## *Reservoir characterization and potential of the old red sandstone around the inner moray firth, NE Scotland*

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# Reservoir Characterization and Potential of the Old Red Sandstone around the Inner Moray Firth, NE Scotland.

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# Abstract

Old Red Sandstone deposition in North East Scotland occurred during Devonian times within the Orcadian Basin, a northeast-southwest elongated structure. The basin formed in an extensional tectonic regime that resulted from gravitational collapse of over thickened Caledonian crust. A half graben topography resulted with the faulting largely controlled by crustal heterogeneities inherited from the Caledonian period of mountain building.

The ORS around the Inner Moray Firth is a red bed sequence deposited on the southern margin of the Orcadian Basin. The succession consists predominantly of braided fluvial sandstones, siltstones and mudstones, with lesser amounts of aeolian sandstones, evaporitic sabkha sandstones (only in the Upper ORS), and lacustrine mudstones and limestones (Middle ORS only). Lithologically the sandstones are coarse to fine grained, moderately to well sorted, and predominantly sublitharenite in composition.

The sandstones show the following diagenetic sequence: (1) Eodiagenesis: formation of clay/Fe oxide rims and the dissolution of lithic fragments; (2) Mesodiagenesis: the precipitation of a blocky, irregularly distributed, calcite cement; and (3) Telodiagenesis: a major dissolution event following inversion during the late Carboniferous involving partial removal of the calcite cement, feldspars (predominantly plagioclase), and lithic fragments (mainly sedimentary and metamorphic) and an associated precipitation of kaolinite.

Intergranular macroporosity is most abundant with lesser amounts of intragranular and microporosity. Porosity values are quite low (an average of 6% for both the Middle and the Upper ORS) and permeabilities are also poor (an average of 17 and 51mD respectively). Porosity reduction has occured mainly through cementation rather than compaction. The low permeabilities are thought to be due to low pore interconnectivity because of the patchy nature of the calcite cement, and to the presence of pore lining/filling kaolinite. Diagenesis has acted to largely overprint the primary permeability characteristics of the different lithofacies identified within the sandbodies. Some fractured samples however, had permeabilities of up to 1400mD similar to the situation in the Buchan Field ORS fluvial sandbody reservoir, where fracturing is the major control on reservoir characteristics.

Reservoir heterogeneities occur on a variety of scales within the ORS and have a marked effect. At the microscale diagenetic heterogeneitites have reduced porosity and permeability to very low levels. Cross-bedding and other sedimentological structures exert directional anisotropies on the permeability. Additionally, the Middle ORS is separated into discrete segments from 10-100m thick by laterally extensive lacustrine deposits, resulting in there being virtually no vertical connectivity between sandbodies. The Upper ORS has a much higher sandbody connectivity with only local horizons of thin, discontinuous fluvial mudstones which are scarce due to syndepositional erosion.

## Declaration

The contents of this thesis are the original work of the author and have not previously been presented for a degree at this or any other university. The work of other people is acknowledged by reference.

## Copyright

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# Contents


- **Title page** i
- **Abstract** ii
- **Declaration and Copyright** iii
- **Acknowledgements** iv
- **Contents** v
- **Chapter 1 Introduction**

| | | |
|---|---|---|
| 1.1 | Aims of the project | 1 |
| 1.2 | Geological history | 3 |
| 1.2.i | Initiation of the Orcadian Basin | 3 |
| 1.2.ii | Dating the ORS | 7 |
| 1.2.iii | Paleogeographic evolution | 8 |
| 1.2.iii.A | Lower ORS | 9 |
| 1.2.iii.B | Middle ORS | 11 |
| 1.2.iii.C | Upper ORS | 14 |
| 1.2.iii.D | Post ORS | 16 |

- **Chapter 2 Sedimentology**

| | | |
|---|---|---|
| 2.1 | Introduction | 18 |
| 2.2 | Lithofacies description | 19 |
| 2.2.i | Massive/crudely bedded gravel (Gms, Gm) | 19 |
| 2.2ii | Cross-bedded sandstone (Sp, St) | 20 |
| 2.2.iii | Low angle laminated sandstone (Sl) | 23 |
| 2.2.iv | Horizontally laminated siltstone/fine sandstone (Sh) | 24 |
| 2.2.v | Ripple laminated siltstone/fine sandstone (Sr) | 24 |
| 2.2.vi | Horizontally laminated mudstone and siltstone (Fl) | 26 |
| 2.2.vii | Aeolian deposits (Spe, She) | 26 |
| 2.2.viii | Sabkha deposits | 29 |

| | | |
|---|---|---|
| 2.2.ix | Lacustrine depsoits | 30 |
| 2.3 | Depositional environments | 32 |
| 2.3.i | Lower ORS | 32 |
| 2.3.ii | Middle ORS | 33 |
| 2.3.iii | Upper ORS | 35 |
| 2.4 | Controls on sedimentation | 38 |
| 2.5 | Summary | 39 |

- **Chapter 3 Petrology and Diagenesis**

| | | |
|---|---|---|
| 3.1 | Introduction | 40 |
| 3.2 | Techniques used | 40 |
| 3.2.i | Petrographic techniques | 40 |
| 3.2.ii | SEM examination | 41 |
| 3.2.iii | X-ray diffraction | 42 |
| 3.3 | Compositional analysis | 43 |
| 3.4 | Grain size | 47 |
| 3.5 | Diagenesis | 49 |
| 3.5.i | Calcretes | 49 |
| 3.5.ii | Eodiagenesis | 58 |
| 3.5.iii | Compaction | 62 |
| 3.5.iv | Mesodiagenesis | 64 |
| 3.5.v | Telodiagenesis | 69 |
| 3.6 | Summary | 77 |

- **Chapter 4 Poroperm Characteristics**

| | | |
|---|---|---|
| 4.1 | Introduction | 79 |
| 4.2 | Porosity | 79 |
| 4.2i | Introduction | 79 |
| 4.2.ii | Classification | 80 |
| 4.2.iii | Prosity measurement | 81 |

| | | |
|---|---|---|
| 4.2.iv | Porosity evolution | 82 |
| 4.2.iv.A | Porosity reduction | 84 |
| 4.2.iv.B | Porosity enhancement | 94 |
| 4.3 | Permeability | 96 |
| 4.3.i | Introduction | 96 |
| 4.3.ii | Permeability measurement | 97 |
| 4.3.iii | Permeability controls | 100 |
| 4.4 | Summary | 106 |

• **Chapter 5 Resevoir Heterogeneities**

| | | |
|---|---|---|
| 5.1 | Introduction | 108 |
| 5.2 | Classification of heterogeneities | 108 |
| 5.3 | Description and quantification of heterogeneities from the ORS | 109 |
| 5.3.i | Faults | 110 |
| 5.3.ii | Fractures | 111 |
| 5.3.iii | Permeability zonation | 115 |
| 5.3.iv | Permeability baffles | 116 |
| 5.3.v | Lamination and cross-bedding | 121 |
| 5.3.vi | Microscopic heterogeneities | 126 |
| 5.4 | Summary | 127 |

• **Chapter 6 Conclusions and Discussion**

| | | |
|---|---|---|
| 6.1 | The Old Red Sandstone lithology and sedimentology | 129 |
| 6.2 | Controls on the reservoir potential of the ORS | 131 |
| 6.2i | Poroperm controls | 131 |
| 6.2ii | Heterogeneities within the ORS | 132 |
| 6.3 | Source rock potential and hydrocarbon generation in the Orcadian Basin | 135 |
| 6.4 | Future exploration areas | 139 |

- **Appendices**


| | | |
|---|---|---|
| Appendix 1 | Point count data | 142 |
| Appendix 2 | Compaction data | 144 |
| Appendix 3 | Porosity loss data | 146 |
| Appendix 4 | Poroperm data | 148 |

- **References** 151

# Chapter 1

# Introduction

## 1.1 Aims of the project

Following the discovery of oil in the Devonian Old Red Sandstone (ORS) of the Buchan Field in the northern North Sea oil province there has been speculation as to the possibility of finding further commercial hydrocarbon accumulations elsewhere within the ORS. Since ORS sedimentation occurred across the whole of the northern North Sea Province large thicknesses of sand, and therefore potential reservoirs, may be present, albeit at great depth, beneath Mesozoic and Tertiary cover.

