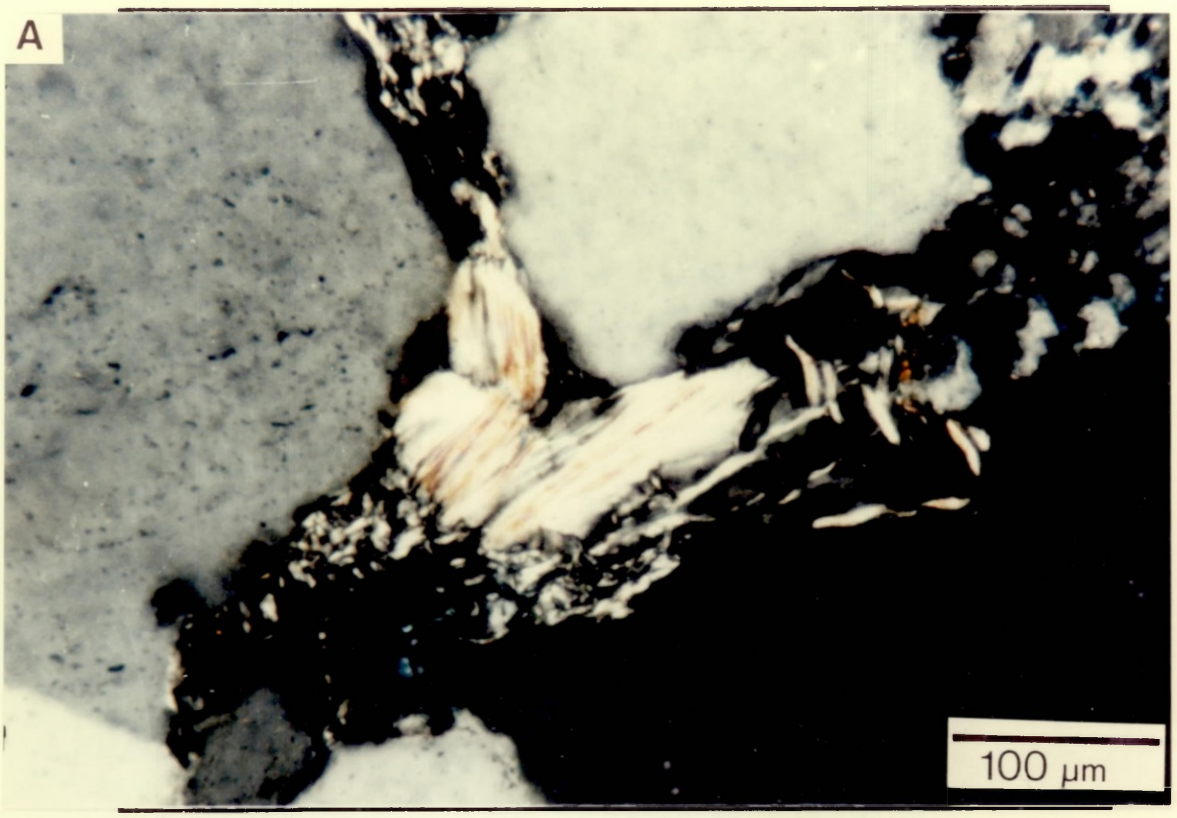
Trumfleet 1, Rough Rock 3164 ft. XPL

Muscovite grain altering to illite and kaolinite.

B

Trumfleet 2 Pule Hill Grit 3405 ft XPL

Detrital (?) chlorite grain. It is not clear whether such grains are detrital or have formed by the diagenetic alteration of biotite. Note anomalous blue interference colours.


A
100 µm

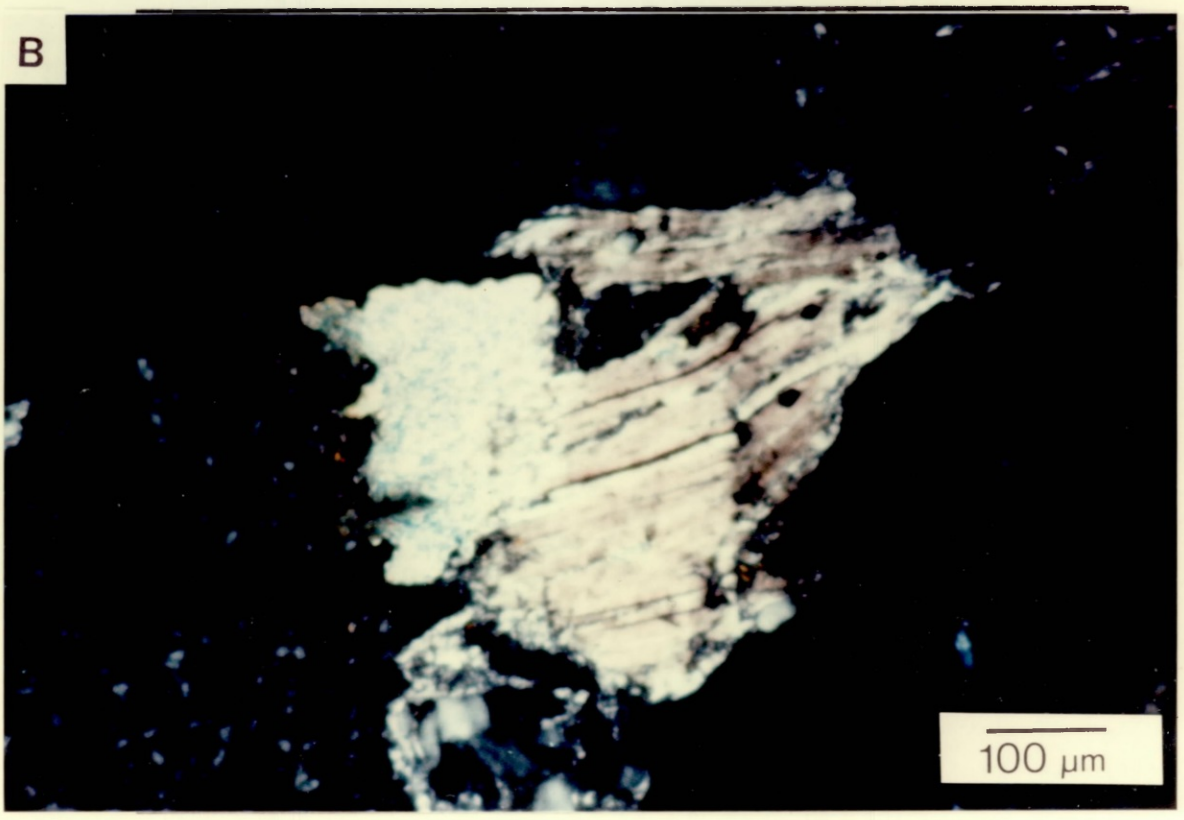

B
100 µm

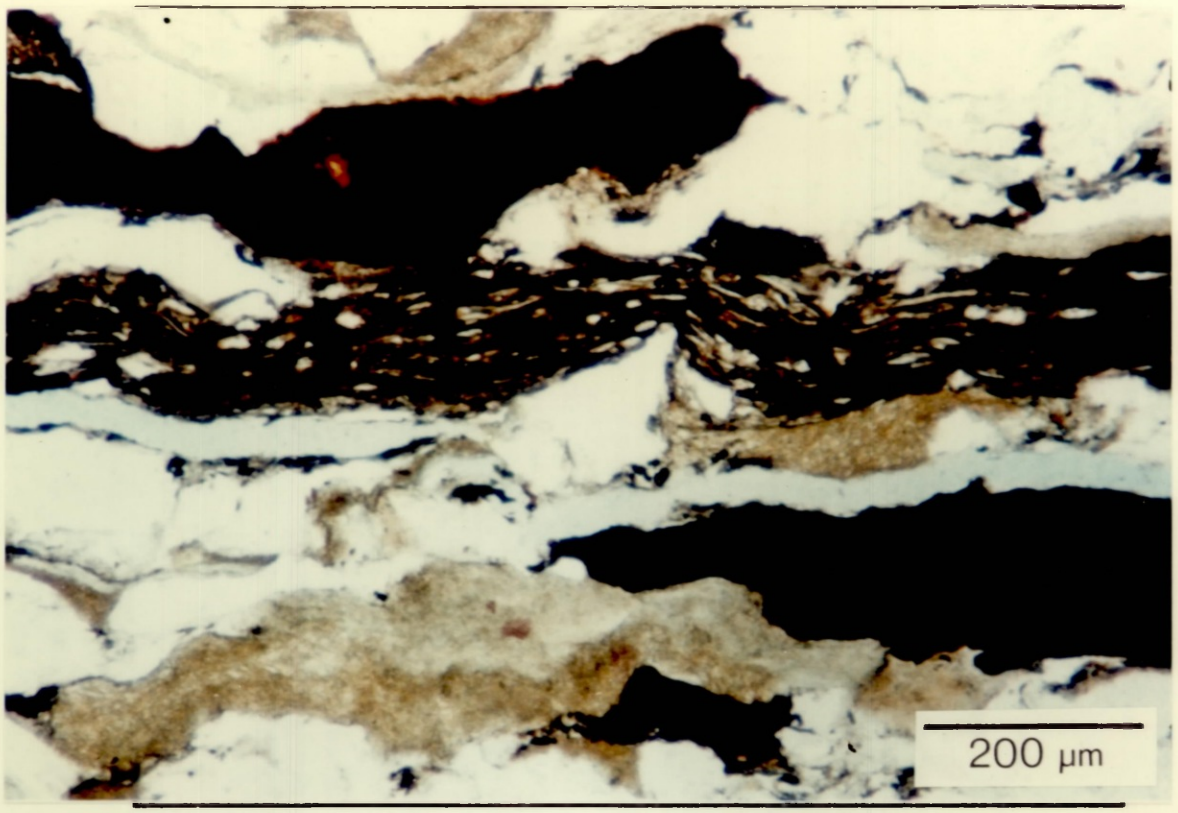
PLATE 6:6

A PPL

B XPL

Moss Farm, Rough Rock 937.75m.

Organic matter (Red/brown-black) in discrete lamina in distributary channel sandstone. This example is from a channel base. The preservation of organic matter, along with the presence of siderite implies that reducing conditions were established soon after deposition.


200 µm

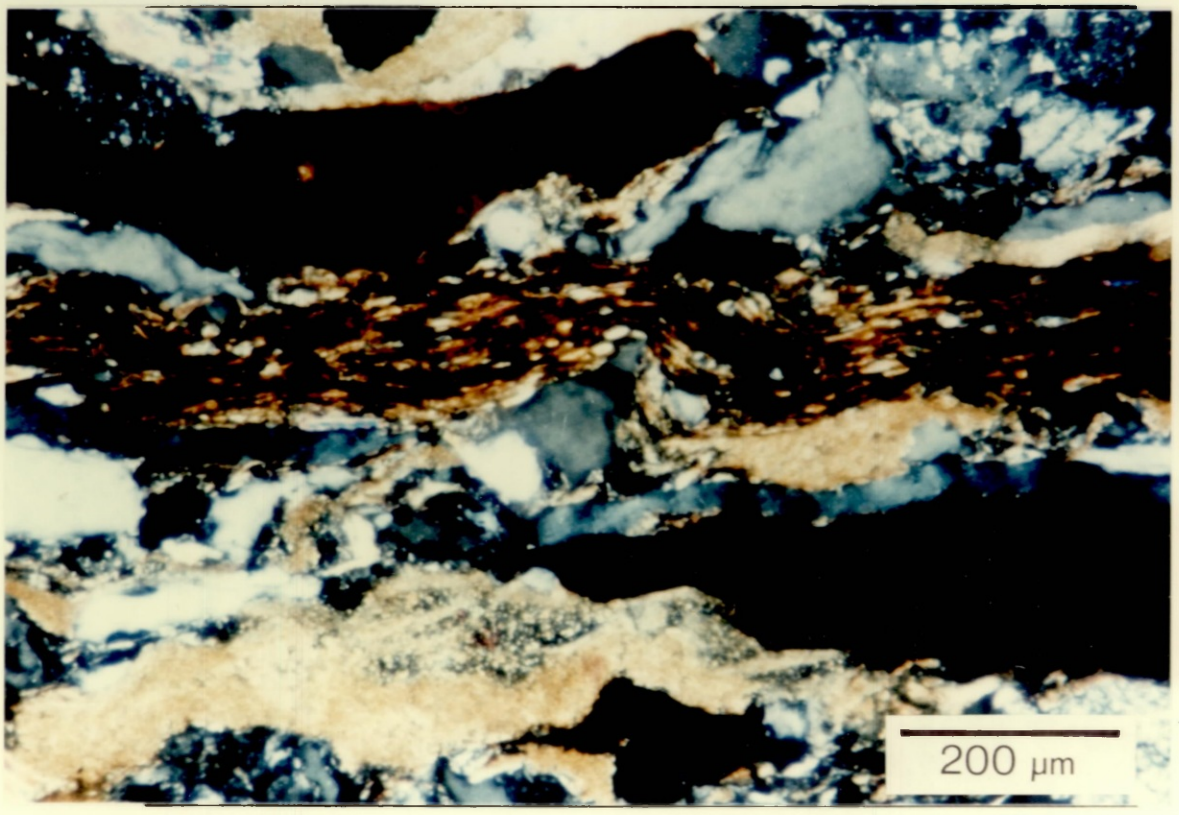

200 µm

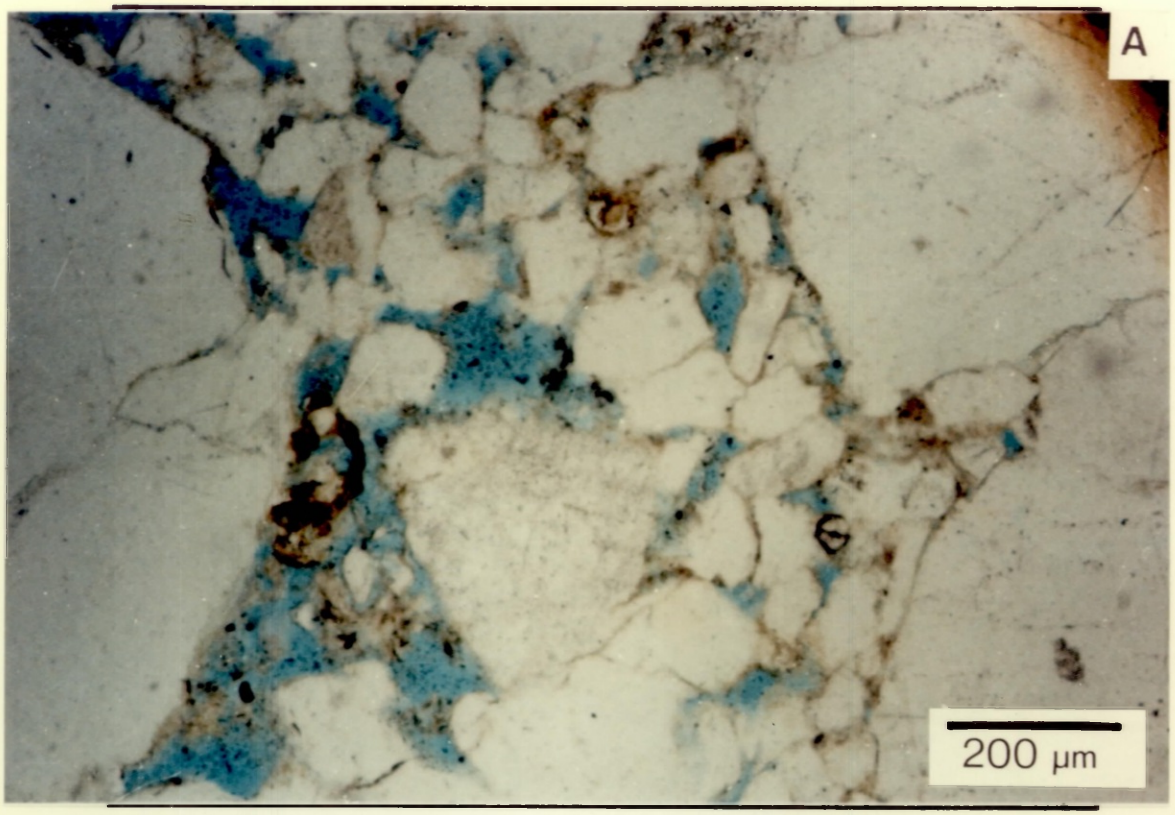
PLATE 6:7

A PPL

B XPL

Trumfleet 1, Beacon Hill Flags, 3858ft.

"Detrital" Illite. The textural position of the illite (between quartz grains) suggests a detrital origin, however SEM examination shows that even illite between quartz grains has undergone recrystalisation during diagenesis.


A
200 µm

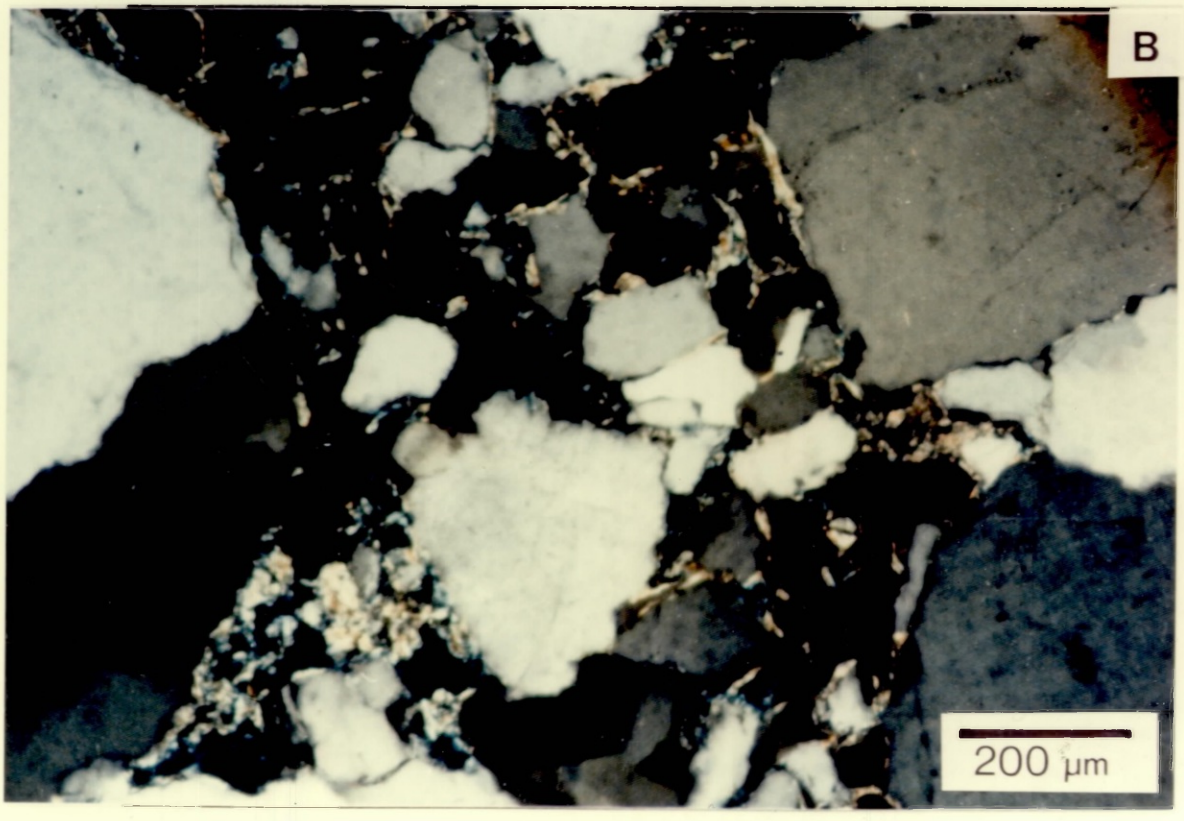

B
200 µm

PLATE 6:8

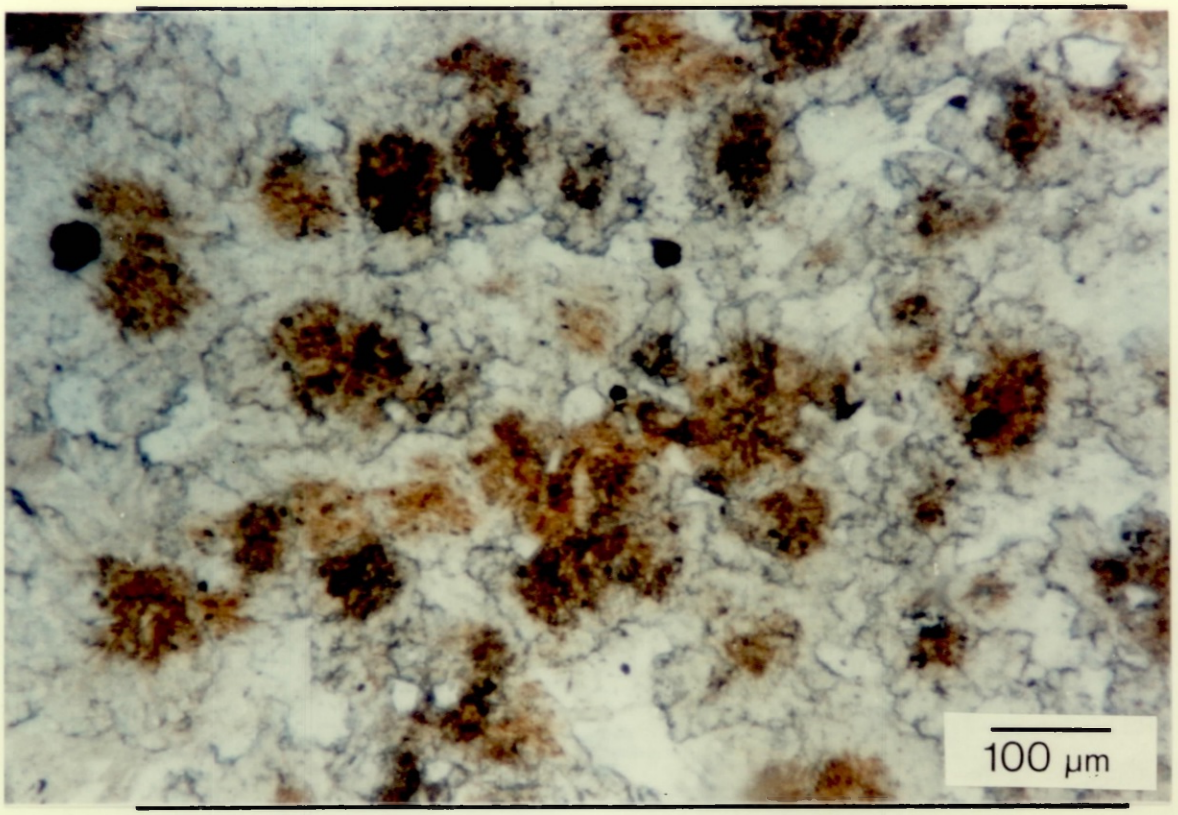
A PPL

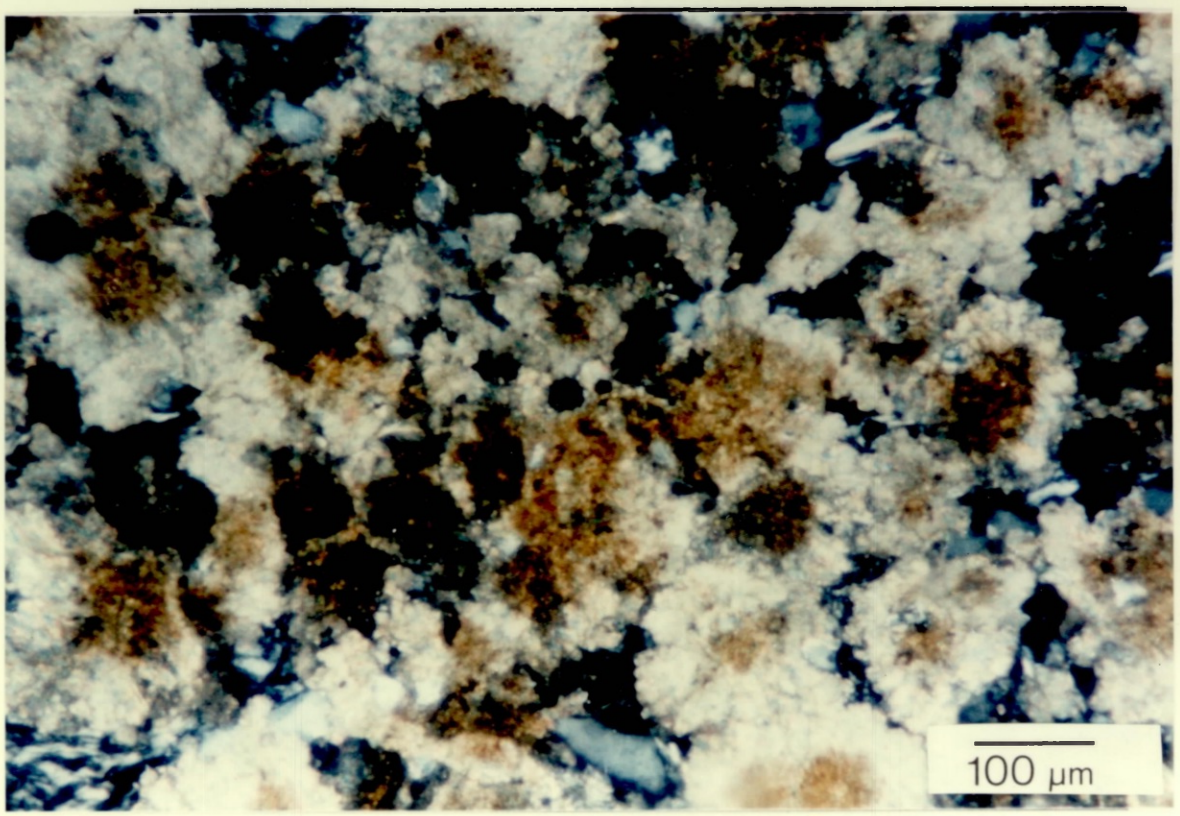
B XPL

Trumfleet 2, Beacon Hill Flags 3306ft

Subaqueous ? seatearth. Rooted horizon above channel sandstone.

Note that over 90% of the rock is authigenic, in the form of sphaerosiderite and kaolinite (See Ch 7).


100 µm


100 µm

CHAPTER SEVEN

SANDSTONE PETROGRAPHY: AUTHIGENIC MINERALS

7:1. INTRODUCTION

Petrographic analysis shows that authigenic mineral growth is the single most important mechanism of porosity reduction in the Trumfleet sandstones. As has been noted in Chapter Six, authigenic minerals can account for over 90% of the volume of the rock and in no samples studied do they account for less than 26% of the rock volume as estimated by point counting (Figs 7:1 to 7:11) In this chapter, the distribution, morphology and formation of the authigenic minerals is considered; the diagenetic implications and the paragenetic sequence will be considered in Chapter Eight. Since the Carboniferous sediments have undergone a long and complex history, interpretation of the modes of formation of the various cements is hampered by the fact that study of progressive diagenesis with depth is not possible. An attempt has been made to explain the observations of the distribution and form of authigenic minerals by reference to modern published data, and to try and integrate it with a burial history for the field.

Figures 7:1 to 7:11 show the distribution of detrital and authigenic minerals over the cored sandstone intervals. For the location of the sections refer to figure 5:4 and figures 5:6 to 5:10.

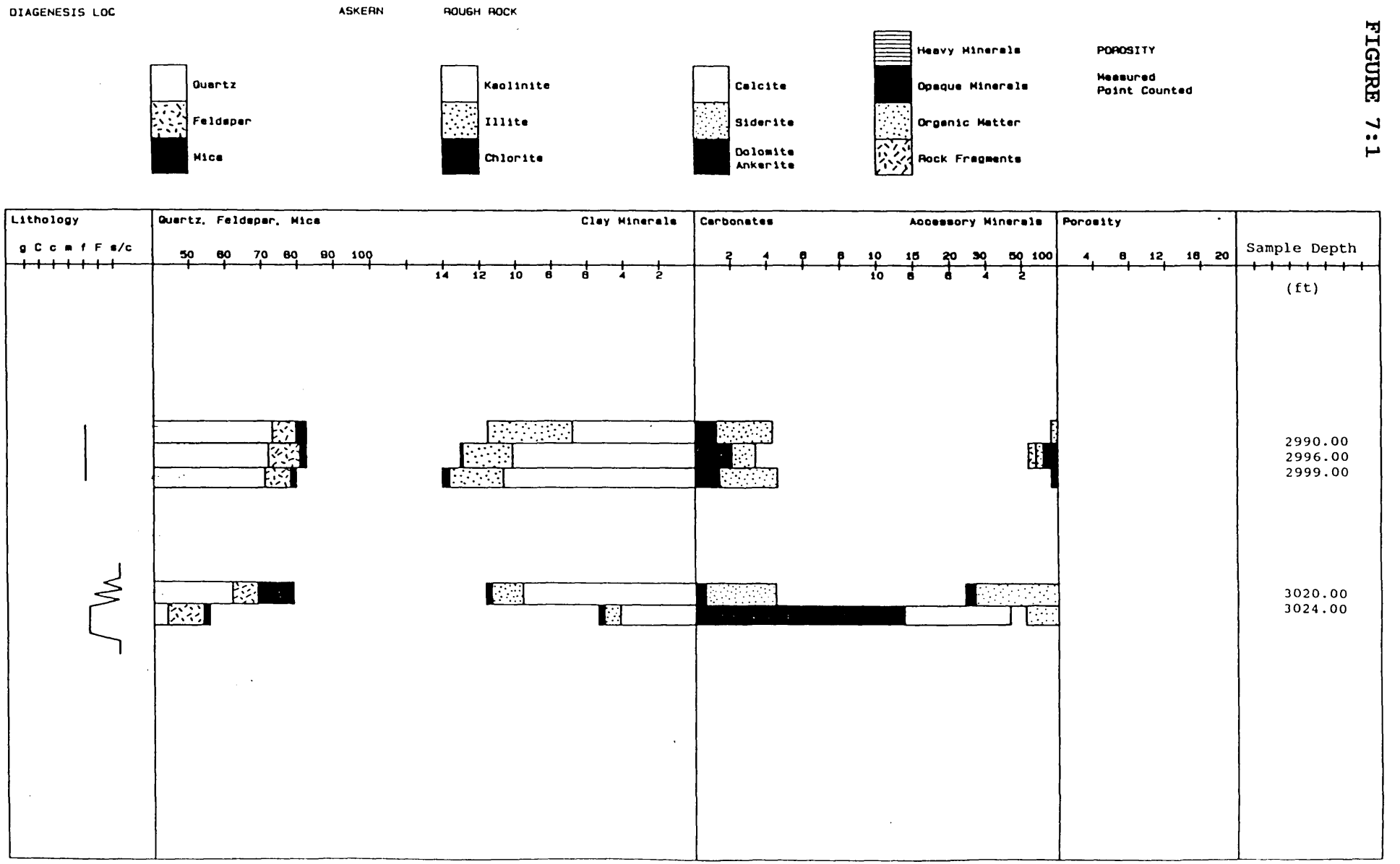

DIAGENESIS LOC
ASKERN
ROUGH ROCK
Quartz
Feldspar
Mica
Kaolinite
Illite
Chlorite
Calcite
Siderite
Dolomite
Ankerite
Heavy Minerals
Opaque Minerals
Organic Matter
Rock Fragments
POROSITY
Measured
Point Counted
Lithology
g C c m f F s/c
Quartz, Feldspar, Mica
Clay Minerals
50 60 70 80 90 100
14 12 10 8 6 4 2
Carbonates
Accessory Minerals
2 4 6 8 10 15 20 30 50 100
10 8 6 4 2
Porosity
4 8 12 16 20
Sample Depth
(ft)
2990.00
2996.00
2999.00
3020.00
3024.00

FIGURE 7:1

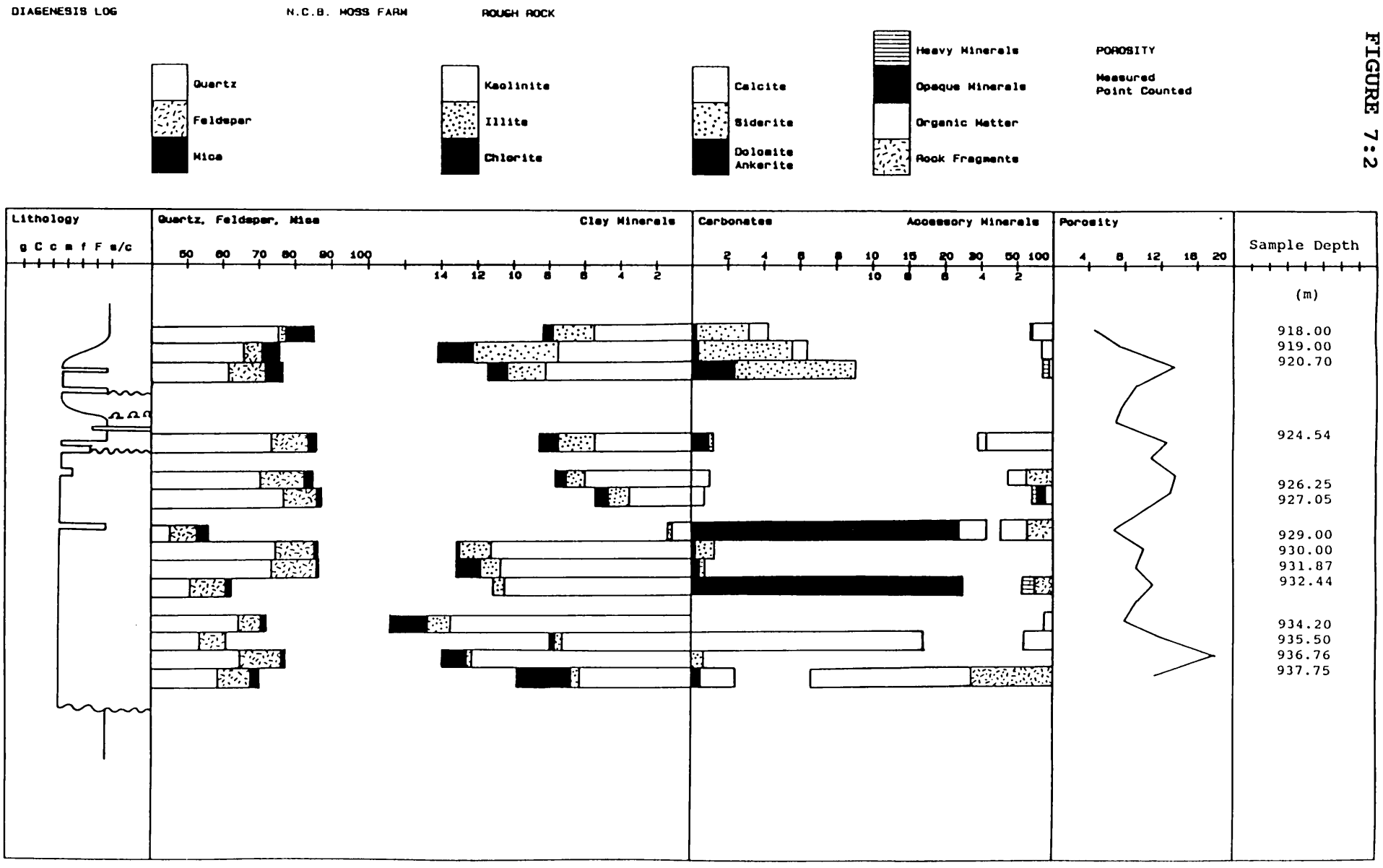

DIAGENESIS LOG
N.C.B. MOSS FARM
ROUGH ROCK
Quartz
Feldspar
Mica
Kaolinite
Illite
Chlorite
Calcite
Siderite
Dolomite
Ankerite
Heavy Minerals
Opaque Minerals
Organic Matter
Rock Fragments
POROSITY
Measured
Point Counted
Lithology
g C c m f F s/c
Quartz, Feldspar, Mica
Clay Minerals
Carbonates
Accessory Minerals
Porosity
Sample Depth
(m)
918.00
919.00
920.70
924.54
926.25
927.05
929.00
930.00
931.87
932.44
934.20
935.50
936.76
937.75

FIGURE 7:2

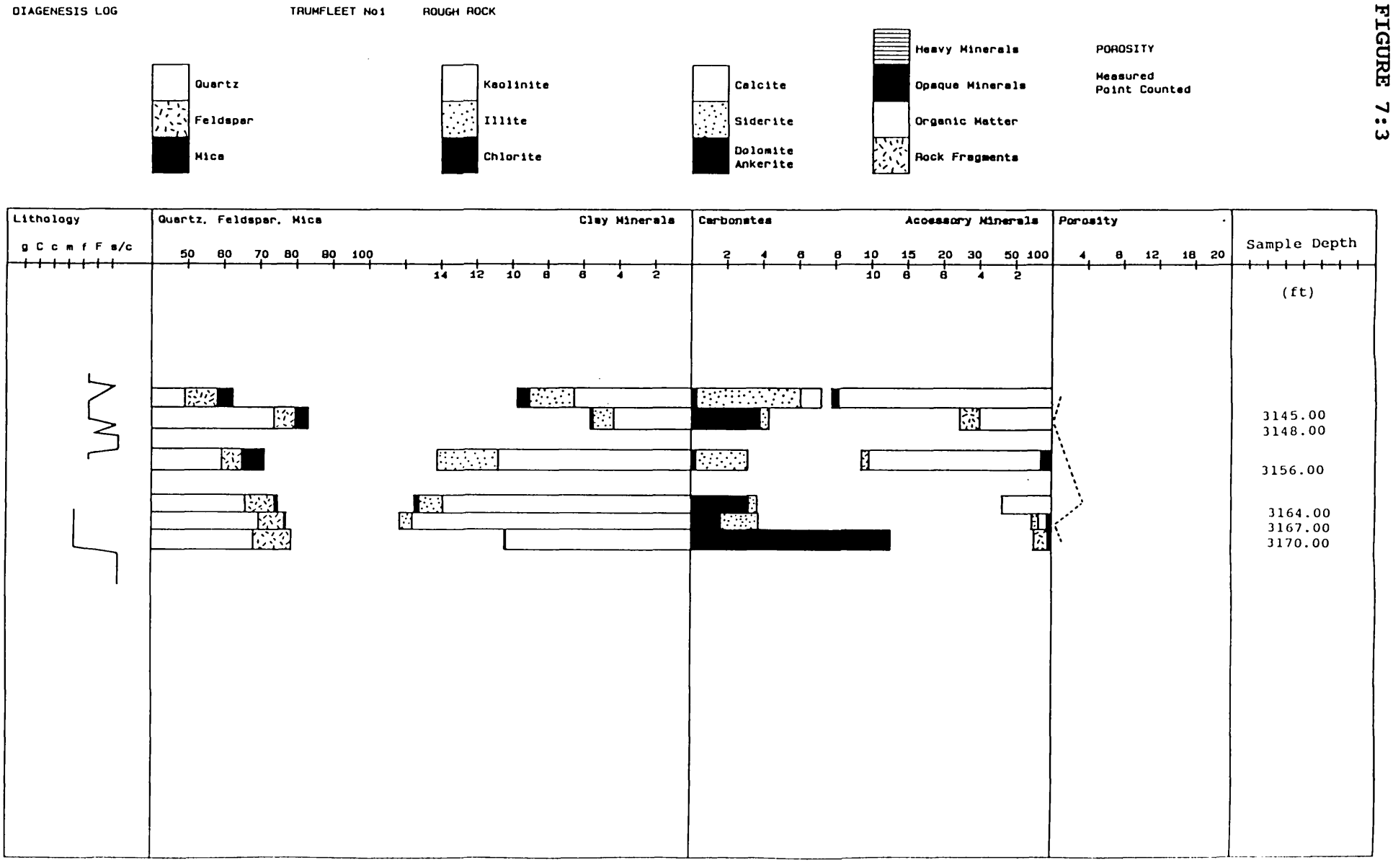

DIAGENESIS LOG
TRUMFLEET No1
ROUGH ROCK
Quartz
Feldspar
Mica
Kaolinite
Illite
Chlorite
Calcite
Siderite
Dolomite
Ankerite
Heavy Minerals
Opaque Minerals
Organic Matter
Rock Fragments
POROSITY
Measured
Point Counted
Lithology
g C c m f F s/c
Quartz, Feldspar, Mica
Clay Minerals
50 80 70 80 80 100
14 12 10 8 6 4 2
Carbonates
Accessory Minerals
2 4 6 8 10 15 20 30 50 100
10 8 6 4 2
Porosity
4 8 12 16 20
Sample Depth
(ft)
3145.00
3148.00
3156.00
3164.00
3167.00
3170.00

FIGURE 7:3

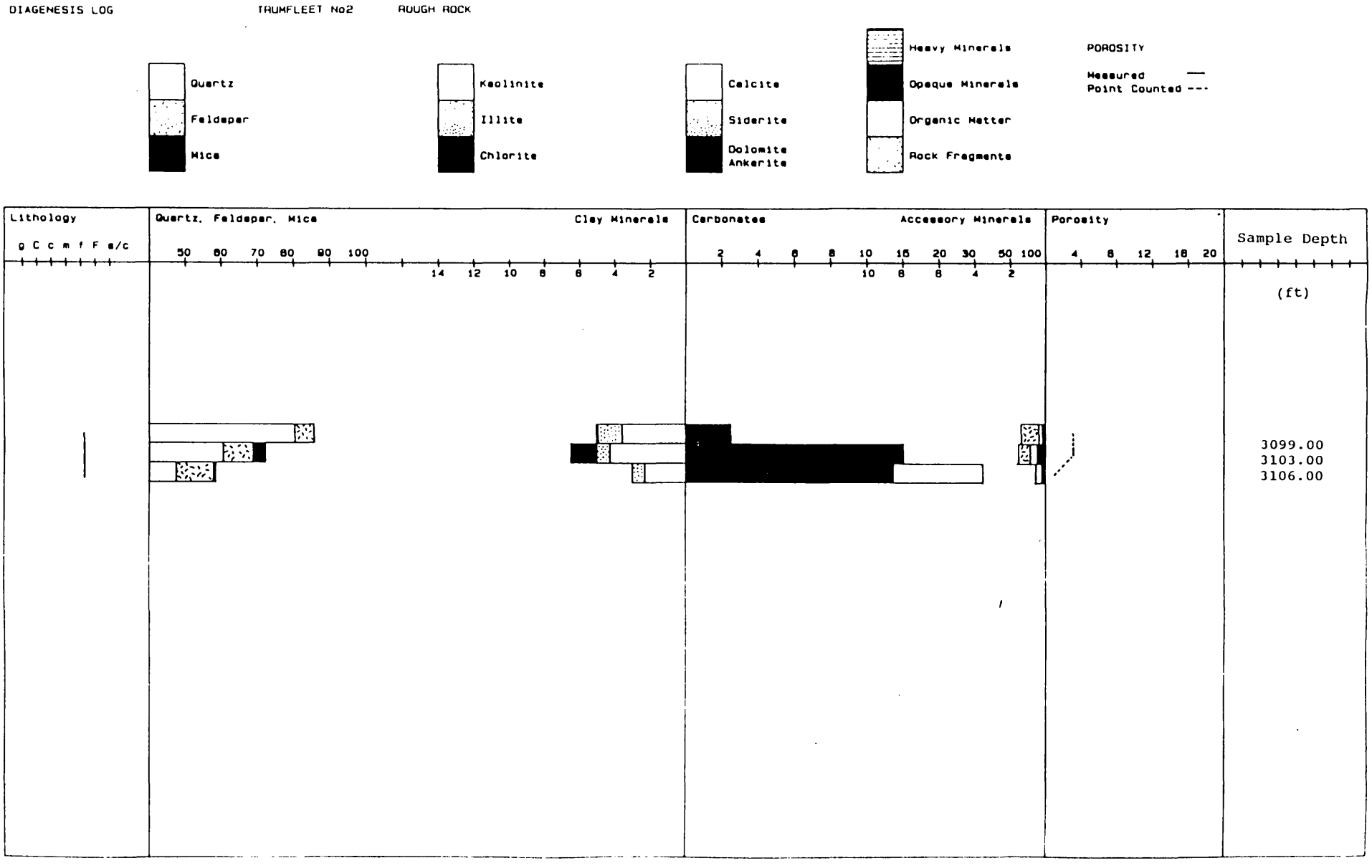

DIAGENESIS LOG
TRUMFLEET No2
ROUGH ROCK
Quartz
Feldspar
Mica
Kaolinite
Illite
Chlorite
Calcite
Siderite
Dolomite
Ankerite
Heavy Minerals
Opaque Minerals
Organic Matter
Rock Fragments
POROSITY
Measured
Point Counted
Lithology
g C c m f F s/c
Quartz, Feldspar, Mica
50 60 70 80 90 100
Clay Minerals
14 12 10 8 6 4 2
Carbonates
2 4 6 8 10 15 20 30 50 100
Accessory Minerals
10 8 6 4 2
Porosity
4 8 12 16 20
Sample Depth
(ft)
3099.00
3103.00
3106.00