The aim of this project is to study the ORS sediments exposed at outcrop around the Inner Moray Firth as a possible model for similar sediments present off-shore, and to describe and quantify their reservoir characteristics and heterogeneities. The techniques employed in this study include:

1) An investigation of the diagenetic and burial history of the sediments from the time of their deposition through to the present day;

2) An evaluation of the porosity and permeability characteristics of the sandstones within the succession and how these characteristics relate to such features as lithofacies, texture, compaction and cementation;

3) An assessment of sandbody geometry and connectivity, and the presence of heterogeneities of different scales within the sediments and their effect on fluid flow.

The field area around the inner Moray Firth is bounded to the south and west by the limit of the ORS outcrop; and extends as far north as Dornoch and as far east as Nairn (Fig. 1.1). It includes the more marginal, fluvially dominated sediments rather than the lacustrine dominated sequence found farther north in Caithness. This

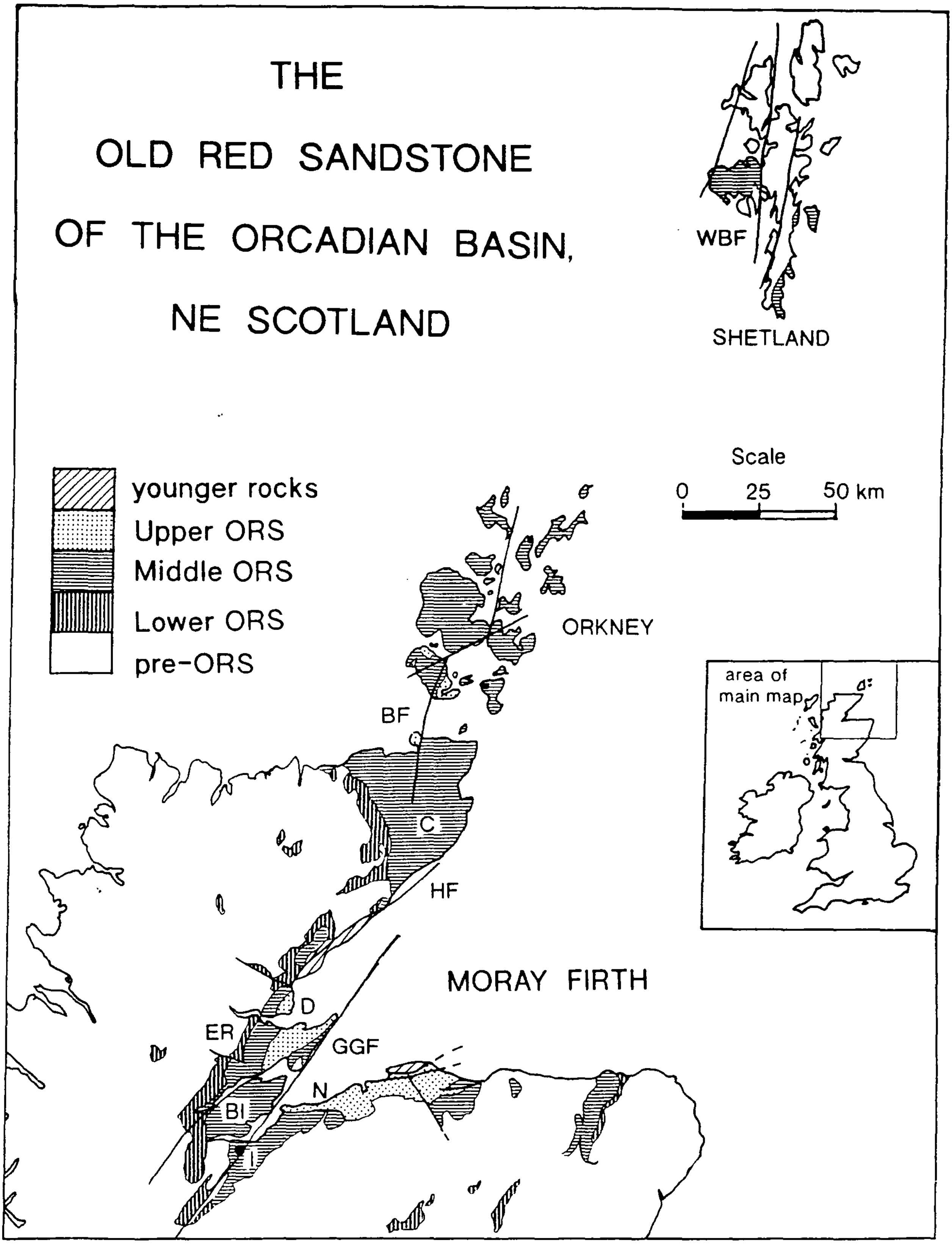



*Fig. 1.1. Geological/location map of northeast Scotland. (I=Inverness, N=Nairn, D=Dornoch, BI=Black Isle, ER=Easter Ross, C=Caithness, GGF=Great Glen Fault, HF=Helmsdale Fault, BF=Brough Fault, WBF=Walls Boundary Fault).*

study is mainly concerned with the Middle and Upper ORS. The Lower ORS has been excluded because it has a limited geographical distribution and a very low ratio of sandstone which makes it exceedingly unlikely that it will contain any suitable reservoir rocks.

## 1.2 Geological history

Before any study can be undertaken into the reservoir characteristics of the Old Red Sandstone, it is necessary to discuss the processes involved in the initiation, development and sedimentary history of the Devonian Orcadian basin and its regional tectonic setting.

The Orcadian Basin is a SW- NE orientated structure (Fig. 1.2a). It has a faulted western boundary, forming part of a more extensive genetically related series of faults running from west of the Moray Firth in the south, to the Strathy Fault west of the Orkneys, in the north. The southern margin is largely unconformable on the Dalradian highlands. The Orcadian Basin is one of several Devonian basins that formed across the northern North Sea from north Scotland to west Norway. These are filled with a succession of continental red beds, which in the Orcadian Basin, are predominantly fluvial sandstones and siltstones, with lesser amounts of aeolian sandstones and lacustrine siltstones, mudstones and limestones.

### 1.2.i Initiation of the Orcadian Basin

The Devonian basins in northern Britain and northwestern Europe were initiated immediately following the Caledonian Orogeny within the Caledonian mountain chain itself. Two theories have been put forward to explain their origin (Fig. 1.2b,c). The first theory attributes their origin to sinistral megashear (Ziegler, 1982; Bukovics *et al.,* 1984), based on the presence of several faults which have been interpreted as having had strike-slip movement (the Great Glen Fault, the Highland

Boundary Fault and the Walls Boundary Fault, in Scotland; the Møre-Trøndelag Fault in Norway; and the Billefjørden Fault in Spitzbergen). The second theory is an

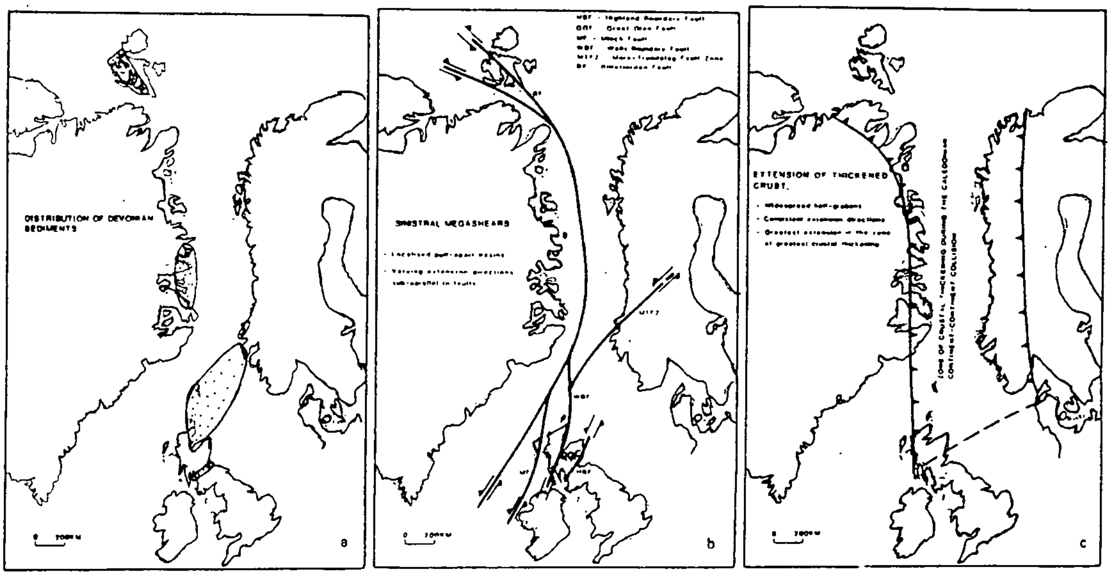



*Fig. 1.2. Pre-rift configuration of the northeast Atlantic region showing a) Distribution of Devonian sediments, b) Sinistral megashear hypothesis, and c) Extensional collapse hypothesis. (After Norton, et al., 1987).*

extensional collapse hypothesis (Dewey, 1982; McClay *et al.,* 1986) which claims that the basin formed as a result of extensional gravitational collapse of crust overthickened by the Caledonian Orogeny.

The fault pattern identified in the Orcadian Basin is one of numerous arcuate faults, dipping to the east. These faults define various sub-basins, bounded to the west by a series of faults (western boundary fault system), which are similar in scale to those found in the Lake Tanganyika region of the East African Rift (Rosendahl *et al.,* 1986). The oblique segments of these arcuate faults can be used to constrain the overall extension direction as approximately northwest-southeast (Enfield and Coward, 1987;

Norton *et al.,* 1987) which is parallel to pre-existing Caledonian structures, but oblique to that expected if sinistral shear was the main causal mechanism. In Shetland the situation is more complex, although Beach (1985) is of the opinion that the Devonian basal marginal fault here is an extensional detachment. Norton *et al.* (1987) have found unequivocal sense of shear indicators on Foula that indicate extension in an approximate east-west direction, that is parallel to the orientation of earlier Caledonian compression.

The Norwegan basins show a similar pattern of arcuate faults forming sub-basins within a larger half graben structure, except that these faults dip to the west and are fault bounded to the east, the exact opposite to that in Scotland. However, the direction of extension again mimics that of the Caledonian structural grain which is also opposite to that in Scotland. Additionally, the Norwegan faults are longer, have a greater spacing and a greater displacement than those in Scotland (Norton *et al.,* 1987). Three models have been suggested for basin development in Norway: extensional half grabens (Hossack, 1984; Norton, 1986, 1987; Seranne & Seguret, 1987); transtension along strike-slip faults (Steel, 1976; Steel *et al.,* 1977; Steel & Gloppen, 1980); and allochthonous basins emplaced by thrusting (Sturt, 1983; Torsvik *et al.,* 1986). The overall increase in size of the basins compared to those in Scotland can be explained by the fact that the immediately post-Caledonian crust was thicker in the Norway region than farther to the west (Norton *et al.,* 1987).

Further, there are several general arguments against the dominance of strike-slip related sedimentation in the Devonian:

1) Basins have developed in a very widespread area, commonly far removed from any known strike-slip structures (*eg.* the Turriff basin in Scotland and the Røragen basin in Norway) (Norton *et al.,* 1987).

2) There is a lack of clear sedimentary signature from Devonian strata near to the major strike-slip faults (Rogers, 1987).

3) The consistency of implied extension directions at high angles to the strike-slip faults (Norton *et al.,* 1987).

4) The correlation between the area of greatest crustal thickening with the largest extensional structures (Norton *et al.,* 1987).

5) The presence of alkaline, calc-alkaline, and rhyolitic volcanics in Greenland, and calc-alkaline volcanics in Scotland, which commonly form during the syn-rifting phase of crustal extension.

From the above evidence a sequence of events leading up to the time of basin initiation can be constructed. The closure of the Iapetus ocean during the Caledonian orogeny led to a thickened crustal welt in the Scottish region due to the overriding of Lewisian basement slices and Moine and Dalradian metasediments over Lewisian foreland. This thickening, which is greater to the east around Norway, was completed by end-Silurian times (420 Ma) (Watson, 1984; Powell & Phillips, 1985). This crust would then spread laterally given sufficient lateral density variation and topographic head (England, 1982; Coney & Harms. 1984). Work by Glazner & Bartley (1985) suggests that this collapse should take place 15-20 Ma after the initial thrusting, which is in agreement with the general timing in the Caledonides (McClay *et al.,* 1986).

Thus, pure shear in the lower crust would lead to overall thinning of the crust to a more stable thickness of 30-40 km, with a surface expression of listric extensional faulting and basin development. This faulting, although not as a result of direct reactivation of Caledonian thrusts, was probably controlled by an overall Caledonian anisotropy in the middle and upper crust (Norton *et al.,* 1987). Despite the dominance of extension, there were probably minor strike-slip movements during the Devonian, but these had no effect on basin formation (McClay *et al.,* 1986; Rogers, 1987).

## 1.2.ii Dating the ORS

The Devonian ORS sediments have been subdivided into three units based on lithostratigraphy and to a lesser extent, biostraigraphy, namely the Lower, Middle, and Upper Old Red Sandstone (Fig. 1.3). It is important to point out that these subdivisions are <u>not</u> synonymous in any way with the chronostratigraphic divisions Early, Middle, and Late Devonian. In fact Lower ORS deposition probably did not begin until Emsian times (late Early Devonian) in most areas. Accurate dating of the ORS succession is difficult because of the continental setting which means that there are few fossils. Lower ORS deposits, which were classified as Middle ORS by early workers, until Westoll's work of 1948 and 1951, have only fairly recently been accurately dated using palynology (Richardson, 1967; Mykura & Owens, 1983). Donovan (1978) also reports the finding of a single fish scale of Lower ORS age.