FIGURE 7:4

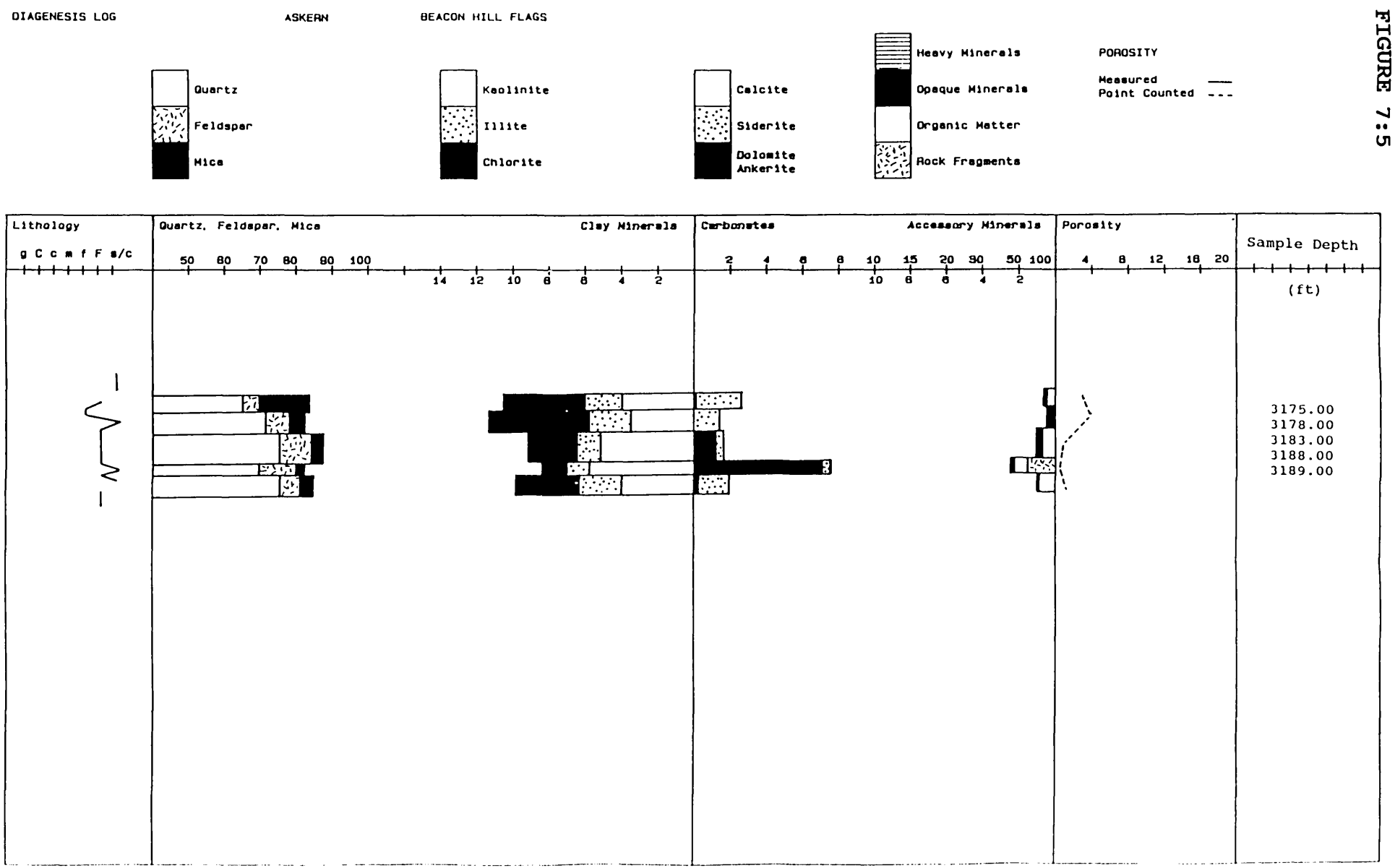

DIAGENESIS LOG
ASKERN
BEACON HILL FLAGS
Quartz
Feldspar
Mica
Kaolinite
Illite
Chlorite
Calcite
Siderite
Dolomite
Ankerite
Heavy Minerals
Opaque Minerals
Organic Matter
Rock Fragments
POROSITY
Measured
Point Counted
Lithology
g C c m f F s/c
Quartz, Feldspar, Mica
Clay Minerals
Carbonates
Accessory Minerals
Porosity
Sample Depth
(ft)
3175.00
3178.00
3183.00
3188.00
3189.00

FIGURE 7:5

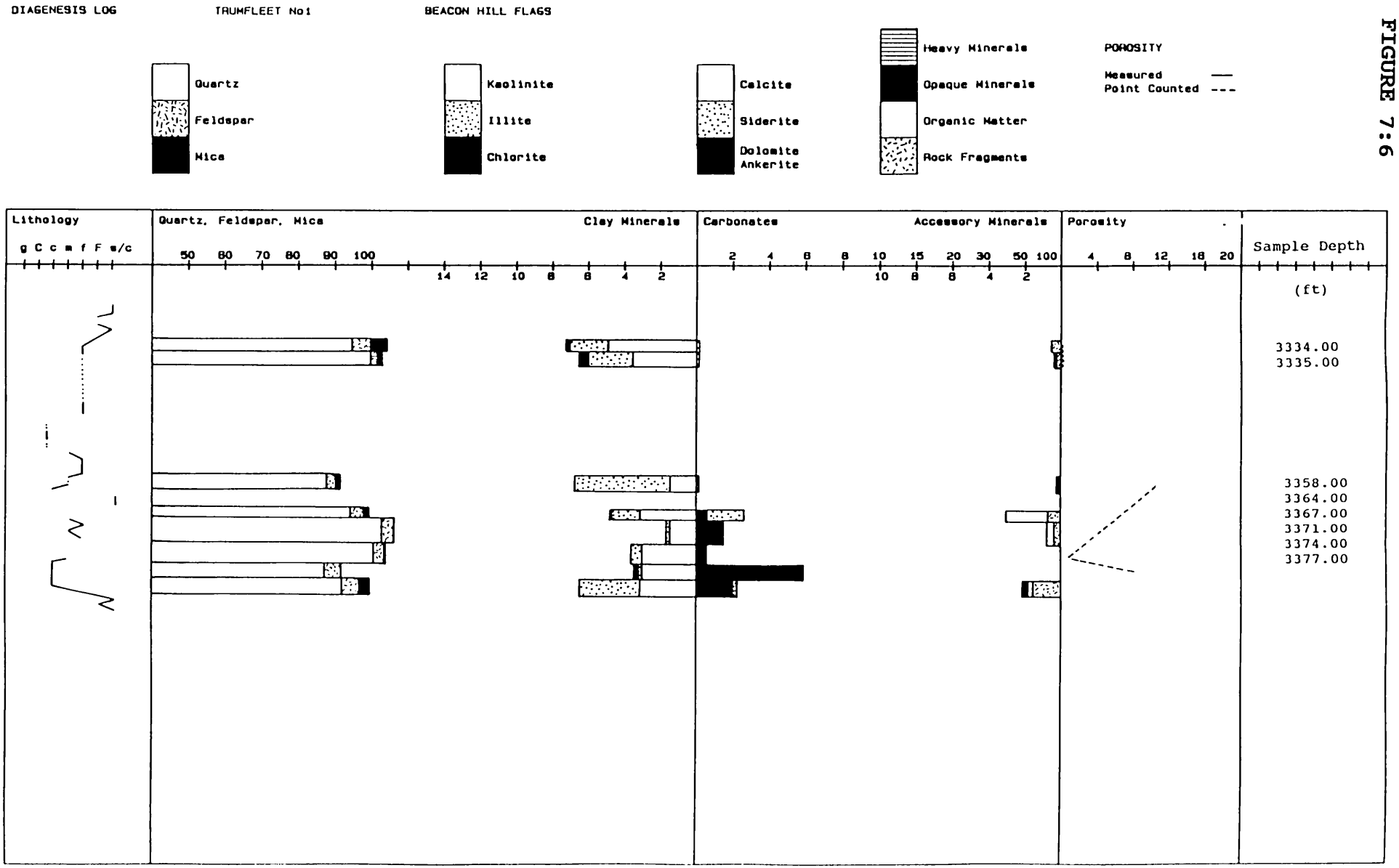

DIAGENESIS LOG
TRUMFLEET No1
BEACON HILL FLAGS
Quartz
Feldspar
Mica
Kaolinite
Illite
Chlorite
Calcite
Siderite
Dolomite
Ankerite
Heavy Minerals
Opaque Minerals
Organic Matter
Rock Fragments
POROSITY
Measured
Point Counted
Lithology
g C c m f F s/c
Quartz, Feldspar, Mica
Clay Minerals
Carbonates
Accessory Minerals
Porosity
Sample Depth
(ft)
3334.00
3335.00
3358.00
3364.00
3367.00
3371.00
3374.00
3377.00

FIGURE 7:6

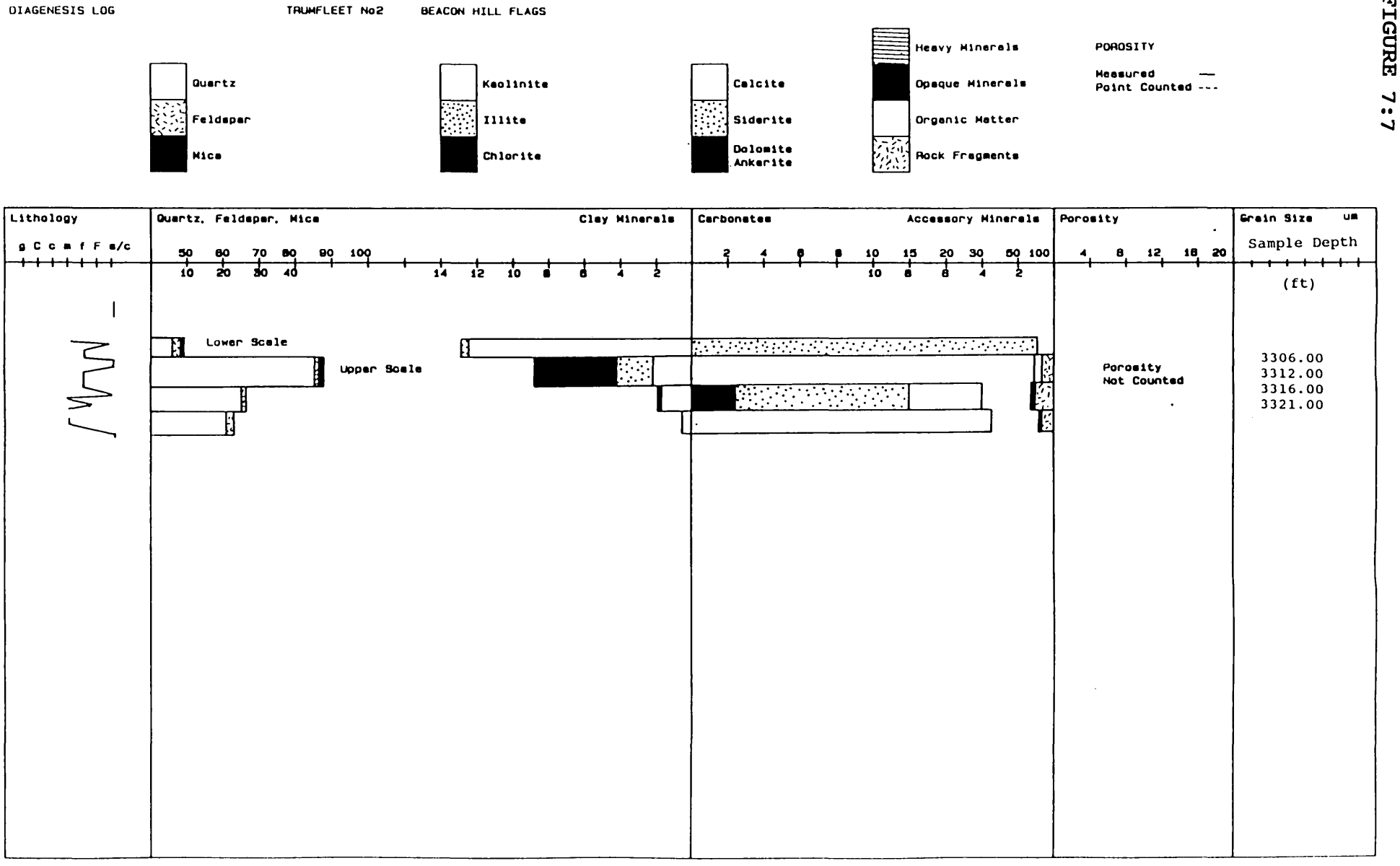
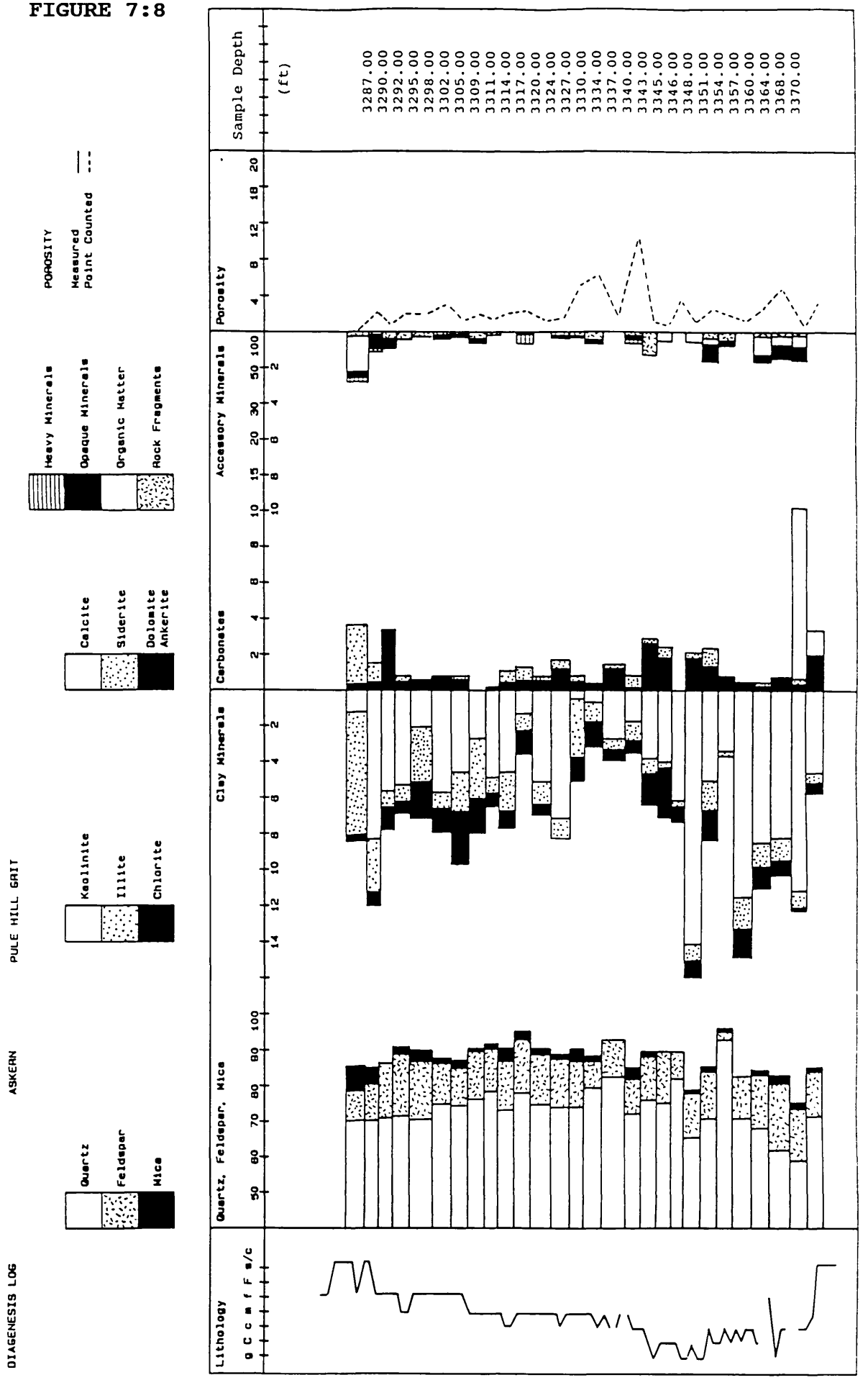

DIAGENESIS LOG
TRUMFLEET No2
BEACON HILL FLAGS
Quartz
Feldspar
Mica
Kaolinite
Illite
Chlorite
Calcite
Siderite
Dolomite
Ankerite
Heavy Minerals
Opaque Minerals
Organic Matter
Rock Fragments
POROSITY
Measured
Point Counted
Lithology
g C c m f F s/c
Quartz, Feldspar, Mica
Clay Minerals
Carbonates
Accessory Minerals
Porosity
Grain Size um
Sample Depth
(ft)
Lower Scale
Upper Scale
Porosity
Not Counted
3306.00
3312.00
3316.00
3321.00

FIGURE 7:7

FIGURE 7:8


DIAGENESIS LOG
ASKERN
PULE HILL GRIT
Quartz
Feldspar
Mica
Kaolinite
Illite
Chlorite
Calcite
Siderite
Dolomite
Ankerite
Heavy Minerals
Opaque Minerals
Organic Matter
Rock Fragments
POROSITY
Measured
Point Counted
Lithology
g C c m f F s/c
Quartz, Feldspar, Mica
Clay Minerals
Carbonates
Accessory Minerals
Porosity
Sample Depth
(ft)
3287.00
3290.00
3292.00
3295.00
3298.00
3302.00
3305.00
3309.00
3311.00
3314.00
3317.00
3320.00
3324.00
3327.00
3330.00
3334.00
3337.00
3340.00
3343.00
3345.00
3346.00
3348.00
3351.00
3354.00
3357.00
3360.00
3364.00
3368.00
3370.00

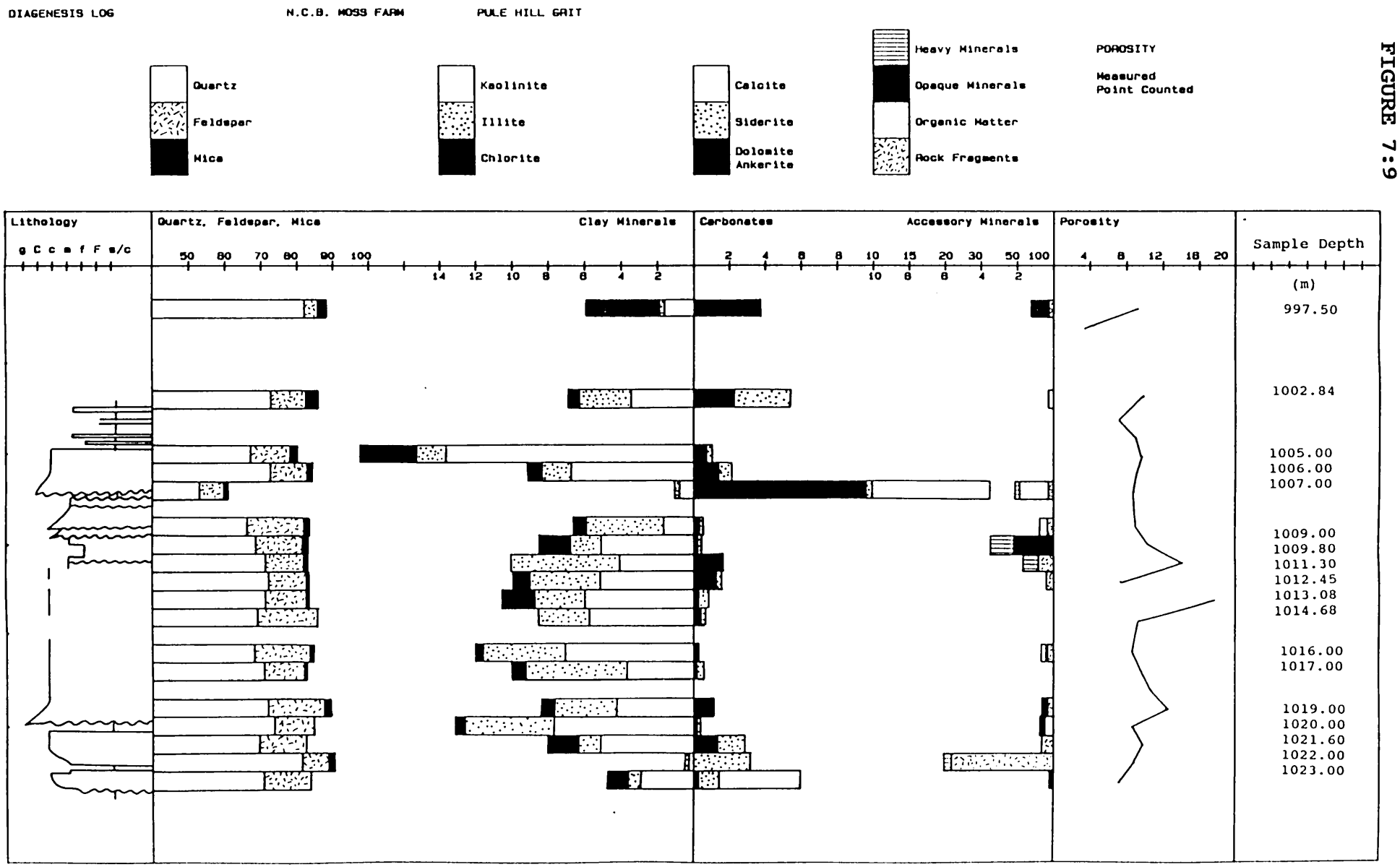

DIAGENESIS LOG
N.C.B. MOSS FARM
PULE HILL GRIT
Quartz
Feldspar
Mica
Kaolinite
Illite
Chlorite
Calcite
Siderite
Dolomite
Ankerite
Heavy Minerals
Opaque Minerals
Organic Matter
Rock Fragments
POROSITY
Measured
Point Counted
Lithology
g C c s f F s/c
Quartz, Feldspar, Mica
Clay Minerals
Carbonates
Accessory Minerals
Porosity
Sample Depth
(m)
997.50
1002.84
1005.00
1006.00
1007.00
1009.00
1009.80
1011.30
1012.45
1013.08
1014.68
1016.00
1017.00
1019.00
1020.00
1021.60
1022.00
1023.00

FIGURE 7:9

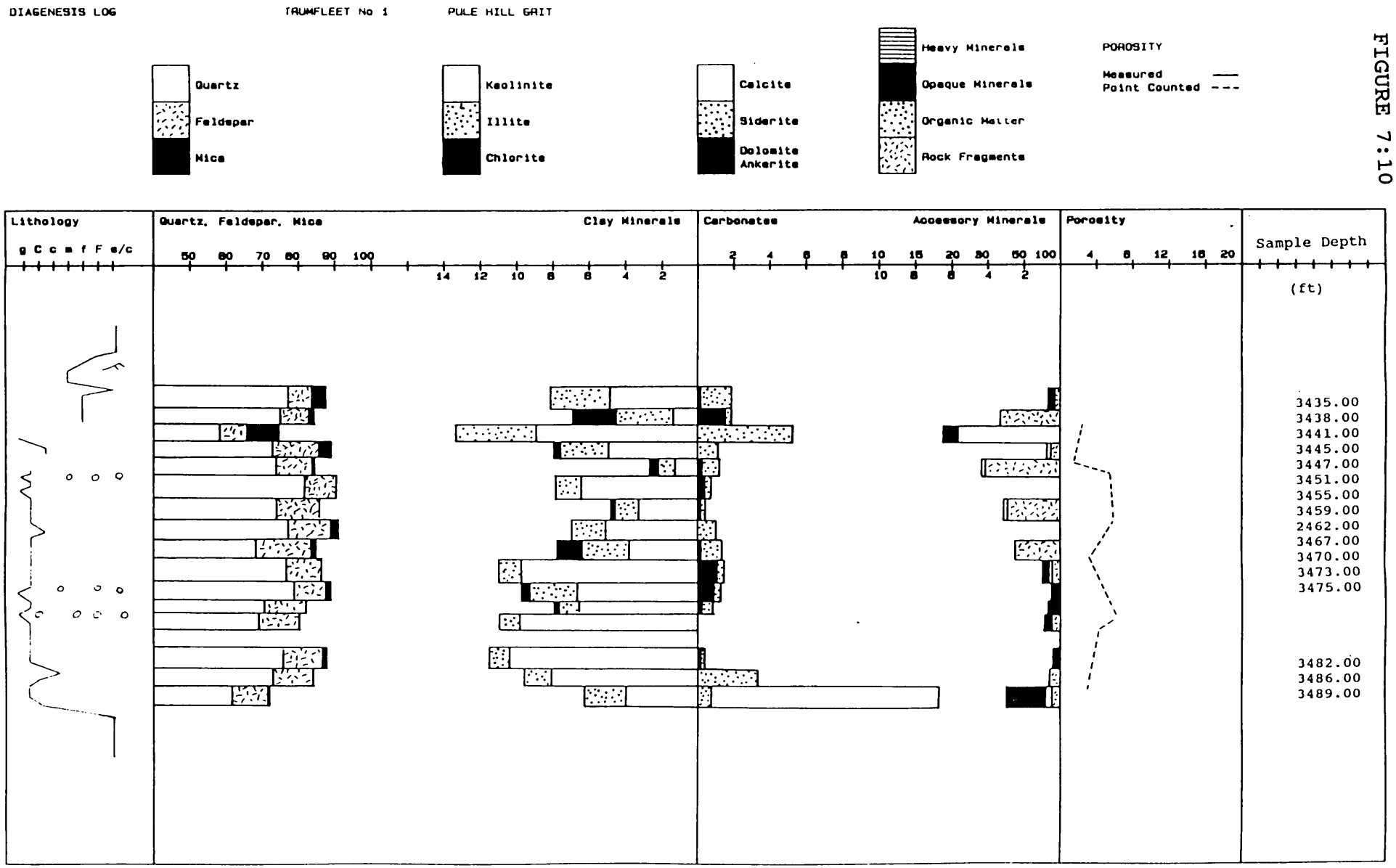

DIAGENESIS LOG
TRUMFLEET No 1
PULE HILL GRIT
Quartz
Feldspar
Mica
Kaolinite
Illite
Chlorite
Calcite
Siderite
Dolomite
Ankerite
Heavy Minerals
Opaque Minerals
Organic Matter
Rock Fragments
POROSITY
Measured
Point Counted
Lithology
g C c m f F s/c
Quartz, Feldspar, Mica
Clay Minerals
Carbonates
Accessory Minerals
Porosity
Sample Depth
(ft)
3435.00
3438.00
3441.00
3445.00
3447.00
3451.00
3455.00
3459.00
2462.00
3467.00
3470.00
3473.00
3475.00
3482.00
3486.00
3489.00

FIGURE 7:10

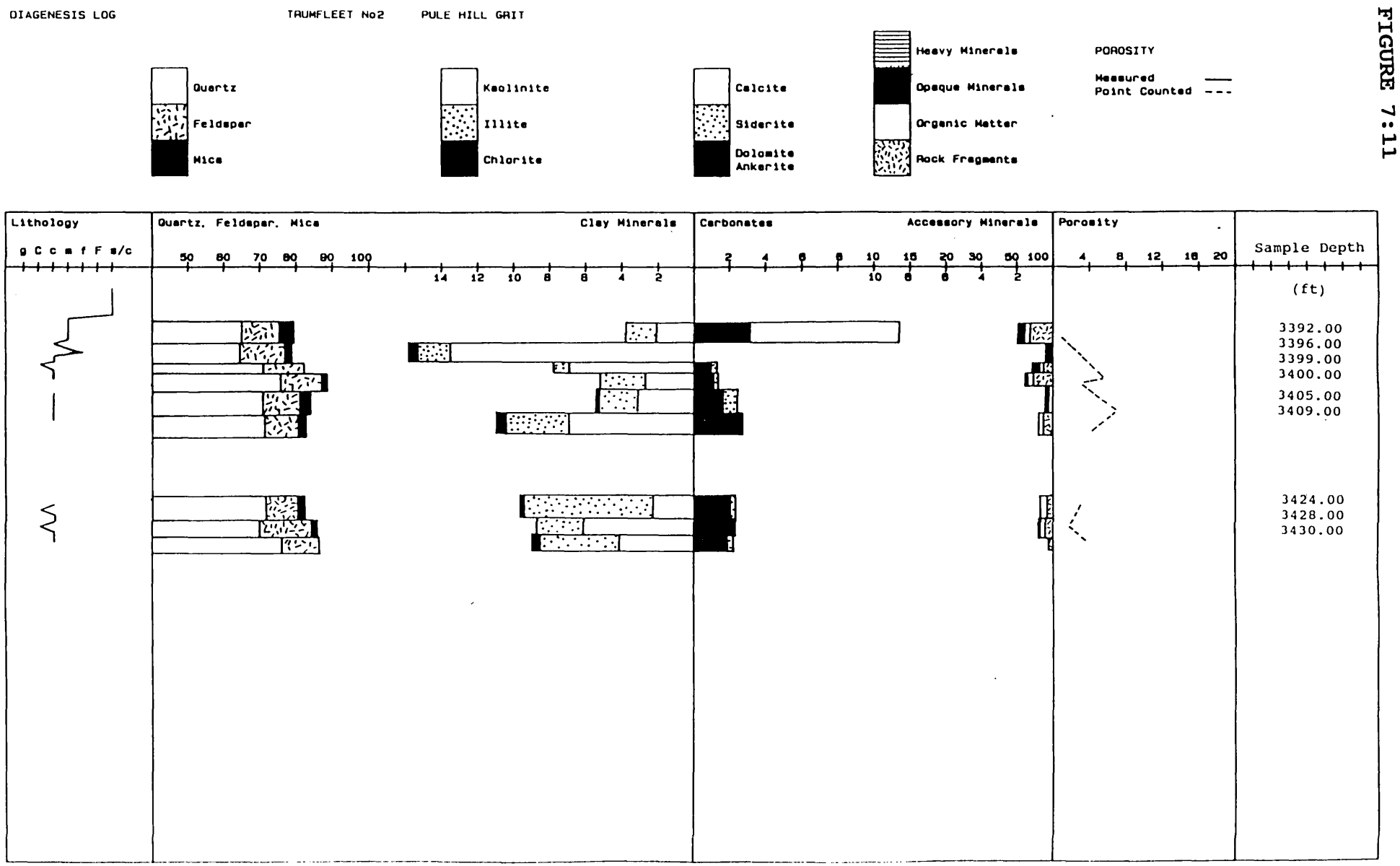

DIAGENESIS LOG
TRUMFLEET No2
PULE HILL GRIT
Quartz
Feldspar
Mica
Kaolinite
Illite
Chlorite
Calcite
Siderite
Dolomite
Ankerite
Heavy Minerals
Opaque Minerals
Organic Matter
Rock Fragments
POROSITY
Measured
Point Counted
Lithology
g C c m f F s/c
Quartz, Feldspar, Mica
Clay Minerals
Carbonates
Accessory Minerals
Porosity
Sample Depth
(ft)
3392.00
3396.00
3399.00
3400.00
3405.00
3409.00
3424.00
3428.00
3430.00

FIGURE 7:11

7:2.A NON CLAY MINERALS

7:2.1. QUARTZ CEMENTS

Well terminated authigenic quartz overgrowths were observed in all the sandstone samples studied (Plate 7:1), however, it was not possible to accurately estimate the amounts of authigenic quartz present due to the absence of "dust rims" on the detrital grain surface. In cases where it was possible to estimate the abundance of quartz cements, they varied from 1% to 23% of the total count. It is quite probable that some sandstones contain more than 23%.

At least two phases of quartz authigenesis can be discerned by the use of textural criteria. The earlier phase predates calcite cementation and probably corresponds to the time when surface waters formed the pore fluids, ie early diagenesis. Early quartz is not volumetrically significant and forms as incipient overgrowths (Plate 7:10b). The great majority of quartz authigenesis took place after calcite cementation, during burial.

Authigenic quartz was most commonly seen developed as well terminated crystal faces partially closing the pore space, (Plate 7:1) and as isolated prisms on the surface of detrital quartz which coalesce to form syntaxial overgrowths. The stages of growth of the quartz overgrowths are similar to those described by Waugh (1970) but, in addition, isolated doubly terminated

crystals were occasionally observed in the pore spaces (Plate 7:2A). The overgrowths are in optical continuity with the host grain and in some cases they coalesce completely to reduce the primary porosity to zero. Porosity reduction by quartz overgrowth is greatest in parts of the sandstone where clay minerals are less abundant (Plate 7:2B). Rhombs of authigenic siderite, which lie on the detrital quartz surface, are common inclusions in the later overgrowths.

Authigenic kaolinite is also present as an inclusion in the overgrowths but is only found close to the boundary between the overgrowth and the pore space. Kaolinite formation therefore postdates the growth of siderite and is in some cases contemporaneous last stages of quartz overgowth.

Scanning electron photomicrographs show that the surface of the authigenic quartz often carries a number of small etch pits (Plate 7:16A), about 1 um across, similar to those described from some Millstone Grit sandstones of the Pennines by Wilson (1980). He attributed this etching to epidiagenetic alteration near the surface by acidic waters derived from local peat deposits. This mechanism cannot explain the etch pits in the Trumfleet sandstones, which have never been exposed to epidiagenetic meteoric water (Downing and Howitt, 1968).

Several sources of silica for the authigenic quartz commonly observed in sandstones have been proposed by various authors. These include; the breakdown of siliceous organisms (Lewin, 1971. Hurd, 1973.), the dewatering of mudstones (Curtis, 1978. Boles

and Franks, 1979. Fuchtbauer, 1981.), the transformation of smectite to illite (Siever, 1962. Towe, 1962. Hower et al, 1979.), feldspar kaolinitisation (Hawkins, 1978) and pressure dissolution (Robin, 1978. Fuchtbauer, 1981.). For most rocks it is possible to estimate the total possible contribution of the last two sources. Bjørlykke (1979), and Blatt (1979) have shown that, owing to the low solubility of SiO_2 in pore fluids (6 ppm at 25° C, and 62 ppm at 100° C), pore fluids will quickly equilibrate with respect to SiO_2 after precipitating an insignificant amount of quartz. Fluid must be passed through the pore to supply fresh silica for quartz authigenesis. Blatt suggests that most silica is precipitated at relatively shallow depth when vertical flow in possible, ie when the surrounding shales are still permeable. However the frequent occurrence of overpressured shales, which halt vertical flow, and the reports of silica cementation at depth (Imam and Shaw, 1985.) have shown that the eogenetic formation of quartz does not operate for all sandstones. It is likely that pore fluid flow in the Trumfleet sand bodies was sub-horizontal for most of the burial history of the rocks, since fluids would have been migrating up dip from more deeply buried sediments of the same age in the Gainsborough Trough. Surdam et al (1984) have suggested that aluminosilicate solubilities can be increased by the action of organic acid complexes (carboxylic acids), which form as a response to maturation of organic matter in mudrocks, and has postulated that this agent may explain the theoretical overabundance of observed silica. The abundance of carboxylic acids in pore fluids is related directly to the action

of bacteria in the sediments which oxidise the acids. As bacteria cannot survive high temperatures, the abundance of carboxylic acid increases at temperatures in excess of 80^{o} C. The affect of thermal decarboxylation reactions also oxidises these acids at higher temperatures.

Feldspar kaolinitisation has been suggested as a source for silica cements in some sandstones (Hawkins, 1972. Huggett, 1982). The amount of silica liberated by the conversion of orthoclase to kaolinite can be calculated and it is found that 220 volumes of of orthoclase will be converted to 100 volumes of kaolinite and 90 volumes of silica (Eq 7:1) will be released.

7:1 (from Huggett 1982)

$$2\ KAlSi_3O_8 + 2H^+ + 2H_2O = Al^{Si_2O_5(OH)_4} + 4SiO_2 + 2K^+$$

Orthoclase	Kaolinite	Silica
220 vols	100 vols	90 vols

Assuming that all the kaolinite in the Trumfleet sandstones was the product of feldspar alteration a maximum of 12% of the total silica could have been sourced in this manner. As has previously been stated, the amounts of authigenic silica in the sandstones can exceed 26%. Kaolinite averages 5% of the rocks studied and therefore kaolinitisation of feldspar could supply an average of 4.5% silica for cement. Fig 7:12 shows that there is no simple relationship between the amounts of silica and kaolinite present in the sandstones, which would be expected if feldspar kaolinitisation was a major source of silica as reported

FIGURE 7:12

TRUMFLEET FIELD

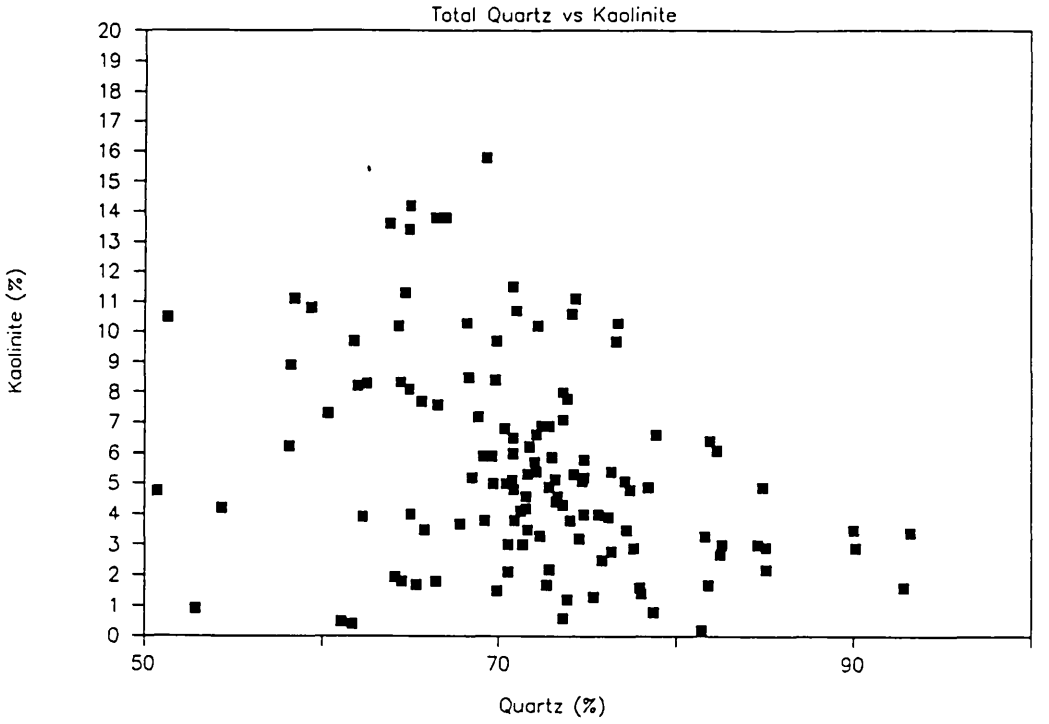
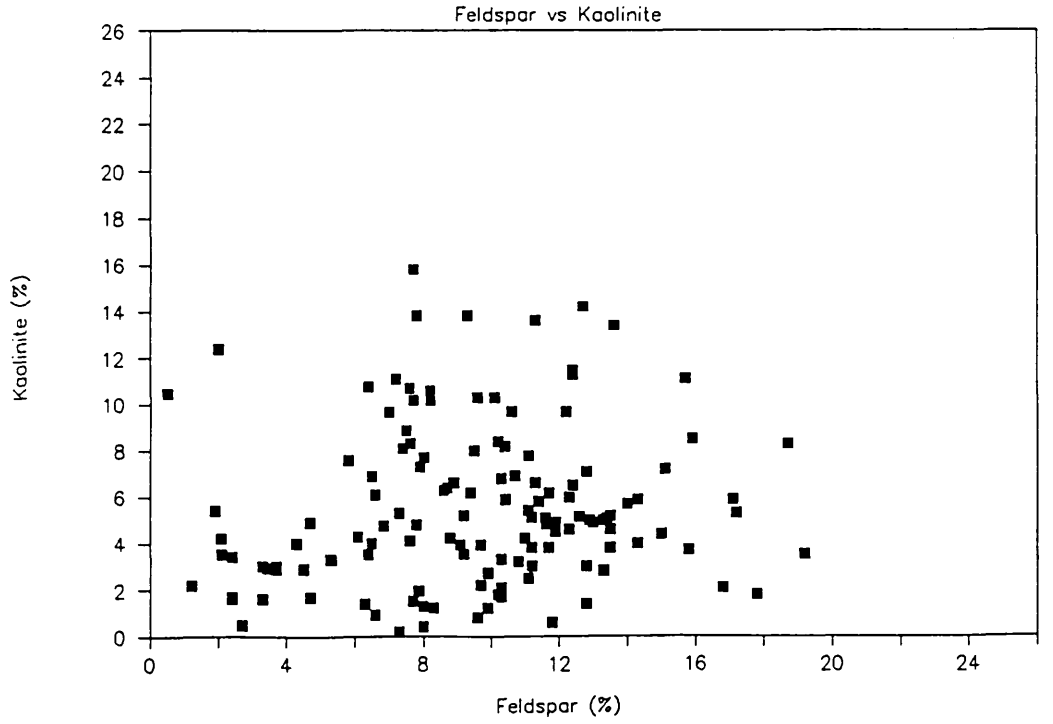


FIGURE 7:13

TRUMFLEET FIELD



for other Carboniferous sandstones by Hawkins (1978) and Huggett (1982). It is possible that the early quartz overgrowths were sourced in this manner, during early diagenesis, via the action of acid, meteoricaly derived pore fluids, possibly abbetted by the action of humic acids, but the precipitation of later cements masks the evidence for this.

Silica is also liberated by the conversion of smectite to illite, and it likely that the surrounding shales, and indeed the sandstones themselves probably contained smectite at deposition. Boles and Franks (1979) have shown that smectite is converted to illite over a large temperature depth range (55 - 210^{o} C) and that for each mole of illite produced, approximately 24 moles of silica could be released.

Fresh water contains 13 ppm of dissolved silica and marine fluids contain only 1 ppm of dissolved silica. Silica could therefore be precipitated by the mixing of fresh and marine pore fluids during early diagenesis, an occurrence highly likely in a deltaic environment. However owing to the low silica solubilities the amounts of silica precipitated would be insignificant.

Pressure dissolution of quartz was noted in this study, and observation using the cathodoluminescent mode of the SEM has shown that the detrital core has suffered some dissolution, and could therefore supply some silica to the quartz cements (Plate 7:3A and 7:3B). The amount of dissolution observed at grain contacts is however, minor and most interlocking grain contacts are formed by intergrowing overgrowths.

Silica is liberated by the transformation of muscovite to

kaolinite and by the alteration of boitite to chlorite (Bjørlykke, 1983.).

Hazeldine et al (1984) have suggested that, since silica solubility is temperature dependent, silica will be precipitated if the temperature of the pore fluids is reduced. They have proposed a convective model of quartz diagenesis in Lower Jurassic sandstones from the North Sea, in which SiO_2 derived from smectite to illite reactions in shales deeper within the sedimentary basin is carried into the sandstones by convective fluid flow, and precipitated there as the fluids cool. They noted that silica cementation took place at between 68 and 94° C. Further study is needed to ascertain if this mechanism could have operated in the Trumfleet sandstones, but it is likely that the pore fluids flowing through the sandstones were at least in part derived from downdip sandstones and mudstones deeper in the Gainsborough Trough.

Of all the sources of silica no single source could supply sufficient silica to explain the amounts of quartz cement observed in the sandstones studied. Textural criteria show that quartz cementation was an active process from deposition of the sediment to relatively deep burial diagenesis, but since it is not possible to study the progressive diagenesis of Carboniferous sediments any interpretation is by nesessity conjectural.

Early quartz was probably sourced by the action of acidic, early diagenetic pore fluids on feldspars and other labile grains.

It is probable that the mudrocks associated with the

sandstones contained smectite when deposited and the transformation of this to illite during burial probably sourced quartz cements. Late, post compactional dissolution of feldspar was observed, and this dissolution most probably sourced some quartz cements. Minor amounts of silica were also supplied by pressure dissolution.

7:2.2. FELDSPAR

Authigenic albite was observed in two forms in the Trumfleet sandstones. Primarily, as has previously been discussed, as an alteration product of detrital grains, and secondly as fresh crystals with a variety of morphologies. Fresh rod shaped overgrowths on detrital "alkali" feldspars (Plate 7:6A) and as isolated rhombs up to 5 μm long usually associated with illite were commonly observed (Plate 7:6B). No K feldspar overgrowths were observed although these have been recorded in the British Carboniferous sandstones by Hawkins (1972, 1978) and Huggett (1982). The overgrowths were observed by SEM examination of the samples but were not detectable in thin section, therefore it was not possible to quantify the amounts of authigenic feldspar present. Limited CL observations show that most feldspars are overgrown, even grains which show extensive, pre-overgrowth etching. The overgrowths do not significantly reduce the porosity.

The overgrowths show no etching and can be seen to postdate the formation of authigenic illite (SEM observation). The low abundance of neoformed albite makes it difficult observe the relationship between it and the other late stage mineral, ankerite, and it is possible that these minerals formed synchronously as in some cases they both show similar relationships with late stage illite. Authigenic albite has been reported by a number of workers, notably Kastner and Seiver (1970), Boles (1979, 1981),

Huggett (1982). In these cases however, albite was formed by the alteration of plagioclase grains or as discrete crystals of neoformed albite. Walker (1984) recorded diagenetic albitisation of potassium feldspar in arkosic sandstones, and noted that that alteration took place by direct pseuddomorphic replacement and via an intermediate phase such as dolomite or anhydrite. Spears and Amin (1981) noted that the most common feldspar type in interbedded turbiditic sandstones and shales of Kinderscoutian age from the Pennines was albite, and they postulate that this albite is detrital in origin since there would have been little opportunity for mineral reactions during diagenesis, as the sandstones and shales would have formed a closed system soon after deposition. However, they based their identification on the basis of XRD, and did not perform chemical analyses on individual feldspar grains.

Evidence of complete albitisation of detrital grains is given by the fact that no potassium feldspars were recognised during routine EDAX analysis of Trumfleet sandstones in the SEM, and microprobe analyses of feldspars show that the most potassic feldspar encountered contained only 4.6% K^+ (Fig 6:2). Backscattered electron images show lighter zones within the feldspars but no satisfactory analyses could be obtained of them, however they do show potassium upon EDAX analysis. Almost all of the feldspars analysed have a pure albite composition (Fig 6:2), and no variation in composition was observed from the core to the edge of the grains. The feldspars are untwinned and were initially mis-identified as alkali feldspars in thin section.

It seems unlikely that pure albite could be exclusively provided from a detrital source, and the untwinned "alkali feldspar" morphology of the grains suggests that the albite has formed from the alteration of detrital alkali feldspars. Walker (1984) noted that diagenetic albitisation of K feldspar can produce textures which can be confused with igneous textures and he states that restricted compositions can be indicative of diagenetic alteration to albite. It is possible that alteration of this kind is more common than has been suspected since no microprobe analyses of sandstones from the Millstone Grit have been published.

It is possible that some of the alteration of alkali feldspar to albite could have occurred due to hydrothermal alteration of feldspars prior to deposition, but the limited range in composition of the feldspars suggests that the process took place after deposition, as a few unaltered grains might be expected if the albitisation was pre-depositional. Walker (1984) has suggested that a source of Na^+ for the albitisation reaction could be the leaching of detrital plagioclase during diagenesis; however, it is unlikely that this mechanism occurred in the Trumfleet sandstones as the twinned plagioclase grains appear fresh; and since they only constitute approximately 10% of the feldspar population, they could not supply enough Na^+, even if totally depleted, to albitise the remaining 90% of the feldspar grains.

A possible mechanism of alteration is leaching of potassium during the dissolution of alkali feldspar. Boles (1982) has

shown that the process of albitisation in the Gulf Coast Tertiary takes place at depths of 3200m, corresponding to temperatures of 100 to 150 degrees C. Vitrinite reflectance data from the Trumfleet rocks suggest a maximum burial temperature of 115_o C, however, Surdam et al (1984) have shown that a zone of intense feldspar leaching takes place at temperatures in excess 80^o C.

Albite overgrowths will form in fluids which have high Na^+/H^+ and low K^+/H^+ ratios (Hess, 1966. Hutcheon, 1982.) During the formation of illite the relative Na^+/K^+ ratio would increase as K^+ was used up to form illite, and if enough illite is formed, the Na^+/H^+ ratio could increase sufficiently to allow the formation of albite.

No direct evidence of the timing or conditions under which feldspar alteration took place can be found by petrographic examination of the sandstones. The following conclusions are therefore conjectural.

The work of Boles (1979), and Surdam et al (1984) suggest that feldspar becomes highly succeptible to diagenetic alteration during burial at temperatures in excess of 80^o C. It is probable that alteration took place in a similar manner to that described by Walker (1984), during burial, possibly synchronously with the observed phase of dissolution.

THe source of the Na^+ for the albitisation was probably smectite/illite reactions in the surrounding mudrocks.

7:2.3. CARBONATES

7:2.3.1. SIDERITE

The volume of siderite observed in the Trumfleet channel sandstones varies from 0 to 12.8% of the rock, although in seatearths siderite can account for 79% of the volume (Fig 7:7. Plate 6:8A, 6:8B and 7:8A).

Siderite occurs in three forms, sphaerosiderite nodules, microcrystalline aggregates and as isolated pore filling rhombs up to 500 um in long. All forms show slight oxidation to haematite.