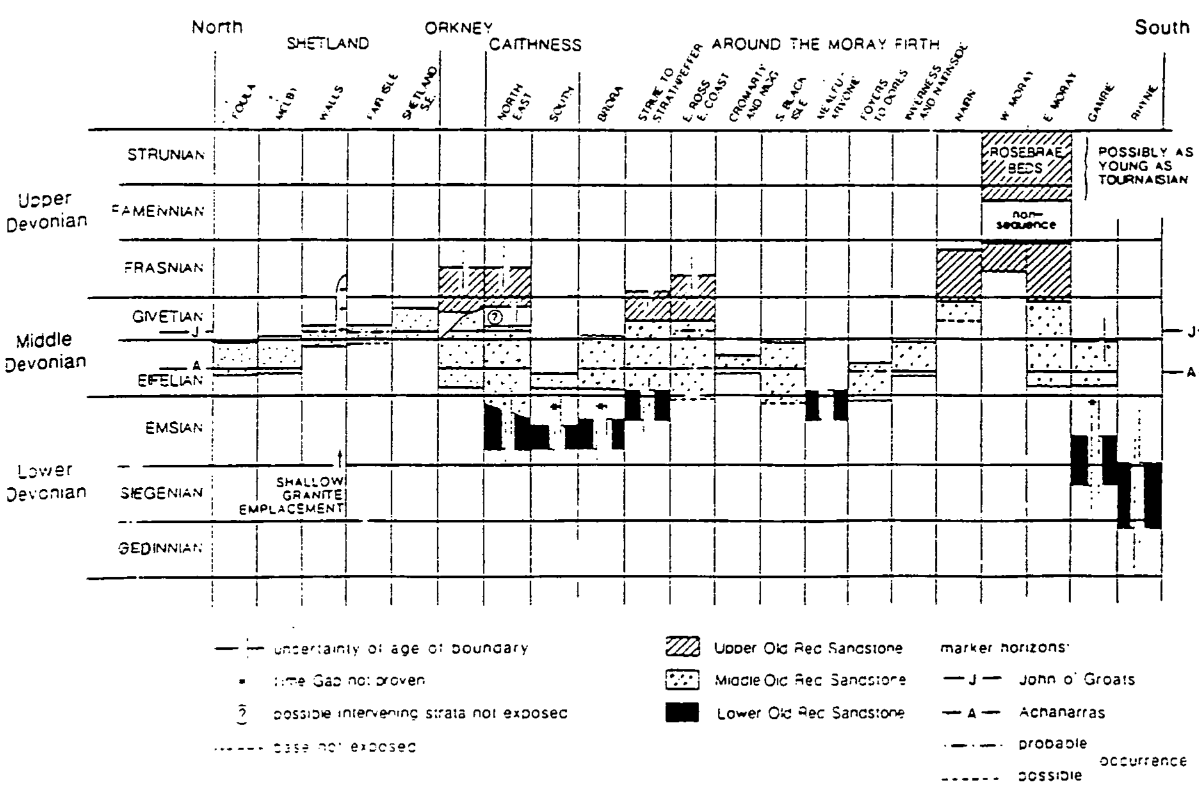



*Fig. 1.3. Chronostratigraphy of the Orcadian Basin. The Achanarras and John O'Groats Horizons are time markers representing widespread lake transgressions. (From Rogers et al., 1989).*

The Middle ORS has proved very difficult to date with respect to the standard Devonian succession (Rogers, 1987). The fish fauna found extensively in lacustrine deposits, despite being useful for correlations, only give an approximate age of Middle Devonian (Traquair, 1896, 1897), with more precise dates given by Westoll (1977) and Mykura (1983) being unconfirmed. Palynological work by Richardson (1960, 1962, 1964) largely agrees with the fish biostratigraphy (Eifelian-Givetian). However the palynological work of Marshall (in Rogers, 1987) suggests that the Middle ORS may be slightly older, some of it being Lower Devonian (Emsian).

The Upper ORS succession in the south of the basin (SE of the Great Glen Fault) is probably the best dated anywhere, using fish fauna. Correlations have been made with the Devonian of Latvia, Estonia, Lithuania and NW Russia, the most recent refinements being by Tarlo (1961), Miles (1968), Westoll (1977) and Mykura (1983). Another accurately dated locality is on Hoy where Upper ORS sandstones overlie a lava that has been dated at 379±10 Ma and 366±8 Ma (Mid/Late Devonian) (Halliday *et al.*, 1977; Halliday, 1982). However, many other areas, especially in the main thesis area, to the north of the Great Glen Fault, are only poorly constrained due to a lack of fossil evidence. In general though Upper ORS ages are from mid Givetian to late Frasnian.

### 1.2.iii Palaeogeographic evolution

Before the reservoir characteristics of the Old Red Sandstone can be discussed, it is necessary to examine the sedimentary history of the Orcadian Basin throughout Devonian times, in order to place the sediments within a proper stratigraphic context and to get an idea of the relationship between the facies present.

Palaeomagnetic data can be used to estimate a palaeolatitude for the Orcadian Basin. Tarling (1983, 1985) has suggested that its position at the base of the Devonian was 20-22°S, moving to 15-16°S by the end of the Devonian, which would place it in trade wind desert conditions. Palaeoenvironments can also be used as climatic indicators. Typical Orcadian environments, ephemeral fluvial systems and playa lakes, together with some calcretes (especially in the Upper ORS), indicate considerable aridity. Rogers and Astin (1991) suggest heavy rainstorms with a frequency of a few to tens of years, but with no seasonality, for Middle ORS playa lakes in Caithness.

The occurrence of deep permanent lakes across large areas of the basin, possibly lasting up to 10,000 years (Rayner, 1963) indicate a change to a much wetter climate (Donovan, 1980). Also the varved laminites which formed in these deep lakes point to a strong seasonal influence (Rayner, 1963; Donovan 1980). This suggests a semi-arid climate with a wet winter and dry summer. Recent work on palaeobotany (Gray, 1985; Banks, 1987) suggests that there was a well developed terrestrial flora by the Middle Devonian, and therefore the presence of aeolian dune deposits within the Middle and Upper ORS advocates the notion of considerable aridity as opposed to a simple lack of stabilising flora. Astin (*pers. comm.*, 1991) suggests a surface temperature of 15-20°C.

## A. Lower ORS

Lower ORS environments have been described in some detail by numerous workers (Armstrong, 1964, 1977; Donovan, 1978; Archer, 1978; Blackbourn, 1981a,b; Mykura, 1978, 1983; Mykura and Owens, 1983; O'Reilly, 1985; Parnell, 1985b; Richards, 1985c; Sweet, 1985) and a fairly detailed picture has been built up. Lower ORS deposits have also been recognised in four boreholes offshore (Andrews *et al.,* 1990).

Deposition began with the onset of rifting during ?late Siegennian to Emsian times, in narrow fault bounded basins, elongated N-S to NE-SW parallel to bounding faults and separated by oblique fault zones, or transfer faults, with the basins fed by locally sourced streams. Small alluvial fans formed along syndepositional marginal faults, whilst within the basins evaporitic lakes developed containing carbonates and organic rich mudstones. Fluvial systems were not fully developed, and sandstones are found only as distal deposits on alluvial fans. The existence of some mature calcretes around the Moray Firth (Donovan, 1982) shows that some of the basins underwent long periods with no subsidence and deposition. Although the exact offshore locations of Lower ORS basins are not known, it is possible that they coincide with the depositional centres of Mesozoic basins (Norton *et al.,* 1987).

Much debate has raged over the nature of the transition from the Lower to Middle ORS. When Westoll (1948, 1951) first proposed the existence of the Lower ORS within the basin he also speculated on the existence of a basin-wide Lower-Middle ORS unconformity, which has indeed been recognised at several localities. Armstrong (1964, 1977) used the presence of intensely deformed Lower ORS in Ross-shire to indicate pre-Middle ORS compression, and hence an unconformity. Pyrolysis data of Parnell (1985a) indicates a higher thermal maturity for the most intensely deformed Lower ORS in Ross-shire, than for the adjacent Middle ORS, thus appearing to support the idea of a compressional event.

There are, however, several factors that argue against the presence of an unconformity. Firstly, at other localities thermal maturities for the Middle and Lower ORS are similar (Parnell, 1985a; Hillier & Marshall, 1992). Also, at various places throughout the Orcadian Basin Middle ORS rocks show high thermal maturities due to the presence of granites within the basement. The deformation of the Lower ORS could be attributed to syndepositional faulting during Middle ORS times. Rogers (1987) has found that at all localities around the Inner Moray Firth, north of the Great Glen Fault, where an unfaulted Lower-Middle ORS contact is exposed, it is conformable. Additionally, there is no biostratigraphic evidence for a long time gap between the Lower and Middle ORS (Rogers, 1987)

Although there is considerable doubt as to the presence of a regional unconformity, there is undisputedly an increase in the proportion of coarse fluvial sediments across the Lower-Middle ORS boundary, which could be related to tectonic activity. Rogers (1987) proposes that the change in facies across the Lower-Middle ORS boundary is due to a combination of increased extension and an increase in the size of the fluvial systems. Rogers (1987) also argues that the Lower-Middle ORS boundary is in fact diachronous, based on lithostratigraphic considerations. He states that a thick Lower ORS sequence overlain by only very thin lower Middle ORS deposits is at least partially contemporaneous with the base of a thicker Middle ORS sequence further east.

## B. Middle ORS

The beginning of Middle ORS times saw the onset of concurrent basin wide deposition (Mykura & Evans, 1985). However, there was still considerable syndepositional faulting, as shown by the presence of thick rudite deposits which represent fault scarp alluvial fans (Stephenson, 1972, 1977; Mykura, 1982). These alluvial fans varied in size from small fans of similar size to those in the Lower ORS (a few km) (Mykura, 1982) to much larger ones, several tens of km or more in extent (Rogers, 1987). Distally these fans deposited sheet flood and channel sands which were intercalated with playa lake mudstones. This is illustrated in Fig. 1.4 which is a more detailed stratigraphic column for the Middle and Upper ORS from the Easter Ross coast, and Fig. 1.5 which is a palaeogeographical reconstruction of the Middle ORS.

Three major fluvial systems occurred in the Moray Firth region during Middle ORS times (Rogers, 1987). One drained the Great Glen region and entered the basin along the central axis of the half graben system. The other two originated in the Central and Northern Highlands respectively, and entered the basin at the intersections of normal and transfer faults, before joining the former axial system. Two other sites of fluvial input into the basin existed along the south coast of the Moray Firth (Mykura, 1976; Archer, 1978; Blackbourn, 1981c); there was no major fluvial input into Caithness and Orkney during early Middle ORS times. The reasons

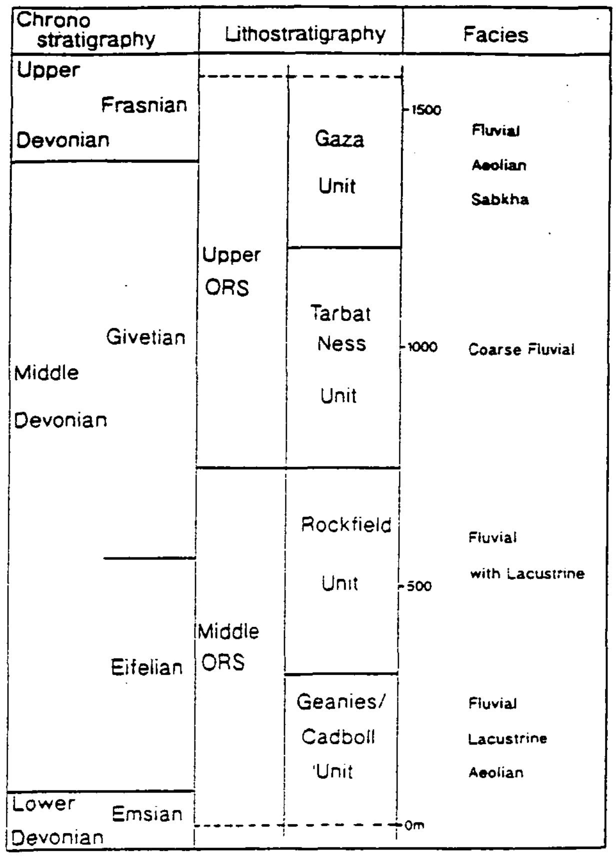



*Fig. 1.4. More detailed stratigraphy for the Easter Ross coast based on work by D.A. Rogers and J. Marshall. Unit names are informal. Unit thicknesses are minimum values and the total thickness may exceed 2000m. (D. Rogers, pers comm., 1992).*

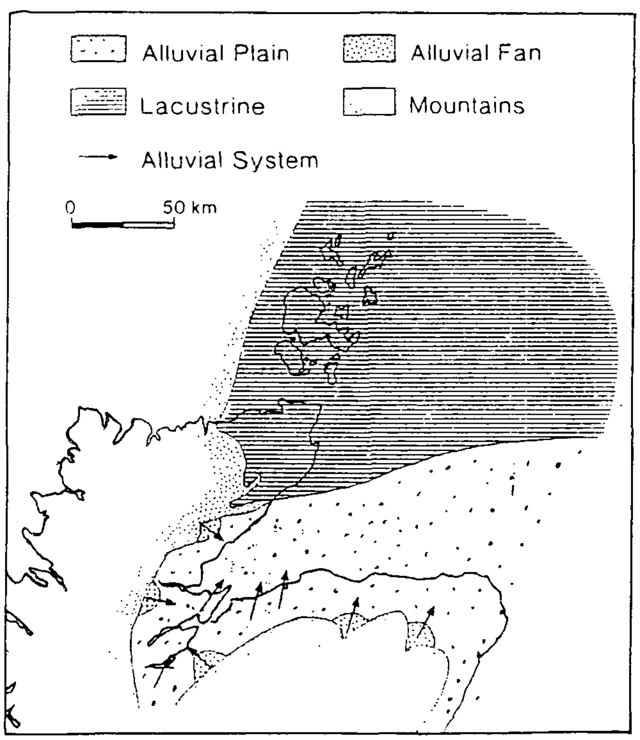



*Fig. 1.5. Palaeogeographic reconstruction for the Middle Old Red Sandstone (early Eifelian) with the Orcadian lake near maximum expansion.*

for this are uncertain, but one possibility is the existence of a basement high along an uptilted footwall block to the west of Orkney (Rogers, 1987).

Throughout the Middle ORS large permanent lakes had been forming every few thousand years, centred around Caithness. Several of these lakes extended into the Moray Firth area and by mid-Eifelian times the most extensive of these had formed, covering almost all of the basin. This is the so called Achanarras lake highstand, which is the best correlated horizon in the basin, based on fish fauna and lithostratigraphy. Synsedimentary faulting was still occurring (Armstrong, 1977; Rogers, 1987), as is shown by overlap and reworking of earlier Middle ORS deposits, and by the presence of further alluvial fans. The axial drainage system was well established by this time.

By the end of the Middle ORS (early-mid Givetian times) the large lake had been almost completely infilled, leaving the Moray Firth area as a large fluvial system flowing northeast along the basin's axis being fed by smaller streams emanating from the eastern and western margins of the basin. A slight upwards coarsening can be seen as a result of increased uplift along the marginal faults and progradation of coarser, more proximal sediment further into the basin (Rogers, 1987). It was at this time that major fluvial systems entered the Orkney-Caithness region for the first time. Astin (1985) has identified two river systems. One flowed northwest across Caithness and into southern Orkney, while the other originated northwest of Orkney and flowed to the south/southeast. These are tectonically controlled which caused parts of the system to be abandoned allowing a large aeolian dune field to form (Astin, 1985). Further dune fields formed around the inner Moray Firth, also due to channel switching, again possibly tectonically controlled.