Plates 7:8B and 7:9A and B show sphaerosiderite nodules. They are up to 2mm in diameter and occur near the tops of the sandstone units, in the overbank and crevasse splay sandstones. Nodular siderite is the type most abundant in the seatearths and is found only at the tops of channel sandstones and in the overbank and crevasse splay sandstones. No sphaerosiderite was observed in the distributary channel sandstones. The nodules are conspicuous in hand specimen and are seen in discrete horizons often associated with detrital illite/smectite and organic matter. In thin section they often appear with hollow centres in some cases partially filled with later diagenetic kaolinite and ankerite. The edges of the nodules appear rough and they usually have a thin rim of tangential illite encasing them. These edges show interpenetrant contacts with the surrounding quartz grains and, although they were not observed by CL, it appears that the interpenetrant contacts are a result of dissolution of the

detrital quartz core during compaction rather than of overgrowths growing around the grains (Plate 7:9B). Some nodules have inclusions of detrital quartz. Mechanical and chemical compaction seems to have had little effect on the nodules and they have maintained their spherical shape even though micas and clay rich seams are often bent around them. These nodules are similar to those described by Deans (1934). Microcrystalline siderite aggregates are common in the distributary channel sandstones, and they occur as masses of rhombs up to 20 μm in diameter intimately associated with illite and organic matter. These aggregates form pore filling masses and in some cases the masses are larger than the surrounding framework grains: it appears that they have formed by the alteration of intraformational claystone fragments.

The formation of siderite requires low Eh and sulphide concentrations, neutral pH and high Fe^{2+}/Ca^{2+} ratios (Curtis and Spears, 1968. Berner, 1980.). The initial pore fluid composition in fluvial and overbank sandstones would have varied between fresh and brackish, probably with a slightly acidic pH but with low sulphide activity. Conditions suitably for the formation of siderite would be satisfied where the oxidation of organic matter caused localised reducing conditions to be established. Early diagenetic fluvially derived pore fluids would have been weakly acidic, but the reduction of ferric oxides increases the pH as hydroxyl ions are released (Curtis, 1977.).

Deans (1934) suggested that sphaerosiderite nodules formed at the sediment water interface, but Curtis and Spears (1968) have shown, from theoretical thermodynamic considerations, that the Eh

of natural surface waters is too high to allow the formation of siderite. Only ferric compounds such as haematite can form on the sediment-water interface. They postulate that siderite must have formed within pores below the sediment water interface. Burley et al (1983) state that nodular siderite commonly occurs in anoxic soil profiles.

Siderite in the Trumfleet sandstones formed during early diagenesis as a consequence reducing conditions established by the oxidation of organic matter in fresh pore fluids. Its formation postdates that of pyrite, and it is often associated with seams rich in organic matter and clays. Nodular siderite formed in sandstones and seatearths where fluid flow was sporadic, allowing reducing conditions to be maintained for longer periods of time than in the channel sandstones where fresh fluid flow inhibited the establishment of reducing conditions. For this reason, when occurring within channel sandstones, siderite is found only as a cement in organic rich intraclasts and in association with laminae rich in organic matter, which produced a reducing micro-environment in the dominantly oxidising environment of the channel sandstones.

7:2.3.2. CALCITE

Calcite cements form from 0 to 51% of the volume of the sandstones and tend to be concentrated at the bases and tops of the sandstone units (Figs 7:1, 7:2, 7:3, 7:8, 7:9, 7:10, 7:11).

Most calcite forms as poikilotopic cements in which the framework grains appear to float (Plate 7:9C, 7:10A). There is some evidence of localised corrosion of framework grains (Plate 7:9C, 7:11A), but in most cases quartz overgrowths are preserved beneath the calcite (Plate 7:10B). These overgrowths have euhedral crystal faces and, consequently, lower surface energies than the detrital cores. They are, accordingly, able to resist corrosion better than the detrital cores, hence the apparent contradictory evidence of corred grains and fresh overgrowths beneath calcite with the same sample. The lack of long grain contacts and the preservation of an open framework texture (allowing for some corrosion) implies that calcite cements formed during early diagenesis prior to major compaction. The amounts of quartz overgrowth preserved on grain surfaces beneath the calcite cements varies from sample to sample, indicating that the timing of calcite cementation was probably variable, even over short distances.

In some cases calcite cements can be seen to replace feldspar grains (Plate 7:11A), often totally replacing the grains and leaving oversized calcite filled pores (Plate 7:11B). It is thought that this is true replacement rather than simply the infilling of dissolution pores, as the bulk of feldspar dissolution postdated quartz cementation and compaction but,

where calcite cements are present, an open, non compacted texture is preserved. It is more probable that this texture is a relict depositional one rather than having formed by extensive etching, since embayed grain contact are rare and where present, quartz overgrowths show euhedral faces.

The individual crystals of calcite forming the poikilotopic cements are up to 6 cm in length and often show repeated twinning. These cements destroy the primary porosity and often show later diagenetic alteration to ankerite.

Calcite cements are common in sandstones from a variety of sedimentary environments (Hurst and Irwin, 1982, Schmidt and MacDonald 1978, Bjørlykke 1983).

Carbonates will precipitate at pH >8.5 where CO_2 is present in solution, and where sufficient Ca^{2+} ions are present.

Calcite can precipitate where fresh and marine waters mix (Bathhust, 1975. Bjørlykke, 1983) as high pH marine fluids carrying Ca^{2+} ions mix with low pH fluvial waters containing CO_2 in solution. Curtis (1978) and Huggett (1982) have suggested that calcite in cemented sandstones had been sourced from Ca^{2+} ions released from the surrounding mudstones.

To satisfy the conditions for calcite precipitation, a pH > than 8.5 and a source of CO_2 and Ca^{2+} is required. During early diagenesis the pore fluid composition would have been acidic and carbonate aggressive. During marine trangressive phases, these fluids would mix with marine derived fluids containing Ca^{2+} ions and would be cut off from their hinterlands by the deposition of a blanket of marine sediment.

Calcite cements tend to concentrated near the bases of the sandstone units and it may be that the underlying shales provided CO_2 derived from the microbial reduction of organic matter. Curtis (1983) has shown that the fluids derived from the early dewatering of of mudrocks are carbonate saturated. However, the present distribution of calcite may be controlled by later dissolution reactions which have been preferentially removed from the channel axes. A late origin for the cements, as proposed by Hawkins (1978) for the Bothhamsall field, is not tenable for the Trumfleet sandstones owing to their demonstrably precompactional origin. It is, therefore, probable that the calcite cements in the Trumfleet sandstones formed in response to two separate events: firstly, the increase of the pH of the pore fluids by the introduction of marine, Ca^{+} bearing, fluids into the system, and secondly, the introduction of carbonate saturated fluids from the underlying mudrocks at low to moderate burial.

7:2.3.3. ANKERITE

Ankerite cement is the least common carbonate in the rocks studied; It makes up from 0 to 25% of the volume of the rock and occurs in three forms. The higher modal percentages form by the wholesale replacement of pre-existing calcite cements and feldspar, and it also occurs as late stage pore filling rhombs up to .2 mm in diameter, associated with authigenic illite (Plates 7:12 and 7:13). Ankerite is often found in association with early siderite (Plate 7:12B) and is generally more common near the centre of the sandstone units (Fig 7:1 to 7:11). Ankerite tends to be concentrated in thin lenses and is often interlaminated on a cm scale with calcite (Plate 7:14A). Limited microprobe data show that the ankerite is iron rich and also contains up to 2% Mn.

Boles (1978) noted the alteration of calcite to ankerite with depth in the subsurface Eocene of south-west Texas and suggested that the alteration took place according to the following equation

Eq 7:2

$$4CaCo_3 + Fe^{2+} + Mg^{2+} = 2CaMg_{0.5}\ Fe_{0.5}\ (Co3)_2 + Ca_{2+}$$

He postulated that the Fe^{2+} and Mg^{2+} for the reaction were supplied by smectite transforming to illite in the enclosing mudstones, and noted that ankerites appear at temperatures of 125 degrees C and at depths greater than 2560m. In the Trumfleet

sandstones, ankerite is also present as a direct precipitate similar to those described by Boles (1977). Curtis (1983) has shown a progressive increase in the Fe^{2+} of precipitated carbonates with depth is common.

Ankerite cements formed as one of the last cements in the Trumfleet sandstones. They formed by the alteration of pre-existing calcite cements and by the replacement of feldspar, as well as by direct precipitation in the pore spaces. Their occurrence in the centre of the sandstone units reflects their late formation, as it is probable at this time that the fluid flow was along the channel axis rather than upwards through it, since the tops and bases of the sandstone units were extensively cemented by calcite.

7:2.4. PYRITE

Pyrite never forms more than 0.3% of the volume of the rock, but it is observed in the majority of the rocks studied.

Pyrite is easily seen in thin section as opaque cubes usually associated with organic matter and often with siderite (Plate 7:7A).

Curtis and Spears (1968) and Berner (1980) state that the factors controlling pyrite precipitation are low Eh and high sulphide activity. If there is a decrease in sulphide activity siderite will form at the expense of pyrite.

As the initial pore fluid composition of the Trumfleet sandstones would have been fresh to brackish, the initial sulphide activity would have been low and siderite would have been the metastable phase. Indeed, in all the sandstones studied siderite is much more common than pyrite. Pyrite could only form locally where anoxic microenvironments formed due to the presence of organic matter which scavenged the available oxygen. Eventually the organic H_2S produced by microbial action was used up in pyrite production and siderite became the stable phase.

In the Trumfleet sandstones pyrite formed locally as an early diagenetic cement where sulphate was supplied by the bacterial decomposition of organic matter.

7:2.5. BARITE

Barite occurs as a pore filling cement, and as a replacement of feldspar in the Pule Hill Grit. It is very rare being observed in only two samples from the Moss Farm well, where it reaches a maximum abundance of 0.9%.

Hawkins (1972) reported similar occurrences of barite cement in the Crawshaw/sub-Alton Sandstone of the East Midlands, and noted that it also replaced kaolinite, and, therefore, that its formation postdated that of the kaolinite.

Hawkins (1972, 1978) proposed that barite cements in sandstones formed from Ba^{2+} released from the surrounding shales by dewatering and sulphate derived from organic matter in the marine bands. The barite observed in this study is found in the channel centre, in sandstones which contain no organic matter. If the barite was derived from the surrounding shales it might be expected that it would be precipitated near the shale/sandstone interface, or sandstones rich in organic matter, where a local supply of sulphate would be present. This is not the case in the Trumfleet sandstones and it is suggested that the barite was derived from barium and sulphate rich pore fluid passing through the sandstones. Barite has been reported as a vein mineral in the British Carboniferous (Inneson and Ford, 1982.). They postulate that it is derived from the Dinantian and that it is derived from brines which have "leaked" up fault planes, however, this is usually associated with galena, and no galena was observed in the Trumfleet sandstones.

It is possible that the sulphate used in the preciptation of barite came from the dissolution of pyrite during burial. No obvious source for the barite presents itself. Unfortunately the low abundance of barite makes its relationships with the other authigenic minerals difficult to establish and its precise mode and time of formation remain unresolved.

7:3.B CLAY MINERALS

7:3.1. INTRODUCTION

Clay minerals comprise from 0.5 to 16.6 % of the total volume of the rock (point count data) and are mainly authigenic in origin. XRD diffactograms show that by far the most common clay mineral is kaolinite with a significant amount of illite. Chlorite was recorded on most diffractograms a very minor phase which proved difficult to find upon SEM examination.

The criteria used by Wilson and Pittman (1978) to differentiate between detrital and authigenic mineral phases proved unreliable in this study. Illite, which appears between grains was seen to have an authigenic aspect upon SEM examination. Illite was commonly observed near the edges of pores and on the pore walls, whilst kaolinite was observed in the centre of the pore; however, SEM examination showed that the kaolinite was overgrown by illite and that therefore the formation of the illite postdated that of the kaolinite.

7:3.2. KAOLINITE

Kaolinite is the dominant clay mineral in the Trumfleet sandstones varying from 0.5 to 15.8% of the volume of the rock (Point count data). All the kaolinite observed in this study was authigenic, since it was always in the form of well crystallised books or vermiculae, or as an alteration product of a precursor detrital mineral. This interpretation does not preclude the possibility that some of the kaolinite may be the recrystallised form of detrital clay minerals, as it is highly probable that some detrital clay minerals were deposited in the sandstones. Spears and Amin (1981) have shown that the amount of kaolinite in interbedded turbiditic sandstones and shales of Namurian age in the Pennines decreases with the distance from the palaeoshoreline.

Within the Trumfleet sandstones kaolinite is observed mainly as a pore filling cement, although it also occurs as a replacement of feldspar and muscovite, and as a fracture filling clay mineral. In some cases kaolinite can be seen to be infilling secondary dissolution pores. It is ubiquitous in sandstones and shales of all environments and is easily identified in thin section owing to the coarse nature of the books. The books are sufficiently coarse (up to 25 μm) to allow the differentiation of kaolinite and dickite to be made in thin section on the basis of their exctinction angles. The books are blocky in appearance with some platy crystals and show euhedral faces. There is no evidence of etching the books and they usually appear "fresh", even when overgrown by illite

(Plates 7:15a, 7:15B, 7:16). The 7.14 Å kaolinite peak identified on the XRD traces is rather stable at high temperature. This peak persists, with reduced intensity, when the sample is heated to 550° C for one hour, and it is thought that this stability is a function of the coarse grain size.

The kaolinite was observed to be partially overgrown by quartz (Plate 7:14B) but no inclusions of kaolinite were observed in the quartz overgrowths. It is proposed that kaolinite crystallisation for the most part postdates that of quartz. The hypothesis of kaolinite crystallisation pre dating quartz and being pushed into the centre of the pores by force of crystallisation was considered but is rejected as kaolinite was never seen enclosed by quartz but the reverse situation was observed frequently. Huggett (1982) suggested that the book like nature of kaolinite is caused by cleavage of crystals which have undergone mechanical compaction. In her study the kaolinite formed at the same time as quartz and in some cases predated its' formation. The bulk of the kaolinite in this study formed after quartz cementation and compaction had set up a rigid framework, and, since they appear in secondary open pores, it is difficult to envisage mechanical compaction as being responsible for the book like nature of the kaolinite in the Trumfleet sandstones (plate 7:15A). It is probable that the morphology of the kaolinite is a function of crystal growth.

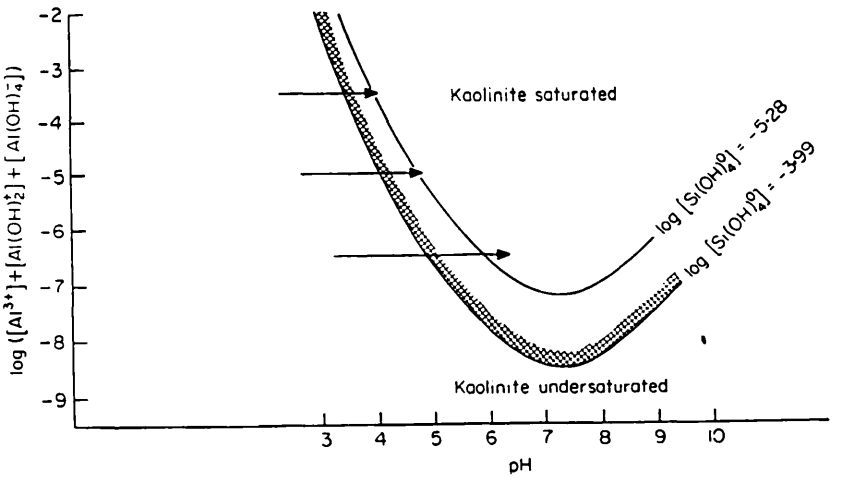
Kaolinite will form by the alteration of feldspars and phyllosilicates in dilute acidic environments at moderate temperatures (Dunnoyer de Segonzac, 1970, Shaw, 1980.) Dunnoyer de Segonzac (1970) has shown that at temperatures which vary from

80 to 180° C, kaolinite will be replaced by dickite and eventually nacrite in an acidic environment, but if the pH is increased illite or chlorite will form. Hurst and Irwin (1982) have suggested that pore fluid movement can affect the morphology of kaolinite books, and have proposed that it is possible to differentiate between kaolinite which has grown as a result of flux controlled reactions (ie, influx of pore fluids of differing compositions) and those crystals which have grown as a result of diffusive processes. They state that kaolinite which has grown in fluvial environments, as a result of feldspar dissolution by meteoric pore fluids, has a skeletal appearance, in contrast to kaolinite of marine environments where diffusive reactions and slower pore fluid flow lead to slower crystal growth and the formation of euhedral kaolinite books. The bulk of the kaolinite in the Trumfleet sandstones resembles the latter diffusion controlled type and the rarer vermicular type (Plate 7:16B) may have formed by a mechanism similar to that proposed by Hurst and Irwin (1982) for fluvial sandstones during early diagenesis. They also state that in fluvial environments, kaolinite authigenesis frequently predates that of quartz, which is not the case in this study.

Curtis (1983) has show that kaolinite will form in response to an increase in pH of pore fluids which carry alumina in solution (Fig. 7:14). He proposes that as the pH increases as pore fluids react with carbonates, to form secondary porosity, kaolinite precipitates as a consequence, since the pore fluids can no longer carry the alumina in solution.

In many published studies of the diagenesis of fluvial sandstones it has been found that kaolinite authigenesis predates silica cementation and is a result of feldspar dissolution by acid fluvial pore fluids (Almon and Davis, 1979. Dutton, 1977. Stanton 1977, Blanche and Whittaker, 1978. Hawkins, 1978, Huggett, 1982. Hurst and Irwin, 1982). In the Trumfleet sandstones it has been demonstrated that most of the kaolinite formed after quartz cementation as a result of precipitation alumina carried into the sandstones from the surrounding mudstones during burial. The presence of kaolinite in fractures indicates that alumina can be transported by diagenetic fluids, and it is proposed that the kaolinite precipitated in response to a pH increase the pore fluids as a consequence of dissolution reactions in the sandstones (Curtis 1983, Figure 7:14).

FIGURE 7:14



Kaolinite stability as a fuction of solubility of Si and Al species and pH.
(Curtis, 1983. Diagram adapted from Burley, 1985.)

7:3.3. ILLITE AND ILLITE/SMECTITE

The term "illite" covers minerals with a 10Å mica structure (Grim, 1952.). This definition includes both dioctahedral and trioctahedral varieties and minerals with a wide range in composition. They differ from true micas in that they are more hydrous and contain lesser amounts of the M^+ cation. Illites also exist in a variety of polytypes based on the nature of the layer stacking (Brindley and Brown, 1980).

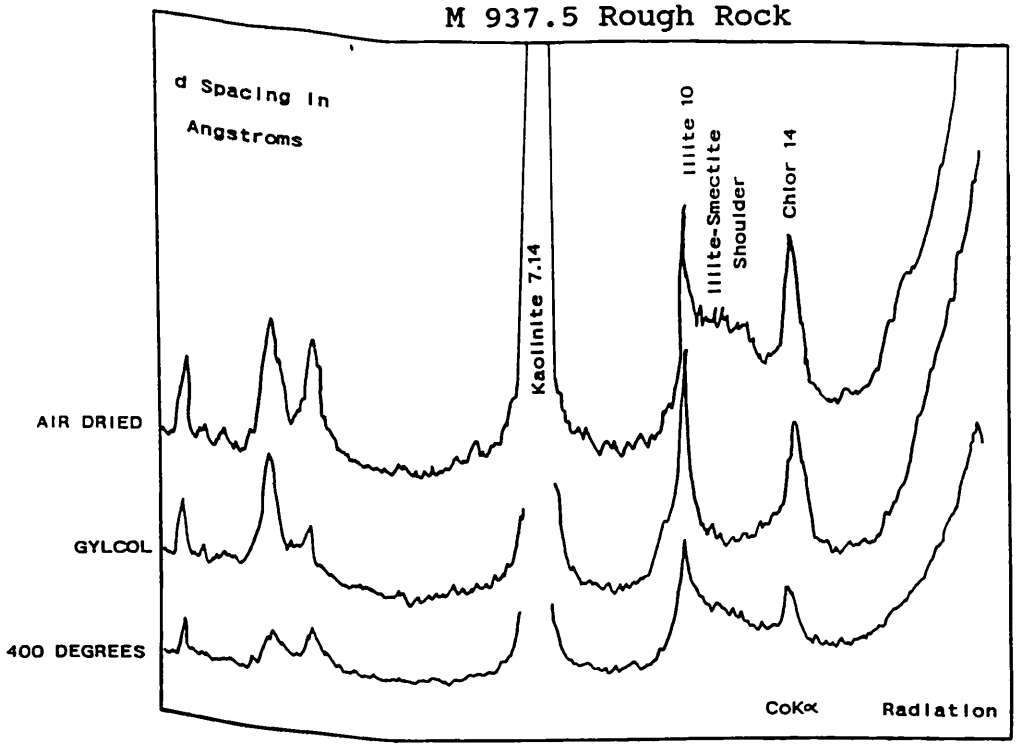
XRD traces of clay mineral fraction samples from the Trumfleet sandstones show a diffuse peak from 9.97Å to 10.8Å (Figs 7:15-7:18). A sharp 10Å peak was frequently observed and is caused by the presence of detrital mica.

In almost all samples where illite was present a "shoulder" was observed on the 10Å illite peak indicating the presence of a mixed layer component too small to quantify (Fig 7:15). Illite and mixed layer illite smectite account for a maximum of 7% (Point count data) of the volume of the rock. They occur as late stage pore lining and pore filling cements, and are also observed coating pressure dissolution surfaces. Detrital illite (lllite/smectite?) was observed between grain boundaries, concentrated into clay and mica rich layers, and in clay rich intraformational mud clasts.

The morphology is highly variable, the most common morphology observed on SEM examination being blades and laths showing collapsed core structures and irregular masses (Plates 7:17A, 7:17B). Illite also exists exhibiting a sheet-like morphology (Plates 7:15B, 7:19B), honeycomb form (possibly mixed

FIGURE 7:15

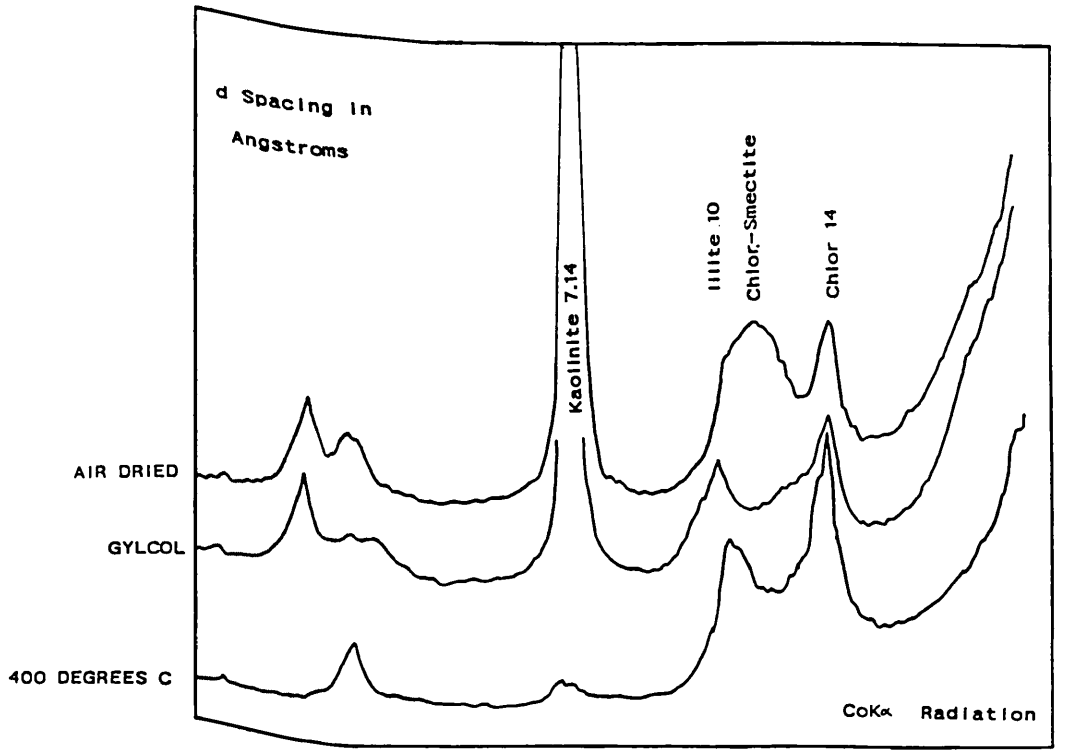
M 937.5 Rough Rock



ORIENTATED CLAY DIFFRACTOGRAM

A 3178

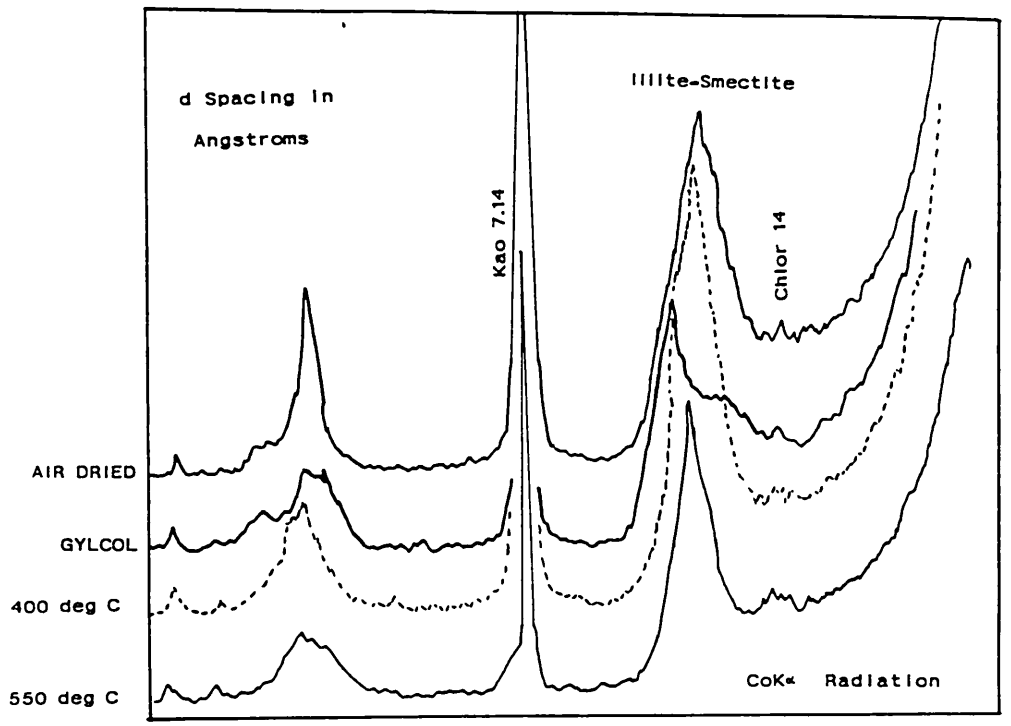
FIGURE 7:16



ORIENTATED CLAY DIFFRACTOGRAM

FIGURE 7:17

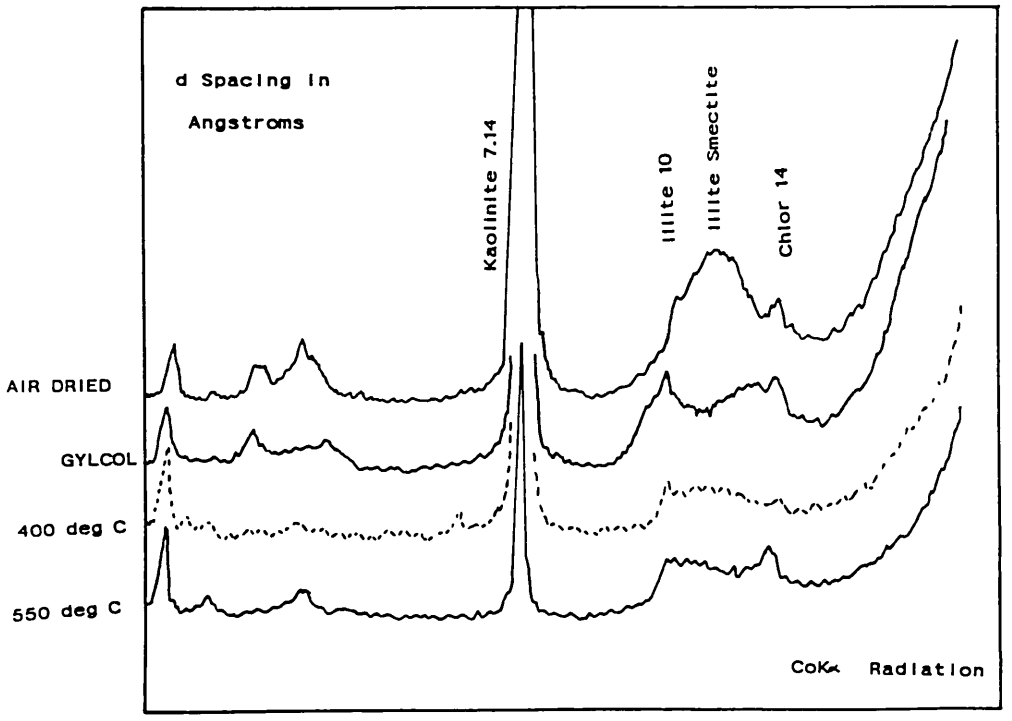
M 1014



ORIENTATED CLAY DIFFRACTOGRAM

M924

FIGURE 7:18



ORIENTATED CLAY DIFFRACTOGRAM

layer, Plate 7:19A), and in a rare lath like pore bridging form (Plates 7:18). It was also observed as an alteration product of muscovite.

Where illite was observed in association with kaolinite, the age of formation of the illite was seen to be later than that of the kaolinite (Plate 7:15B). The illite forms sheets coating the kaolinite but only limited evidence of illite replacement of kaolinite was observed, and the kaolinite showed no signs of etching. It is often found near the pore walls whilst the pore centres are totally free of illite.

Illite is also observed coating detrital feldspar and probably formed by replacement of detrital feldspar, but it is difficult to tell the difference between pre-depositional hydrothermal alteration and diagenetic alteration. Authigenic feldspar is not illite coated and isolated crystals of albite can be seen apparently on top of illite; the evidence suggests that illite formation predated that of albite (Plate 7:6B).

The honeycomb form of (mixed layer?) illite is developed on stylolitic surfaces and on pressure dissolution surfaces (Plate 7:20). It is not clear whether this is recrystallised detrital clay or wholly neoformed illite which has formed by diffusion during the pressure dissolution process. Rutter (1983) has shown that aqueous films can exist on stressed mineral interfaces, allowing the possibility of the removal and introduction of material by diffusive processes. Volkmar Schmidt (pers comm) has shown that lamellar porosity exists between interlocking grain

interfaces and has postulated the possibility of mineral dissolution and growth along these boundaries.

It has been suggested that illite may form via illite-smectite, by the dehydration of smectite at temperatures as low as 60° C. (Burst, 1977. Boles and Franks, 1979. Boles, 1982, Dunnoyer de Segonzac, 1970.). The degree of ordering of the interstratification and the crystallinity of the illite increases with increasing temperature (Perry and Hower, 1970. Dunnoyer de Segonzac, 1970.), random mixed layering being characteristic of lower temperatures (Hoffman and Hower, 1979). The nature of the smectite illite "transformation" has been recently re-assessed, and a dissolution/precipitation reaction seems likely, rather than solid state transformations (Nadeau 1984).

Illite can also form by the alteration of kaolinite in an alkali pore fluid environment, when K^+ ions are available at temperatures from as low as 60° C (Dunnoyer de Segonzac, 1970. Shaw, 1980.)

Illite and mixed layer illite/smectite exist in a variety of morphologies and compositions in the Trumfleet sandstones. They formed by transformation of detrital illitic clay material, replacement of detrital feldspar and muscovite, and as a true neoformed phase precipitated directly from the pore fluids. Core and lath morphologies were formed by direct precipitation and honeycomb varieties are well developed along pressure dissolution surfaces. The smectite component of the mixed layer illite/smectite component of the illite is probably a relict detrital phase.

7:3.4. CHLORITE

Chlorite is the least abundant clay mineral present in the Trumfleet sandstones and, though it was commonly detected during XRD analysis, it was often difficult to find in thin section and during SEM examination. When the backscattered mode of the SEM was used later in the study the true nature of the chlorite was revealed. It was observed that the majority of the chlorite in the Trumfleet sandstones was present as an alteration product of biotite (7:21A). The values of chlorite estimated by modal analysis tend therefore to be underestimated as partially altered biotite grains were counted as biotite rather than chlorite. The values estimated for chlorite vary from 0 to 5.6 % of the volume of the rock.

Pore filling and rimming chlorite was rare in the Trumfleet sandstones; the bulk of it was in the form of chloritised grains and chlorite mica stacks similar to those described by Huggett (1982).

Hoffman and Hower (1979) noted that chorite will form in sandstones over a wide range of pressure and temperature conditions, from weathering profiles to the onset of greenschist metamorphism. Dunnoyer de Segonzac (1970) has proposed that early diagenetic chlorite forms by aggradation of Mg^{2+} and that chlorite will form by the alteration of kaolinite when Mg^{2+} ions are available. No evidence of kaolinite alteration to chlorite was observed in the Trumfleet sandstones; the bulk of

the pore filling variety was seen to consist of poorly crystallised masses of chlorite, probably of detrital origin.

Chlorite exists in the Trumfleet sandstones as an alteration product of biotite, and as rare pore filling aggraded detrital masses. Pore rimming forms are very rare and were only observed by SEM examination of the rocks. It is not clear whether the alteration of biotite to chlorite is a pre-depositional or diagenetic feature but, as biotite is relatively unstable mineral under most depositional and diagenetic environments, it is quite likely that chloritisation started prior to deposition and continued during diagenesis. No unaltered biotite was observed in the Trumfleet sandstones.

7:3.5. CHLORITE/SMECTITE

A mixed layer mineral, tentatively identified as mixed layer chlorite smectite was identified from XRD traces. It was not recognised in thin section or SEM examination. An 11.5 A peak was identified on air dried samples, which expanded to 12.8A on solvation with ethylene glycol. The peak shifted to 10.8A after heating to 400^{o} C. and disappeared completely upon heating to 550^{o} C (Fig 7:16). The temperature stability suggests a chlorite component and the fact that similar results were obtained when the sample was saturated with an MgCl solution, suggests that smectite rather than vermiculite is responsible for the expansion. The behaviour upon solvation with gylcerol was similar to that with ethylene glycol. The higher order reflections appeared to be unaffected by solvation. The mineral most likely to be responsible for such behaviour is ordered mixed layer chlorite-smectite (Brindley and Brown, 1980, p 287. H Shaw, pers comm.).

The fact that this mineral appears to be more common in the shales than in the sandstones suggests that it is probably of detrital origin.

7:3.6. GLAUCONY

Glaucony is an extremely rare mineral in the Trumfleet sandstones and was only recorded as single grains in two thin sections of sandstones of the Beacon Hill Flags. These occurrences are however significant as the presence of glaucony is somewhat unusual in fresh water sediments (Bjørlykke, 1982), and its presence is often cited of glaucony as being proof of a marine origin for sediments in which it occurs (Selley, 1974), however, Rhorlich et al (1969) have recorded chamosite glaucony from a Scottish fresh water loch (Odin and Matter, 1981.).

The term "glaucony" refers to any green pelletal, iron rich clay mineral found in sediments. The loose definition means that a wide range of mineral structures and compositions are covered by the term.

Odin and Matter (1981) consider that the formation of glaucony is controlled by the availability of Fe^{2+} and K^{+} and the sedimentation rate. Slow sedimentation rates are needed to form glaucony at the sediment/water interface. Rhorlich et al (1969) show that glaucony can form diagenetically below the sediment/water interface, and a similar origin is proposed for the glaucony in the Trumfleet sandstones.

The extreme paucity of glaucony in the Trumfleet sandstones makes its paragenesis difficult to establish. It is widely accepted that glaucony cannot form under fresh water early diagenetic conditions, but there is no reason why it cannot form in sediments of non marine origin if the correct pore fluid

conditions for its formation can be satisfied. It is proposed that the traces of glaucony in the Beacon Hill Flags below the sediment/water interface, possibly after the sandstones had been flushed by saline water after the Reticulocers superbilingue marine transgression. The possibility that the glaucony was reworked from a marine environment as considerd by thought unlikely since there is little evidence to suggest that the Pennine Basin was truly marine, and the glaucony would have been deposited down palaeoslope from its present position.

PLATE 7:1

A

Trumfleet 1, 3286ft. Beacon Hill Flags. PPL

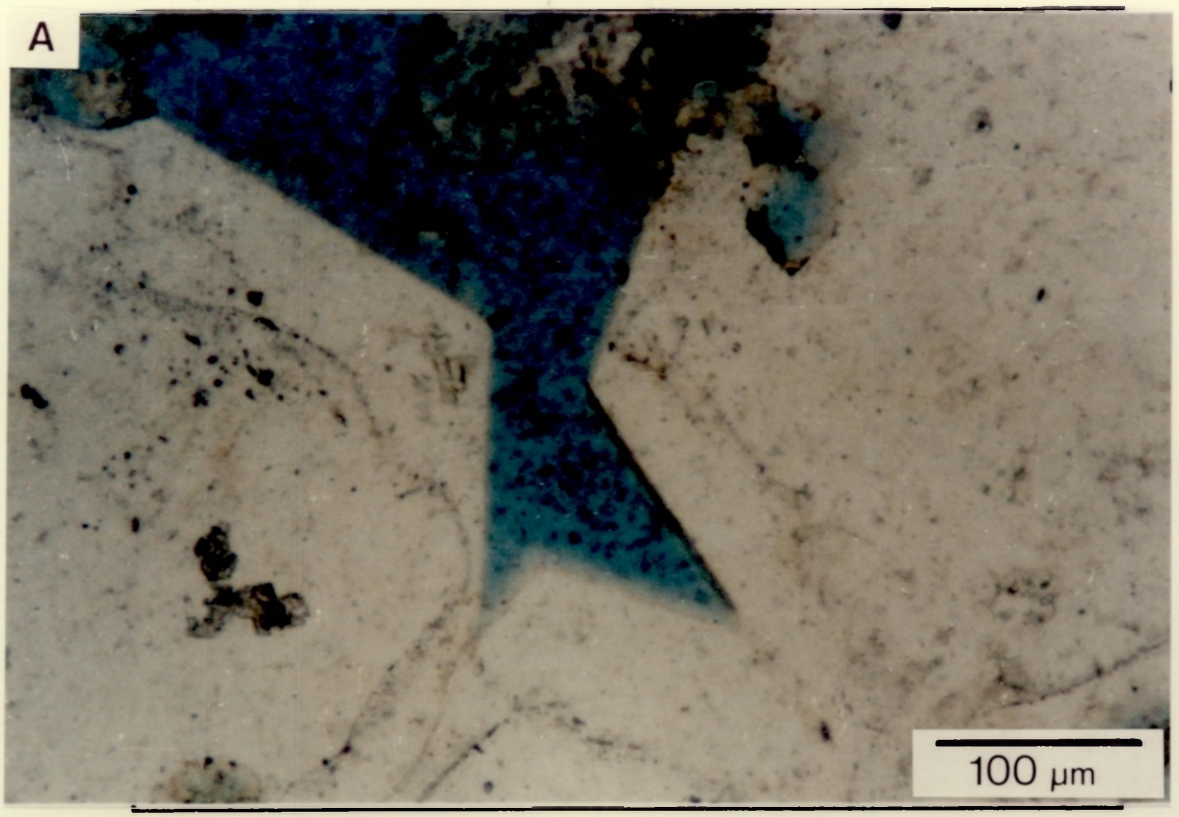
Well developed syntaxial quartz overgrowths occluding pore space. Note incomplete "Dust rims" on detrital surfaces.

B

Moss Farm, 1010m. Pule Hill Grit, SEI

SEI view of quartz overgrowths. Note included kaolinite books and zoned appearance of overgrowths which probably reflects variations in growth rate. Textural evidence suggests that quartz overgrowth was taking place from early diagenesis to late burial.

Plate 7:1



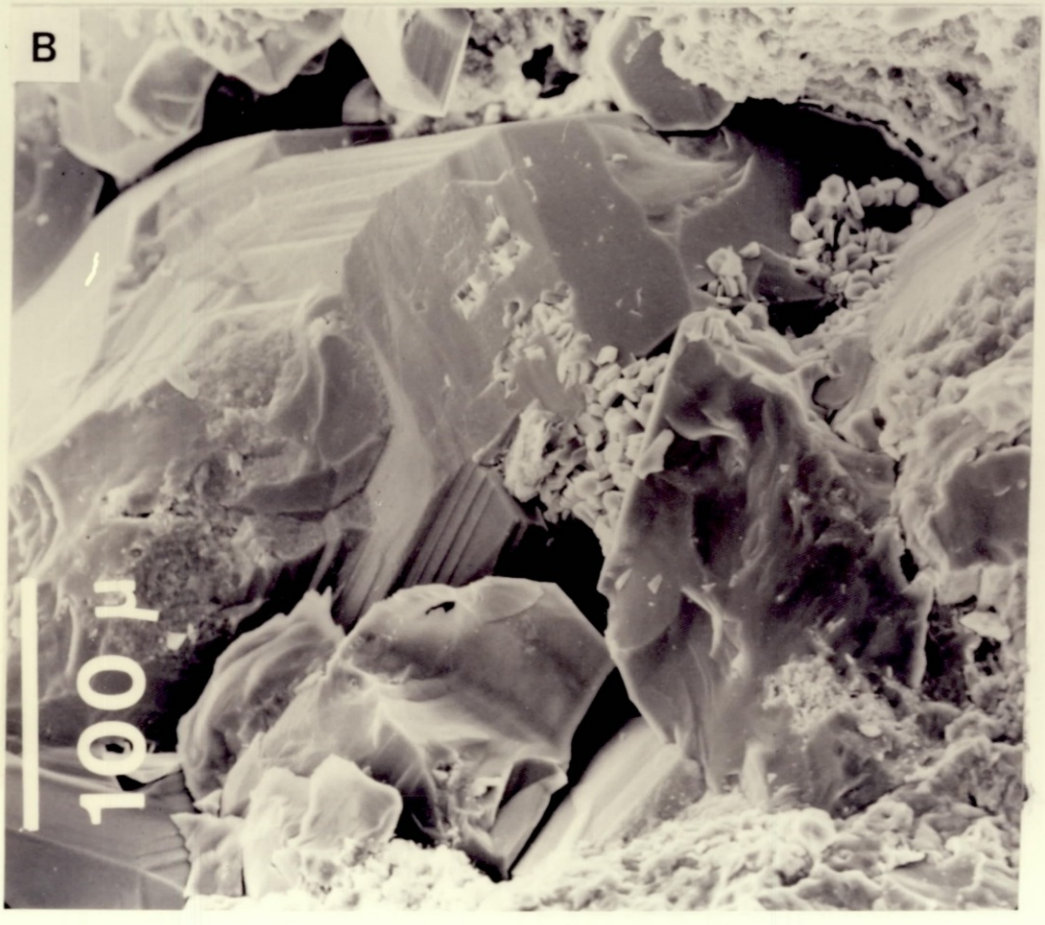


PLATE 7:2

A

Trumfleet 1, 3100ft Rough Rock, SEI

Stubby doubly terminated quartz crystals in pore space. This form is not as common as the syntaxial overgrowth observed in the bottom third of the plate.

B

Trumfleet 1, 3397ft. Beacon Hill Flags. XPL

Fully interlocking quartz overgrowths which have completely destroyed porosity. Concavo-convex contacts are formed by the overgrowths with only minor contributions from pressure dissolution. Very well developed overgrowths of this type only form where feldspars and clays are absent, and are only localised in their distribution.


A
25 26 2
10 µm

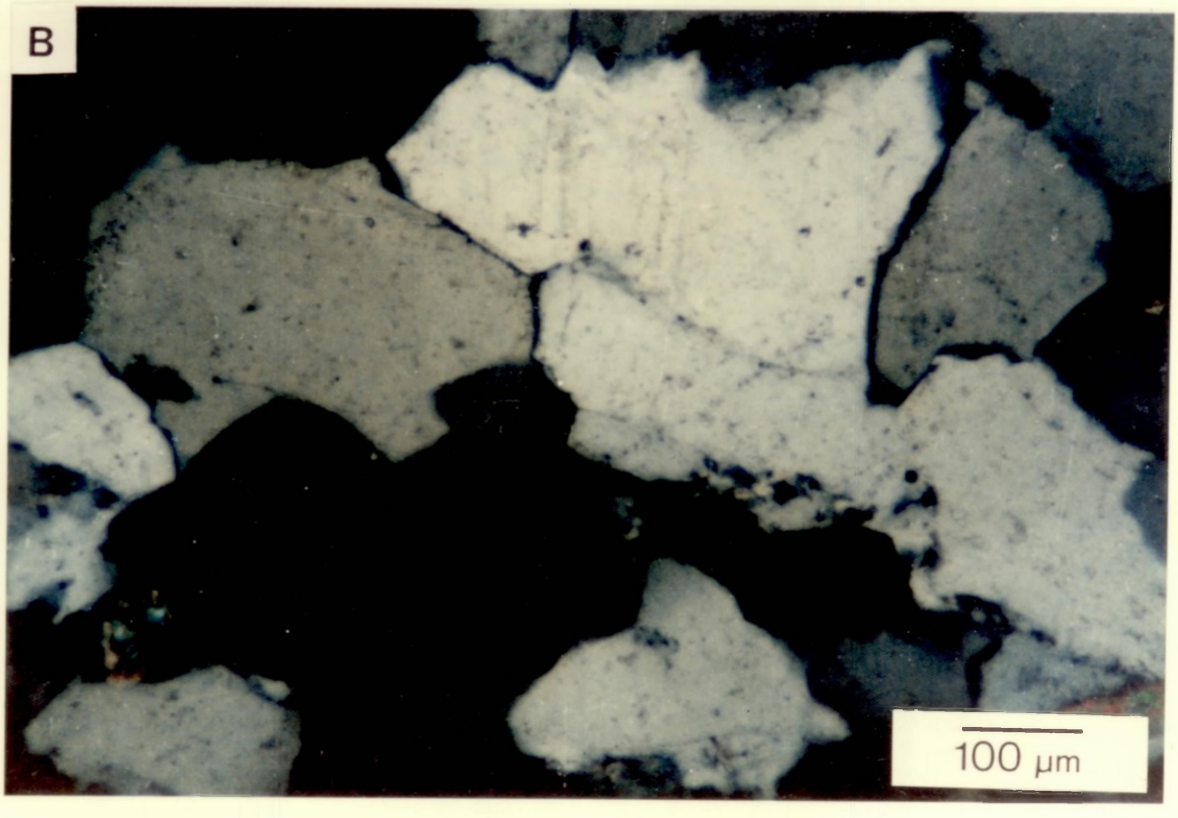
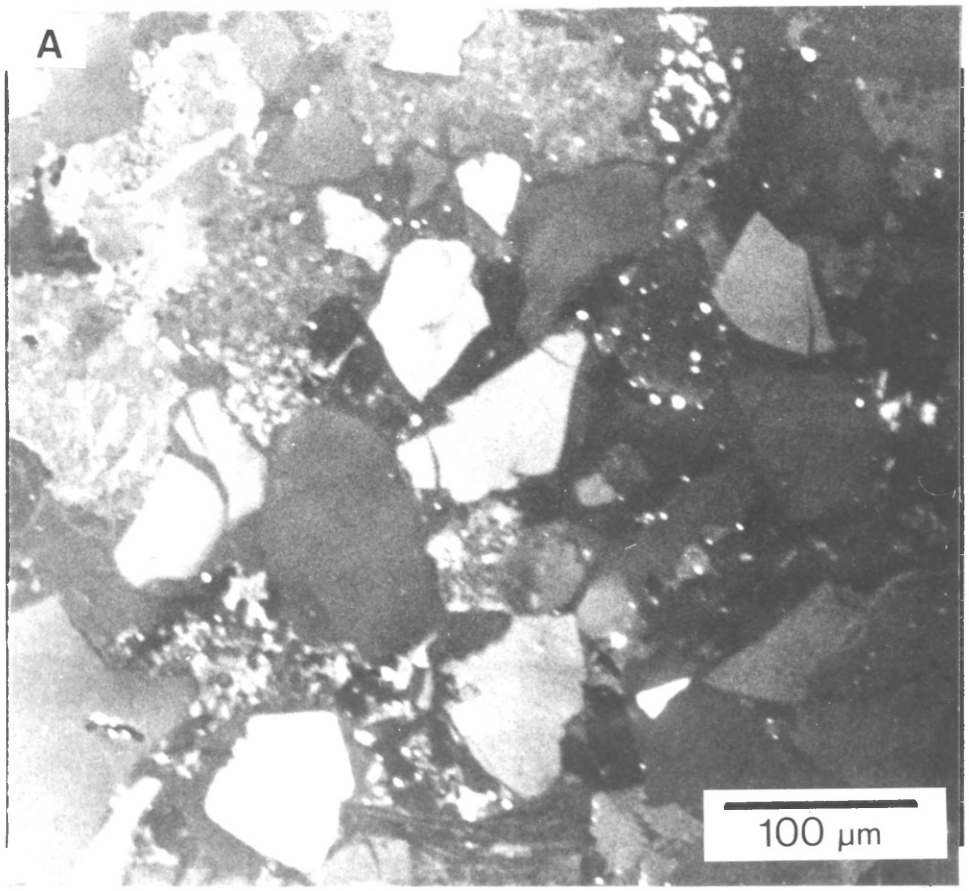

B
100 µm

PLATE 7:3

A AND B

Moss Farm, Pule Hill Grit, 1005m. CL

Cathodoluminescence photomicrograph showing detrital cores (light grey to black), and authigenic overgrowths which show no luminescence and therefore appear black. Note healed fractures in quartz grains and lack of significant detrital core interpenetration which would indicate that pressure dissolution had occurred.