Offshore the Middle ORS has been identified in several boreholes (Andrews *et al.,* 1990). The succession resembles that seen onshore and in one borehole the spore assemblages are very similar to those of the Lower Caithness Flags and the Achanarras Limestone (Richardson, 1964).

As with the Lower-Middle ORS transition, the Middle-Upper transition is also surrounded by controversy. Malcolmson (1859) was the first to suggest an unconformity around the Moray Firth. Geikie (1878) used the existence of an unconformity at the base of the Upper ORS in Orkney, and the fact that elsewhere in Britain the Upper ORS unconformably overlays Lower ORS, to propose a basinwide unconformity. The Geological Survey mapped the boundary as such even though in many areas around the Moray Firth the boundary is not exposed (Black & MacKenzie, 1957).

The only place south of Caithness where the boundary is not conjectural is along the Easter Ross coast. Many early workers (Sedgwick & Murchison, 1828; Murchison, 1859a; Gordon & Joass, 1863) could all trace a conformable transition there, despite the presence of several small faults, along one of which Armstrong (1973) placed the boundary. Some workers, however, still support the idea of a regional unconformity (Westoll, 1977; Mykura, 1983; Trewin, 1985). Other evidence against the regional unconformity, apart from a lack of outcrop and one conformable transition, includes the lack of any biostratigraphic support for a significant time gap and the use of onlap as an explanation for an absence of Middle ORS between Upper ORS and basement rocks (Rogers, 1987).

## C. Upper ORS

The base of the Upper ORS is distinctly coarser than the underlying Middle ORS in many areas, despite the gradual coarsening upward trend seen towards the top of the Middle ORS. In addition there is a sudden increase in the amount of reworked Middle ORS sandstone fragments, suggesting a possible phase of increased tectonic activity. However, the active fluvial systems at this time were essentially the same as those in the Middle ORS, which suggests that the tectonic pattern was largely unchanged (Rogers, 1987). The presence of an unconformity in Orkney can be simply explained as localised uplifting of a fault block, given the continuity of palaeocurrent directions in the Upper ORS of Orkney and Caithness (McAlpine, 1978).

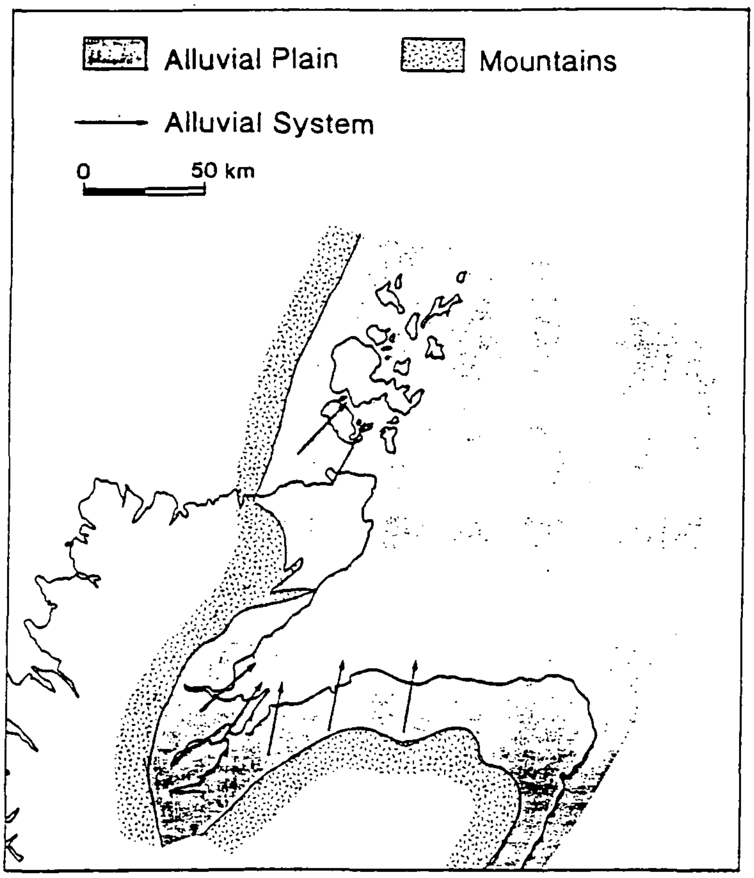



*Fig. 1.6. Palaeogeographic reconstruction for the Upper Old Red Sandstone (Givetian/Frasnian). Aeolian dunes can occur anywhere on the alluvial plain; sabkhas occur distal to fluvial systems.*

The only possible Upper ORS lacustrine sediments are found around Shetland. Across the rest of the Moray Firth area there was a large alluvial plain with fluvial, aeolian and siliciclastic sabkha environments (Figs. 1.4 and 1.6). Aeolian dune deposits were present throughout the Upper ORS times (McAlpine, 1978; Astin, 1985; Rogers, 1987). These would have formed between fluvial systems as well as distal to them. The evaporitic siliciclastic sabkha deposits are a feature unique to the Upper ORS and probably formed as a distal equivalent of the fluvial systems (Rogers, 1987). They consist of flat bedded, fine to coarse sandstones which show pseudomorphs of calcite after halite and/or gypsum (Parnell, 1985b). The relative abundance of mature calcretes in the Upper ORS around the Moray Firth (Peacock *et al.*, 1968; Donovan, 1982; Parnell, 1985a,b) suggest that rates of subsidence and fault motion had slowed down considerably from earlier times. Rogers (1987) suggests variable rates of tectonism as the major control on lateral variations in the depositional environment.

Marshall *et al.* (1990) report the first findings of a marine influence in the Orcadian basin in the form of two transgressive mudstone units of Frasnian (Late Devonian) age which contain acritarchs and scolecodonts. Field investigations in Orkney and Easter Ross apparently revealed a similar marine influence in slightly older Givetian rocks. The marine transgressions probably originated in the arm of the Devonian proto Tethys Ocean that existed in the North Sea at this time (Pennington, 1975).

Despite being found onshore all round the Moray Firth, and in Caithness and Orkney, no Upper ORS sediments have been recognised offshore in the Inner Moray Firth (Andrews *et al.,* 1990), probably as a result of Carboniferous uplift and erosion (see below). However, they are present in the outer Moray Firth, for example, in the Buchan Field, where fining-up cycles of conglomerate, sandstone and siltstone have been dated as Late Devonian-Early Carboniferous (Hill and Smith, 1979; Richards, 1985b)

## D. Post ORS

Carboniferous age coal-bearing strata have been found in the Outer Moray Firth Basin (Andrews *et al.,* 1990) and the sandstones in the Buchan Field are of Devonian-Carboniferous age. In fact, palynological dating in that region indicates a Tournaisian age, while some microflora as young as Viséan have been found in the Buchan Field (Hill and Smith, 1979) However, onshore and offshore around the inner areas of the Moray Firth, the only Carboniferous age rocks are volcanic (Andrews *et al.,* 1990). This, together with the absence of Upper ORS sediments offshore in the inner regions of the Moray Firth, indicates that there has been a period of either non-deposition or erosion.

Rogers *et al.* (1989, and references therein) summarise the evidence pertaining to the timing of movements along the Great Glen fault system. They conclude that there has been approximately 15-20 km of dextral displacement along the fault system between the Frasnian cessation of Orcadian extension, and the onset of Permian rifting in the Inner Moray Firth Basin. The orientation of compressive structures identified within the Orcadian succession (Donovan *et al.,* 1974; Rogers 1987) imply that these fault motions were associated with basin inversion. Palaeomagnetic readjustment of the Middle ORS by meteoric waters in the Permo-Triassic (Robinson, 1985) is consistent with this timing (Rogers *et al.,* 1989).