A
100 µm

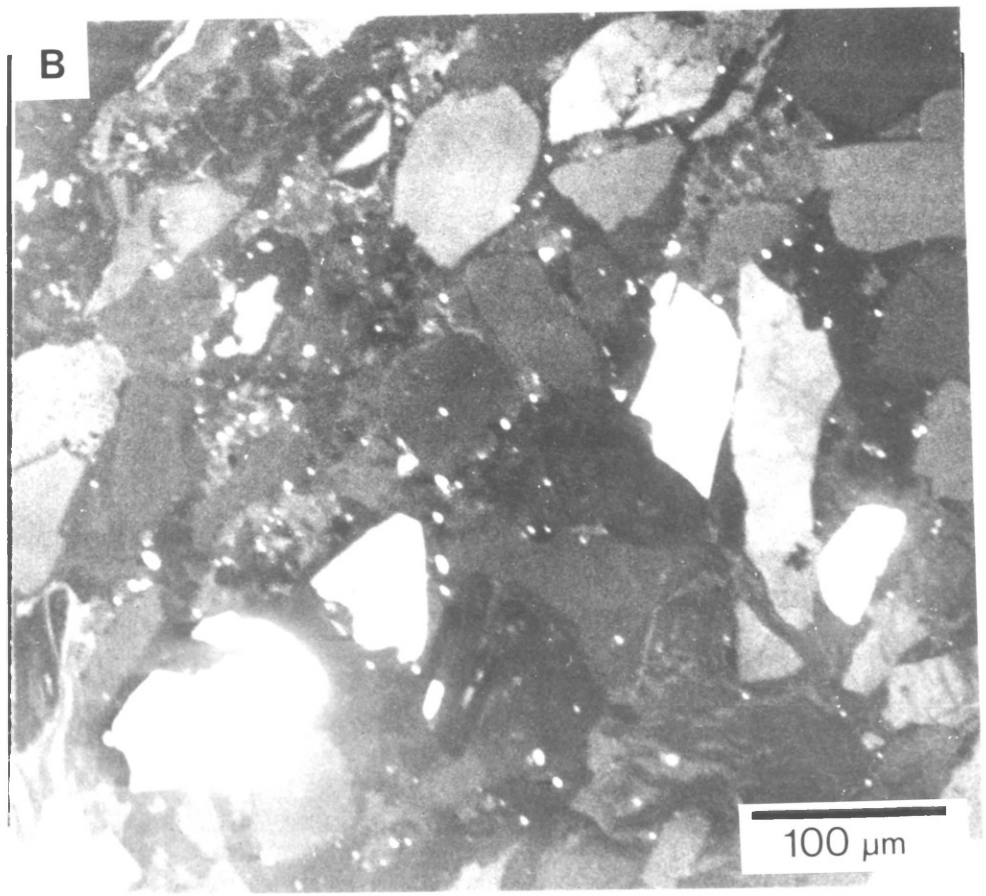

B
100 µm

PLATE 7:4

A

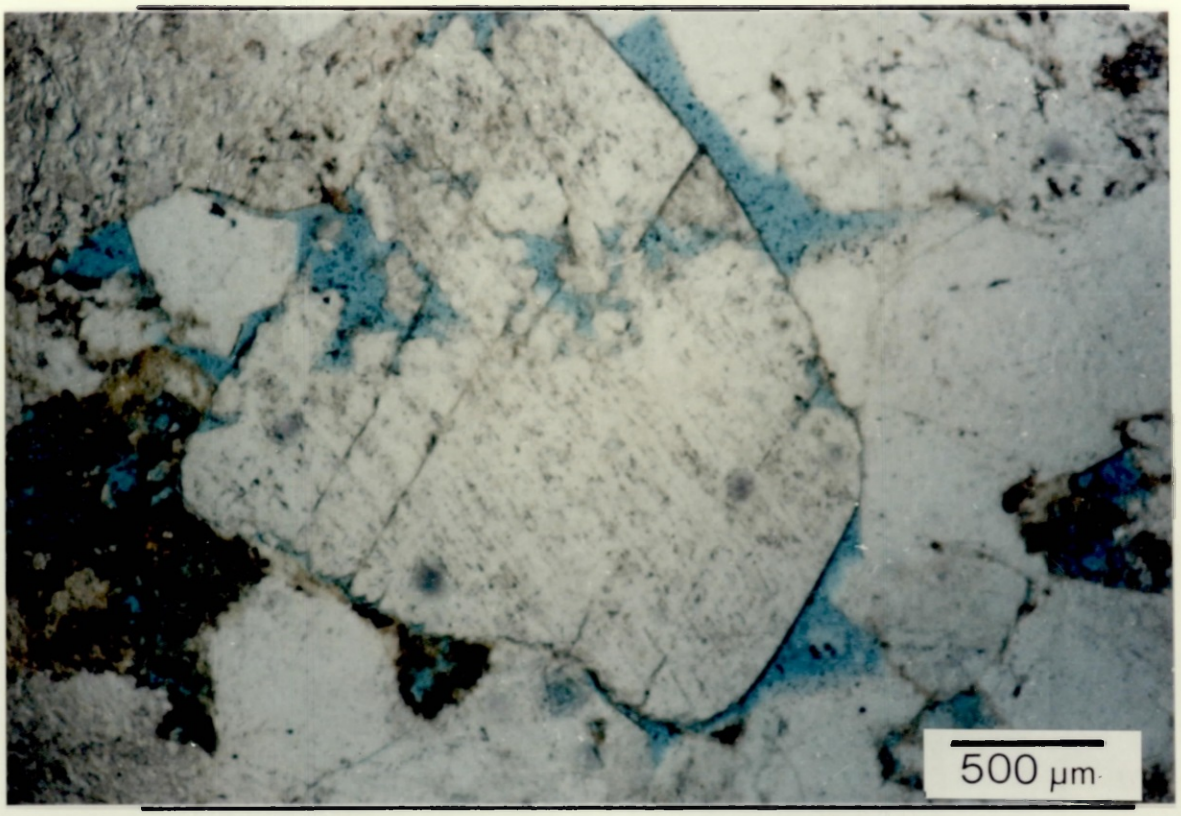
Trumfleet 1, 3455 ft Pule Hill Grit. PPL

Partial dissolution of feldspar forming secondary porosity. Such porosity is common but is unlikely to improve reservoir characteristics significantly, since the individual feldspar grains are not in communication.

B

Trumfleet 1, 3475 ft Pule Hill Grit PPL

Partial dissolution of feldspar feldspar. Note brown material (lllitic clays) in secondary pores. Feldspar dissolution is often preferentially orientated along crystalographic planes, and in some cases the grain is totally removed, leaving only a network of orientated sericite partially filling the pore space.


500 µm

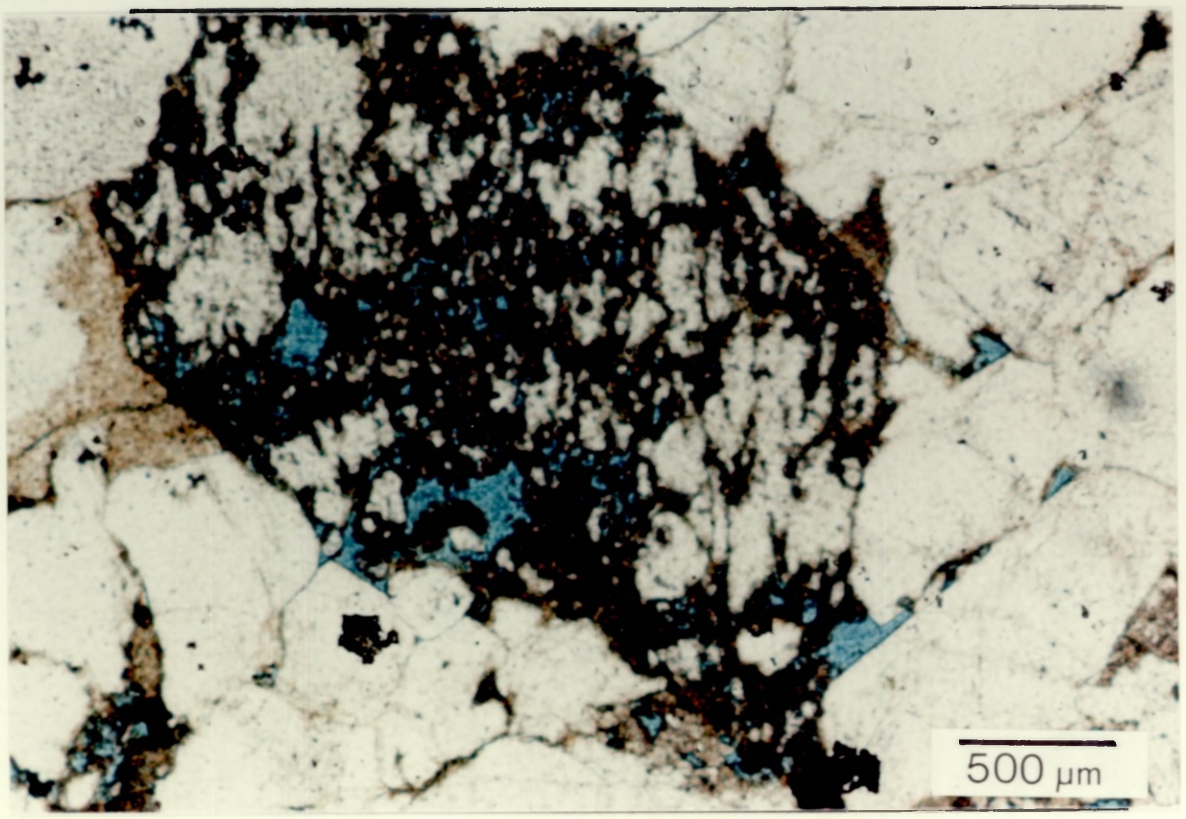

500 µm

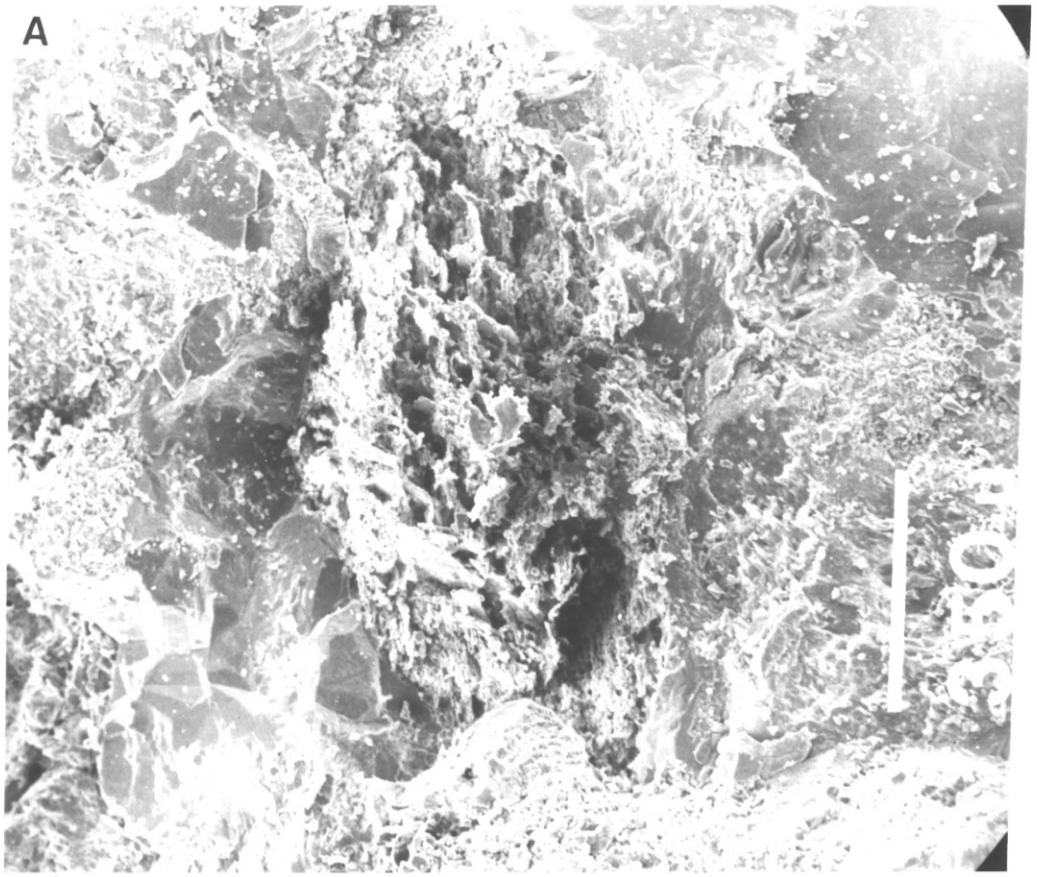
PLATE 7:5

Moss Farm. Pule Hill Grit, 1005m. SEI.

Partially dissolved feldspar grain. Such delicate structures could not have survied the compactional events which fractured quartz and plagioclase grains as well as bending micas. The dissolution event which reduced the grain to this state is clearly therefore post compactional

Moss Farm, Pule Hill Grit, 1014m SEI

Feldspar showing lesser dissolution than the above example. The degree of dissolution suffered by feldspars is highly variable.


A
10 µm

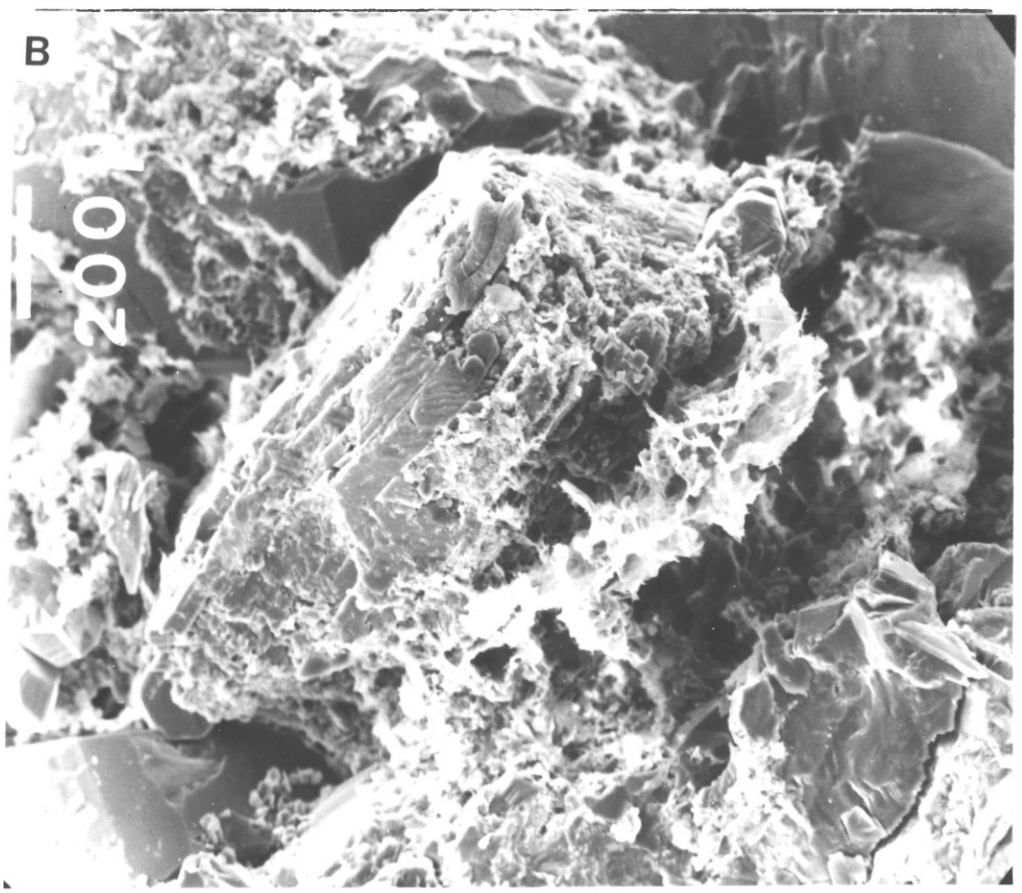

B
200

PLATE 7:6

A

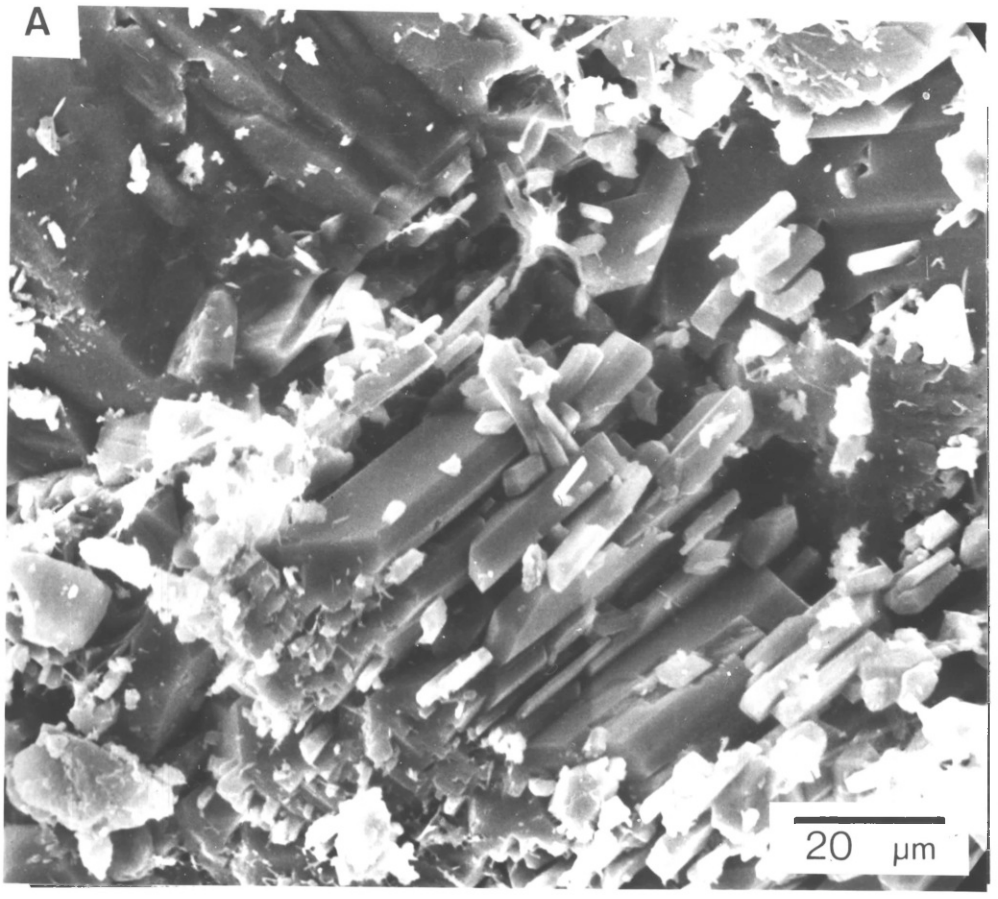
Moss Farm, 1017m. Pule Hill Grit. SEI

Rod shaped authigenic albite crystals which have overgrown partially dissolved feldspars.

B

Moss Farm, 924m. Rough Rock. SEI

Doubly terminated albite crystals associated with illite in pore spaces, The relationship between the two is difficult to discern but as the albite only occurs near the edge of the illite cemented area it is probable that albite authigenesis postdates the formation of illite.


A
20 µm

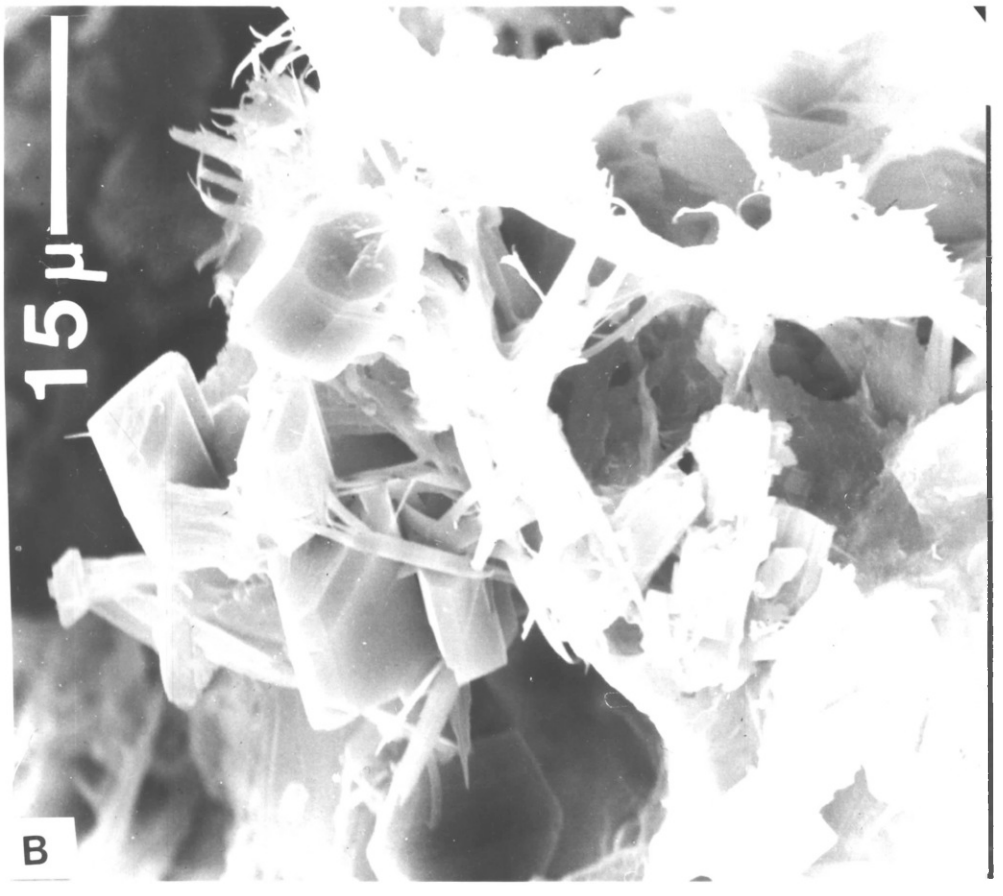

15µ
B

PLATE 7:7

A

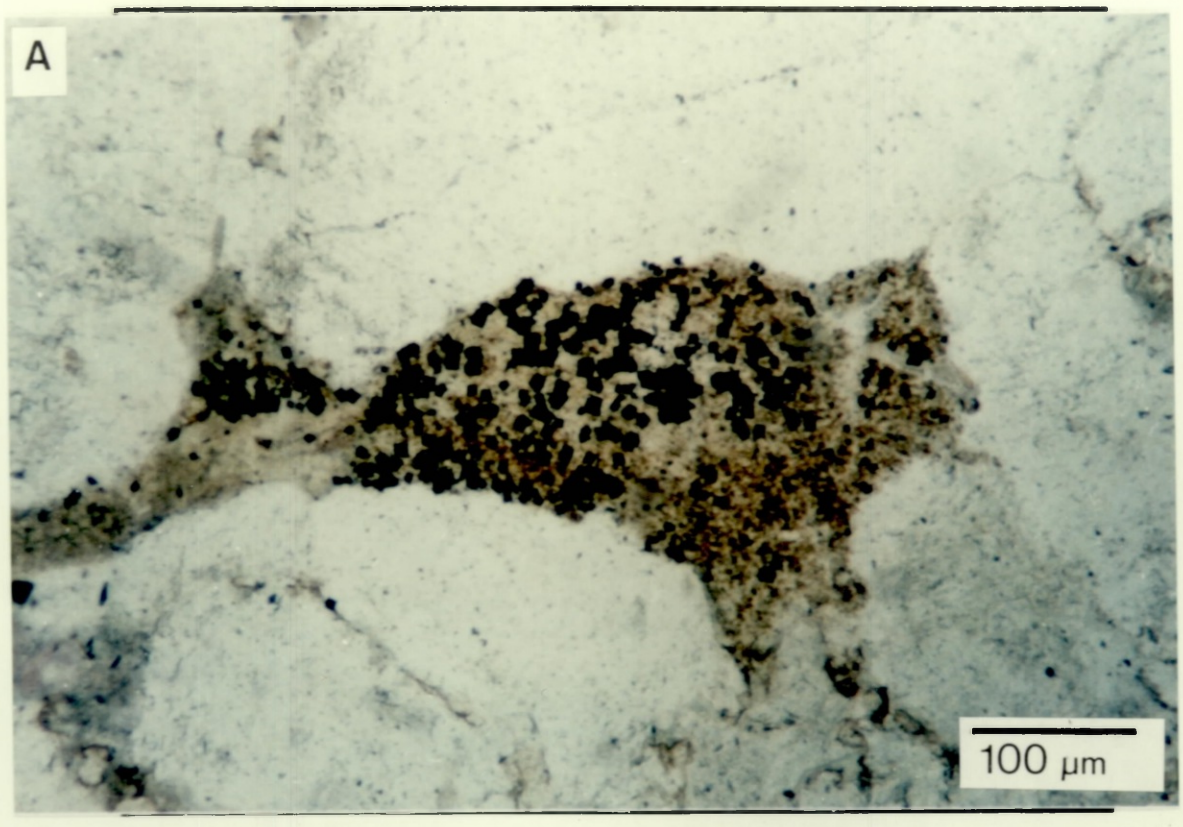
Askern, 3357 ft. Pule Hill Grit PPL

Early diagenetic pyrite overgrown by microcrystalline siderite. Pyrite formed under reducing conditions using available sulphate, when the source of sulphate is exhausted, siderite precipitates in its place, using available Fe^{2+}

B

Moss Farm, 922.65m. Rough Rock. PPL

Microcrystalline siderite cemented mudstone clast Such clasts are often 3-4 times the grain diameter and cementation has prevented the formation of pseudomatrix during compaction.


A
100 µm

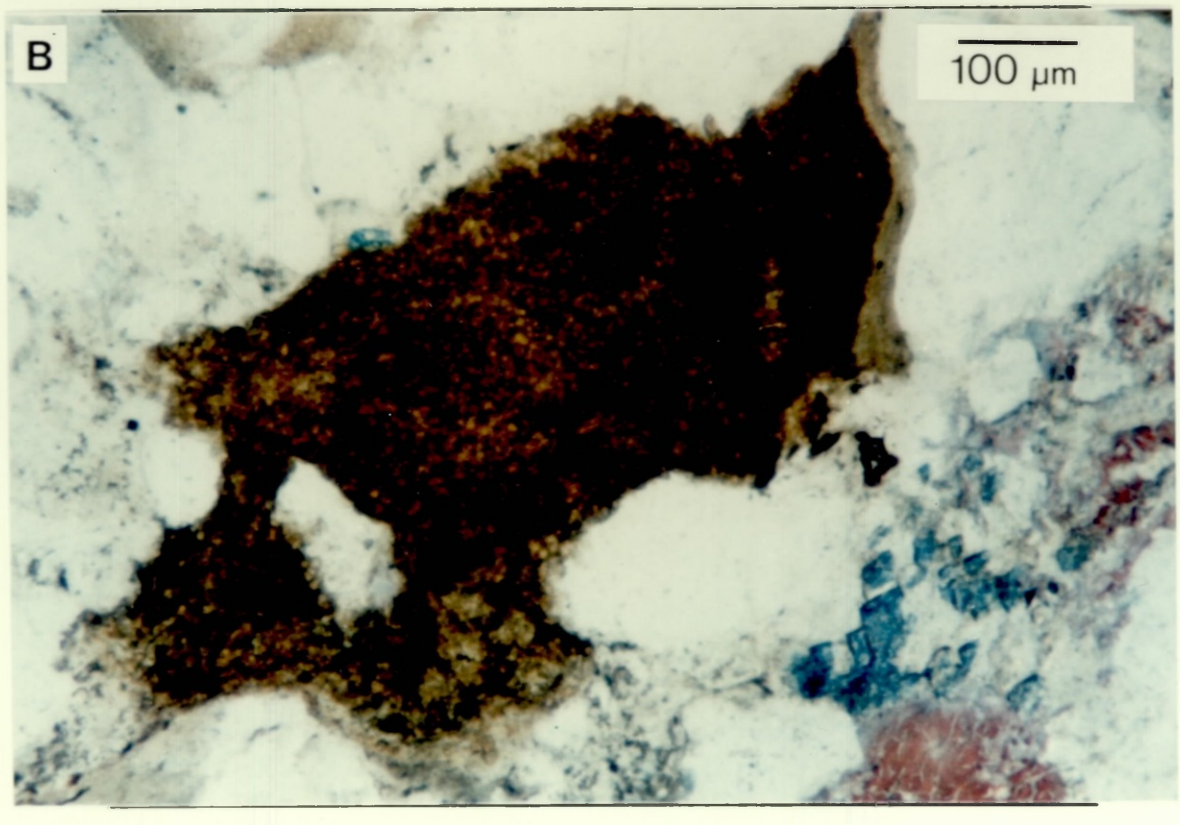

B
100 µm

PLATE 7:8

A

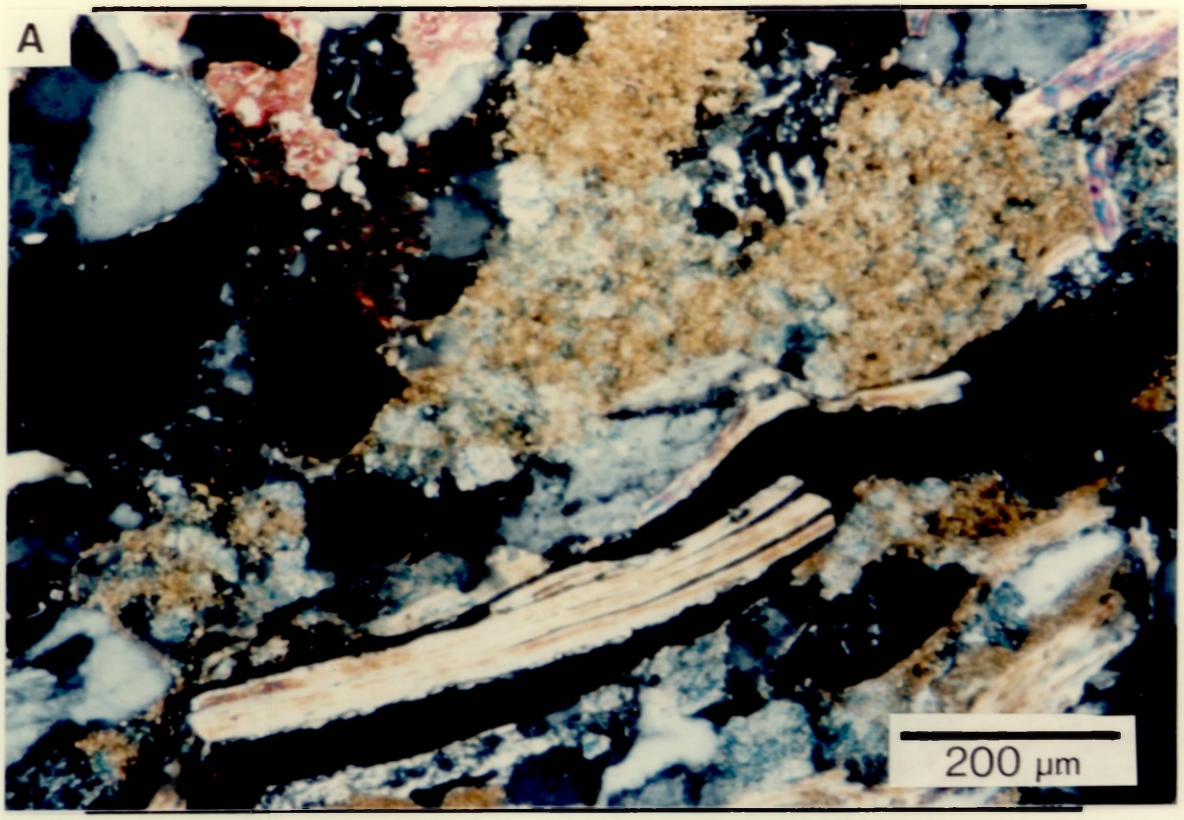
Moss Farm, 929m. Rough Rock. XPL

Micaceous/argillaceous rich horizon showing the association of organic matter (opaque, surrounding mica) with siderite and ankerite.

B

Moss Farm, 918m. Rough Rock PPL

Sphaerosiderite nodule in overbank sandstone.


A
200 µm

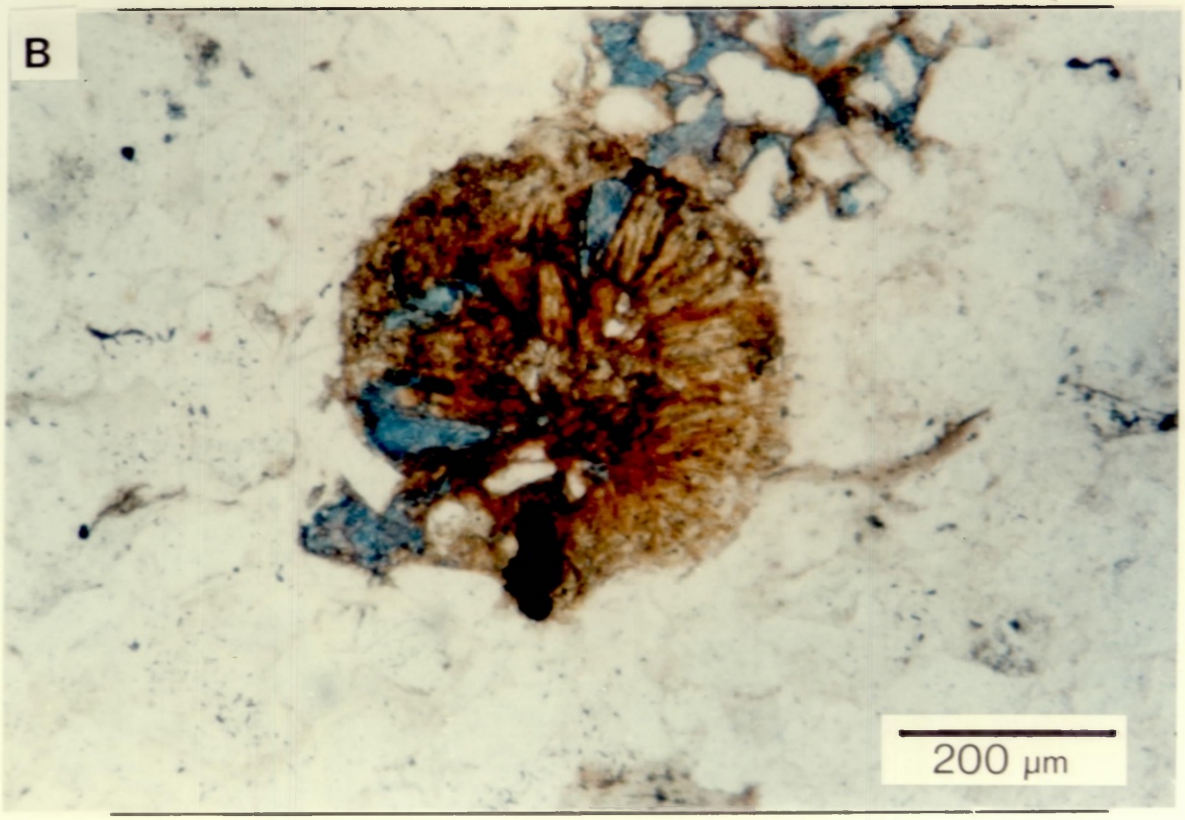

B
200 µm

PLATE 7:9

A+B

Moss Farm, 918m. Rough Rock. BSE

Sphaerosiderite nodules in overbank sandstones. These nodules formed during early diagenesis and often contain inclusions of calcite and clay minerals. Often pores are preserved in the centres, but these are often cemented by late stage cements such as ankerite.

C

Moss Farm, 929m. Rough Rock. BSE

Poikilotopic calcite cemented sandstone. Note angularity of quartz grains and occasional embayments caused by etching as well as the zoned nature of the calcite.

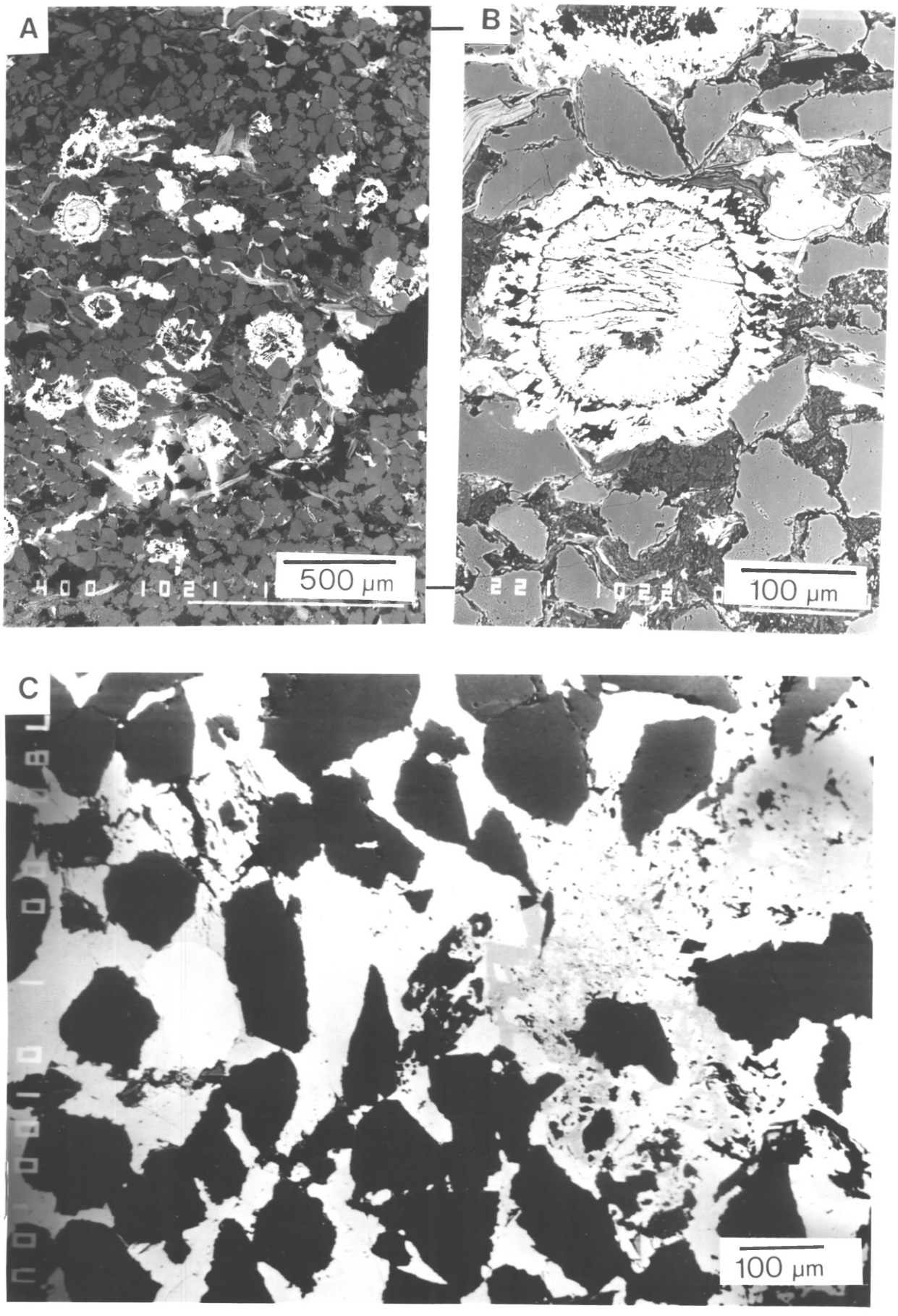

A
500 µm
B
100 µm
C
100 µm

PLATE 7:10

A

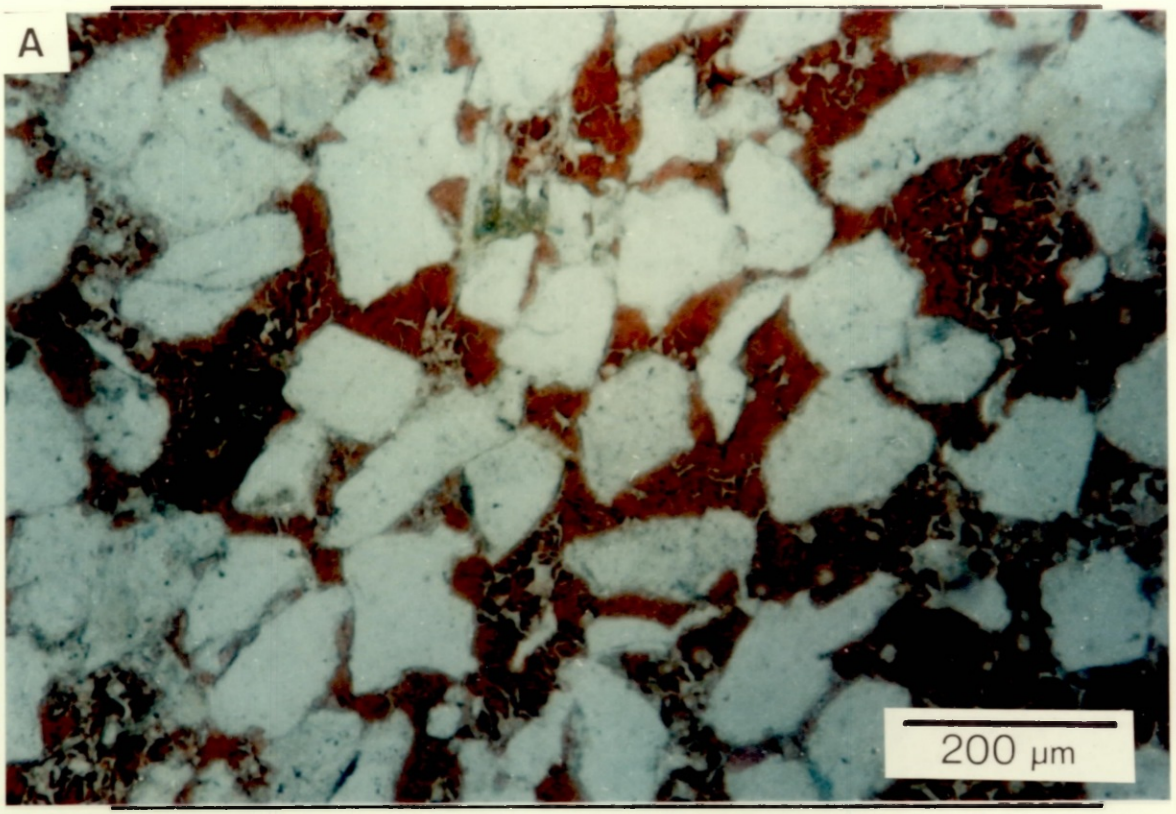
Moss Farm, 929m. Rough Rock. PPL

Poikilotopic calcite cement. Note open, uncompacted, texture of framework grains and partial replacement of feldspar (upper centre). Such cements occur locally and are usually observed near the tops and bases of the channel sandstones

B

Trumfleet 2, 3321ft. Beacon Hill Flags. XPL

Close up of calcite cement showing preserved incipient quartz overgrowths.