Rogers *et al.* (1989) have estimated that the Upper ORS in Orkney was probably only buried to a depth of 1 km, and used this to suggest that Orcadian inversion occurred sometime in the mid-late Carboniferous, following late Devonian-early Carboniferous deposition. Hillier and Marshall (1992) have used spore colour and vitrinite reflectance to assess levels of organic maturation, and have estimated that 2-3 km of late Devonian/Carboniferous rocks are absent basin wide from the succession. While this is a greater amount than suggested by Rogers *et al.* (1989) it does provide strong evidence in support of the erosion theory.

Hall (1991) has reviewed the post Carboniferous history of the Orcadian Basin, and the following summary is based on his review and on information from Andrews *et al.* (1990). Following the Carboniferous inversion, there was deposition of a thin Late Permian to Triassic sequence which ended with extensive calcrete development. This was followed by deposition of a thin but continuous series of Jurassic rocks, with evidence of contemporaneous fault movement near the top of the succession. Reworked shallow water Jurassic corals suggest the presence of Jurassic deposits to the west on the highlands. Cretaceous deposits are represented by abundant greensand clasts in glacial drift from Grampian, and flint pebbles in Tertiary gravels attesting to chalk cover over most of the Scottish highlands. The final phase of inversion was during the Tertiary.

Chapter 2

# Sedimentology of the ORS

## 2.1 Introduction

A great deal of work has been carried out on the sedimentology of the ORS over the years such that a great deal of confidence can be placed in the interpreted depositional environment (*e.g.* Donovan *et al.,* 1976; Armstrong, 1977; Westoll, 1977; Mykura, 1976, 1982, 1983; Mykura & Owens, 1983; Astin, 1985; Rogers, 1987; Rogers & Astin, 1991). The ORS exhibits facies and sedimentary structures typical of a variety of environments: low-sinuosity braided fluvial; permanent and temporary lacustrine; aeolian dune and interdune; and evaporitic siliciclastic sabkha.

Facies analysis is an important tool in aiding the description, correlation and interpretation of depositional environments. Significant works in this field include Allen (1970), Middleton (1973), Boothroyd & Nummedal (1978), Miall (1977, 1978, 1985) and Walker (1984). The work of Miall (1977, 1978) is of particular importance because he introduced a code system that covered lithofacies descriptions for fluvial environments (Table 2.1). Boothroyd & Nummedal (1978) added to this scheme by including codes for aeolian dune cross-stratification.

In the first part of this chapter the lithofacies, as classified by Miall (1977, 1978) and Boothroyd & Nummedal (1978), that occur in the ORS are catalogued and described. The lacustrine and sabkha deposits which are not covered by the lithofacies coding scheme are also described. In the second part of the chapter the facies associations and the depositional environments that they represent are described. The Lower ORS is examined in much less detail than either the Middle or Upper ORS because of the nature of its constituent facies which make it a very poor potential reservoir.

| Facies Code | Lithofacies | Sedimentary Structures | Interpretation |
|---|---|---|---|
| Gms | massive, matrix supported or crudely bedded gravel | grading, horizontal bedding | debris flow deposits |
| Gm | massive or crudely bedded gravel | horizontal bedding, imbrication | longitudinal bars, lag deposits |
| St | sand, fine to very coarse, may be pebbly | solitary or grouped trough crossbeds | dunes (lower flow regime) |
| Sp | sand, fine to very coarse, may be pebbly | solitary or grouped planar crossbeds | lingoid, transverse bars, sand waves (lower flow regime) |
| Sr | sand, very fine to coarse | ripple marks of all types | ripples (lower flow regime) |
| Sh: i) Shu | i) sand, very fine to very coarse, may be pebbly, silt | i) horizontal lamination, parting or streaming lineation | i) upper flow regime plane bed flow |
| Sh: ii) Shl | ii) sand, very fine to coarse, silt | ii) horizontal lamination | ii) lower flow regime plane bed flow; suspension fallout |
| Sl | sand, fine | low angle laminations (<10°) | scour fills crevasse splays, antidunes |
| Spe,She | sand | analogous to Sp, Sh | aeolian dunes |
| Fl | sand, silt, mud | fine lamination, very small ripples | overbank bartops or waning flood deposits |

***Table 2.1.*** *Lithofacies coding and interpretation. Adapted from Miall (1977, 1978) and Boothroyd & Nummedal (1978)*

## 2.2 Lithofacies description

### 2.2.i Massive/crudely bedded gravel (Gms, Gm)

Conglomerate bodies are common around the margins of the Orcadian Basin. They are predominantly matrix supported and are massive to crudely bedded (Fig. 2.1). Clast sizes range from 10cm to >1m (pebbles to boulders), with an average size of 15-20 cm (cobble), while the matrix is a coarse to very coarse immature feldspathic sandstone. There are frequently lenses and layers of fine to coarse sandstone interbedded with the conglomerate. These are usually only a few cm's to a few tens of cm's thick, and show horizontal laminations (Sh) or trough cross-bedding (St).

***Fig. 2.1.*** *Crudely bedded marginal conglomerate (Gm) from the Middle ORS near Cromarty. (Hammer length = 40cm).*

These conglomerates are interpreted to have formed as small alluvial fans adjacent to active faults around the basin margins. The lithofacies represent deposits from a variety of depositional processes: debris flow (Gms), channellised stream flow and sheet flood (Gm) (Miall, 1977).

### 2.2.ii Cross-bedded sandstone (Sp, St)

Cross-bedding is a common feature of the ORS. Cross-bedded sandstones are medium to coarse, sometimes pebbly, and may vary in colour from red to buff to yellow. Trough cross-bedding (St) is by far the most common type (Fig. 2.2). Individual sets are up to 1m thick, with an average of 50 to 80 cm, and up to 20m

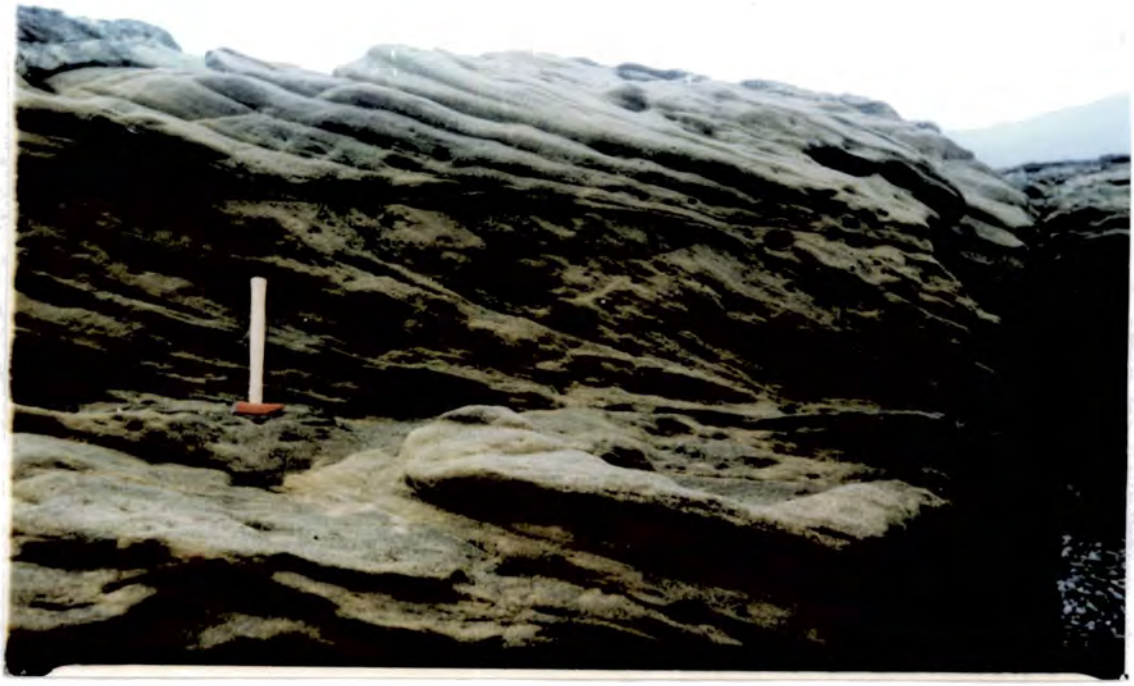


*Fig. 2.2. Example of trough cross-bedding (St) from the Upper ORS of Easter Ross. (Hammer length = 40cm).*