A
200 µm

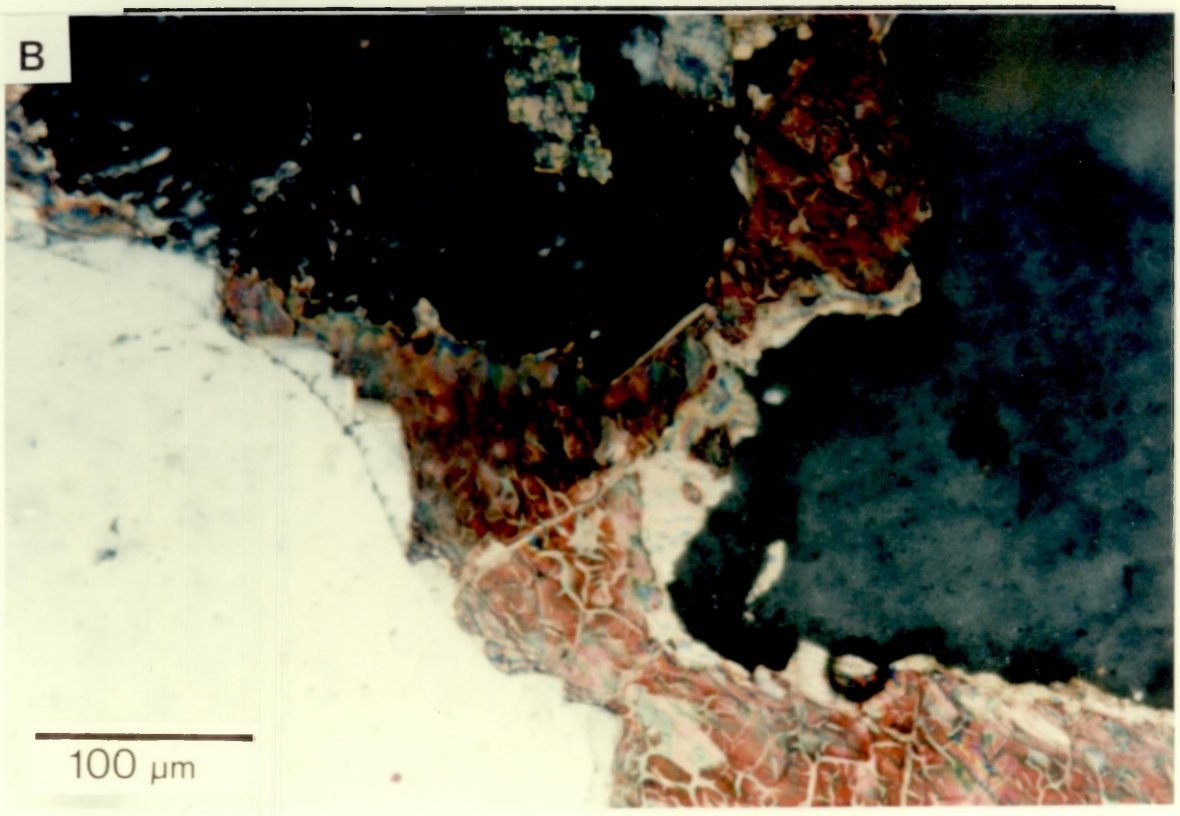

B
100 µm

PLATE 7:11

A

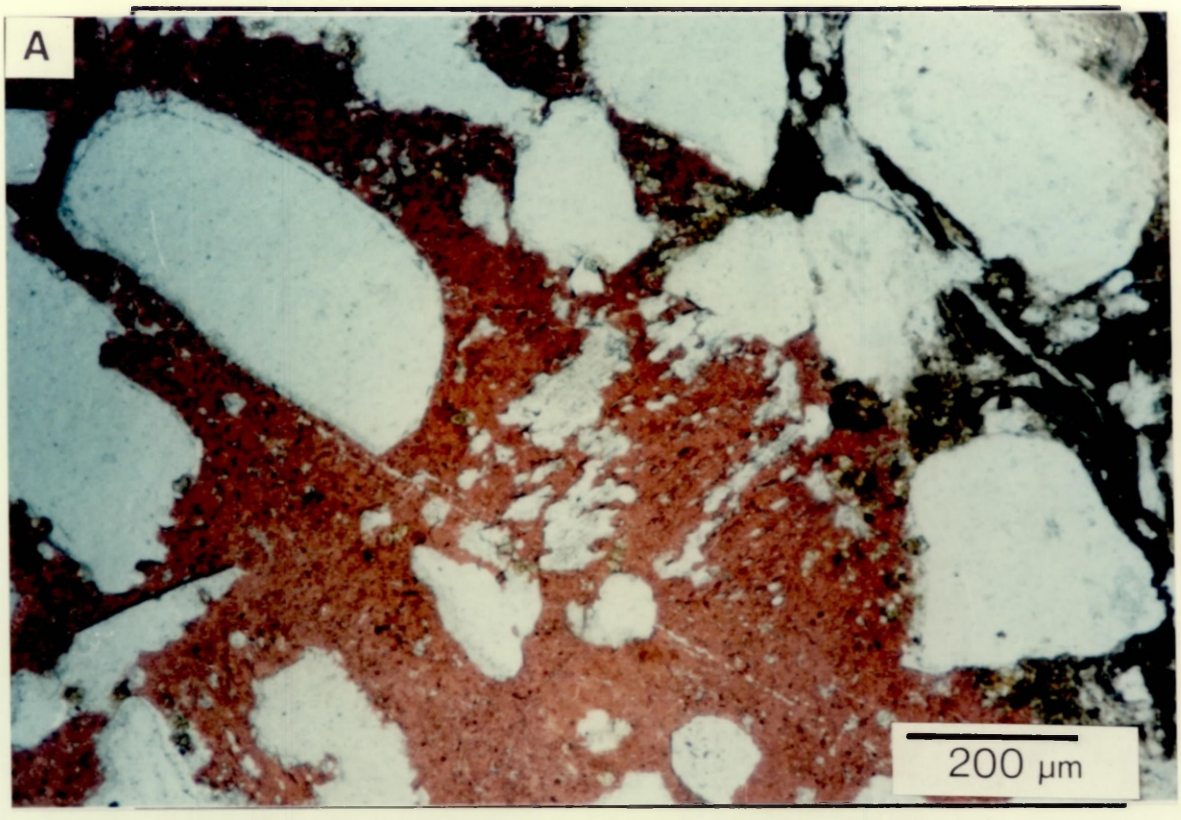
Trumfleet 2, 3316ft. Beacon Hill Flags PPL

Poikilotopic calcite cement replacing feldspar grain, note quartz overgrowth preserved under cement on quartz grain in the upper left of the photomicrograph.

B

Trumfleet 2, 3316ft. Beacon Hill Flags XPL

Large oversized "pore" with twinned calcite crystal infill. This may have formed by the wholesale replacement of feldspar. Note also scattered rhombs of siderite.


A
200 µm

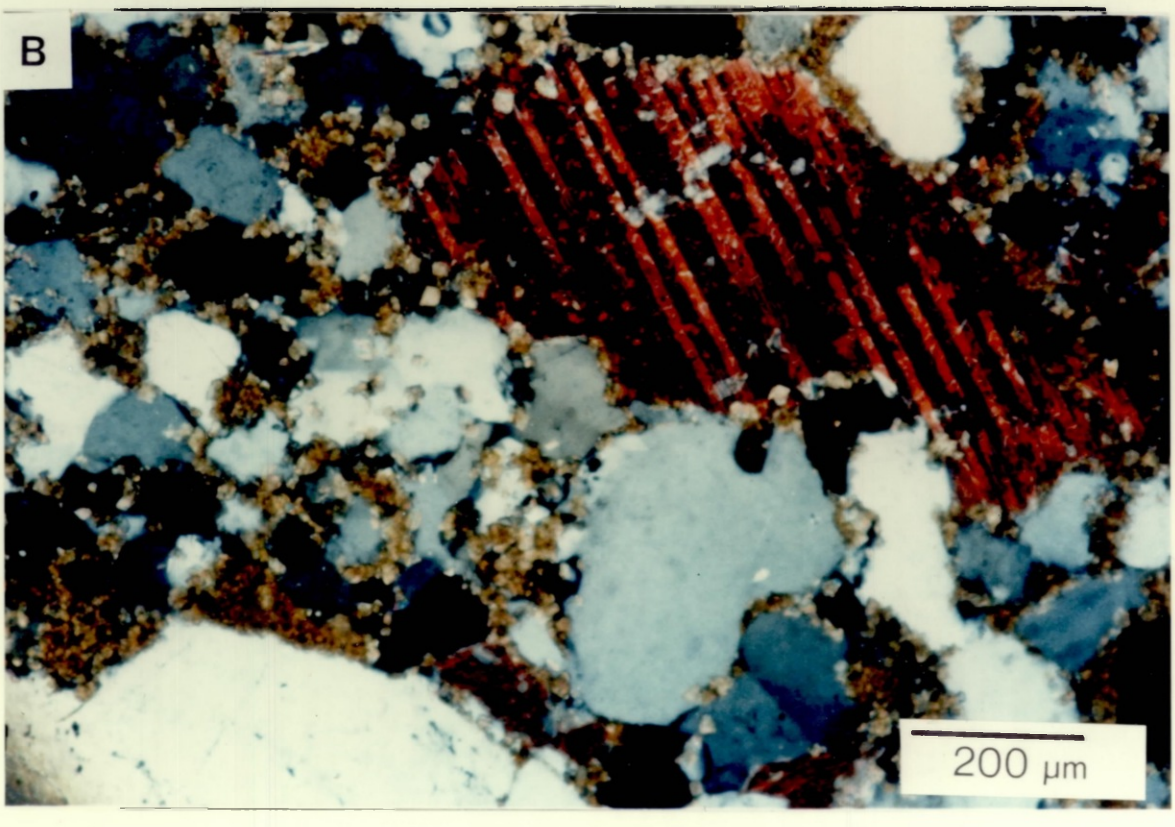

B
200 µm

PLATE 7:12

A

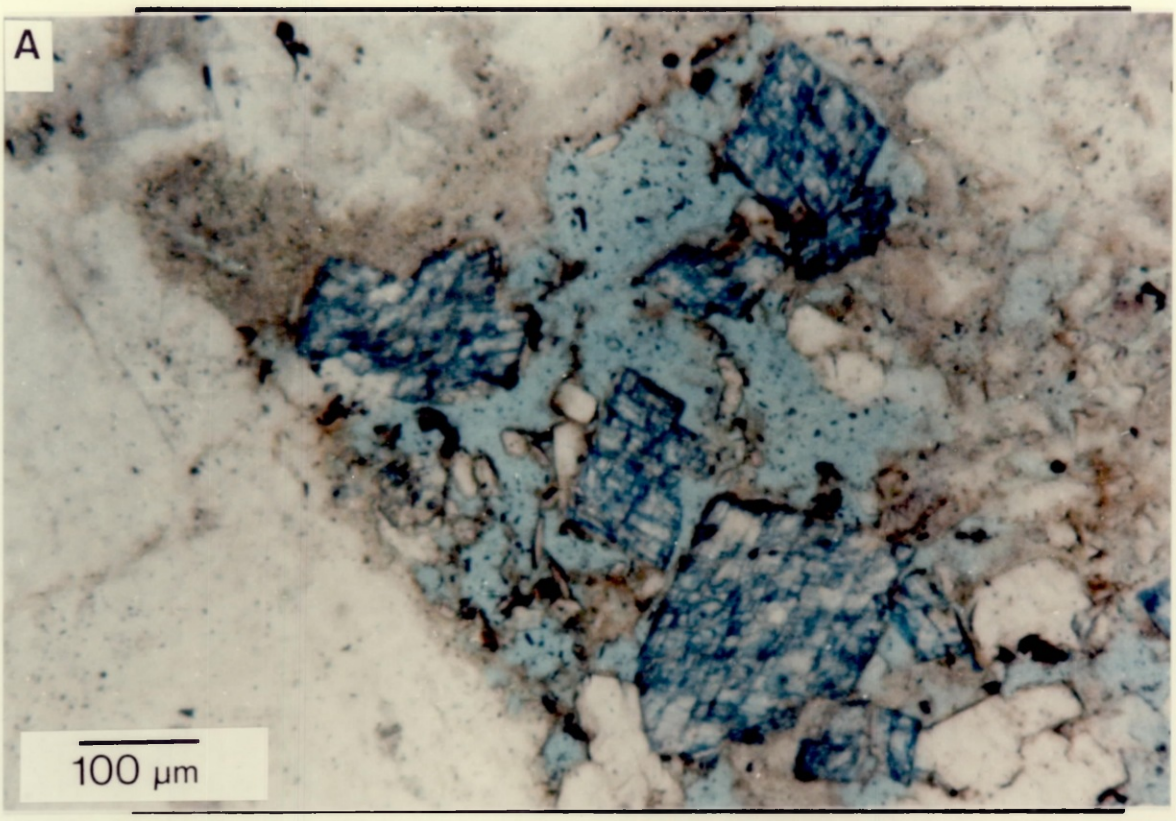
Trumfleet 2, 3405ft. Pule Hill Grit. PPL

Late stage pore filling ankerite rhombs.

B

Moss Farm, 929m. Rough Rock PPL

Microcrystalline ankerite cement associated with siderite and calcite.


A
100 µm

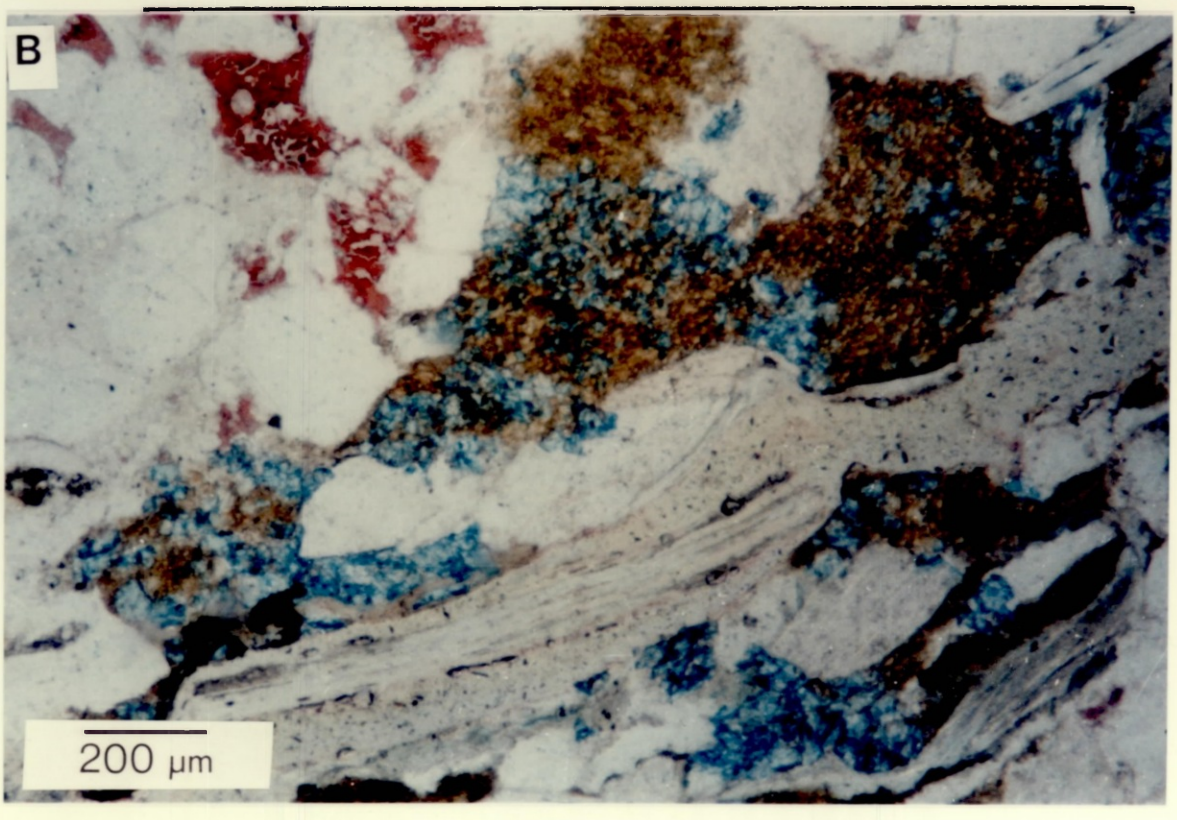

B
200 µm

PLATE 7:13

A

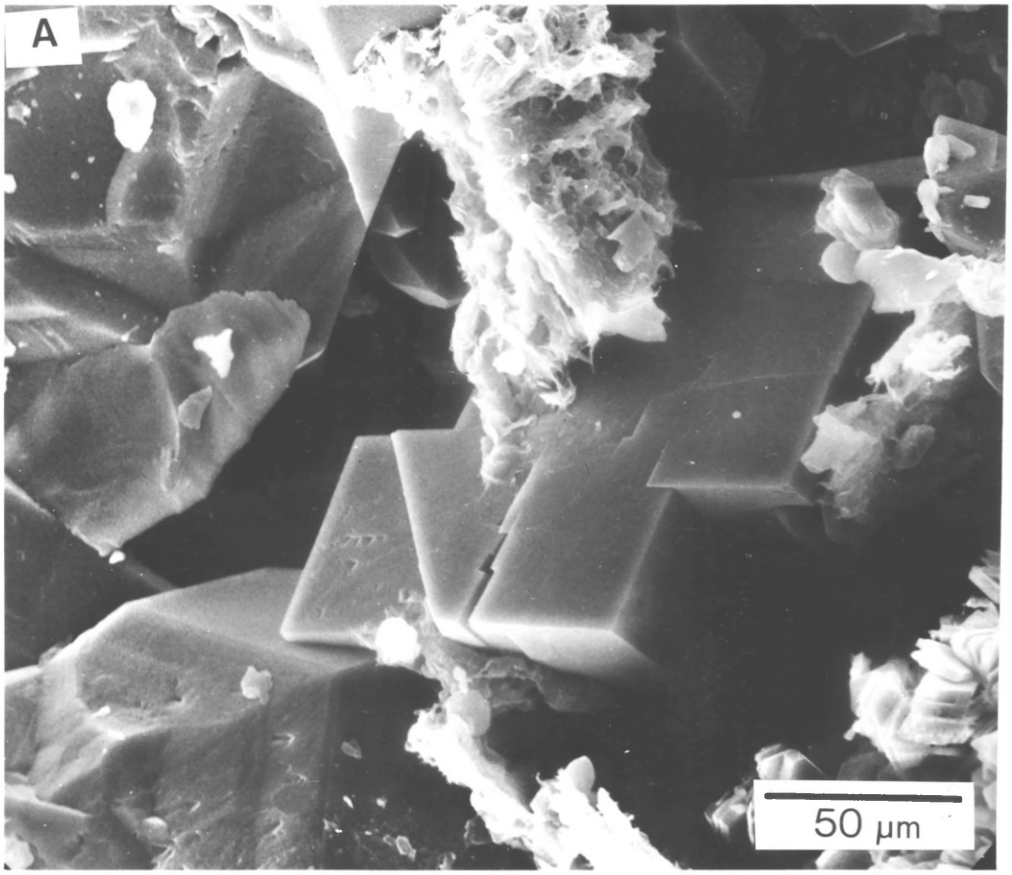
Moss Farm, 1021m. Pule Hill Grit. SEI

Isolated pore filling rhomb of ankerite. Such cements formed late in association with albite and illite.

B

Moss Farm, 927m. Rough Rock, BSE

Poikilotopic calcite cement in the process of being dolomitised and altering to ankerite. Darker grey ankerite rhomb can be seen in bottom left.


A
50 µm

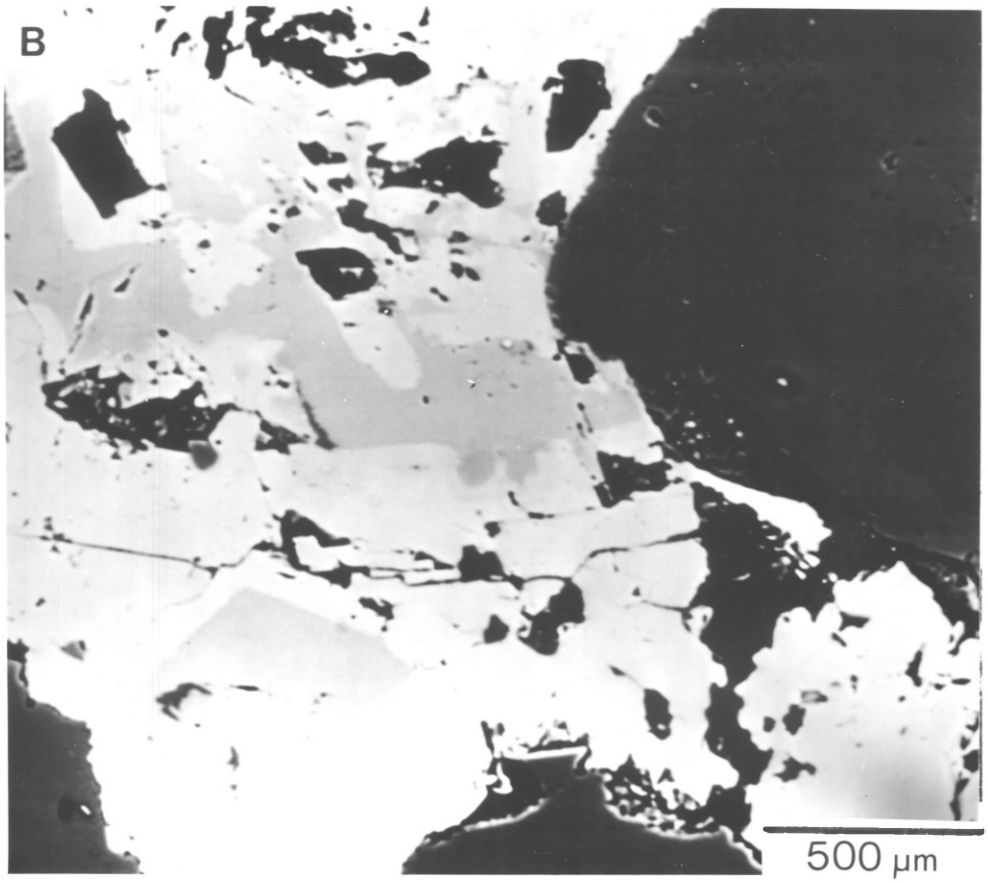

B
500 µm

PLATE 7:14

A

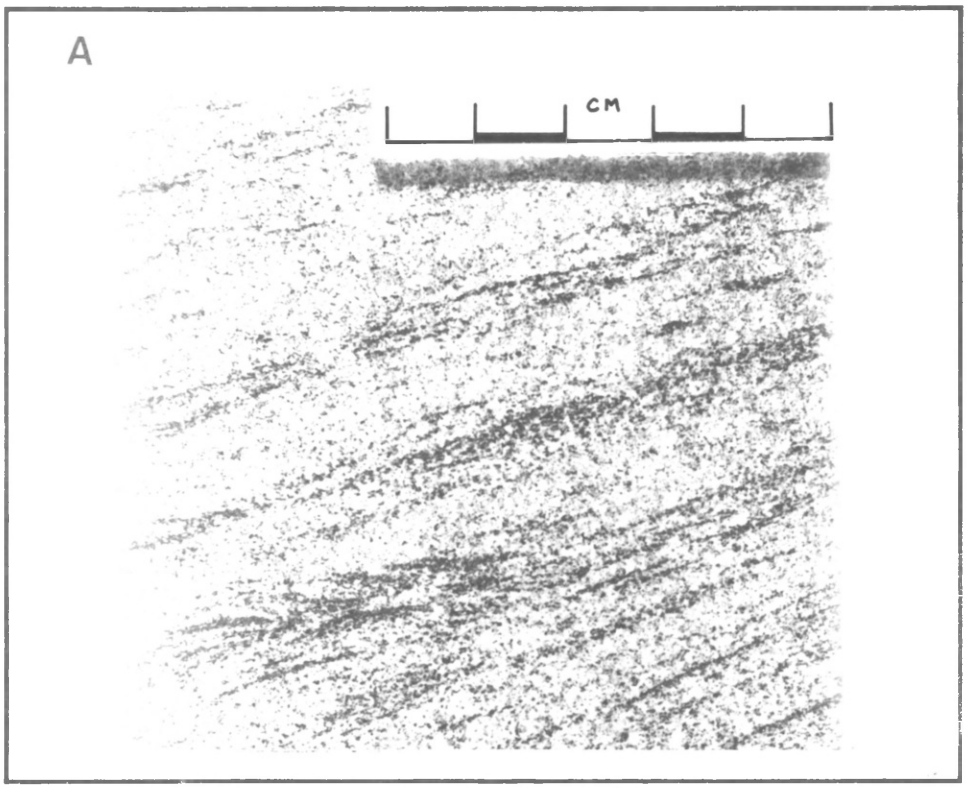
Moss Farm, 924ft. Rough Rock

Photograph of slabbed and stained core showing distribution of ankerite in hand specimen. The entire rock is carbonate cemented. Dark areas contain ankerite cements and lighter areas are calcite cemented. Ankerite cements are concentrated in argillaceous laminae and associated with siderite and organic matter, suggesting a localised source for the iron.

B

Moss Farm, 1014m. Pule Hill Grit SEI

Well crystalised kaolinite books partially overgrown by quartz. It is envisaged that quartz authigenesis was virtually a continuous process, and that the bulk of kaolinite authigenesis occurred after the bulk of the quartz authigenesis.


A
CM

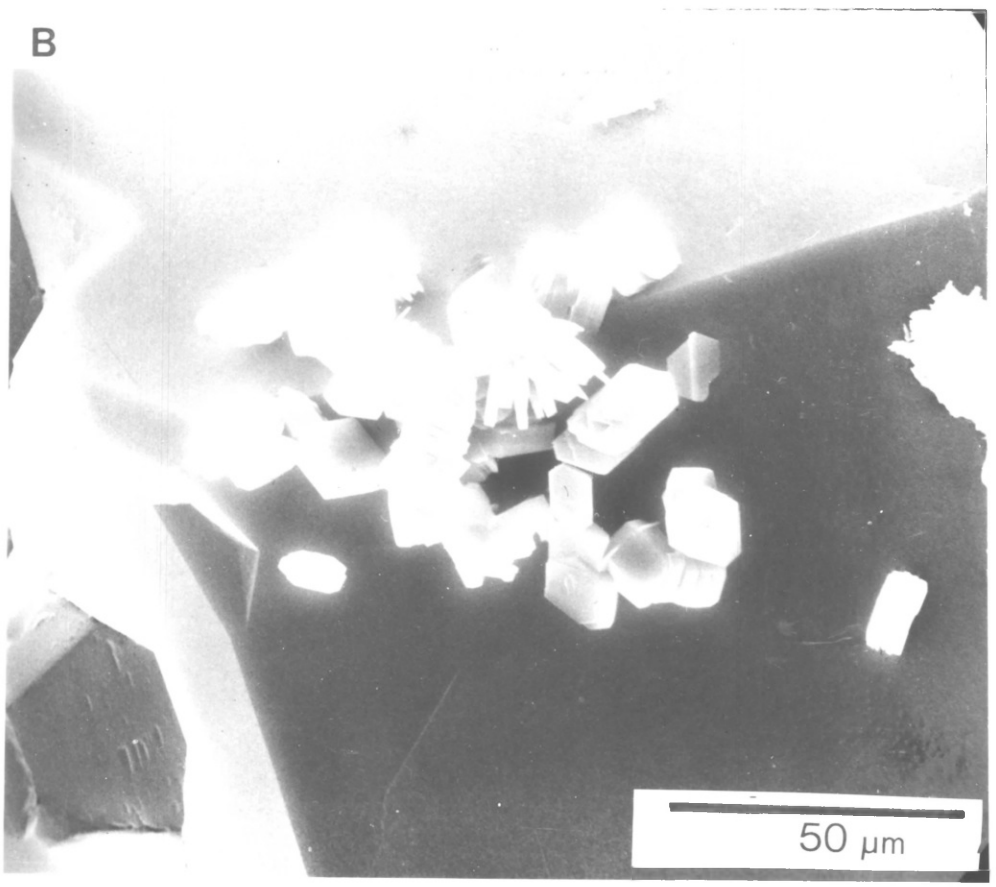

B
50 µm

PLATE 7:15

A

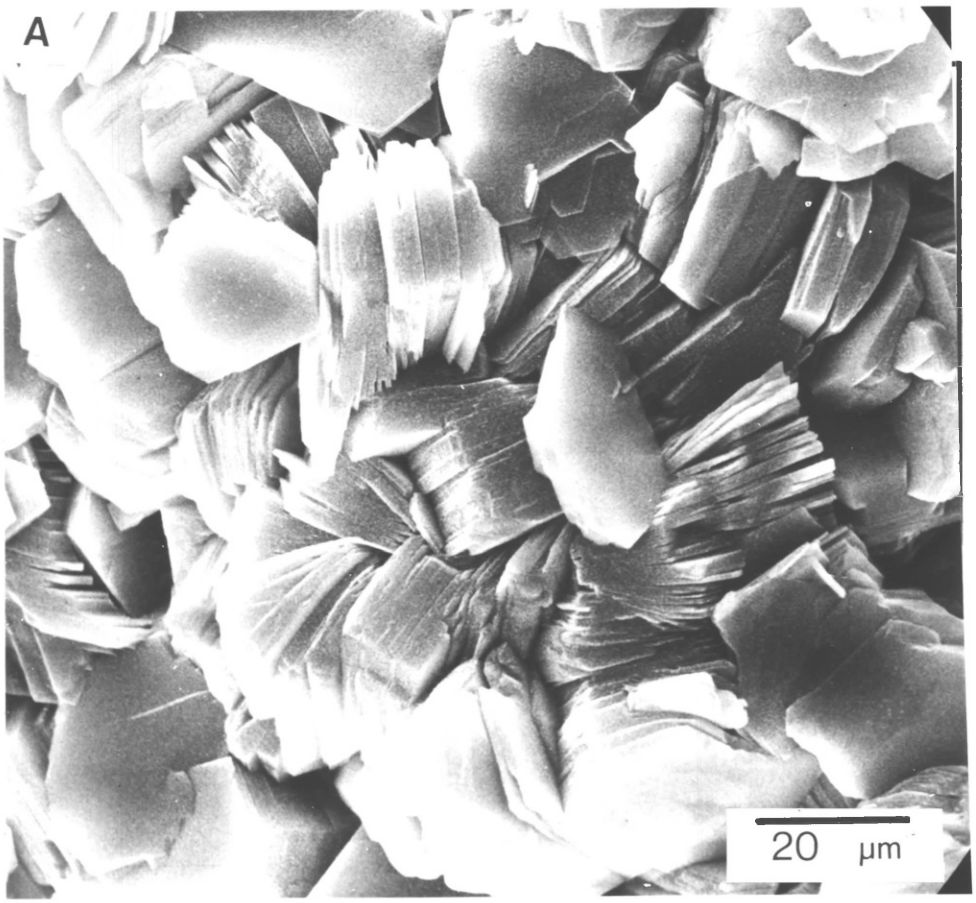
Moss Farm, 1017m. Pule Hill Grit SEI

Tightly packed well crystalised kaolinite books totally occluding porosity in the pore. The packing of kaolinite is highly variable, and some pores show abundant intra and intercrystaline microporosity.

B

Moss Farm, 1005m. Pule Hill Grit

Kaolinite partially overgrown by sheet like and lath like illite. There appears to be little alteration to illite in this case, but kaolinite can show extensive alteration to illite, and fresh kaolinite near the pore centre can coexist with highly illitised kaolinite at the edges of the same pore.


A
20 µm

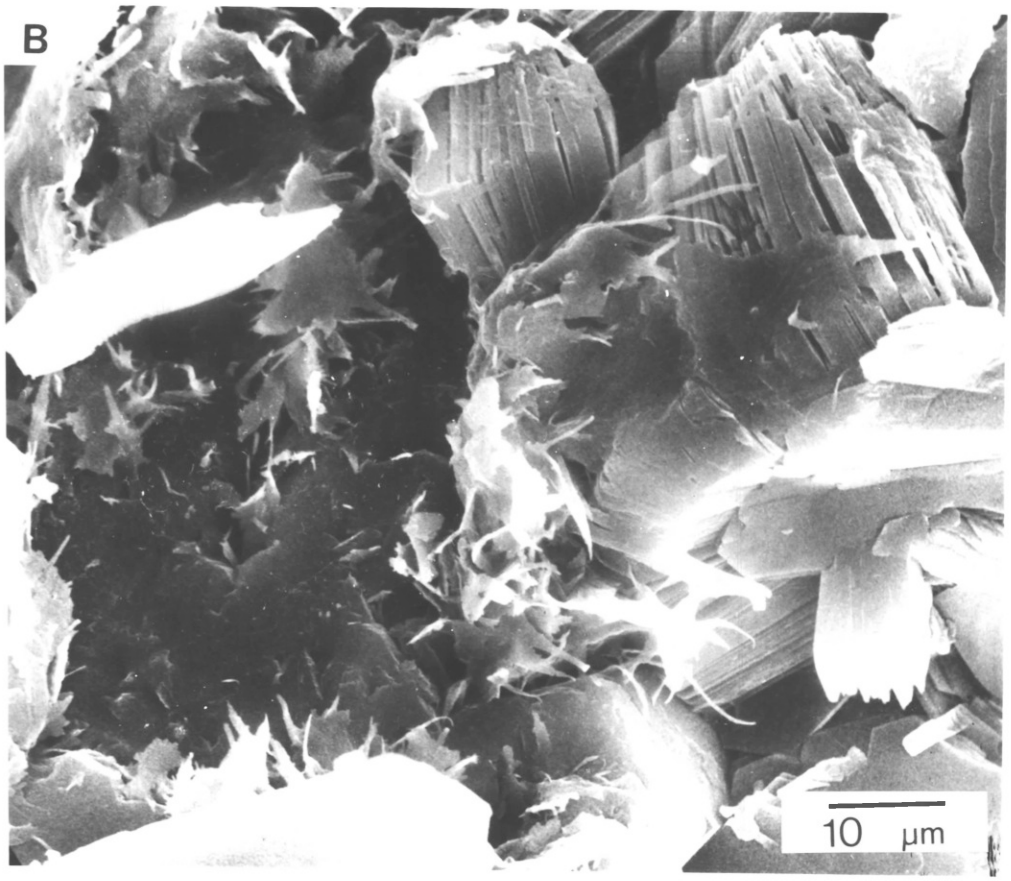

B
10 µm

PLATE 7:16

A

Moss Farm, 1005m. Pule Hill Grit SEI

Well crystalised kaolinite books in pore space, partially overgrown by quartz and illite. Note fresh appearance of individual books.

B

Moss Farm, 1017m. Pule Hill Grit SEI

Vermicular kaolinite showing etched appearance. This type is only found on pore edges and probably predates the better crystallised varieties.


A
25 µ

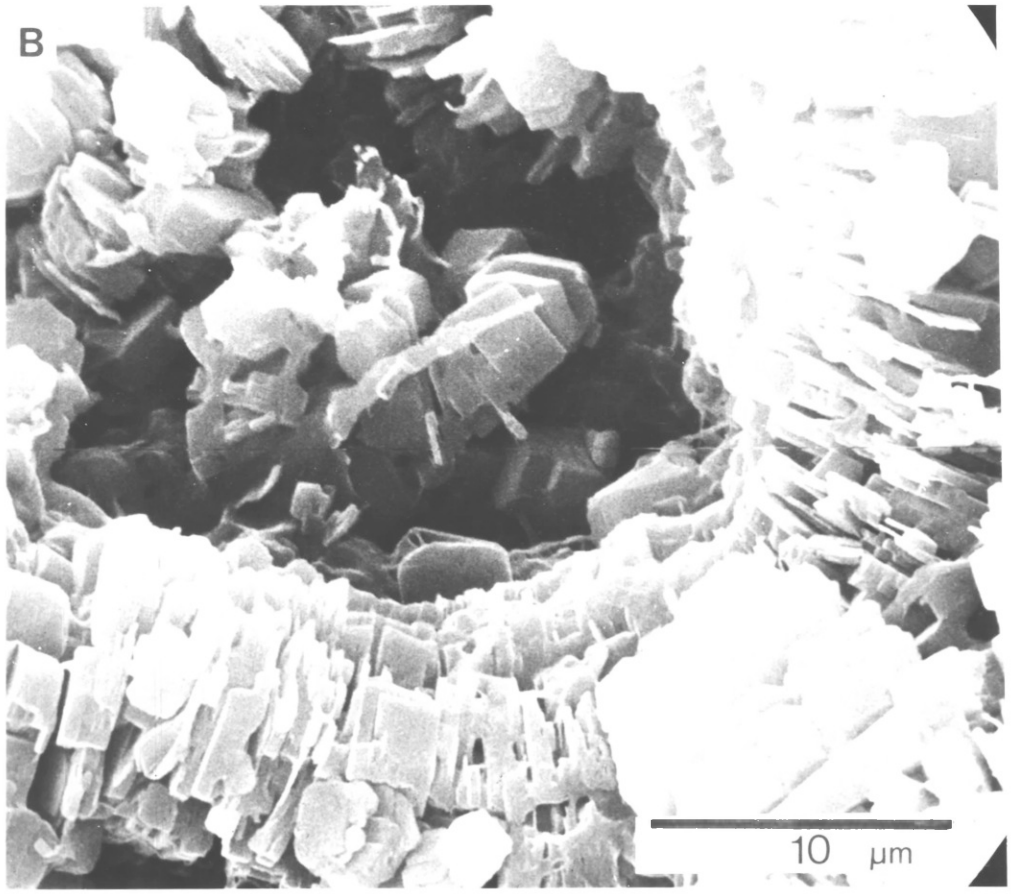

B
10 µm

PLATE 7:17

A

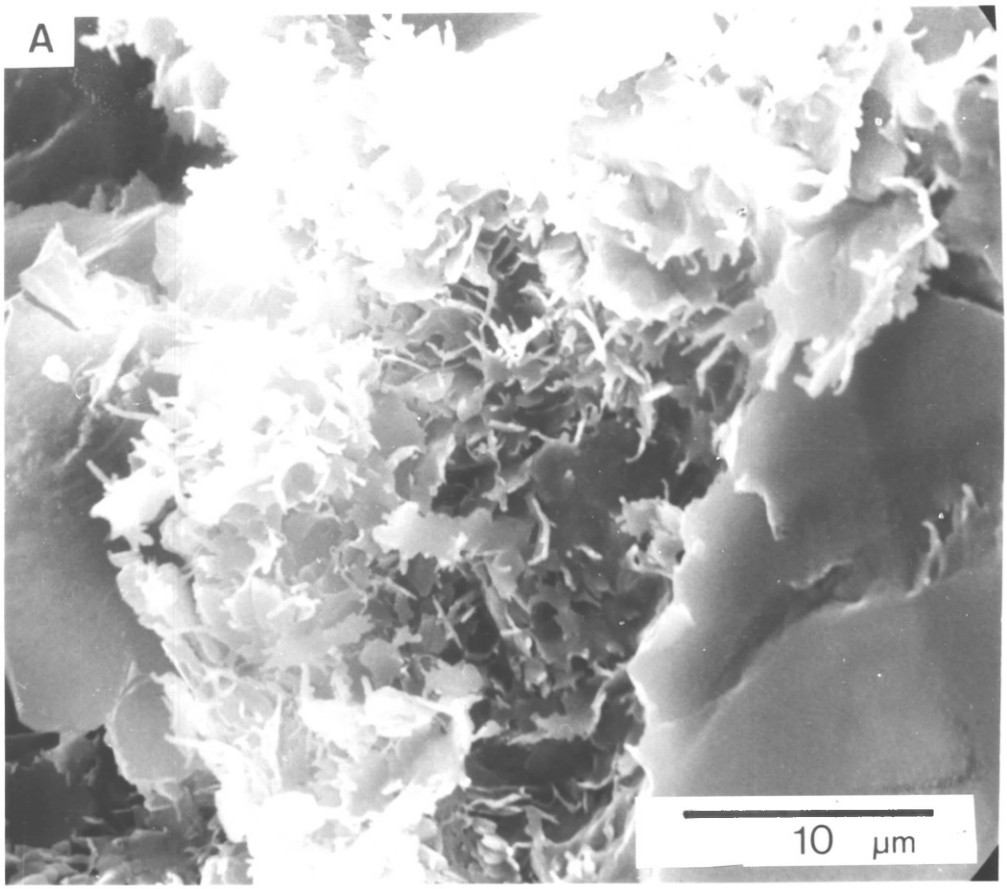
Moss Farm, 924m. Rough Rock. SEI

Bladed and sheet like illite coating pore walls. Illite shows a high variability of morphology reflecting its varied chemistry and mode of formation. Illite of this type may have formed by the aggradation of a detrital precursor.

B

Moss Farm, 1009m. Pule Hill Grit. SEI

Bladed illite in pore centre. Bladed illite probably formed by direct precipitation.


A
10 µm

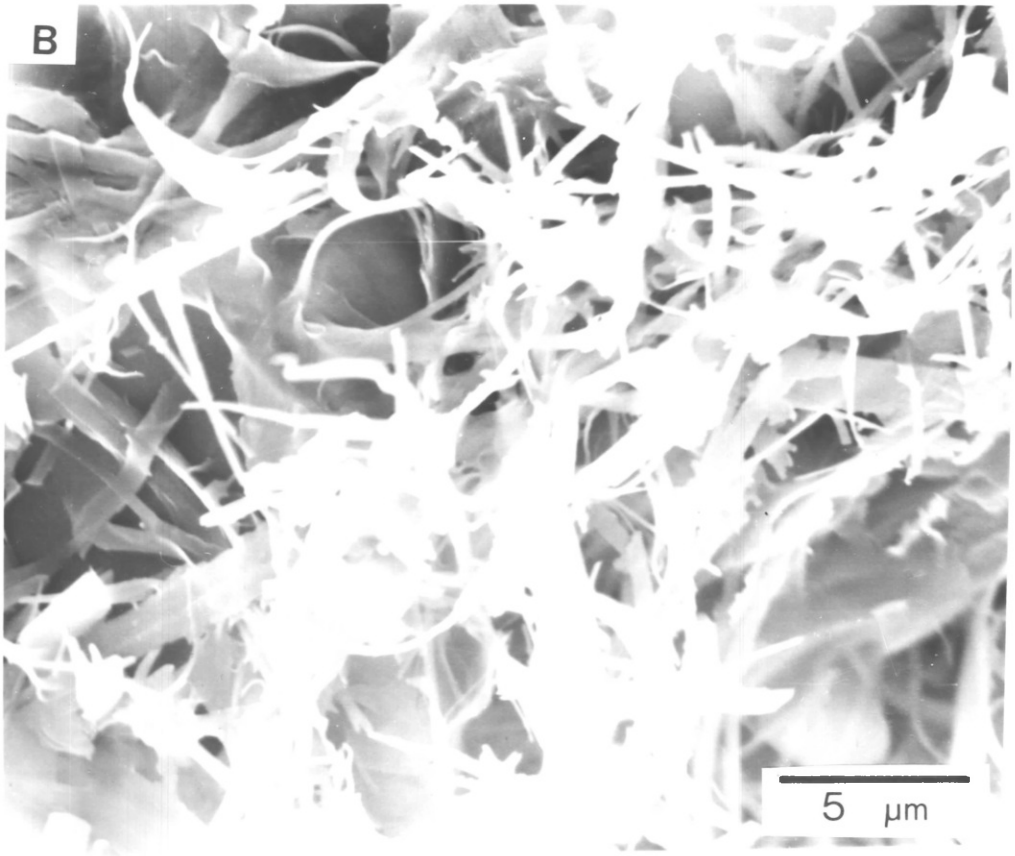

B
5 µm

PLATE 7:18

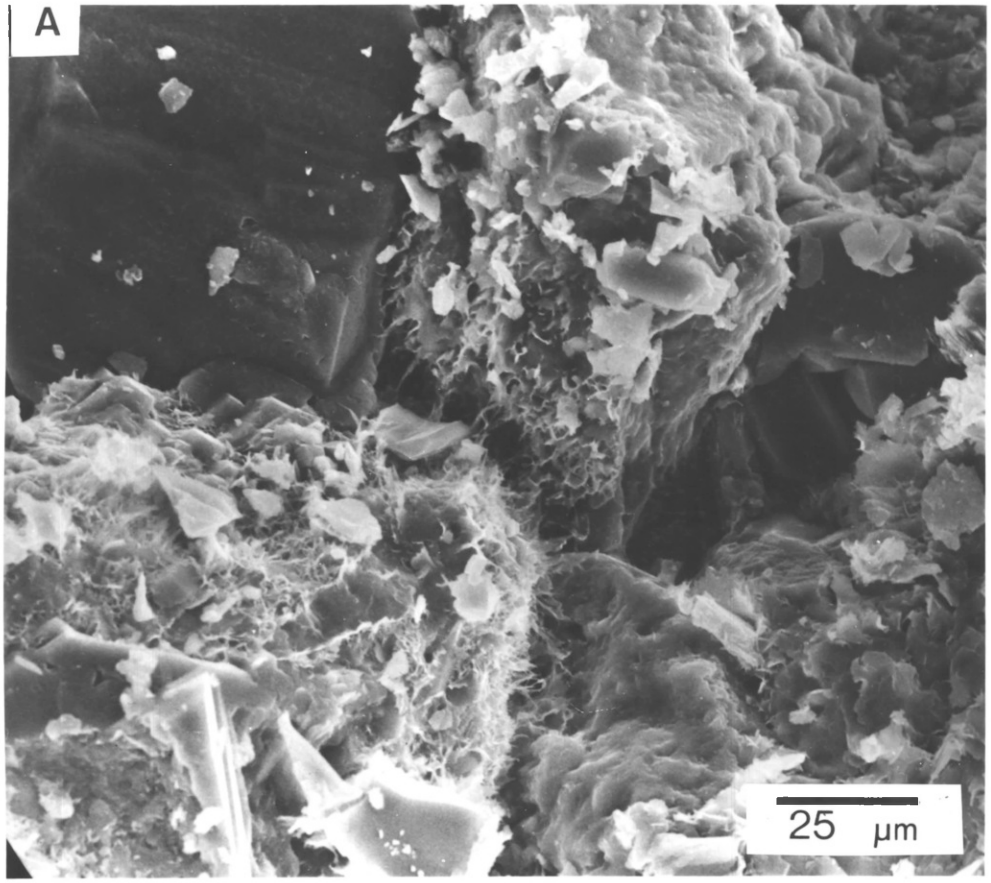
A

Moss Farm, 937m. Rough Rock SEI

Pore bridging bladed illite. This form is quite rare, note the lithic grain in the bottom left of plate A which is coated with illite.

B

Close up of central portion of above image.


A
25 µm

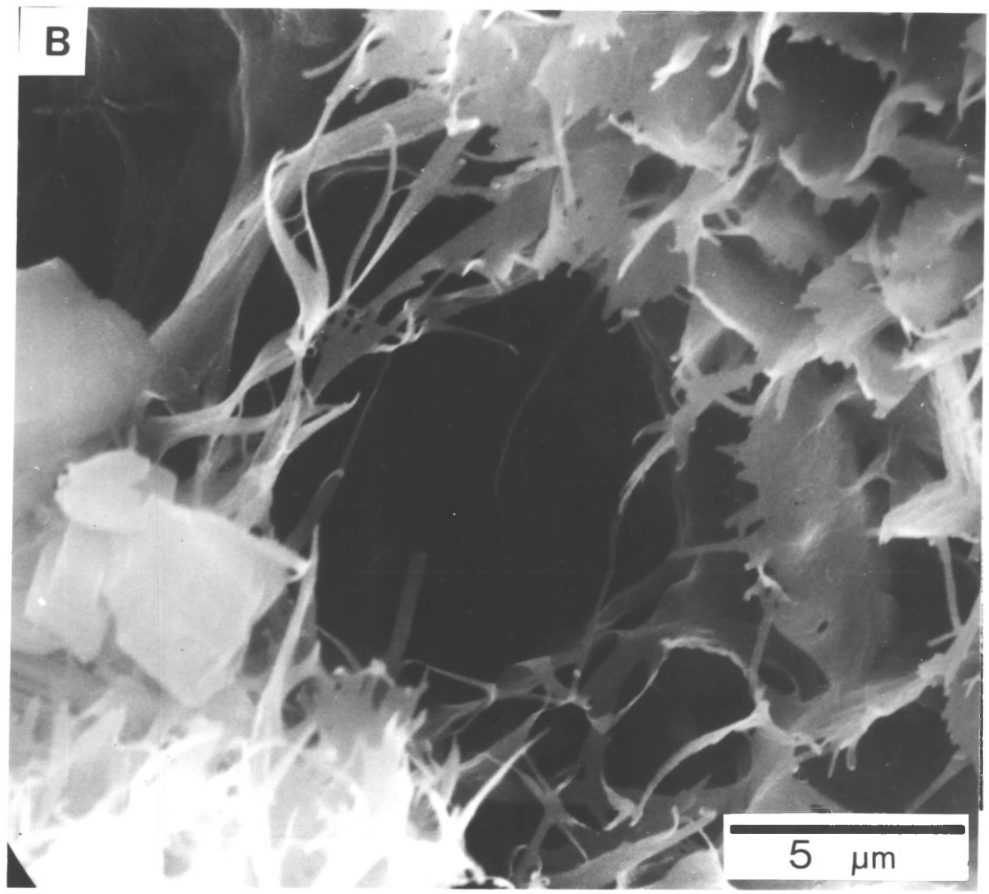

B
5 µm

PLATE 7:19

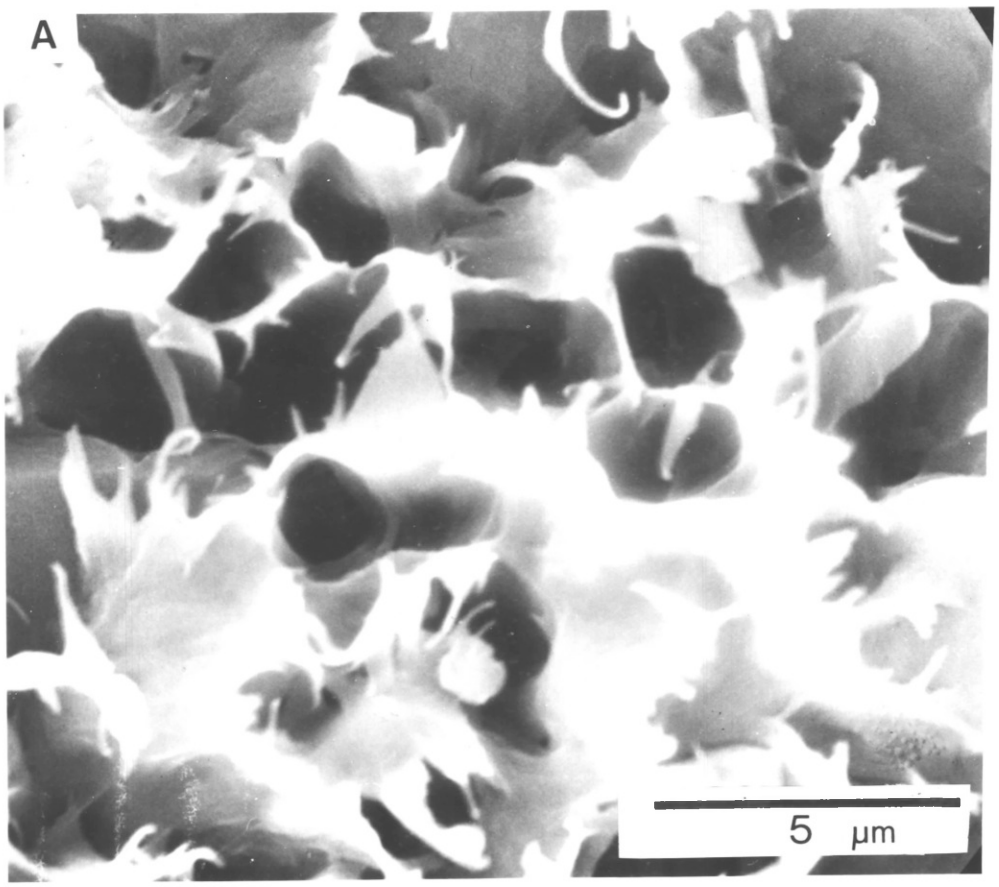
A

Moss Farm, 1009m. Pule Hill Grit. SEI

Pore lining boxwork illite showing a lath like/bladed morphology at the edges, such illite can severely reduce effective porosity by increasing irreducible water saturations.

B Trumfleet 1, 3340 Beacon Hill Flags. SEI

Sheet like illite in left of photo flaring out into pore space forming incipient laths.


A
5 µm

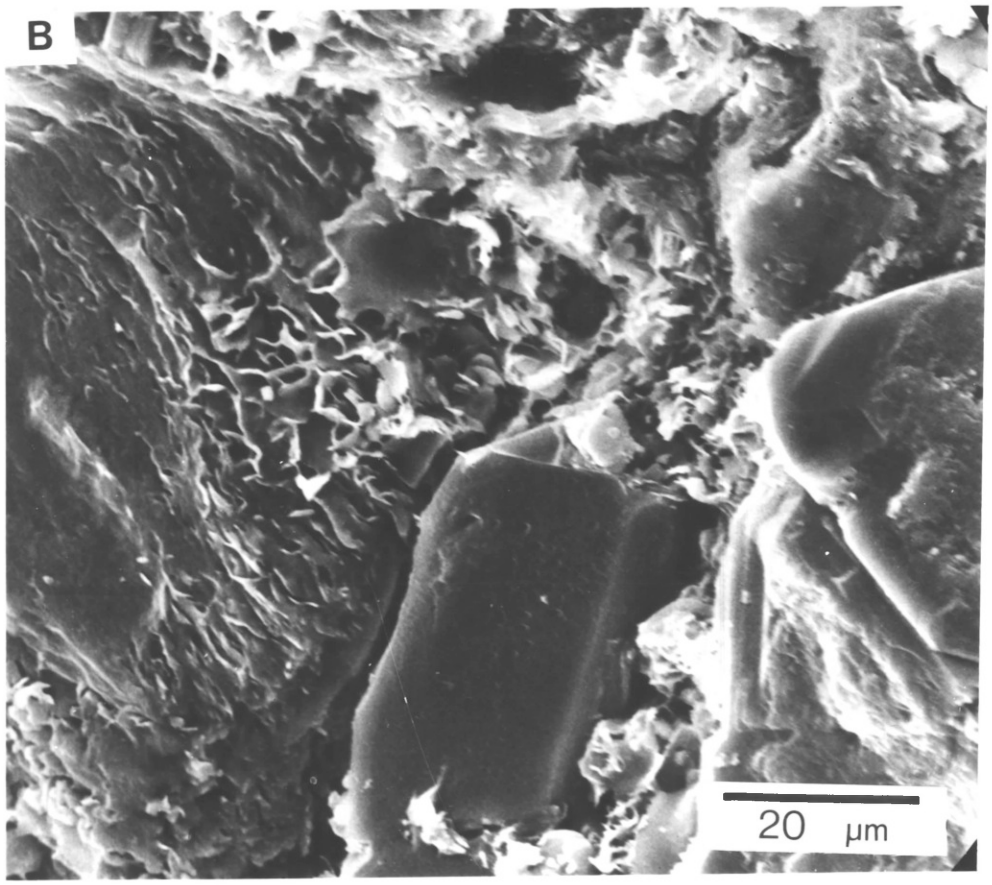

B
20 µm

PLATE 7:20

A

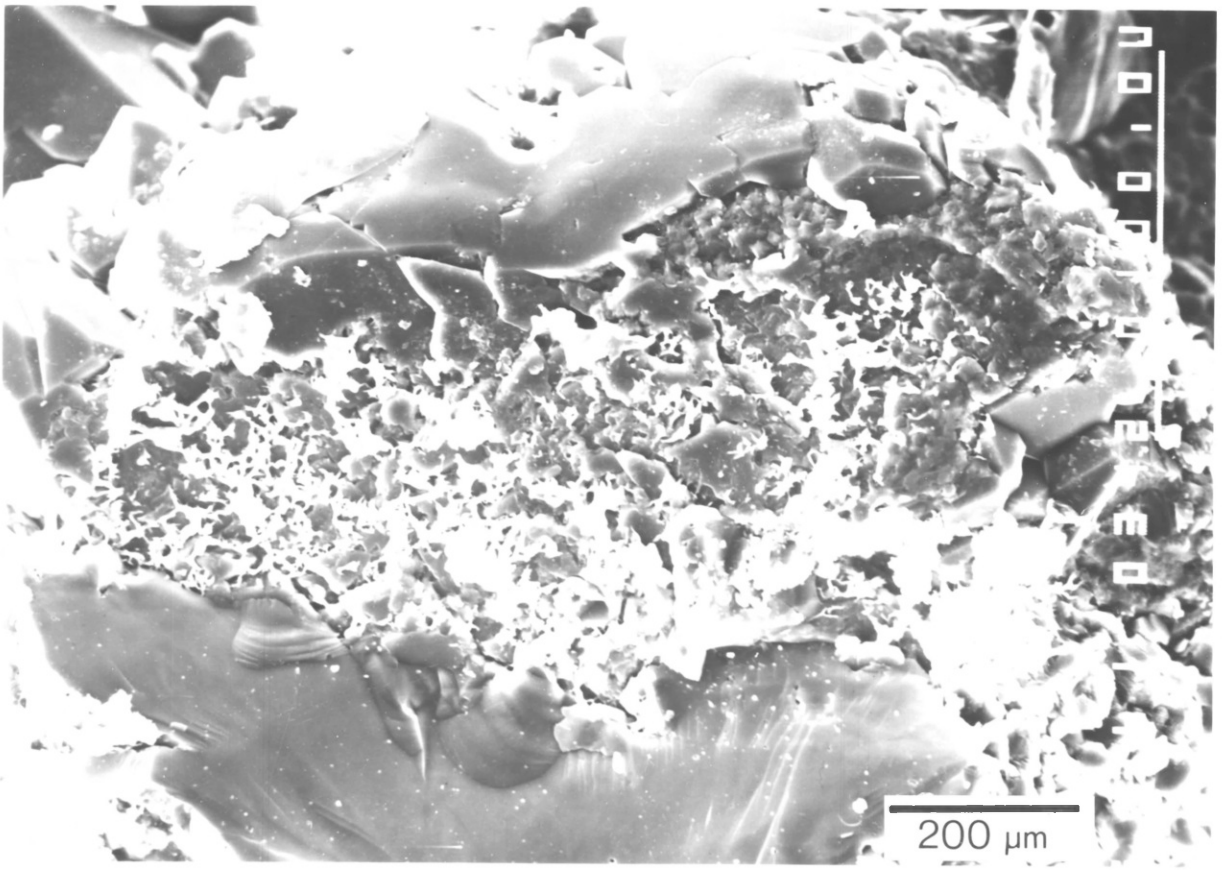
Moss Farm, 1015m. Pule Hill Grit. SEI

Boxwork illite coating pressure dissolution surface.

B

as above, SEI

Close up showing illite morphology.


200 µm

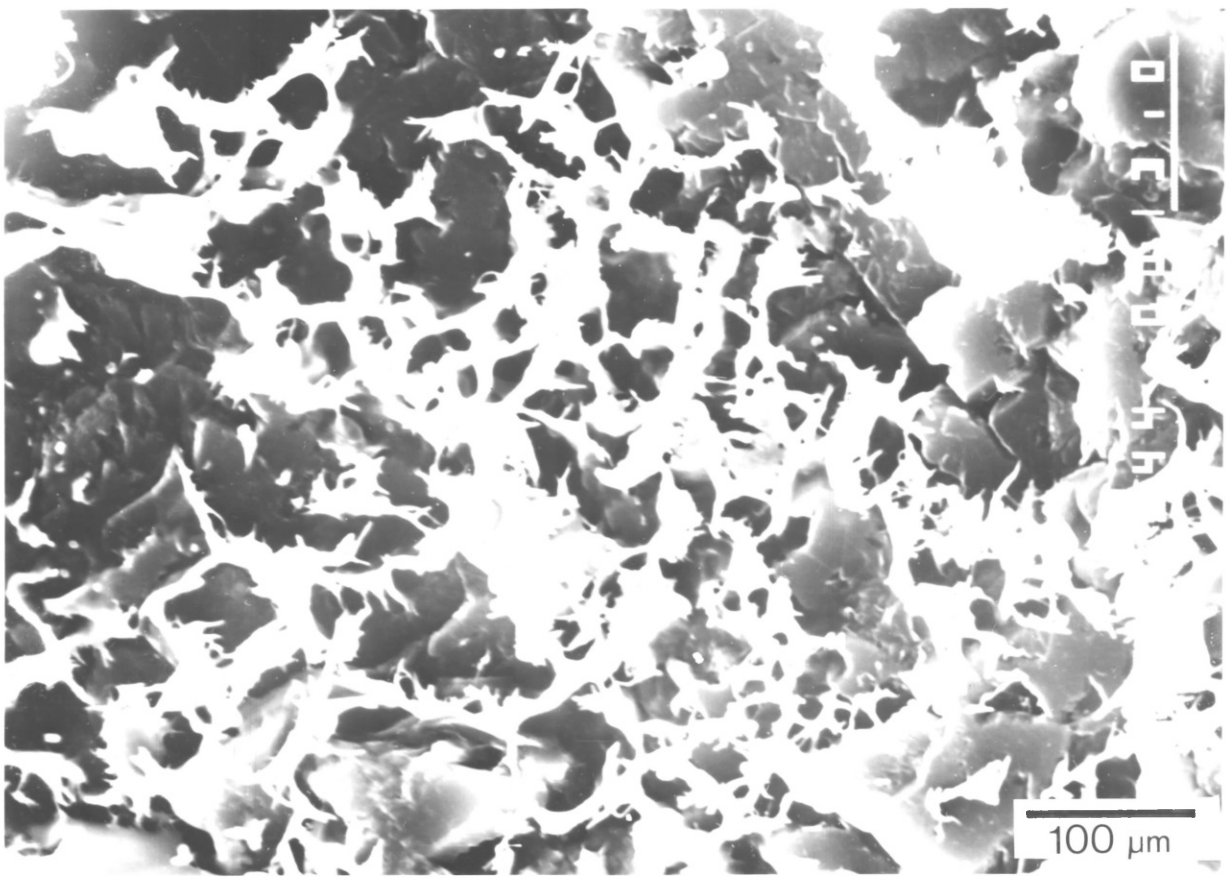

100 µm

PLATE 7:21

A

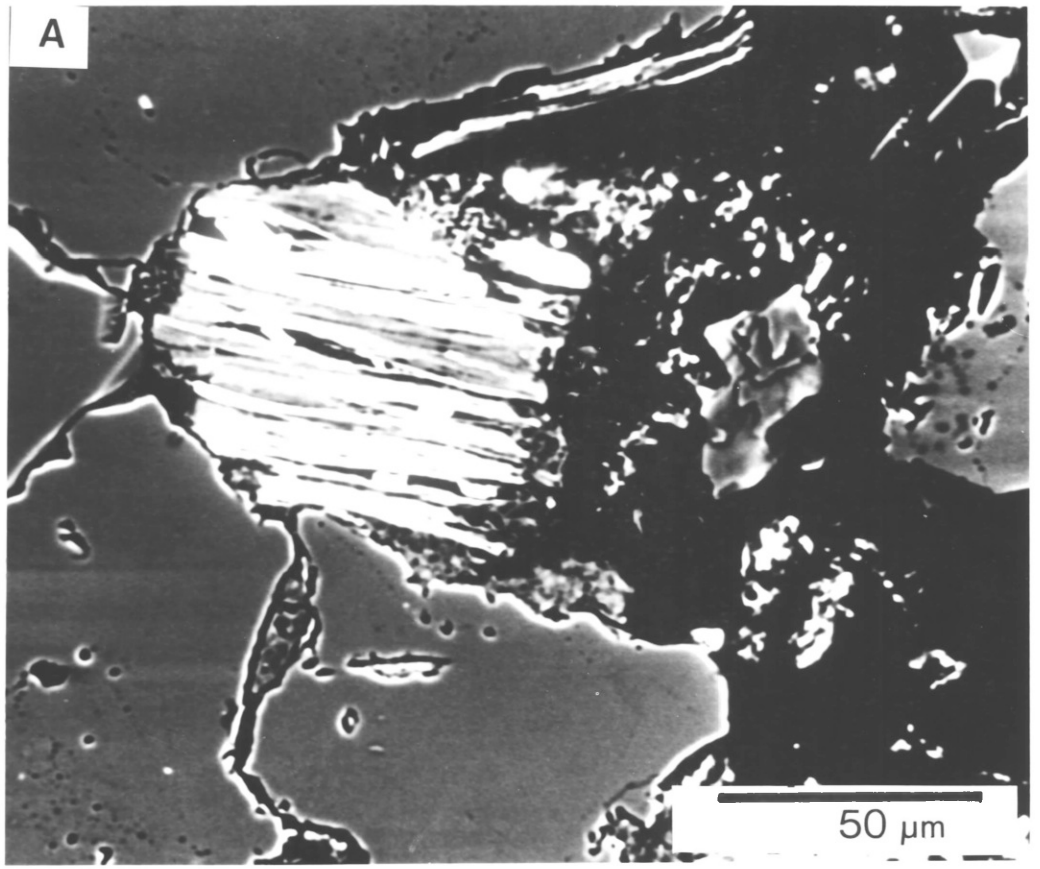
Moss Farm, 1015m. Pule Hill Grit, BSE

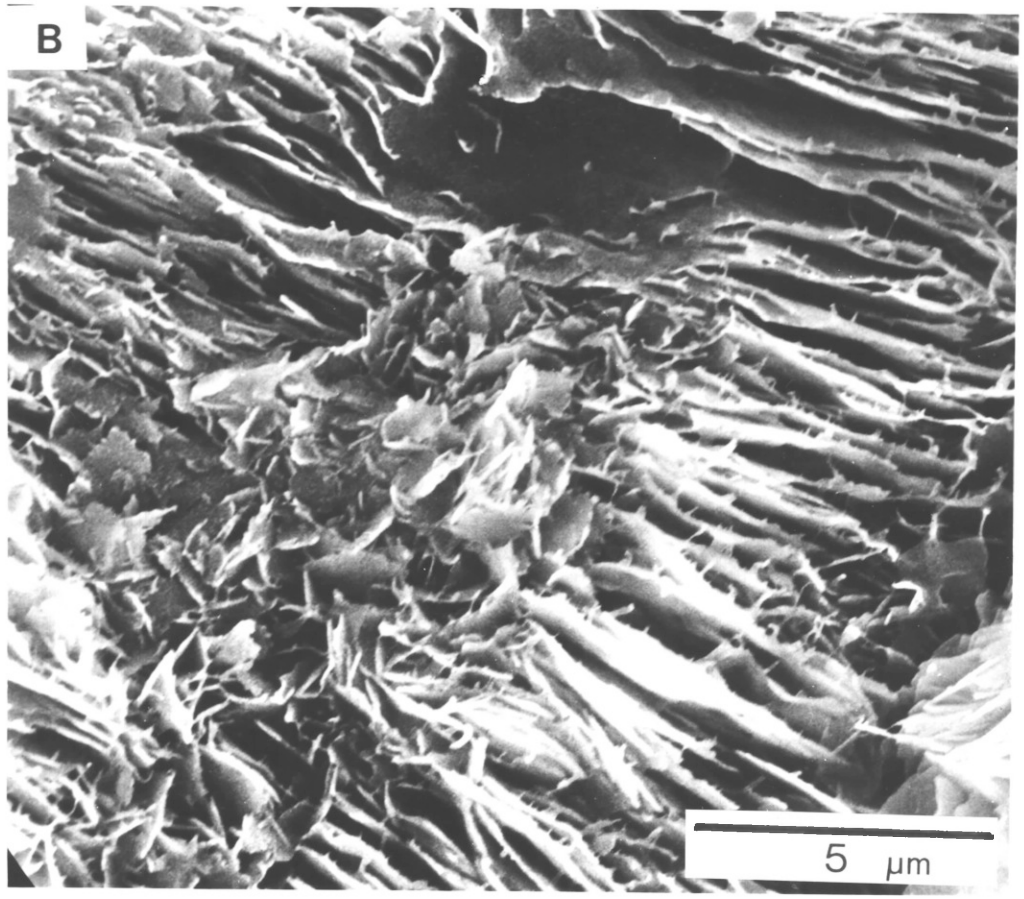
Detrital biotite grain showing layer/layer replacement by chlorite.

B

Moss Farm, 931m. Rough Rock. SEI

Boitite grains showing alteration to Illite? or chlorite. EDAX analysis proved unreliable due to the small size of the clay platelets.


A
50 µm


B
5 µm

CHAPTER EIGHT

DIAGENESIS

8:1. INTRODUCTION

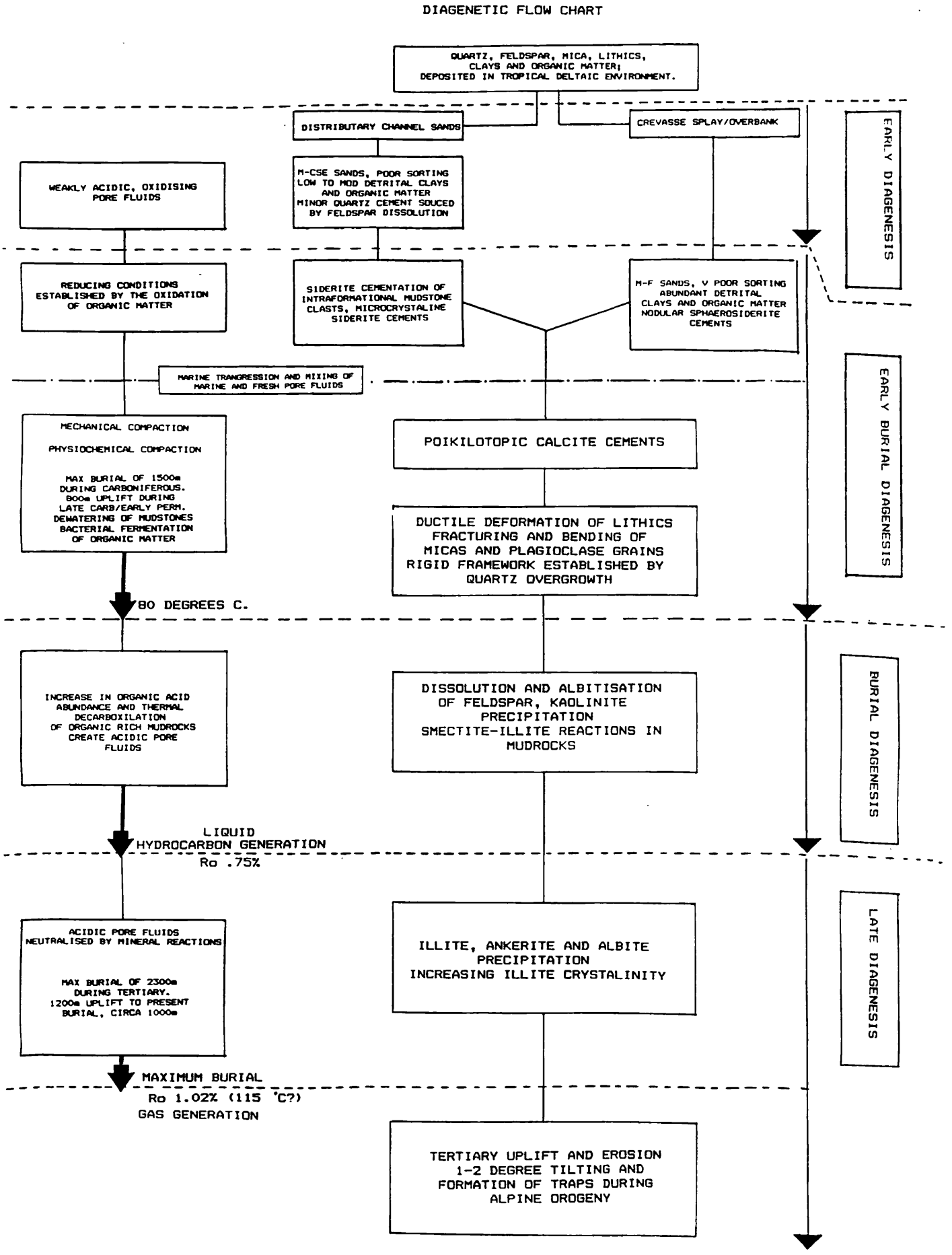
Three types of diagenetic process shall be considered in this chapter; A, physical processes (grain bending, fracturing and rearrangement); B physiochemical processes (pressure dissolution) and C; chemical processes (mineral precipitation and dissolution). Physical processes are often referred to as "mechanical compaction" and the term is used here to refer to bending, fracturing and rearrangement of grains without any other proceeses operating. In practice the three processes operate together, pressure dissolution will allow a grain to move (rearrangement) in response to stress and the growth of authigenic minerals may cause grains to move. Fractured and deformed grains often show a greater tendency towards chemical alteration, and pressure dissolution may be enhanced by compositional (chemical) differences between grains, for example, in the rocks studied, pressure dissolution is best developed where quartz abutts muscovite. Major diagenetic events are summarised schematicaly in Fig 8:1.

8:2. PARAGENETIC RELATIONSHIPS

The paragenetic relationships observed within the rocks studied are based upon the integration of data from the study of a large number of samples.

The aim was to establish a representative paragenetic sequence of the relationships observed (Figure 8:2). Inductive data of this type cannot be used to accurately predict the diagenetic history of a particular pore or area within the rock units. For example, calcite cements show a limited spatial distribution and are not present in every part of the sandstone unit, however it can be demonstated that they were precipitated prior to any mechanical compaction and, therefore, predate post compactional feldspar dissolution. The purpose of this, and the next chapter, is to attempt to build a diagenetic model, making use of the recorded inductive data.

FIGURE 8:1


DIAGENETIC FLOW CHART
QUARTZ, FELDSPAR, MICA, LITHICS, CLAYS AND ORGANIC MATTER; DEPOSITED IN TROPICAL DELTAIC ENVIRONMENT.
DISTRIBUTARY CHANNEL SANDS
CREVASSE SPLAY/OVERBANK
EARLY DIAGENESIS
WEAKLY ACIDIC, OXIDISING PORE FLUIDS
M-CSE SANDS, POOR SORTING LOW TO MOD DETRITAL CLAYS AND ORGANIC MATTER MINOR QUARTZ CEMENT SOUCED BY FELDSPAR DISSOLUTION
REDUCING CONDITIONS ESTABLISHED BY THE OXIDATION OF ORGANIC MATTER
SIDERITE CEMENTATION OF INTRAFORMATIONAL MUDSTONE CLASTS, MICROCRYSTALINE SIDERITE CEMENTS
M-F SANDS, V POOR SORTING ABUNDANT DETRITAL CLAYS AND ORGANIC MATTER NODULAR SPHAEROSIDERITE CEMENTS
MARINE TRANGRESSION AND MIXING OF MARINE AND FRESH PORE FLUIDS
EARLY BURIAL DIAGENESIS
MECHANICAL COMPACTION
PHYSIOCHEMICAL COMPACTION
MAX BURIAL OF 1500m DURING CARBONIFEROUS. 800m UPLIFT DURING LATE CARB/EARLY PERM. DEWATERING OF MUDSTONES BACTERIAL FERMENTATION OF ORGANIC MATTER
POIKILOTOPIC CALCITE CEMENTS
DUCTILE DEFORMATION OF LITHICS FRACTURING AND BENDING OF MICAS AND PLAGIOCLASE GRAINS RIGID FRAMEWORK ESTABLISHED BY QUARTZ OVERGROWTH
80 DEGREES C.
INCREASE IN ORGANIC ACID ABUNDANCE AND THERMAL DECARBOXILATION OF ORGANIC RICH MUDROCKS CREATE ACIDIC PORE FLUIDS
DISSOLUTION AND ALBITISATION OF FELDSPAR, KAOLINITE PRECIPITATION SMECTITE-ILLITE REACTIONS IN MUDROCKS
BURIAL DIAGENESIS
LIQUID HYDROCARBON GENERATION
Ro .75%
ACIDIC PORE FLUIDS NEUTRALISED BY MINERAL REACTIONS
MAX BURIAL OF 2300m DURING TERTIARY. 1200m UPLIFT TO PRESENT BURIAL, CIRCA 1000m
ILLITE, ANKERITE AND ALBITE PRECIPITATION INCREASING ILLITE CRYSTALINITY
LATE DIAGENESIS
MAXIMUM BURIAL
Ro 1.02% (115 'C?)
GAS GENERATION
TERTIARY UPLIFT AND EROSION 1-2 DEGREE TILTING AND FORMATION OF TRAPS DURING ALPINE OROGENY

PARAGENESIS OF TRUMFLEET SANDSTONES

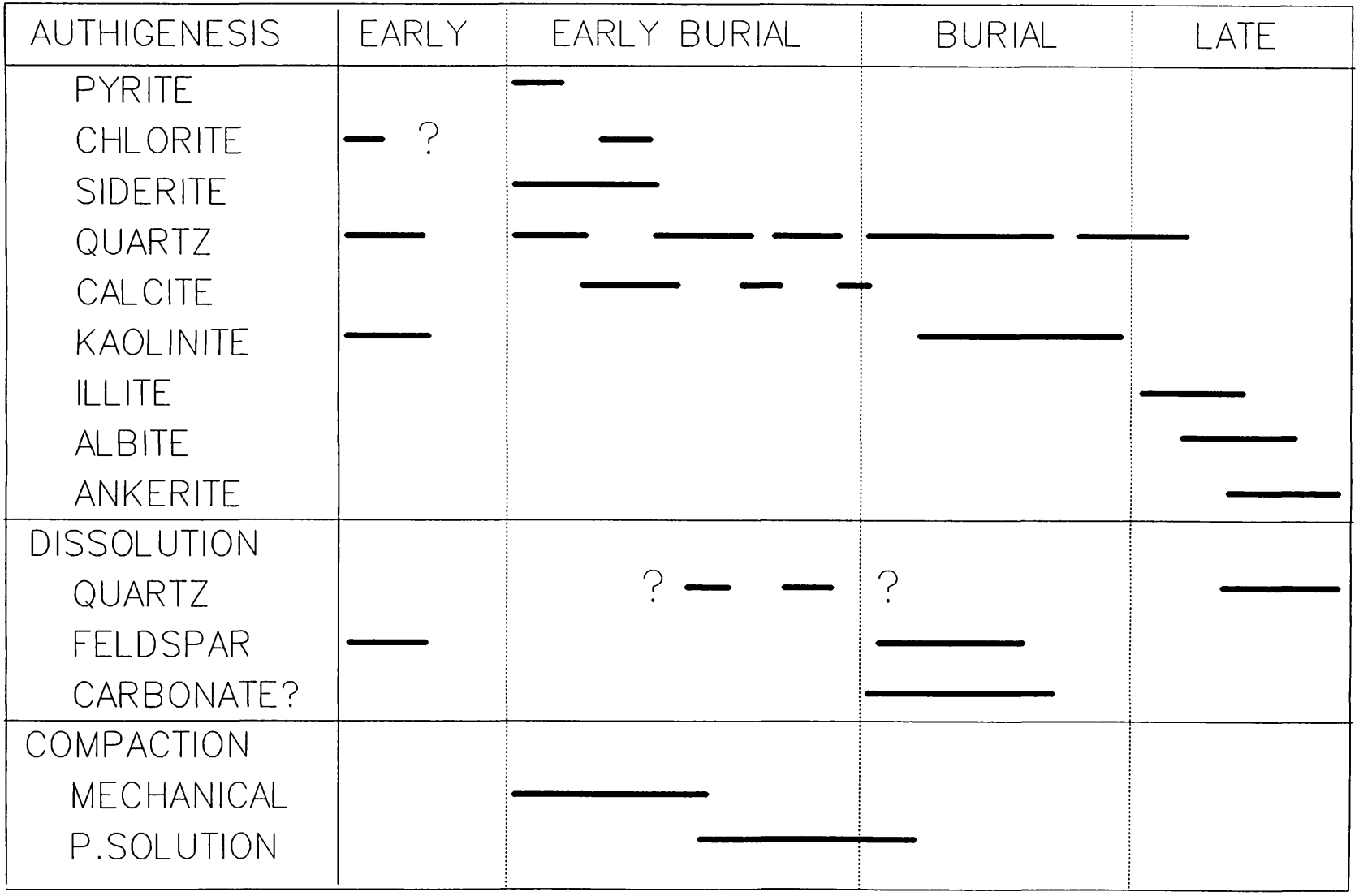


FIGURE 8:2

8:3. PHYSICAL PROCESSES

8:3.1. GRAIN REARRANGEMENT

The absolute amount of grain rearrangement is difficult to quantify since the initial packing of the grains is unknown, and attempts to estimate it by observing the packing of early poikilopicaly cemented sands proved unreliable, since the loose packing of these sandstones is in part a function of grains being moved by the force of crystalisation (Hawkins, 1972), as well as grian dissolution. By analogy with modern fluvial sands the intial porosity was probably in the order of 35 to 40% (Pettijohn, Potter and Siever, 1973)

Fuchtbauer (1967) has shown that in extreme cases porosity can be decreased by 50% by a combination of grain rearrangement and the dissolution of only 1.5% of the framework grains.

It is probable that in the Trumfleet sandstones, mechanical rearrangement of grains occurred in conjunction with other forms of compaction such as ductile grain deformation, fracturing and pressure dissolution, and that purely mechanical rearrangement did not have a significant effect on porosity; however it is not possible to accurately quantify the contribution of grain rearrangement to porosity reduction in these sandstones.

8:3.2. GRAIN FRACTURING

Grain fracturing is most noticeable in twinned plagioclase feldspar grains, but careful examination of untwinned feldspar grains shows that they too have sufferred fracturing, but that the fractures have been healed, probably during later diagenetic

alteration to albite. Mica grains also show some limited fracturing where they have been intensely deformed. Grain fracturing must have occurred during burial as a response to increasing overburden pressure. The fact that individual grains are fractured and that the fractures do not pass through grain boundaries implies that grain fracturing took place before a rigid framework was developed by interlocking quartz overgrowths (Plate 6:3B). Later, tectonically controlled fractures do pass through grain boundaries, but these show kaolinite infill suggesting that they formed during deeper burial.

8:3.3. GRAIN DEFORMATION

Organic matter and detrital intraformational claystone grains show ductile deformation as they are squeezed between framework grains to form a pseudo-matrix. Mica grains exhibit bending sub-perpendicular to the "C" crystalographic axis. The bending involves slippage of the mica plates which in small grains can be accommodated easily without disruption of the grain but in large grains, where the crystal is bent in more than one direction grain fracturing and splitting can occur (Plates 6:4A 6:5A). Some large grains show no bending at all and grain rearrangement is accommodated by pressure dissolution of quartz grains along the length of the grain (Plate 6:4B). As is the case with grain fracturing, grain bending occurs in response to overburden pressure and can only occur when grains are free to rearrange themselves to decrease the grain to grain stress.
For this reason, most grain bending probably occurred

before a rigid framework was established in the rock by quartz overgrowth.

8:4. PHYSIOCHEMICAL PROCESSES

8:4.1. PRESSURE DISSOLUTION

Pressure dissolution was one of the earliest compactional features to be described in sedimentary rocks with the use of the petrographic microscope (Sorby, 1863, 1865.)

The mechanism of compaction was thought to be one of dissolution at stressed grain interfaces, the material being removed by diffusion via an aqueous film. It was assumed that the material removed in this manner reprecipitated in a part of the rock where stress was lower, ie in the pore space (Griggs, 1940, Weyl, 1959, Bjørlykke, 1983.)

Rutter (1983) provided a succinct review of the development of pressure dissolution theory and attempted to explain the observation that pressure dissolution is often better developed where clay minerals are present (McEwan, 1978.). Material is removed in solution via an aqueous film, and the ability of this film to transport ions is dependant on the thickness of the film, the width of the film, and the grain to grain stress, as well as the temperature. It has been suggested that since the effective grain boundary width increases with the presence of non diffusible material on the grain interfaces (clay minerals in this case) then the ability of the rock to transfer material increases because the surface area of the aqueous diffusion film is increased. Rutter (1983) has shown experimentaly that aqueous

films can exist across stressed mineral interfaces, but at high stresses the film is thinned and its ability to transport ions isdecreased. These observations can explain why the best developed pressure dissolution in the Trumfleet Sandstones was observed where quartz abutts fresh muscovite grains, and up to one third of the original quartz grain appears to be missing (Plate 6:4B). Although the grain/grain stress is reduced more by the dissolution of quartz against mica, rather than quartz against quartz, due to the greater area of contact between the platy mica and quartz, the abilty of the film to transfer ions is increased as the film is thicker, and the surface area is greater.

Interpenetrant grain contacts were commonly observed in the Trumfleet sandstones, (Plate 6:1B) however, they can be produced either by pressure dissolution or by interlocking authigenic quartz intergrowths. CL photomicrographs show that there is not signifcant removal of material from the detrital cores (Plate 7:3A;B), indicating that only limited pressure dissolution has taken place. Less ambiguous evidence can be taken from the presence of stylolitic surfaces developed between quartz grains, these surfaces usually have an illitic clay showing a honeycomb morphology developed at the grain interfaces (Plate 7:20 A+B). However, stylolites were rare. It was not possible to quantify the contribution of pressure dissolution to the compaction of the sediments as only a limited number of samples were studied using CL, and the quality of the photmicrographs was poor. The grains in the Trumfleet sandstones showed no foreshortening

perpendicular to bedding, and although the presence of clay minerals seems to enhance the development of pressure solution (clay films were ubiquitous on stylolitic surfaces) it was also observed that where clay minerals are absent the primary porosity is greatly reduced by the action of silica precipitation (Plate 8:2B). It is, therefore, probable that mineral authigenesis had a much greater effect on porosity reduction than pressure dissolution.

8:5. CHEMICAL PROCESSES

8:5.1. INTRODUCTION

The paragenetic relationships of the authigenic minerals present within the rocks studied is shown schematically in Fig 8:3, and the diagenetic history is summarised in Fig 8:1. Four main assemblages of authigenic minerals can be identified on the basis of mineralogy and paragenesis:

I. Early Diagenesis: reflecting the effects of meteoric diagenesis under acidic oxygenated conditions; characterised by incipient quartz overgrowths.

II. Early Burial Diagenesis: under reducing alkali conditions characterised by siderite and calcite cements.

III. Burial Diagenesis; characterised by the reduction of porosity by authigenic quartz overgrowths, and culminating in the dissolution of feldspar and precipitation of kaolinite cements by acidic solutions generated at depth.

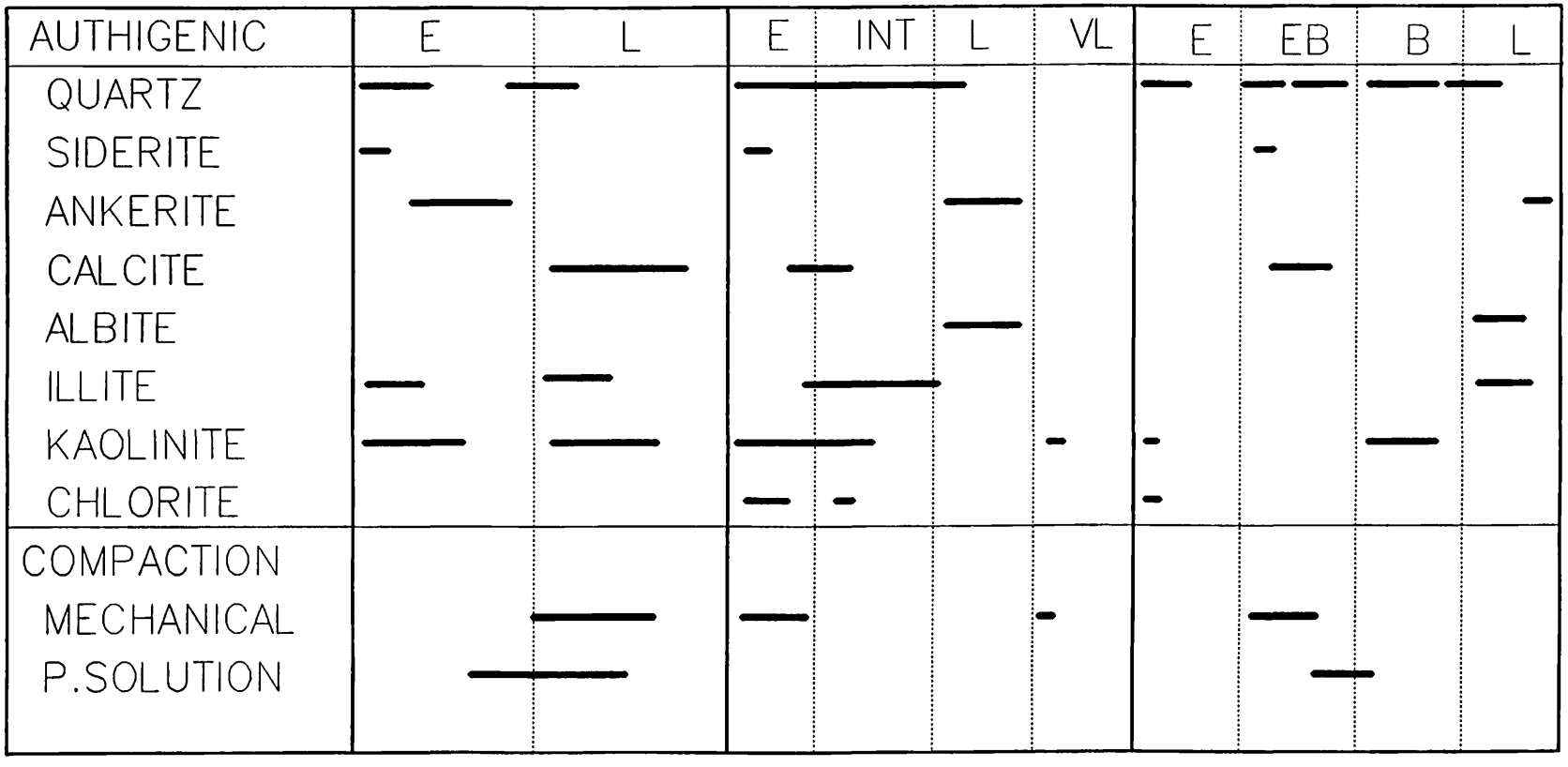
IV. Late Diagenesis: characterised by the formation of minor amounts of albite, ankerite and illite cements, in an alkali pore fluid.

Although the divisions between the various stages of diagenesis have been designated as sharp "instantaneous" lines on Fig 8:3, it is envisaged that there would have been considerable overlap between them due to the effects of chemical bufferring and of chemical gradients within the sandstone bodies. Nor can it be said that the complete sequence of

mineralisation will be observed in every part of a sandstone, as the locallised nature of the calcite and dolomite cements and testifies.

Fig 8:3 compares the paragenetic sequence observed in this study with those of Namurian and Westphallian Sandstones in the East Midlands studied by Hawkins (1972, 1978), and Huggett (1982, 1984). The main difference beween the studies concerns the timing of kaolinite cementation. Both Hawkins (1972) and Huggett (1982) consider that most kaolinite precipitation predates that of quartz, and they postulate that acidic, meteorically derived early diagenetic fluids caused feldspar dissolution and the precipitation of kaolinte. In this study,the bulk of kaolinite precipitation occurred after major quartz precipitation.

PRINCIPAL DIAGENETIC REACTIONS IN CARBONIFEROUS SANDSTONES, FROM HAWKINS (1978), HUGGETT (1984) AND THIS STUDY



E EARLY
INT INTERMEDIATE
L LATE

EB EARLY BURIAL
B BURIAL
VL VERY LATE

FIGURE 8:3

8:6. EARLY DIAGENESIS

8:6.1. INTRODUCTION

Fluvial waters in humid climates, similar to those of the Carboniferous delta (Johnson, 1982.) commonly have low ionic strengths, notably of sulphate, and are as a conseqence undersaturated with respect to most mineral species (Bjørlykke, 1983.).

The pH of these waters is commonly weak to slightly acidic, due to the presence of meteorically derived CO_2, and they are well oxygenated. The effect of humic acids in soil profiles may also locally decrease the pH.

It might be expected from the statements above that the channel sandstones would show a diagenetic assemblage similar to Berners (1980) "Oxic" sediments (Page 241), since they would have been supplied with hydrostatically driven fresh oxygenated waters, but the presence of comminuted organic matter in the channel sandstones, as well as early siderite, and minor pyrite in the channel sandstones suggests that the effects of this stage were minimal. The Namurian sandstones were probably isolated from their hinterlands early in their history, since the repeated tectonoeustatic sea level changes would intoduce marine fluids into the channel sandstones and the deposition of marine horizons during the height of the transgression would have the effect of cutting off fresh water drive through the channels, if the transgression was well enough established.

8:6.2. DIAGENESIS

The minerals precipitated and their sequence of mineralisation is shown on figure 8:3.

The effects of the Early Diagenetic phase are not volumetrically significant in the Trumfleet sandstones, and are in fact only recognised in a few samples where the precipitation of poikilotopic calcite cements have preserved them in isolation.

The first minerals to precipitate in the sandstones were minor amounts of quartz and pyrite. The quartz forms as isolated prisms on quartz detrital grains but its relationships with the other minerals is difficult to establish since they are masked by later quartz precipitation. Variable amounts are seen preserved under early poikilotopic calcite cements but the amounts are usually small (less than 2%). The relationship of the early quartz to other cements is, therefore, difficult to establish and it is tentatively postulated that it formed from early diagenetic fluvially derived waters, possibly in association with dissolution of feldspar as an effect of chemical weathering (Curtis, 1983.). Later growth (and aggradation ?) of kaolinite, mask these relationships and make interpretation difficult.

Chlorite was recognised on XRD traces and was seen to be developed mainly as a decomposition product of biotite. Since biotite is known to be unstable under surface conditions, and it was not clear whether the alteration was pre or post depositional, it is probable that chloritisation of boitite was an early diagentic event.

8:7. EARLY BURIAL DIAGENESIS

8:7.1. INTRODUCTION

In fine grained, organic rich, sediments oxygen is removed by the microbial oxidation of organic matter in the first few tens of centimetres, below the bioturbation zone. Sulphate reducing bacteria are active below this depth, to be replaced by methane producing organisms as the sulphate is used up. (Berner, 1980. Gautier and Claypool, 1984.). Berner, (1980) has proposed a classification of fine grained sediments during early diagenesis based upon the presence or absence of measurable dissolved oxygen or sulphate (Table 8:1).

Table 8:1, BERNERS CLASSIFICATION OF EARLY DIAGENESIS

Environment	Characteristic Phases
I. Oxic	Haematite, Goethite No Organic Matter.
II. Anoxic	
A Sulphidic	Pyrite, Marcasite, Rhodochrosite, Organic Matter
B Nonsulphidic	
1 Post Oxic	Glauconite and other Fe silicates Little organic Matter
2 Methanic	Siderite, Vivianite, Rhodochrosite. Earlier sulphide minerals, Organic Matter

The methanic zone can extend as deep as 1km, and corresponds to the major phase of mudstone dewatering, and can, therefore,

strongly influence the diagenesis of adjacent sandstones (Curtis, 1983, Gautier and Claypool, 1984.). Although Berner's (1980) classification was intended to be applied to fine grained rocks it is useful to consider it when refering to sandstones, since not only does the diagenesis of enclosing mudstones to a large extent control the diagenesis of sandstones, but the processes described by Berner (1980) can be demonstrated to have been operating in the Trumfleet sandstones. The major difference between fine grained sediments and the sandstones is the greater ability of the sandstones to allow fluid to flow through them (Bjørlykke, 1983.).

The overbank and crevasse splay sandstones were, however, only occasionaly flushed through with oxygenated waters during flood events, and they occur interlaminated with impermeable mudstones. The behavior of these sediments during early diagenesis can, therefore, be considered to be similar to that of fine grained sediments.

8:7.2. DIAGENESIS

Pyrite formed locally where anoxic conditions were initiated by the oxidation of organic matter. It is for this reason that it is only found where organic matter is abundant, in micaceous and clay rich laminae within the sandstones. The pyrite is usually observed as a few scatterd rhombs surounded by later sphaerosiderite or microcrystalline siderite cements (Plate 7:6A) and is never abundant.

As the initial sulphide content of the sediments would have been low, because of their non-marine nature, the sulphate would

quickly have been used up by pyrite precipitation and siderite is precipitated in place of pyrite.

Sphaerosiderite nodules are common in the crevasse splay, overbank and seatearth facies, but are absent from the channel facies which contains microcrystalline siderite/clay agregates, concentrated in organic rich laminae. Shpaerosiderite nodules are uncompacted and clay and micas can be seen to bend around them. Compaction is accommodated by pressure dissolution of quartz around the nodules (Plate 7:9A). The poikilotopic calcite cements observed in the Trumfleet sandstones preserve an open framework texture in which quartz and feldspar grains appear to float. Quartz grains show only limited dissolution textures, although feldspar is often extensively replaced, and it is postulated that these cements formed prior to any mechanical compaction of the rock. Hawkins (1972) has postulated that framework grains found in similar Westphalian sediments had been moved by the force crystalisation, and this as well as grain replacement, enhances the apparent "floating" of framework grains. Calcite cements are usually preserved just above the bases and just below the tops of the sandstone units (Figs 7:1 to 7:10) and it is postulated that they were sourced by $C0_2$ expelled from the surrounding mudstones, mixing with Ca^{2+} in solution in the channel sands. $C0_2$ is derived from the microbial reduction of organic matter in the mudstones (Curtis, 1983.). Although similar reactions produce $C0_2$ in the sandstones, the amounts produced in the mudstones would be much greater due to their higher organic content.

Irwin and Hurst (1983) have recorded early calcite cement

sourced in this manner from the Jurassic of the North Sea, and they postulate that they formed from CO_2 derived from the action of sulphate reducing bacteria as the sandstones are deficient in iron. Whist sulphate reducing bacteria are acting any free iron is fixed by H_2S to form pyrite, and any carbonate minerals formed are consequently iron poor (Curtis, 1978.). The mixing of fresh and marine pore fluids can also have the effect of increasing the pH of the pore fluids causing the precipitation of carbonate cements, and this undoubtedly occurred during marine trangressions. The early calcites in the Trumfleet sandstones were probably sourced by both these meachanisms, and, given the varaible amounts of authigenic quartz pre-dating the calcite cements, probably occurred at differing times. Limited microprobe analyses of the Trumfleet calcite cements show that they are almost entirely free of iron, in addition, they show strong red colours upon staining with alizarin red but remain unaffected by potassium ferricyanide.

8:8. BURIAL DIAGENESIS

8:8.1. INTRODUCTION

The leaching of carbonate, and to a lesser extent feldspars, by acid solutions at depth has been postulated by a number of authors (summarised in Schmidt and MacDonald, 1979.) as the agent resposible for secondary porosity formation in sandstone reserviors. Secondary dissolution porosity has been recorded from sandstones of a wide variety of tectonic settings, sedimentary environments and ages (Schmidt and McDonald, 1979.).

Surdam et al (1984) have noted that aluminosilicate dissolution may play an important role in secondary porosity formation. The low mobility of aluminium, due to its low solubility, has long been believed to hinder any major dissolution of aluminosilicates, but Surdam et al (1984) have shown that aluminium solubility is increased by the formation of alumina-organic complexes, and they have demonstrated that these can form at depth in diagenetic environments.

The acidic pore fluids responsible for the dissolution of carbonate are believed to be derived from the thermal maturatiom of organic matter (kerogen decarboxylation) combined with the thermal transformation of smectite to illite/smectite and illite (Powers, 1968, Schmidt and McDonald, 1979, Curtis, 1983.)

Curtis (1983) pointed out that the fluids migrating from the mudstones to the sandstones during this phase of diagenesis contain metalic cations, including alumina. When the fluids

react with either calcite or feldspar the pH is increased, the pore fluid becomes oversaturated with respect to alumina and kaolinite can precipitate as a consequence (Fig. 7:14).

8:8.2. DIAGENESIS

The most obvious evidence of the presence of so called "aggressive" pore fluids in the Trumfleet sandstones is the post compactional dissolution of feldpar, in some cases leaving oversized pores as the only trace of the existence of a former feldspar. Surdam et al (1984) have shown that humic kerogen, closest to the types found in East Midlands (Dobson, 1984) can generate significant amounts of carboxylic acids during early maturation, which would have the effect of increasing the solubilities of feldspar grains, if they migrated into the sandstones. The generation of carboxylic acid is however, strongly depedant on the pH of the pore fluids, and they are destroyed during diagenesis by thermal decarboxylation reactions, but can exist at temperatures up to 140^{o} C. (Carothers and Kharaka, 1978, 1980.); a higher temperature than that to which the Trumfleet sandstones were subjected.

The effects of these acidic solutions on the calcite cemented zones is variable, some sediments seem relatively uneffected by dissolution, whilst in others only a few scattered remnants of calcite remain. Surdam et al (1984) have shown that under conditions of high ρCO_2 with a carboxylic acid buffer, it is possible to obtain dissolution textures in silicates whilst leaving carbonates unaffected.

The main differences between this study and those of Hawkins (1972, 1978) and Huggett (1982, 1984) concerns the timing of the precipitation of kaolinite cements; the two previous studies consider that most kaolinite formed during early diagenesis by the action of acidic, fluvially derived pore fluids, however, the presence of unoxidised organic matter in the Trumfleet sediments indicate that this process could not have occurred in these sandstones, since it is probable that if sufficent meteoric fluid were allowed to pass through the pores to dissolve framework grains, then the more volatile organic matter would have been quickly oxidised.

Silica released from clay mineral reactions in the mudstones and from feldspar dissolution continued to be precipitated as quartz overgrowths, Surdam et al (1984) have postulated that silica may also be released as an organic complex, perhaps explaining the apparent overabundance in silica when its low solubility is considered (Bjørlykke, 1979, Land and Dutton, 1979, Boles and Franks, 1979.)

It is not clear from textural evidence whether the albitisation of feldspar occurred during this phase diagenesis, or later cogenetic with the precipitation of authigenic albite. Boles (1982) showed that albitisation of plagioclase grains in the Gulf Coast Tertiary occurred at temperatures between 110 and 120^{o} C. Since the depletion of K^{+} from feldspars along with a source of Na^{+}, (probably released from Na^{+} smectite to illite transformations), controls albitisation, it seems probable that potssium was removed from feldspar during dissolution, and

added to the pore fluids, therby decreasing the relative abundance of Na^+ in the pore fluids. When K^+ was removed by the precipitation illite the Na^+/K^+ ratio would increase making the precipitation of albite more likely.

Pore fluid chemistry was controlled by mineral reactions within the sandstones and surrounding mudstones, responding to increasing burial temperatures. Post compactional dissolution of feldspar grains, by organic acid complexes, ocurred during the burial phase of diagenesis. Kaolinite was precipitated from alumina released by feldspar dissolution and clay mineral reactions in the surrounding mudstones.

Silica cements were sourced by silica released by the same reactions.

8:9. LATE DIAGENESIS

8:9.1. INTRODUCTION

The final assemblage of authigenic minerals found in the Trumfleet sandstones comprises illite, ankerite and albite. The mudstones surrounding the sandstones in the Trumfleet Field, show vitrinite reflectance measurements from 0.75 to 1.02, which correspond to a maximum temperature of 115^{o} C, indicating that they were in the oil zone during the period of maximum burial. Fluids migrating into the sandstones from deeper in the basin, ie from the Gainsborough Trough, were probably generated at higher temperatures. Curtis (1983) states that acid generated in mudrocks will tend to be consumed by reactions in sandstones, mainly carbonate dissolution. The presence of undissolved early calcite cements in the Trumfleet sandstones indicates that not enough acid pore fluids were generated to dissolve all the carbonate present, and the presence of a late ferroan dolomite/ankerite cements in the sandstones indicates that the pH must have increased sufficently for the precipitation of carbonates.

It would seem, therefore, that acid comsumption reactions in the sandstones occurred faster than acid generating reactions in the mudrocks.

A similar sequence of diagenesis, ie albitisation and the precipitation of ferroan dolomite cements after kaolinite precipitation, has been reported from the Frio Formation in the Texas Gulf Coast (Loucks et al 1977; Curtis, 1982.), as well as N Sea Jurassic sediments (Burley, 1986) and S N Sea Rotliegendes sediments (Goodchild and Whittaker, 1986).

8:9.2. DIAGENESIS

Figure 8:4 shows the equilibrium diagram for illite, kaolinite and K^+ feldspar in terms of temperature and activity of K^+ and H^+. It can be seen that for constant log K^+/H^+ activites and pressure, illite becomes the stable phase instead of kaolinite, with increasing temperature . In addition the dissolution of feldspar and carbonate have the effect of increasing the pH of the pore fluids, and since illite becomes stable at lower temperatures in alkaline solutions the precipitation of illite occurs as a consequence of increasing temperature and dissolution reactions in the sandstones.

Where illite overgrows kaolinite, there is limited alteration of kaolinite, mainly on the pore edges, where illite tend to be concentrated, despite its late origin. The majority of the kaolinite however, appears unaffected and the illite is in the form of neoformed sheets and laths. Illitisation of kaolinite was not a commonly observed phenomenon.

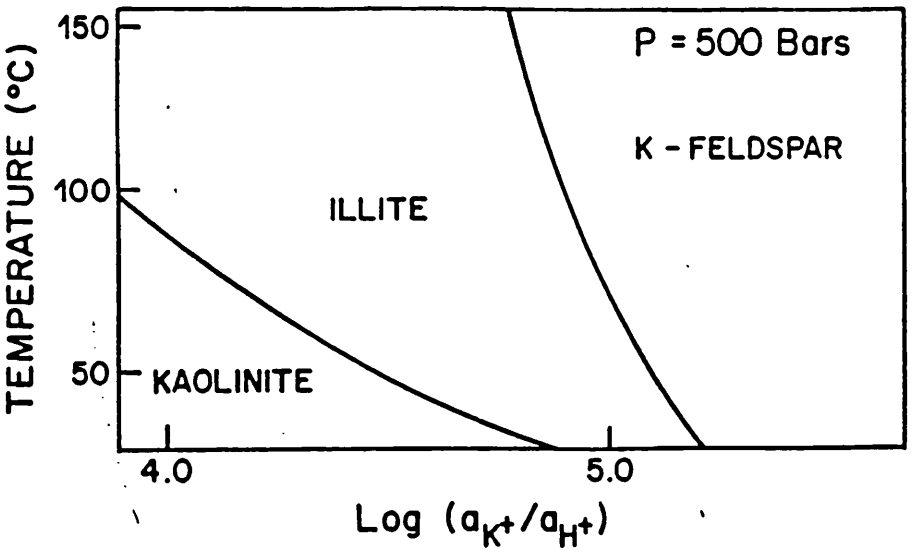
Ankerite was the last mineral to form in the sediments. Late stage iron rich carbonate cementation is a commonly observed phenomenon in Gulf Coast sandstones (Dutton, 1977; Louck et al, 1979; Franks and Forester, 1984.). Spatially, ankerite is more common near the centres of the sandstone bodies rather that at the tops and bases, and this suggests a late origin since early diagenetic cements reduced porosity of the tops and bases, leaving the centres as the main conduits for fluid flow.

Boles and Franks (1979) suggest that iron and magnesium for ankerite cements is released from mixed layer clays during the transformation from smectite to illite at temperatures in excess of 100° C.

By analogy with similar cements in the Gulf Coast (Franks and Forrester, 1984.), the illite, albite and ankerite in the Trumfleet sandstones formed at high temperatures (ie near maximum burial) using ions sourced from clay mineral transformations and CO_2 produced by thermal decarboxylation reactions within the enclosing mudrocks.

FIGURE 8:4

Stability of illite as function of temperature and K_+/H^+ activity.



from Hutcheon (1981)

8:10. PORE FLUID ANALYSES

Downing and Howitt (1969) have published pore water analyses from Namurian sandstones of the East Midlands. A large area of low salinity, which exists to the south east of the Derbyshire Dome, is caused by the post Tertiary infiltration of meteoric water through Millstone Grit rocks which now outcrop in the Pennines. The highest salinities recorded occur in the Trumfleet area, Downing and Howitt (1969) cite the high salinities in this area as evidence that there has been no influx of meteoric waters, and conclude that the pore waters in the Trumfleet Field represent connate Namurian pore fluids which have evolved to a brine like composition during burial. They propose that the shale units enclosing the sandstones acted as semi permeable membranes and enhanced the ionic composition of the fluids by holding back cations.

The petrographic analysis of the sandstones carried out in this study reinforces this view, since no evidence of meteoric influx could be found (for example, siderite oxidation or the oxidation of organic matter). Two phases of uplift could have introduced meteoric fluids into the sediments, late Carboniferous and Hercynian movements or Alpine earth movements. The effects of the Hercynian produce Carboniferous red beds, the so called Barren Red Measures which can be of any stratigraphic age and are caused by the late and post Carboniferous infiltration of oxidising fluids during uplift. The so called Barren Red Measures should not be confused with late Carboniferous sediments which were deposited in an arid climate and show typical arid

early diagenetic effects, such as the Etruria Formation.

Table 8:2 shows the pore fluid analyses of sandstones from the Trumfleet Field published by Downing and Howitt (1969) and Burley (1982).

Table 8:2

Well	Trumfleet. 1		Trumfleet 2
Formation	Rough Rock	Pule Hill G.	Pule Hill G.
Na	82360	64610	69460
K	490	2170	253
Ca	8840	24400	14000
Cl	146230	152330	134900
SO4	425	10	20
HCO3	55	0	29
Mg	na	na	925
Total	146230	246053	219573
pH	6.3	5.6	6.3

Analyses in Mg/L

Both the pH and the low HCO_3 content of the pore fluids suggest that the present day pore fluids are out of equilibrium with the last phases precipitated in the sandstones, namely dolomite. The Dolomite appears fresh and unaltered. However, the analyses were carried out in the 1950's and no information on the techniques used for sampling at the time is available. Oil field water analyses are notoriously unreliable, degassing and scaling frequently alter the chemistry of the fluids before they reach the surface, and it is virtually impossible to obtain un-contaminated samples. The analyses pesent above, therefore, do not reflect the observed mineralogy of the samples (especially with regard to the carbonate content) and are probably not representative of the pore fluids at depth at present in the reservoir.

8:11. CONCLUSIONS

The observed relationships of the authigenic mineral species present in the sandstones and their inferred origin allows a relative time sequence to be established. It is known that the sediments were deposited approximately 315 my, and that the effects of the early diagenetic, ie meteorically derived fluvial pore fluids were limited. Oxidation of the abundant organic matter in the sediments led to the establishment of reducing conditions early in the history of the rocks. In the overbank and crevasse splay sediments it seems that the establisment of reducing conditions was essentially a geologicaly instantaneous event. The preservation of organic matter in the sandstones implies that reducing conditions were established in these sediments soon after deposition also.

The dissolution of feldspar at depth implies that the sandstones were subjected to temperatures in excess of 80^{o} C during burial (Surdam et al, 1984), and the vitrinite reflectance data from the interbedded mudstones from the Trumfleet one well imply that that the maximum temperature to which the sediments were subjected was in the order of 115^{o} C. Geological consideration suggest that this maximum burial occurred during the late Cretaceous.

The assemblages and types of burial diagenesis recorded show certain similarities to the diagenesis of sediments of the Gulf Coast of the USA.

With the above in mind, the next Chapter presents an attempt to define the absolute timing of these key diagenetic events by

integrating this data with a simple burial and temperature model.

CHAPTER NINE

BURIAL HISTORY AND THERMAL MATURATION

9:1. INTRODUCTION

Waples (1980) published a method of estimating the maturity of organic matter in sediments using the Lopatin Time Temperature Index technique, the basis of which is outlined in Chapter 4.

Although this technique is intended for use a predictive geochemical modeling tool, it can be used constructively in diagenetic studies since mineral transformations and precipitation often occur within known (if poorly defined) temperature/depth ranges, and the thermal maturation of organic matter exerts a strong control on pore fluid composition. Integrating the temperature history of a particular horizon with the diagenetic history allows estimates of the absolute timing of diagenetic events to be made.

Dobson (1984) applied this technique to the Walkeringham borehole, some 15 km from Trumfleet. He noted that in order to obtain the temperatures required to produce the vitrinite reflectance values observed in the well it was necessary to postulate that a palaeogeothermal gradient of 44^{o} per km was acting on the sediments at the time of maximum burial during the Cretaceous, and he assumed that the geothermal gradient would have been constant through geological time. A more detailed model than that used by Dobson (1984) was constructed for this study, by attempting to estimate palaeogeothermal gradients and in-putting them into a commercially available geochemical modeling computer programme; "Mature" developed by IGI Ltd.

9:2. BURIAL MODEL

The tectonic history of the area is set out in Chapter 2, and the burial history has been estimated for the Trumfleet No 1 well, since it contains a full sequence between the G. subcrenatum and R. gracile marine bands, and the vitrinite reflectance data were estimated from rocks from this well.

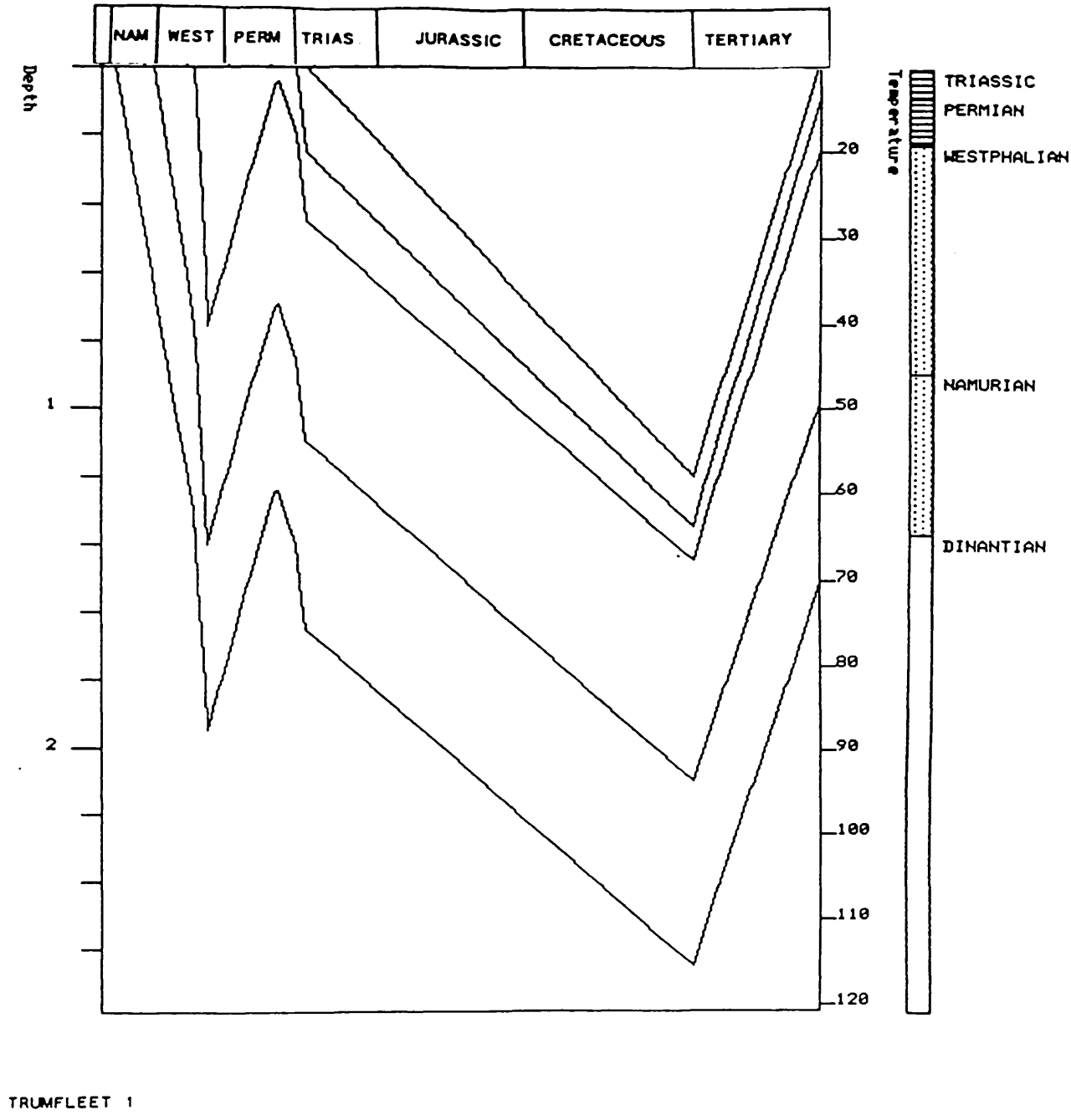
Construction of the burial history is hampered by the complex nature of post Carboniferous uplift and by the fact that the Sherwood sandstone forms the outcrop over the area. There are, therefore, no data from 230 my to the present day, and the amount of Mesozoic sediment deposited, and later eroded, has to be estimated using data from adjacent areas. Unfortunately it is not possible to use the Bunter Shale Velocity method described by Marie (1975) to estimate the amount of pre Tertiary burial, since this formation occurs near enough to the surface to have suffered from the effects of weathering, and sonic logs were not available through the relevant section of the wells.

Kent (1980) has described the burial history of the East Midlands; the major complications being a two stage uplift at the end of the Carboniferous, and uplift during the Tertiary Alpine Orogeny. Regular subsidence is assumed for the rest of the time.

Approximately 1500m of Coal Measures sediment was deposited over the Gainsborough Trough, and since the effects of block and basin topography were minimal during the Westphalian,

(Bedrock, 1984) a similar amount was probably deposited over the Trumfleet area. Taking this figure for the amount of Westphalian deposition, (1500m) it is estimated that 800m of Westphalian sediment was removed from the Trumfleet area. Approximately 1500m of post Permian sediment was deposited over the area (Kent, 1978.), and some 1200m of sediment were removed during the Tertiary. The maximum burial depths of the G. subcrenatum and the R. gracile marine bands, at the end of the Mesozoic, were therefore roughly 2150 and 2300m respectively, (ignoring compaction of mudstones). The effects of late Carboniferous and early Permian orogenic events have been simplified into a single "Hercynian" for the purpose of the modeling. The burial model is shown in Figure 9:1, the top Namurian horizon is shown in yellow.

FIGURE 9:1 BURIAL MODEL



TRUMFLEET 1

9:3. TEMPERATURE MODEL

Heat flow, and therefore geothermal gradient, variations through time are difficult to estimate but a few generalisations can be made.

The present day gradients can be estimated from borehole data. No corrected temperature data was available for the Trumfleet Field, but data from the Askern well has been published by the BGS (Burley et al, 1984). A temperature of 68° C was recorded at a depth of 1457m from a DST taken in the Dinantian. A Bottom hole temperature of 39° C at 1089m was recorded during logging of the Moss Farm Well, some three hours after circulation stopped. Assuming a present day average surface temperature of 10° C, present day geothermal gradient of 39.8° C per km is obtained for the Askern well. Since the Askern data is from a drillstem test it was used in preference to the less accurate log temperature estimate from Moss Farm.

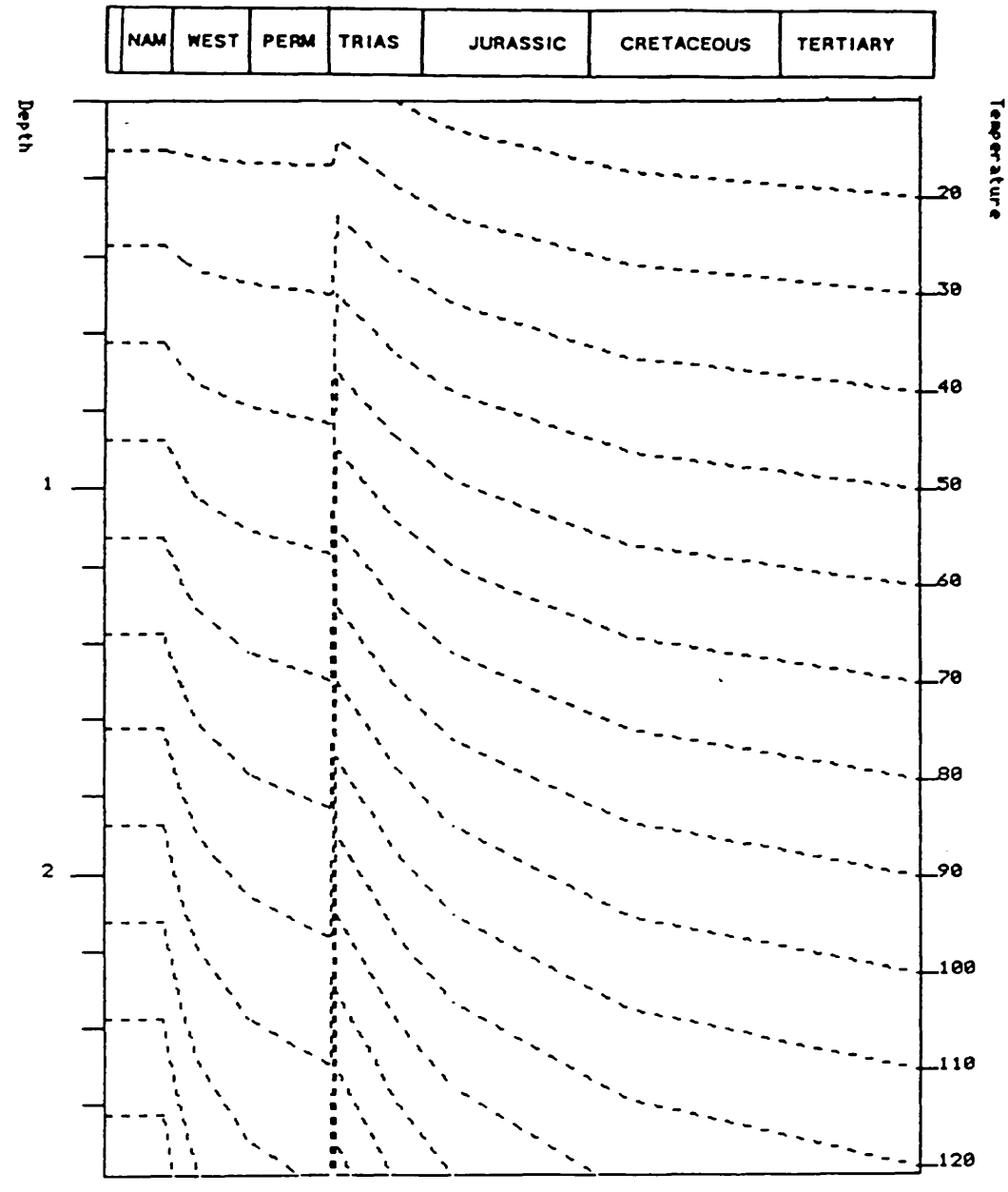
It is probable that palaeo-heatflows were highest during the period of crustal thinning which led to the formation of the Permo-Triassic basins. A conservative estimate of 50° C per Km was taken as representative gradient for the Triassic, with a surface temperature of 25°. This gradient was then exponentially decayed to the present day gradient of 40° C per Km (rounded up from 39.8). Estimating the Carboniferous gradients is more difficult. Leeder et al (1982) have attempted to use a MacKenzie (1978) type thermal subsidence model to estimate stretching of the crust during the Carboniferous.

Since the major period of tectonic subsidence, and therefore crustal thinning took place during the Dinantian (Leeder, 1982), it is likely that increased heat flows were at a maximum during the this time (MacKenzie, 1978). If very high gradients were in effect during the Carboniferous it is likely that the vitrinite reflectance gradients would preserve evidence of this. Cope (1987) has published vitrinite reflectance gradients from the Carboniferous of the Southern North Sea basin and showed that the vitrinite reflectance measurements could be matched to areas of post Triassic inversion (as determined by the Bunter Shale velocity technique described by Marie (1975)). Only in areas of deep Carboniferous erosion during the Hercynian Orogeny was there an anomalous relationship. He suggested that these effects may be due to Carboniferous heating. He also found that a modern geothermal gradient of around 30^{o} C per Km could account for the vitrinite reflectance gradients observed in the Southern North Sea.

A geothermal gradient of 40^{o} C per km has therefore been assumed for the Dinantian, decaying to an average 30^{o} per km during the thermal subsidence period (Namurian to Permian) prior to the Triassic rifting phase.

The temperature model is shown in Figure 9:2.

FIGURE 9:2 TEMPERATURE MODEL



TRUMFLEET 1
Depth in 1000 meters sub sediment surface.Temperatures in centigrade

9:4. TTI MODELING

Figure 9:3 shows the combined burial/temperature model with the superimposed iso-TTI values at 6 (yellow), 15 (green) and 60 (Blue). Table 9:1 gives approximate equivalent TTI/Ro estimates.

Table 1

TTI VALUE	Ro (%) WAPLES	approx temp ^{o}C	Hydrocarbon Windows
6	0.55	70	Onset of Oil Generation
15	0.65	90	Peak Oil Generation
60	0.95	110	
70	1.00	115	Gas generation
450	1.70	140	

Waples (1980)

The gas generation threshold is more commonly thought to correspond to an Ro of 1%, and corresponding TTI's and temperatures are given in (Table 1) for this value according to Waples (1980).

The relationship between Ro and TTI is given by the equation:

EQ 9:1,

$$\text{Log (TTI)} = 6.1841 * \text{Log(Ro)} + 2.656$$

(Estimated from worldwide data, Waples 1980)

Data which would allow an Ro/TTI relationship to be estimated specificaly for the Carboniferous of the East Midlands basin was

not available.

The horizon of interest, the Namurian/Westphalian boundary, reached a maximum burial temperature of 70° C during the Carboniferous and the maximum temperature to which the base of sediment pile (R, gracile) was subjected was almost 100° C. The estimated vitrinite reflectance value is 0.94%, less than the maximum recorded value (1.02%) but a close approximation. The weakest part of the model is the estimation of the amount of uplift due to the Alpine Orogeny. Increasing the amount of uplift (and therefore thickening the amount of pre Tertiary sediment) by 200m was found to give an estimated maximum temperature of 110° C and an estimated Ro of 1.08% (Figure 9:4). Since this empirically amended data better reflects the observed parameters it will be used in the following discussion.

FIGURE 9:3, TTI MODEL, 1200m Tertiary uplift

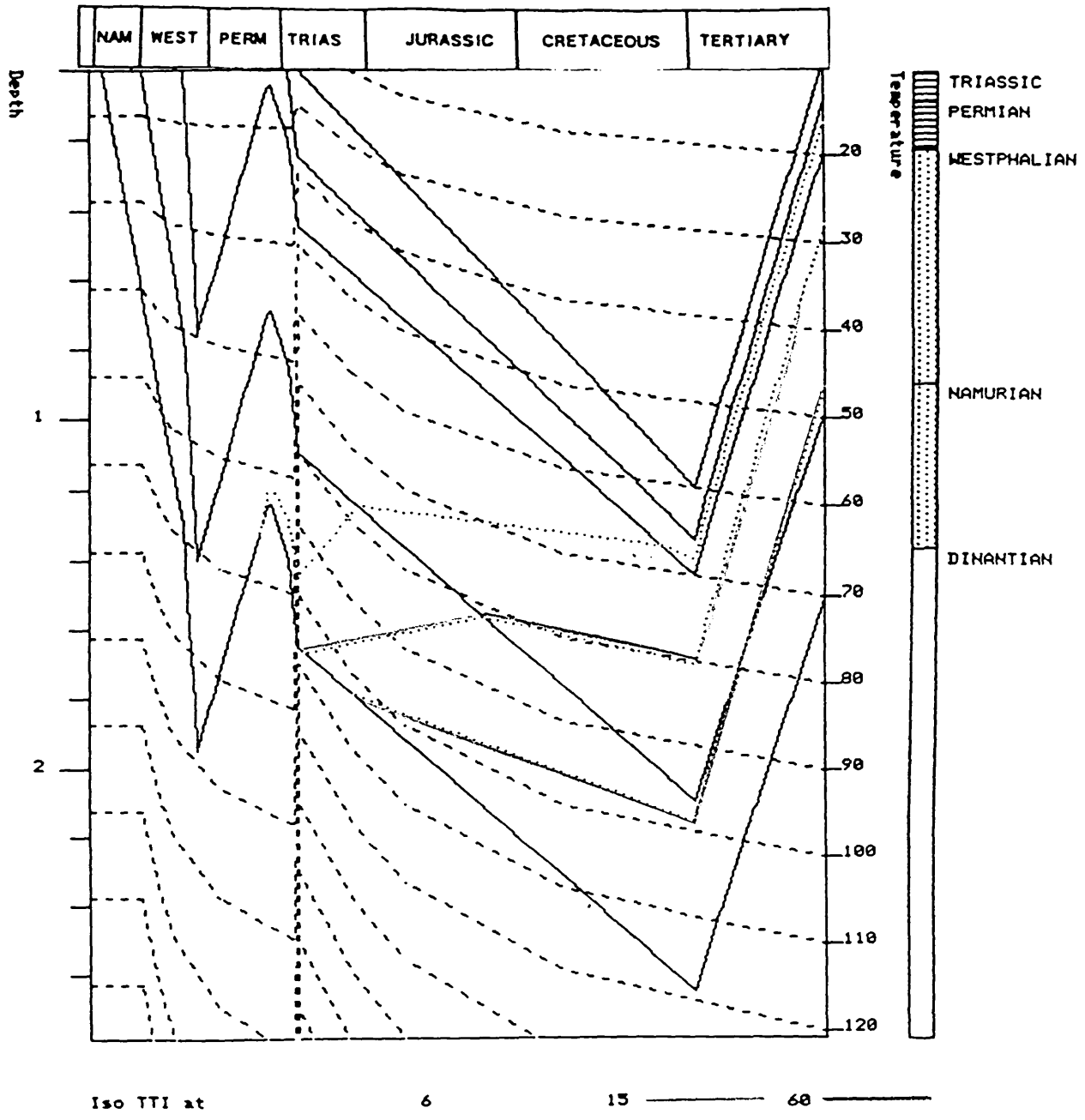
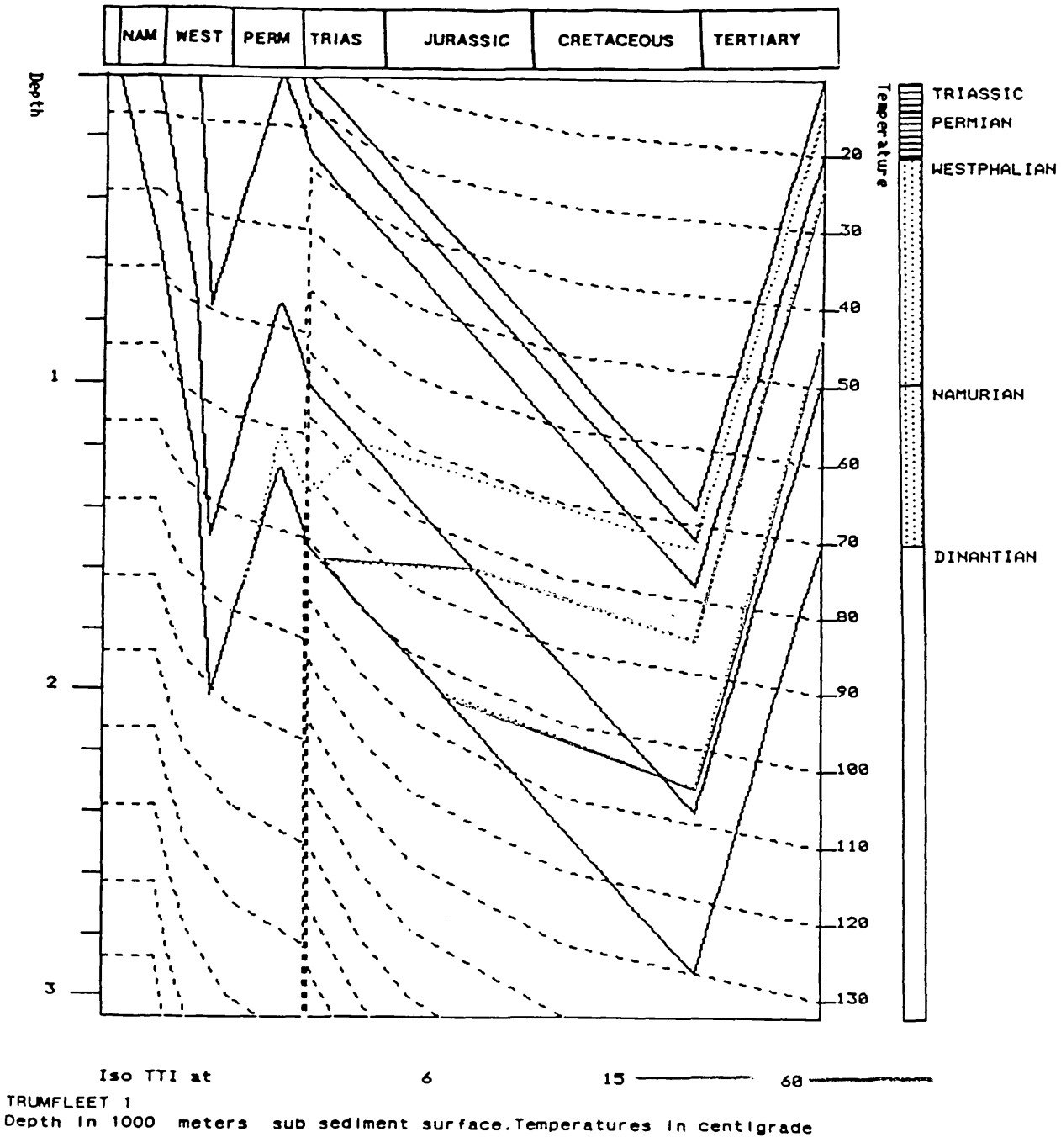


FIGURE 9:4, TTI MODEL, 1400m Tertiary uplift



9:5. MATURATION

Figure 9:5 shows that the Namurian/Westphalian boundary (G. subcrenatum) passed into the oil zone approximately 220 my, during the late Triassic. Peak oil generation occurred approximately 170my, during the Middle Jurassic and gas generation occurred approximately 70 my (late Cretaceous) immediately prior to Tertiary uplift.

The top Dinantian passed into the oil window during the Carboniferous, prior to the Hercynian Orogeny, and peak oil generation occurred about 240my during the Early Triassic. Gas generation began in the Late Triassic around 215 my.

The organic rich Lower Namurian source rocks therefore reached the various stages of maturity some time between these times, and the more deeply buried Namurian sediments of the Gainsborough Trough would have started oil and gas generation earlier.

Cope (1987) states that an Ro of 1(%) will be reached at a depth of 3500m in the Southern North Sea basin, 1000m deeper than depths to which the Trumfleet sediments have been subjected. The difference can be accounted for by the fact that Cope (1986) used a lower geothermal gradient in his modeling than was used in this study.

The age of oil, peak oil and gas generation are appreciably earlier than those postulated by Dobson (1984), and these discrepancies can be accounted for by the fact that Dobson used a uniform geothermal gradient through time. The geological history suggests that high gradients have been in effect for much of geological time, and that the present day gradients should, if anything, be lower due to uplift during the Tertiary.

The thermal model presented here is simplistic, no account has been taken of relative thermal conductivities of rock strata, and better estimates of the palaeogeothermal gradients could be achieved by quantifying the amounts of lithospheric thinning and thickening during basin formation and orogenic events. It is outside the scope of this study to attempt the detailed basin modeling, and maturation modeling required to better define these parameters, since the object of this modeling was to provide a rough temperature/burial framework on which to hang the diagenetic history outlined in Chapter Eight.

9:6. DIAGENETIC IMPLICATIONS OF THE MATURATION MODEL

The modes of formation of the various mineral species have been described in Chapter Seven, and the relative timing of diagenetic events given in Chapter Eight.

For the purposes of this discussion it is convenient to consider the study interval as a single unit between G. subcrenatum and R. cancellatum, highlighted in Figure 9:5.

Early diagenesis occurred at the surface, and only effected the sediments for a very short period of time, probably only thousands or perhaps tens of thousands of years. An insignificant time considering the scale of this study. We can therefore define the timing of early diagenesis at 315my.

Fixing the transition from the early burial stage to the burial stage is somewhat more problematic. Curtis (1982) has shown that fluids derived from mudrock diagenesis are carbonate saturated up to temperatures of up to 40° C. Since calcite and siderite cements are characteristic of this phase, the boundary between the carbonate dominated early burial phase and silicate dominated burial phase was placed at 40° C.

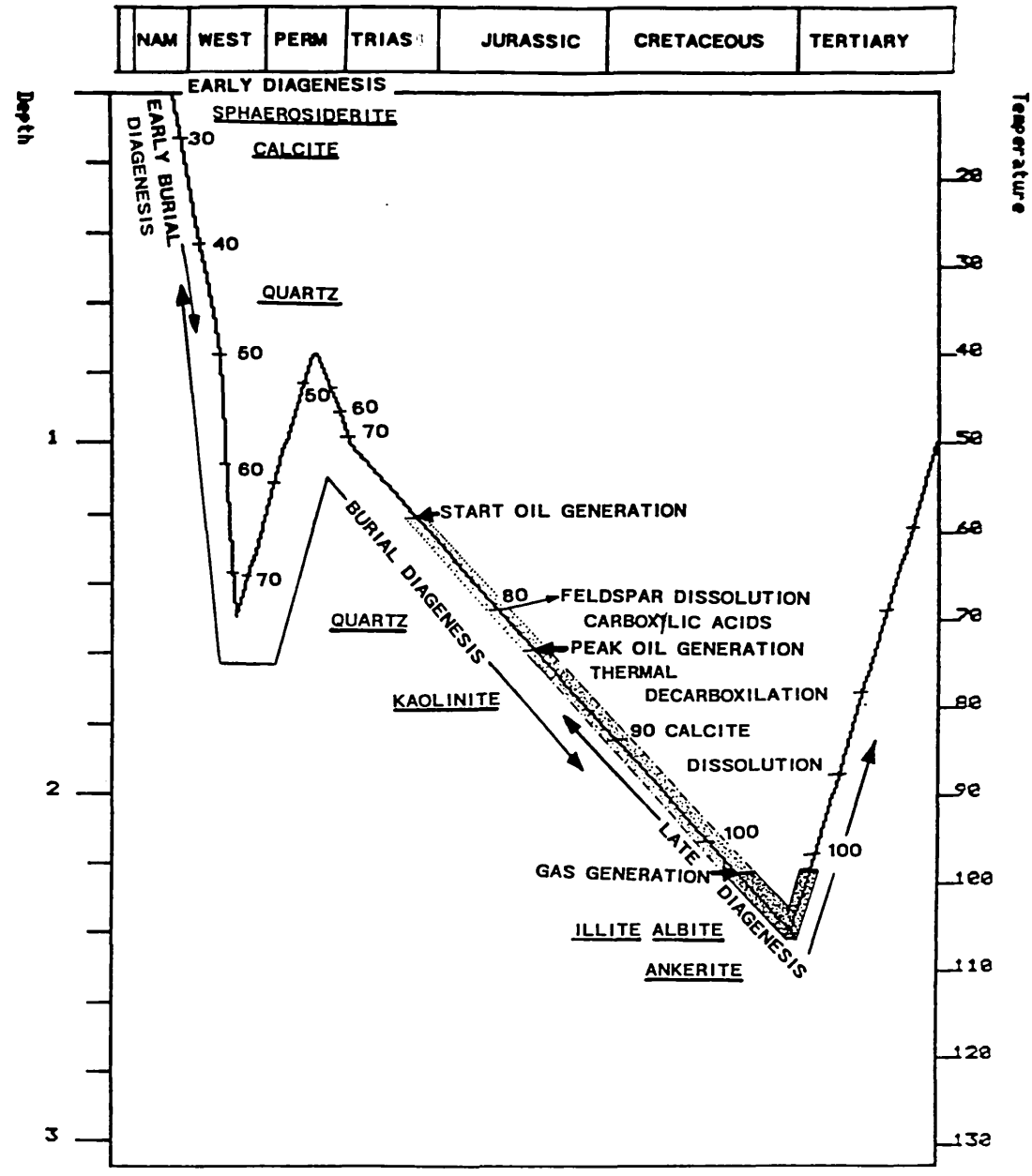
It is interesting to note that temperatures would have been reduced to around 45° C, from a maximum of 75° C, during the Carboniferous uplift phase (Hercynian). It is likely that this event had little effect on the diagenesis of the sediments, since the zone of maximum diagenetic change (80-100° C) was not reached during burial, and the uplift was not sufficient to allow meteoric waters to infiltrate the sediments. Mineral reactions probably continued as prior to uplift, but at a slower rate.

Burial diagenesis can be divided into two distinct phases. The first dominated by silica precipitation, and the second dominated by feldspar and carbonate dissolution. As has been discussed in chapters Seven and Eight, there is good evidence to suggest that feldspar dissolution occurred at temperatures of around 80° C, and the thermal decarboxylation reactions which controlled carbonate dissolution occurred at between 80 and 90° C (Curtis, 1982).

Boles (1981) and Loucks et al (1974) recorded albitisation and precipitation of iron carbonate and illite cements at temperatures in excess of 100° C in the Frio Sandstone of the Texas Gulf Coast, a similar assemblage as was precipitated in the Trumfleet sandstones during late diagenesis. The boundary between burial and late diagenesis was taken as being close to this temperature.

Uplift would have brought the temperature of the study interval to around 50° C. No evidence of retrogressive diagenesis can be seen in the last phases precipitated. Since the solubility of silica increases with temperature and the solubility of carbonates decreases, it might be expected that silica would precipitate, and ankerite would suffer dissolution during uplift. The ankerite appears unaltered by this uplift, and there is no evidence of silica precipitation (although the amounts precipitated would be insignificant given the low solubilities). It appears therefore that uplift had little effect on the diagenesis of the sandstones, even although the present day formation fluids appear to be out of equilibrium with with last phases precipitated (Downing and Howitt, 1969).

FIGURE 9:5 DIAGENESIS RELATED TO BURIAL



TRUMFLEET 1
Depth in 1000 meters sub sediment surface.Temperatures in centigrade

(NOTE: since writing this Chapter, unpublished fission track data, (Dr Paul Green, pers comm) re-inforces the interpretation herein, namely, of 1.5 km Tertiary uplift, with maximum burial at the end of the Cretaceous.)

CHAPTER TEN

CONTROLS ON RESERVOIR PROPERTIES: DISCUSSION AND CONCLUDING REMARKS

10:1. INTRODUCTION

The diagenetic history outlined in the previous two chapters defined some of the factors controlling porosity, but it useful to consider the effects on permeability also, and to try and quantify the relative importance of the various mechanisms of porosity reduction. It was only possible to take core plugs for petrophysical examination from the Moss Farm Well and the following data refer to that well. Porosity and permeability measurements were taken a approximately 1m intervals where suitable samples were available. Only 34 plugs were cut for study.

10:2. PRIMARY CONTROLS-DEPOSITIONAL ENVIRONMENT

Figure 10:1 shows averages of total cement, total clay and measured porosity plotted as a stacked bar graph for sandstones of the overbank/crevasse and distributary channel facies. Since only five plugs were available for the overbank and crevasse facies, these facies have been combined. Twenty nine plugs were available for the distributary channel facies.

Table 10:1 shows the same data with the addition of arithmetic average permeability measurements for the same facies.

Figure 10:1 shows that there is very little difference in

FIGURE 10:1

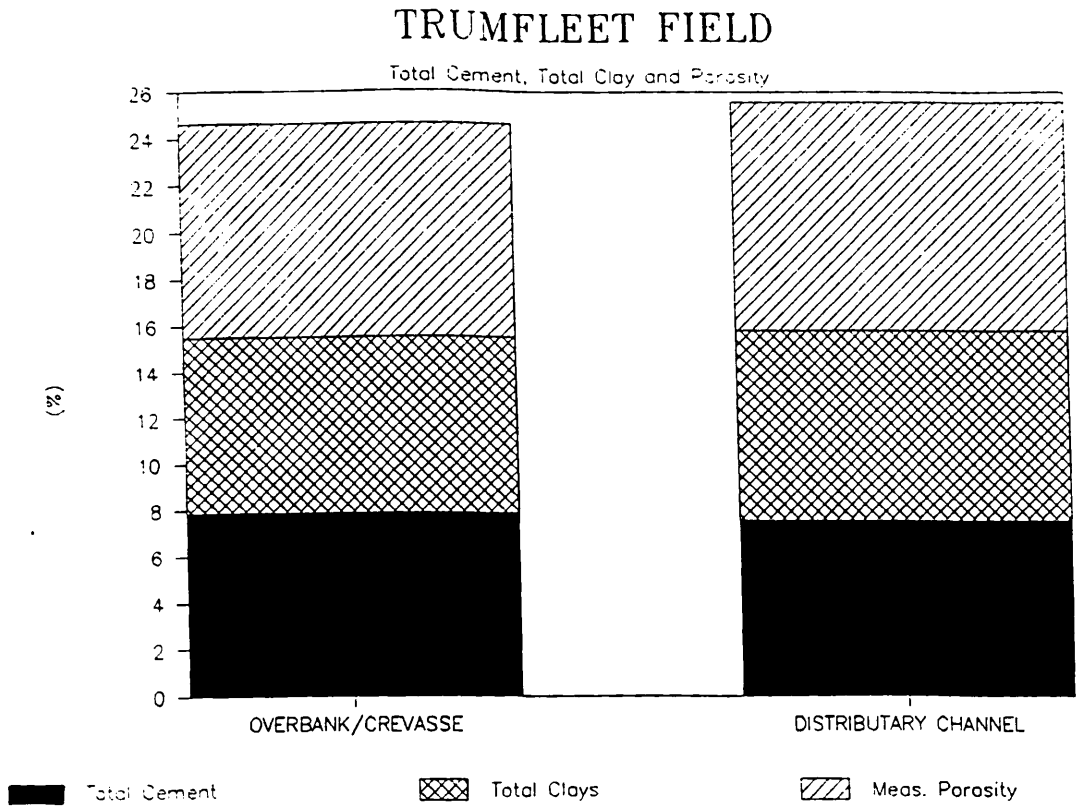


TABLE 10:1

	TOTAL CEMENT	TOTAL CLAYS	MEASURED POROSITY	MEASURED PERMEABILITY
OB/CREVASSE	7.84	7.66	9.12	0.143 (mD)
DIST.CHANNEL	7.52	8.27	9.79	0.380 (mD)

terms of total cement, clay content or measured porosity for distributary channel sands and overbank/crevasse sands. The permeability values are also very similar, with a slight decrease in the overbank and crevasse facies which can be accounted for by poorer sorting in that facies, however, a variation in permeability of less than an order of magnitude is probably not significant. More significant, and not accounted for in the above discussion, is the well laminated nature of the crevasse sandstones, which will affect vertical permeability, and their limited areal distribution. Although the petrophysics of the core plugs may be similar the crevasse/overbank facies must be considered a poorer reservoir than the thicker and more areally extensive sandstones of the distributary channel belt, which have sheet like geometries (Bristow, 1988)

From the limited data above, however, it appears that the depositional environment does not have a significant effect reservoir properties on the microscopic scale, even although the grain size and sorting are significantly different. This probably reflects the fact that most porosity is secondary, and the original depositional rock texture has been overprinted to such an extent by diagenetic processes that it has little effect on the presently observed reservoir properties.

10:3. SECONDARY CONTROLS-DIAGENETIC PROCESSES

Figure 10:2 shows the porosity/permeability relationship for the measured samples from the Moss Farm well, and Figure 10:3 shows the total cements versus measured porosity for the available core plugs from the same well. Figure 10:4 plots the total authigenic cements versus visible porosity for all the samples studied.

Most granular (sandstone reservoirs) show log linear relationship between porosity and permeability, but this is clearly not the case for the data shown in Figure 10:2. Some, but not all, of this scatter can be accounted for by the problems of measuring low permeability core plugs.
The fact that porosity and permeability can partially be explained by the abundance of microporosity in the rock, as well as the preponderance of secondary, inter-grain dissolution porosity which is observed in the rocks. This exists as oversized, isolated pores (often filled with kaolinite), which have low connectivity. In addition, stereoscopic SEM examination of pore casts shows that there is also a significant proportion of lammellar porosity between grains and overgrowths.

The high degree of scatter in the porosity/permeability plot is a caused by the fact that the original intergranular porosity has been lost, in this case by diagenetic mineral precipitation and dissolution.

FIGURE 10:2

TRUMFLEET FIELD

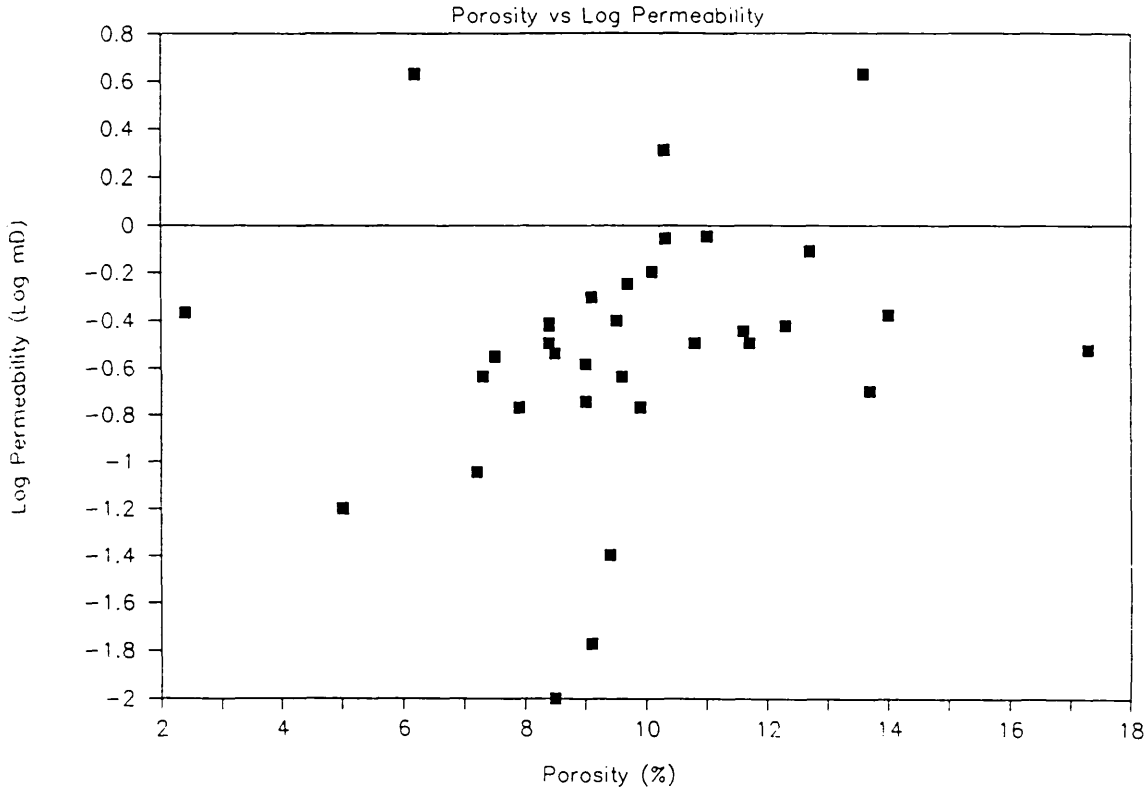


FIGURE 10:3

NCB MOSS FARM

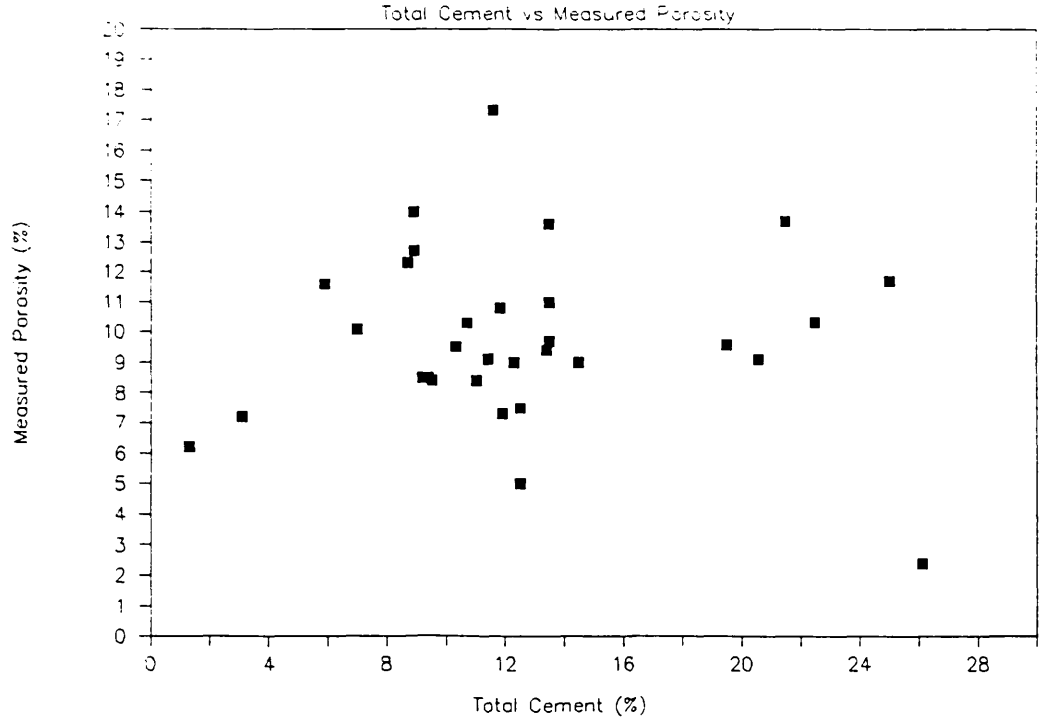
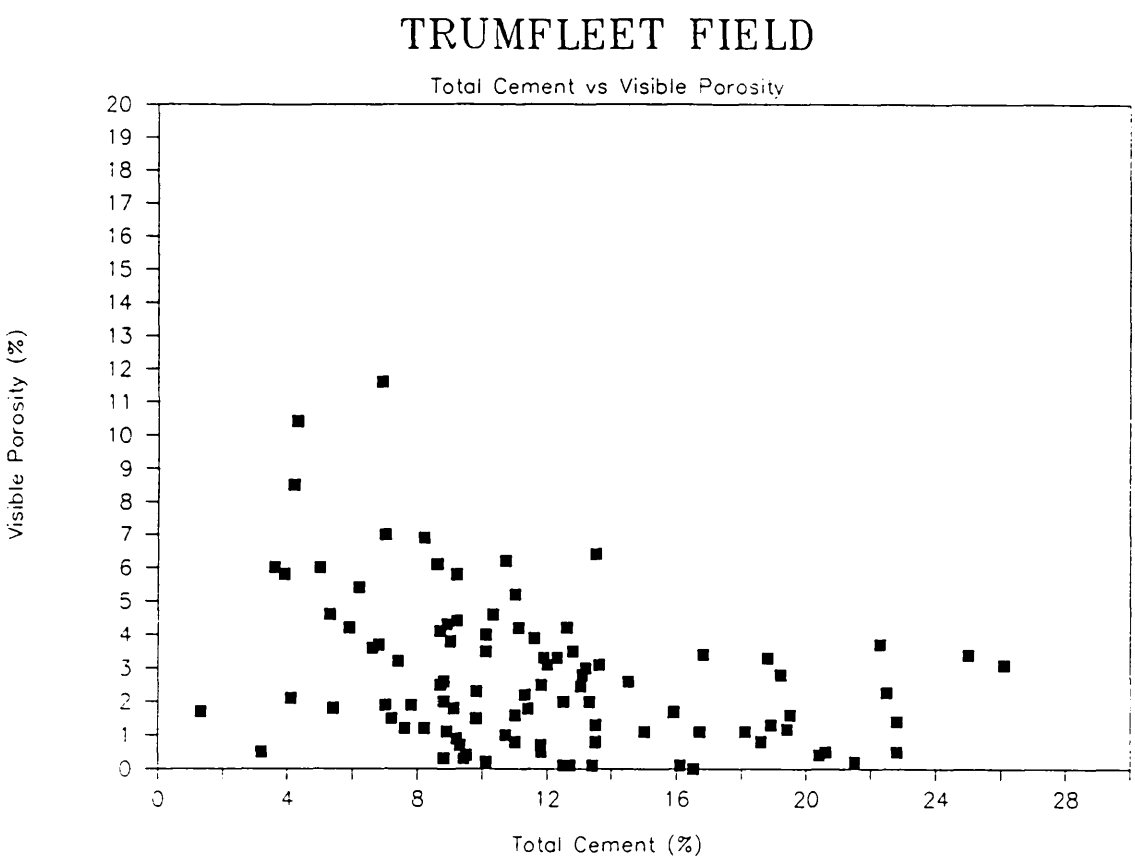


FIGURE 10:4


TRUMFLEET FIELD
Total Cement vs Visible Porosity
Visible Porosity (%)
20
19
18
17
16
15
14
13
12
11
10
9
8
7
6
5
4
3
2
1
0
0
4
8
12
16
20
24
28
Total Cement (%)

10:3.1. DISCUSSION: THE SHORTCOMINGS OF MODAL ANALYSIS DATA.

If the reservoir properties were controlled entirely by authigenic cements a straight line relationship should exist between porosity and total authigenic cement (clays and mineral cements). Figure 10:4 shows that there is a vague trend between total cement and porosity, however, this data is somewhat flawed since it was not possible to quantify the amounts of authigenic quartz present. The role of compaction should also be considered: it is difficult (if not impossible) to estimate the absolute amount of compaction in a 3 dimensional granular solid from a 2 dimensional thin section. At deposition the sediments would have had up to 40% original porosity, yet it is normal to find uncemented reservoir sandstones in which, apparently, no compaction has occurred, with porosities of around 20 %. The first 50% reduction in porosity must have occurred by compaction yet it leaves no trace in a thin section. In a hexagonally packed set of spheres it is highly unlikely that any randomly cut section will intersect a grain contact, ie, in an uncompacted sediment "floating" grains should be the norm. When we consider poorly sorted sediments and angular grains as well as undetected quartz overgrowths, the difficulty in estimating compaction is magnified. What is clear is that porosity in the Trumfleet sandstones has been reduced by a combination of authigenic mineral growth and compaction, and increased by the dissolution of framework grains and pore filling cements. There

is clearly a complex inter-relationship between these diagenetic events and the initial naive aim of this study to quantify the relative importance of these various phenomena has not been realised. The evidence suggests, however, that authigenic mineral precipitation and dissolution are the most important controls on reservoir properties in the Trumfleet sandstones for the following reasons:

1 The sheer volume of cements recorded (see figures 7:1 to 7:11), bearing in mind that it was not possible to quantify quartz cements.

2 CL photomicrographs show only minor long grain contacts, with abundant quartz overgrowths, indicating that advanced compaction has not taken place.

3 Almost all the porosity observed in thin section is secondary, caused either by the dissolution of carbonate or framework grains (mainly feldspar). It is clearly very difficult to estimate the amount of cement removed.

No simple answers have been provided by the crossplotting of data in this study, and the fit of the data is too poor to allow statistical analysis to be performed. Given the care taken to gather this data the poor correlations give information about the very complex nature of grain-cement-clay-porosity systems in rocks which have undergone complex diagenesis like these and they also give information on the nature of the

data gathering, specifically the shortcomings of point counting.

The main problems encountered in trying to quantify the modal percentages are outlined below.

CLAY MINERALS: No reliable method exists for quantifying the modal percentages of clay minerals in thin section. Image analysis of BSE images may prove useful, but the geometrical problems caused by the presence of platy or fibrous minerals will be difficult to resolve.

MICROPOROSITY: The measurement of microporosity will not be possible until the problems of measuring clay mineral modes are resolved.

COMPACTION: No reliable method of estimating compaction has been published to my knowledge. The methods outlined by Pettijohn (1972), Calver (1971) provide relative indexes of compaction but cannot provide a measure of the amount of compaction which has taken place. Modern image analysis techniques combined with CL and SEM imaging are clearly required to resolve these problems.

QUARTZ OVERGROWTH: The problem of differentiating between detrital and authigenic quartz in not unique to the study. The problem can be resolved by the of CL images. This is, however, clearly not practicable for a large number samples given the time taken to generate CL images on the present generation of equipment.

CEMENT REPLACEMENT: It is impossible to estimate the amount of grain replacement which has taken place.

When the above problems are resolved the full potential of modal analysis in petrographic analysis can be realised.

10:4. CONCLUDING REMARKS

The main conclusions of this study are listed in the first chapter, and there is little point in recounting them here. What I hope that this study has shown is not only that careful collection of modal analysis data is essential in any diagenenetic study, but that it is equally important to be aware of its' limitations.

The burial model outlined in Chapter 9 is an attempt to produce a model of the diagenetic history for the Trumfleet Field which has a wider, predictive function. Like all good models it is capable of being tested: use of isotope and fluid inclusion studies to delimit the temperatures of precipitation of the authigenic phases present, and fission track analysis to better define the thermal history of the basin will demonstrate the validity or invalidity of the model. It will be interesting to see if the model presented here will stand the test of time without major alteration. The science of diagenesis is moving ahead quickly and data from the East Midlands which will confirm or refute this model must soon be published.

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Addendum.

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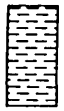
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KEY TO LITHOLOLOGICAL SYMBOLS

LITHOLOGY


SANDSTONE


SILTSTONE


SHALE


MARINE BANDS


COAL

SEATEARTH

SEDIMENTARY STRUCTURES

Parallel Lamination


Planar Cross Lamination


Trough Cross Lamination


Ripple Lamination


Ripple Drift Lamination

Wavy Lamination


Slumped Bedding


Load Casting


Bioturbated

Fractures

BED BOUNDARIES

Sharp

Gradational

Erosive


Load Cast

QUALIFIERS

Calcareous


Dolomitic

Carbonaceous


Noduler


Pebbly

Silty


Macrofossils


Plant Fragments

GRAIN SIZE

g Gravel

C V. Coarse

c Coarse

m Medium

f Fine

F V. Fine

s/c Silt & Clay

POINT COUNT DATA; TRUMFLEET FIELD

			DETRITAL GRAINS					CLAYS				CARBONATES			VISIBLE			GRAIN SIZE			TEXTURE			MEASURED POR		
WELL	FM	DEPTH	QTZ	FELD	MICA	LITH	ORG	OPD	KAO	ILL	CHLOR	CALC	DOL	SID	POR	OTHERS	MAX	MEAN	MIN	SRT	R	S	POR	PERM	FACIES
NCB MOSS FARM	ROUGH ROCK	918.00	76.30	1.90	7.60		1.10	0.10	5.40	2.50	0.50	1.00	0.30	2.80	0.10		VFU	VFL	SLT	M	SA	L	5.00	0.06	OVERBANK
NCB MOSS FARM	ROUGH ROCK	919.00	66.50	5.80	4.80		0.70		7.60	4.30	2.30	0.80	0.40	5.20	0.50		VFU	VFL	SLT	M	A	L	9.10	0.02	OVERBANK
NCB MOSS FARM	ROUGH ROCK	920.70	62.00	10.40	5.00		0.10		8.20	2.10	1.00		2.50	7.70	0.20	0.30 ZIRCON	CRSL	FU	VFL	P	SA	L	13.70	0.20	OVERBANK
NCB MOSS FARM	ROUGH ROCK	924.54	74.20	7.30	1.70		3.70		5.30	2.00	0.10		1.10	0.20	4.10	0.30 ZIRCON	FL	VFL	SLT	G	SA	L	12.30	0.38	DISTRIBUTARY CHANNEL
NCB MOSS FARM	ROUGH ROCK	926.25	70.80	12.30	2.30	3.40	1.40		6.00	1.10	0.50	1.30			1.10		FU	VFU	VFL	G	SA	L	12.70	0.78	DISTRIBUTARY CHANNEL
NCB MOSS FARM	ROUGH ROCK	927.05	77.20	9.20	1.90	0.50	0.60	0.40	3.50	1.10	0.50	0.80			4.20	0.20 SKF	FU	FL	VFL	G	SA	L	11.60	0.36	DISTRIBUTARY CHANNEL
NCB MOSS FARM	ROUGH ROCK	929.00	44.80	8.30	3.20	2.00	3.10	0.10	1.20	0.10					1.70		FU	FL	VFL	G	SA	L	6.20	4.27	DISTRIBUTARY CHANNEL
NCB MOSS FARM	ROUGH ROCK	930.00	74.30	7.20	1.30	0.20			11.10	1.30	0.20	10.80	23.80		2.80		FU	FL	VFL	M	SA	L	9.90	0.17	DISTRIBUTARY CHANNEL
NCB MOSS FARM	ROUGH ROCK	931.87	74.10	8.20	1.40	0.20			10.60	1.10	1.30		0.20	1.30	2.60		FU	FL	VFL	M	SA	L	9.00	0.18	DISTRIBUTARY CHANNEL
NCB MOSS FARM	ROUGH ROCK	932.44	51.30	0.50	1.60	0.20			10.50	0.60			0.60	0.10	0.50	0.80 SKF							10.80	0.32	DISTRIBUTARY CHANNEL
NCB MOSS FARM	ROUGH ROCK	934.20	63.80	11.30	1.50	2.30	0.50		13.60	1.00	2.00	0.10	24.70		3.50								7.90	0.17	DISTRIBUTARY CHANNEL
NCB MOSS FARM	ROUGH ROCK	935.50	60.30	7.90	1.00	0.70	1.70		7.30	0.50	0.20	17.00			3.40								11.70	0.32	DISTRIBUTARY CHANNEL
NCB MOSS FARM	ROUGH ROCK	936.76	64.70	12.40	0.50	1.50	0.10		11.30	0.10	1.50			0.60	6.40								13.60	4.25	DISTRIBUTARY CHANNEL
NCB MOSS FARM	ROUGH ROCK	937.75	58.20	9.40	2.10	4.60	10.50		6.20	0.40	3.00	3.40	0.50		0.80								11.00	0.90	DISTRIBUTARY CHANNEL
NCB MOSS FARM	HWR	961.00	36.60	6.30	1.60		1.40	0.30	1.40	0.10	0.30	50.60	0.40												CREVASSE
NCB MOSS FARM	HWR	962.90	64.90	7.40	6.50		3.30	1.00	8.10	4.70	0.60	0.10	0.20			0.20 ZIRCON									CREVASSE
NCB MOSS FARM	PHG	997.50	81.80	4.70	2.30	0.40		0.90	1.70	0.10	4.10		3.60		0.40		FL	VFU	SLT	M	A	L	8.40	0.38	CREVASSE
NCB MOSS FARM	PHG	1002.84	72.30	10.30	3.60	0.60	0.10		3.30	3.10	0.60		3.20	3.20	0.10		FL	VFU	SLT	M	SA	M	9.40	0.04	OVERBANK
NCB MOSS FARM	PHG	1005.00	67.00	9.30	2.30		0.30		13.80	1.70	3.00		0.70	0.30	1.60								9.60	0.23	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1008.00	72.40	10.70	1.20		0.30		6.90	1.60	0.80		1.30	0.80	1.80								9.10	0.50	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1007.00	52.90	6.60	0.80	0.40	1.50		0.90	0.20		26.20	9.60	0.30		0.20 TOUR+							8.40	0.39	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1009.00	66.40	17.80	2.00	0.40	0.40		1.80	4.30	0.40		0.30	0.20	7.00	ZIRCON	G	CRSL	SLT	VP	SA	M	8.70	0.22	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1009.80	68.50	13.50	1.20			2.10	5.20	1.70	1.70		0.20	0.10	4.30	1.40 RUTILE							10.10	0.64	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1011.30	71.50	11.00	0.50	0.90			4.20	6.00			1.70		3.30	0.90 BARITE							14.00	0.42	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1012.45	72.10	11.10	0.80	0.40			5.40	3.90	0.70		1.20	0.40	3.90								7.30	0.23	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1013.08	71.70	11.70	0.60				6.20	3.70	1.70		0.30	0.40	3.30								17.30	0.30	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1014.68	69.10	17.10	0.10				5.90	2.70			0.40	0.20	4.40								9.00	0.26	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1016.00	68.80	15.10	1.00	0.30		0.10	7.20	4.60	0.40		0.30		2.00	0.20 BARITE							8.50	NA	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1017.00	70.90	11.20	0.80				3.80	5.60	0.70		0.20	0.40	6.20								7.50	0.28	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1019.00	73.20	15.00	1.50	0.40		0.20	4.40	3.40	0.50		1.00	0.10	0.30								10.30	2.04	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1020.00	73.80	11.10	0.20		0.50	0.10	7.80	5.00	0.40		0.20	0.10	1.30								8.50	0.29	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1021.80	69.70	12.90	0.40	0.60			5.00	1.50	1.60		1.30	1.60	5.20								9.70	0.57	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1022.00	81.40	7.30	2.10	5.70			0.20	0.20	0.10			2.60		0.30 SRF							8.40	0.32	DISTRIBUTARY CHANNEL
NCB MOSS FARM	PHG	1023.00	71.30	12.80	0.60				3.00	0.70	1.10	4.10	0.30	1.10	4.60								7.20	0.09	DISTRIBUTARY CHANNEL
ASKERN	ROUGH ROCK	2990.00	72.80	6.50	2.50		0.50		6.90	4.60	0.10		1.20	3.10	1.70		MU	FU	SLT	G	SR	L			
ASKERN	ROUGH ROCK	2996.00	72.20	8.20	1.50	0.40	0.40	0.90	10.20	2.80	0.20		2.00	1.30			MU	FU	SLT	G	SR	M			DISTRIBUTARY CHANNEL
ASKERN	ROUGH ROCK	2999.00	71.00	7.60	1.60			0.30	10.70	3.00	0.40		1.30	3.20	0.80	IR GLAUCONY	ML	FL	SLT	M	SA	M	NA	NA	DISTRIBUTARY CHANNEL
ASKERN	ROUGH ROCK	3020.00	61.80	7.00	9.80		4.70	0.40	9.70	1.70	0.30		0.50	3.90	0.10		FL	VFU	SLT	G	SA	M			DISTRIBUTARY CHANNEL
ASKERN	ROUGH ROCK	3024.00	44.30	8.80	1.70		2.00		4.20	0.90	0.30	23.70	13.80				FL	VFU	VFL	G	SA	M			DISTRIBUTARY CHANNEL
ASKERN	BHF	3175.00	65.00	4.30	14.00		0.40	0.10	4.00	2.00	4.50		0.10	2.60	3.00		FU	VFU	SLT	G	SA	M			DISTRIBUTARY CHANNEL
ASKERN	BHF	3178.00	71.60	6.40	4.50			0.40	3.50	2.30	5.60			1.40	3.50		FU	VFU	SLT	G	SA	M			DISTRIBUTARY CHANNEL
ASKERN	BHF	3183.00	74.80	9.20	3.30		0.80	0.20	5.20	1.30	2.70		1.10	0.40	1.00		MU	FL	VFU	M	SA	M			DISTRIBUTARY CHANNEL
ASKERN	BHF	3188.00	69.20	11.70	2.00	1.60	0.70	0.20	3.80	1.30	1.40		7.10	0.40			MU	FU	VFU	G	SA	M			DISTRIBUTARY CHANNEL
ASKERN	BHF	3189.00	75.60	6.50	3.50		1.00		4.00	2.30	3.50		0.20	1.80	0.70	0.30 SPHENE	MU	FU	VFU	G	SA	M			DISTRIBUTARY CHANNEL

POINT COUNT DATA; TRUMFLEET FIELD

ASKERN	PHG	3287.00	69.90	7.70	6.80	0.10	2.30	0.10	1.50	6.70	0.90		0.30	3.30	0.10	0.20 AP +	FL	VFU	SLT	G	SA	M	DISTRIBUTARY CHANNEL
ASKERN	PHG	3290.00	69.80	10.20	4.50	0.20	0.40	0.50	8.40	2.90	0.50		0.40	1.10	2.00	0.10 ZIRCON	CRSU	MU	VFL	VP	A	L	DISTRIBUTARY CHANNEL
ASKERN	PHG	3292.00	72.00	14.00	1.30	0.60		0.30	5.70	0.90	1.00		3.40		0.80		CRSU	MU	VFL	P	SA	L	DISTRIBUTARY CHANNEL
ASKERN	PHG	3295.00	71.60	17.20	2.30			0.10	5.30	1.10	0.50		0.50	0.20	1.20		CRSU	MU	SLT	P	SA	L	DISTRIBUTARY CHANNEL
ASKERN	PHG	3298.00	70.50	16.80	2.20		0.50		2.10	3.10	2.00		0.60		1.90	0.40 ZIRCON	CRSU	ML	VFL	P	A	L	DISTRIBUTARY CHANNEL
ASKERN	PHG	3302.00	74.80	11.40	1.80		0.40	0.20	5.80	0.90	1.30		0.80		2.60		FU	FU	VFL	G	SA	M	DISTRIBUTARY CHANNEL
ASKERN	PHG	3305.00	73.30	11.90	2.30	0.10		0.20	4.50	2.90	2.90		0.60	0.10	1.60		ML	FU	SLT	G	A	M	DISTRIBUTARY CHANNEL
ASKERN	PHG	3309.00	76.30	13.30	0.90	0.50		0.10	2.80	2.40	1.80				1.90		ML	FL	VFU	M	A	M	DISTRIBUTARY CHANNEL
ASKERN	PHG	3311.00	78.40	11.90	1.50	0.10			4.90	0.90	0.50		0.20	0.70	1.50		CRSU	ML	VFU	M	A	L	DISTRIBUTARY CHANNEL
ASKERN	PHG	3314.00	73.30	13.50	1.30				4.60	2.20	0.90		0.40	0.70	2.00		G	CRSU	VFU	VP	A	M	DISTRIBUTARY CHANNEL
ASKERN	PHG	3317.00	78.00	12.80	1.70			0.20	1.40	0.90	1.10		0.60	0.10	2.10	0.50 ZIRCON	MU	FU	VFU	P	SA	L	DISTRIBUTARY CHANNEL
ASKERN	PHG	3320.00	74.70	13.40	2.00				5.10	1.30	0.60		0.60	0.60	1.20		G	CRSL	VFL	P	SA	M	DISTRIBUTARY CHANNEL
ASKERN	PHG	3324.00	73.60	12.80	1.70	0.20		0.10	7.10	1.20			1.20	0.30	1.50		CRSU	CRSL	FL	P	SA	M	DISTRIBUTARY CHANNEL
ASKERN	PHG	3327.00	73.60	11.80	4.10			0.20	0.60	3.20	1.20		0.30		4.60		MU	FU	SLT	P	A	L	DISTRIBUTARY CHANNEL
ASKERN	PHG	3330.00	78.70	9.60	1.50	0.50		0.10	0.80	1.00	1.40		0.40		6.00		MU	FU	SLT	P	A	L	DISTRIBUTARY CHANNEL
ASKERN	PHG	3334.00	82.50	9.90	0.50				2.70	0.80	0.50	0.10	1.30		1.80		CRSU	MU	VFU	P	A	L	DISTRIBUTARY CHANNEL
ASKERN	PHG	3337.00	72.70	10.30	1.80	0.20		0.10	1.70	1.10	0.60		0.10	0.80	10.40	0.20 ZIRCON	CRSU	FL	SLT	M	SA	M	DISTRIBUTARY CHANNEL
ASKERN	PHG	3340.00	74.00	13.50	1.20	1.30			3.80	0.80	1.80		2.70	0.10	0.90		CRSU	CRSL	VFU	P	SA	M	DISTRIBUTARY CHANNEL
ASKERN	PHG	3343.00	74.80	14.30	0.40	0.10	0.40		4.00	0.40	2.70		1.80	0.40	0.70		CRSU	MU	VFL	P	A	M	DISTRIBUTARY CHANNEL
ASKERN	PHG	3345.00	82.30	6.60	0.10			0.50	6.10	0.30	0.40				3.70		CRSU	MU	FL	G	SA	M	DISTRIBUTARY CHANNEL
ASKERN	PHG	3346.00	65.80	19.20	3.10	0.50			3.50	1.80	0.40		0.30	0.20	5.40		CRSU	CRSL	VFU	P	A	L	DISTRIBUTARY CHANNEL
ASKERN	PHG	3348.00	65.00	12.70	1.60		0.50		14.20	0.90	0.90		1.90	1.00	1.30		CRSU	CRSL	VFU	P	A	L	DISTRIBUTARY CHANNEL
ASKERN	PHG	3351.00	70.40	13.30	1.50	0.50	0.30	0.90	5.00	1.70	1.70		1.40		2.30		CRSU	MU	VFU	VP	A	L	DISTRIBUTARY CHANNEL
ASKERN	PHG	3354.00	93.20	2.40	0.10	0.10			3.40	0.10			0.80				CRSU	ML	SLT	VP	A	H	DISTRIBUTARY CHANNEL
ASKERN	PHG	3357.00	70.80	12.40	0.10	0.40		0.30	11.50	1.80	1.40		0.30		1.10		CRSU	CRSU	VFU	P	A	M	DISTRIBUTARY CHANNEL
ASKERN	PHG	3360.00	68.30	15.90	0.70	0.30	1.00	0.30	8.50	1.30	1.10		0.30	0.10	2.20		CRSU	CRSL	VFU	P	A	M	DISTRIBUTARY CHANNEL
ASKERN	PHG	3364.00	62.50	18.70	2.20	0.60	0.50	0.30	8.30	1.20	0.70		0.80	0.10	4.20		CRSU	ML	VFU	P	SA	L	DISTRIBUTARY CHANNEL
ASKERN	PHG	3368.00	58.50	15.70	1.00	0.80	0.60	0.10	11.10	1.00	0.30	9.80	0.50	0.10	0.50		CRSU	FU	VFL	P	SA	M	DISTRIBUTARY CHANNEL
ASKERN	PHG	3370.00	71.50	12.30	1.00	0.80	0.60	0.50	4.60	0.60	0.50	1.40	3.00		3.50		CRSU	FU	VFU	P	SA	M	DISTRIBUTARY CHANNEL
TRUMFLEET 1	ROUGH ROCK	3145.00	49.10	8.60	6.70		14.80	2.60	6.30	2.60	0.70	1.10	0.20	5.80	1.10	0.20 ZIRCON	VFU	VFL	SLT	M	A	L	DISTRIBUTARY CHANNEL
TRUMFLEET 1	ROUGH ROCK	3148.00	73.60	6.10	4.70	1.30	4.00		4.30	1.20	0.20		3.80	0.60	0.20		FL	VFL	SLT	M	SA	L	DISTRIBUTARY CHANNEL
TRUMFLEET 1	ROUGH ROCK	3156.00	59.40	6.40	6.60	0.40	9.60	0.50	10.80	3.40			0.20	2.50			L	VFL	SLT	G	A	L	DISTRIBUTARY CHANNEL
TRUMFLEET 1	ROUGH ROCK	3164.00	66.40	7.80	0.40		2.80		13.80	1.40	0.40		3.20	0.40	2.80		CRSL	MU	FL	G	SA	L	DISTRIBUTARY CHANNEL
TRUMFLEET 1	ROUGH ROCK	3167.00	69.30	7.70	0.90	0.40	0.60	0.20	15.80	0.70			1.70	2.20	0.40		MU	FU	VFL	M	SA	M	DISTRIBUTARY CHANNEL
TRUMFLEET 1	ROUGH ROCK	3170.00	68.20	9.60	0.20	0.70	0.30		10.30		0.10		12.40		1.40		MU	FL	FU	VG	SR	H	DISTRIBUTARY CHANNEL
TRUMFLEET 1	BHF	3334.00	84.90	4.70	2.30	0.50			4.90	2.20	0.10			0.20			ML	FL	VFL	P	A	L	DISTRIBUTARY CHANNEL
TRUMFLEET 1	BHF	3335.00	90.00	2.10	1.00	0.20	0.10		3.50	2.50	0.40			0.20			ML	FL	VFL	G	SA	H	DISTRIBUTARY CHANNEL
TRUMFLEET 1	BHF	3358.00	77.90	2.40	1.00			0.10	1.60	5.10				0.20	11.60		CRSU	ML	VFL	VP	A	L	DISTRIBUTARY CHANNEL
TRUMFLEET 1	BHF	3364.00	84.60	3.30	1.50	0.60	2.30		3.00	1.60	0.10		0.60	1.20			MU	FU	VFL	P	SA	M	DISTRIBUTARY CHANNEL
TRUMFLEET 1	BHF	3367.00	92.80	3.30	0.30				1.60	0.10			1.50		0.50		MU	FU	SLT	M	A	L	DISTRIBUTARY CHANNEL
TRUMFLEET 1	BHF	3371.00	90.10	3.70	0.70	0.30	0.20		2.90	0.70			0.60		8.50		MU	FL	SLT	M	SA	L	DISTRIBUTARY CHANNEL
TRUMFLEET 1	BHF	3374.00	77.60	4.50	0.10		0.30		2.90	0.10	0.10		6.00				CRSU	ML	VFU	VP	R	H	DISTRIBUTARY CHANNEL
TRUMFLEET 1	BHF	3377.00	82.60	3.70	2.90	1.50		0.30	3.00	3.40			2.10	0.20			CRSU	ML	VFU	VP	R	H	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	3435.00	77.40	7.80	4.20	0.30		0.30	4.80	3.30			0.20	1.70			CRSL	ML	VFL	M	SA	P	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	3438.00	75.30	8.00	1.60	3.40	0.10		1.30	3.20	2.40		1.60	0.20	2.50		FU	FL	SLT	VG	SA	P	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	3441.00	58.30	7.50	9.30		5.70	0.60	8.90	4.40				5.30			FU	MU	VFU	P	A	P	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	3445.00	72.80	13.00	2.70	0.50	0.10		4.90	2.70	0.30		0.10	1.10	1.80		CRSU	FU	SLT	VP	A	P	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	3447.00	73.80	9.90	1.40	4.10	0.10		1.20	0.90	0.50		0.30	1.00	5.80		G	CRSL	VFU	M	SA	M	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	3451.00	81.90	8.70	0.70				6.40	1.40			0.50	0.30			CRSU	CRSL	VFU	P	SA	P	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	3455.00	74.50	10.80	0.80	2.90	0.20		3.20	1.30	0.10		0.30	0.10	6.00		CRSU	CRSL	VFU	P	SA	P	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	3459.00	77.10	11.60	2.70				5.10	1.90				1.10			CRSU	ML	VFL	P	SA	P	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	2462.00	67.80	15.80	1.20	2.40			3.70	2.70	1.30		0.30	1.00	3.80		G	CRSU	FL	P	SA	P	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	3467.00	76.60	10.60	0.50	0.40	0.10	0.30	9.70	1.30			1.20	0.40			G	CRSL	VFU	P	SA	P	DISTRIBUTARY CHANNEL

POINT COUNT DATA; TRUMFLEET FIELD

POINT COUNT DATA; TRUMFLEET FIELD

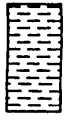
TRUMFLEET 1	PHG	3470.00	78.80	8.90	1.50			0.40	6.60	2.60	0.40		0.40	0.50			G	CRSL	VFU	A	SA	P	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	3473.00	70.80	12.40	0.60	1.50			6.50	1.20	0.20		0.70		6.10		P	CRSL	VFU	P	SR	P	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	3475.00	69.90	12.20	0.40	0.40			9.70	1.00			1.90		4.20		P	CRSL	VFU	VP	R	P	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	3482.00	76.70	10.10	1.20			0.40	10.30	1.10			0.30	0.20			CRSU	CRS	VFU	P	SA	P	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	3486.00	73.60	9.50	0.50	0.50			8.00	1.60			3.40	0.10	2.80		G	MU	FL	P	SA	P	DISTRIBUTARY CHANNEL
TRUMFLEET 1	PHG	3489.00	62.30	9.70	0.90	0.40	0.30	2.10	3.90	2.30			0.80	0.10			CRSU	MU	VFU	P	SA	P	DISTRIBUTARY CHANNEL
TRUMFLEET 2	ROUGH ROCK	3099.00	81.60	5.30	0.40	1.20	0.20	0.20	3.30	1.40	0.10		2.60		3.20		FL	VFU	SLT	G	SA	H	DISTRIBUTARY CHANNEL
TRUMFLEET 2	ROUGH ROCK	3103.00	61.70	8.00	3.40	0.60	0.30	0.50	0.40	0.80	1.50		16.10		3.30		FL	VFU	VFL	G	AS	M	DISTRIBUTARY CHANNEL
TRUMFLEET 2	ROUGH ROCK	3106.00	49.00	10.30	0.90		0.60	0.30	2.10	0.60		21.90	14.00		0.70		FU	VFU	VFL	VG	A	L	DISTRIBUTARY CHANNEL
TRUMFLEET 2	BHF	3306.00	6.00	2.00	1.60				12.40	0.40				79.60			CRSU	VFL	CLAY	VP	A	L	SEATEARTH
TRUMFLEET 2	BHF	3312.00	85.10	1.20	2.30	0.70	0.50		2.20	2.00	4.60				0.30		FU	VFU	SLT	VG	A	L	DISTRIBUTARY CHANNEL
TRUMFLEET 2	BHF	3316.00	65.30	2.40		1.20		0.10	1.70	0.10		14.40	2.30	12.80			CRSL	ML	SLT	P	SA	L	DISTRIBUTARY CHANNEL
TRUMFLEET 2	BHF	3321.00	61.10	2.70		0.70		0.10	0.50			34.90					CRSL	ML	VFU	M	SA	L	DISTRIBUTARY CHANNEL
TRUMFLEET 2	PHG	3392.00	64.50	10.20	4.40	1.10	0.30	0.30	1.80	1.50		11.60	3.20		1.10		CRSL	FU	SLT	M	SA	L	
TRUMFLEET 2	PHG	3396.00	64.90	13.60	1.10			0.20	13.40	2.20	0.20		0.70	0.30	3.40		G	MU	SLT	P	SA	L	
TRUMFLEET 2	PHG	3399.00	72.10	11.30	0.70	0.40	0.30	0.20	6.60	0.90	0.20		1.20	0.30	5.80		G	MU	FU	P	SA	L	DISTRIBUTARY CHANNEL
TRUMFLEET 2	PHG	3400.00	75.80	11.10	1.60	1.00	0.20	0.10	2.50	2.30	0.10		1.50	0.20	3.60		P	MU	SLT	P	A	L	DISTRIBUTARY CHANNEL
TRUMFLEET 2	PHG	3405.00	70.50	11.20	2.90		0.10	0.20	3.00	2.20	0.50		2.00	0.50	6.90	0.20 HAEM	G	MU	VFL	VP	SA	L	DISTRIBUTARY CHANNEL
TRUMFLEET 2	PHG	3409.00	70.30	10.30	1.10	0.30	0.20	0.10	6.80	3.50	0.10		3.20		3.10		CRSU	MU	VFU	P	A	L	DISTRIBUTARY CHANNEL
TRUMFLEET 2	PHG	3424.00	72.80	9.70	1.90	0.20	0.40		2.20	7.10	0.10		2.40	0.20	3.10		CRSU	MU	VFU	P	SA	M	DISTRIBUTARY CHANNEL
TRUMFLEET 2	PHG	3428.00	69.60	14.30	1.20	0.30	0.20	0.10	5.90	2.70	0.40		2.80		2.50		G	CRS	VFU	P	SA	M	DISTRIBUTARY CHANNEL
TRUMFLEET 2	PHG	3430.00	76.20	9.10	0.50	0.10			3.90	3.50			2.30	0.40	4.00		CRSL	ML	FU	M	SA	M	DISTRIBUTARY CHANNEL

KEY TO LITHOLOLOGICAL SYMBOLS

LITHOLOGY



SANDSTONE



SILTSTONE



SHALE



MARINE BANDS



COAL

SEATEARTH

SEDIMENTARY STRUCTURES



Parallel Lamination



Planar Cross Lamination



Trough Cross Lamination



Ripple Lamination



Ripple Drift Lamination



Wavy Lamination



Slumped Bedding



Load Casting



Bioturbated

Fractures

BED BOUNDARIES

Sharp

Gradational

Erosive



Load Cast

QUALIFIERS



Calcareous



Dolomitic



Carbonaceous



Nodular



Pebbly

Silty



Macrofossils



Plant Fragments

GRAIN SIZE

g Gravel

C V. Coarse

c Coarse

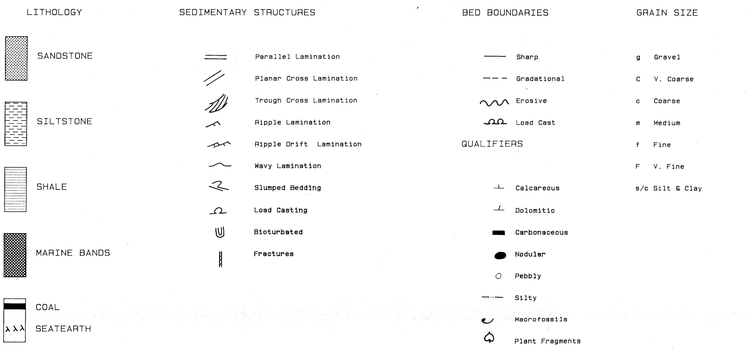
m Medium

f Fine

F V. Fine

s/c Silt & Clay

KEY TO LITHOLOLOGICAL SYMBOLS



SEDIMENTOLOGICAL LOG

N.C.B. MOSS FARM