*Fig. 2.3. Planar cross-bedding (Sp) in the Upper ORS near Dornoch, resulting from transverse bar migration. (Hammer length = 40cm).*

wide, with an average of 2-8m. Foresets have tangential bases and dip at around 10-20°. Their thickness varies from <1cm -3cm. Planar cross-bedding (Sp) (Fig. 2.3), is much less common. Sets are <2m thick, typically 20-60cm. The foresets are planar and dip at up to 30°.

Trough cross-bedding occurs in cosets anything from 2 to 45m thick, either erosively bounded or interbedded with a variety of other facies: lacustrine, aeolian or sabkha. Planar cross-beds usually occur as solitary sets with trough cross-bedding above and below. Soft sediment deformation in the form of upright folds and overturned foresets is common within cross-bedded intervals (Fig 2.4).

***Fig. 2.4.*** *Deformed fluvial cross-bedding in Upper ORS braided fluvial deposits from Easter Ross. (Hammer length = 40 cm).*

Both trough cross-bedding and planar cross-bedding form under lower flow regime conditions, although at different velocities, with planar cross-bedding forming at the lower velocities. The actual velocities will be dependant on the stream depth and sediment grain size (Reineck and Singh, 1973). Trough cross-bedding is the result of migrating chains of 3-dimensional in-channel dunes with curved crests, while planar cross-bedding results from straight crested 2-dimensional bedforms (Allen, 1963; Miall, 1977).

## 2.2.iii Low angle laminated sandstone (Sl)

This lithofacies is fairly common and is usually seen in fine to medium grained sandstones with individual laminae being between 0.5-1cm thick. It typically occurs in erosively based 0.5-1m thick units which pass gradationally upwards into rippled or horizontally laminated fine sandstones and

*Fig. 2.5. Upper ORS low angle scour fills from Easter Ross. (Hammer length = 40cm).*

siltstones. These latter deposits however are commonly absent due to their erosion during deposition of succeeding lithofacies (Fig 2.5).

These low angle laminations form in conditions transitional between the upper and lower flow regimes. These probably occur in fast flowing water during flood events. They are often found intercalated with trough cross-bedded sandstones or lacustrine deposits.

## 2.2.iv Horizontally laminated siltstone/fine sandstone (Sh)

This lithofacies is confined to fine sandstones and siltstones where it occurs in two forms. The most common form (Shu) which is found associated with facies Sl is characterised by a parting lineation (Fig. 2.6) and formed by plane bed deposition under upper flow regime conditions. The other, less common, form (Shl) is typically very micaceous, with mica concentrated along bedding planes, and is of limited extent, both laterally and vertically. This formed by a mixture of lower flow regime plane bed deposition and suspension fall out.

*Fig. 2.6. Primary current lineation in horizontally bedded siltstone from the Middle ORS of Easter Ross. (Tape measure length = 5cm).*

## 2.2.v Ripple laminated siltstone/fine sandstone (Sr)

This lithofacies is not very common and only occurs in fine sandstones and siltstones. Two types of ripple cross-lamination occur in the ORS. The first type occurs gradationally above horizontally laminated sands of upper flow regime plane

*Fig. 2.7. Current ripples (Sr) in a Middle ORS fine sandstone, Black Isle (Pencil length = 10cm).*

*Fig. 2.8. Wave ripples (Sr) in a Middle ORS siltstone, Easter Ross. (Pencil length = 10cm).*

bed origin (Shu). Characteristically it consists of small-scale asymmetric ripples (Fig. 2.7) with ripple lengths of between 4.5 and 15cm and ripple heights of 0.5-1cm. Ripple index values of 8-13 are similar to the range of 6-10 given by Harms (1969) as typical of current ripples. These ripples will have formed under lower flow regime conditions in fine sediment (Allen, 1968; Miall, 1978).

The other ripples are more symmetrical with rounded crests (Fig. 2.8) and are associated with lacustrine deposits. Ripple lengths are about 6cm and ripple heights 0.25cm, giving a ripple index value of 24. These are wave ripples which typically have ripple index values above 25 (Harms, 1969). These form in standing water from waves that are possibly wind generated (Rogers, 1987).

### 2.2.vi Horizontally laminated mudstone and siltstone (Fl)

The Fl lithofacies consists mainly of thin red mudstones, <50cm thick, with minor amounts of siltstone (Fig. 2.9). The mudstones are mainly associated with trough cross-bedded sandstones, capping some cosets, and with the Sl facies. These laminated mudstones were probably deposited from suspension on floodplains from overbank/unconfined flood events.

### 2.2.vii Aeolian deposits (Spe, She)

Aeolian deposits are dominated by buff, medium to coarse sandstones, showing concave foresets (2-10cm thick) and tangential bottomsets (Spe) (Fig. 2.10). The sets are thin (<2m), and normally isolated, although occasional thicker intervals with cosets between 12-15m thick do occur. Intercalated with the bottom-sets of the aeolian cross-bedding are lesser amounts of horizontal to sub-horizontal laminae (She) (Fig. 2.11), composed of interbedded (on a scale of a few cm's) wind ripple laminae, grain fall laminae and sand flow laminae.

*Fig. 2.9. Thin fluvial mudstone with calcrete nodules from Upper ORS braided sandstone deposits, Easter Ross. (Hammer length = 40cm).*

*Fig. 2.10. Upper ORS aeolian cross-bedded sandstone (Spe) from Easter Ross. (Rucksac = 80cm).*

*Fig. 2.11. Horizontally laminated interdune deposits (She) with wind ripples visible on a bedding surface. Upper ORS of Easter Ross. (Hammer length = 40cm).*

*Fig. 2.12. Laminated mudstone and siltstone forming interdune deposits in the Middle ORS of Easter Ross. (Hammer length = 40cm).*

Aeolian deposits are often found interbedded with other facies; usually fluvial sandstones or sabkha deposits. Interdune deposits are locally abundant, occuring between sets and cosets and include horizontally laminated sandstones, siltstones and mudstones, with structures evident of dry, damp and wet conditions (Fig. 2.12). These include aeolian plane bed laminations, granule rich horizons, adhesion ripple cross-laminations and dissipation structures.

### 2.2.viii Sabkha deposits

These appear in the field as flat bedded buff-red or buff-green, mottled, very fine to coarse sandstones. They are characterised by an irregular or wavy lamination defined by colour variations and/or silty partings (Fig. 2.13). They were first recognised by Donovan (1978) in Caithness, although they were not identified within the thesis area until more recently (Parnell, 1985b; Rogers, 1987).


*Fig. 2.13. Upper ORS horizontally laminated siliciclastic sabkha deposits from Easter Ross. (Lens cap = 5cm).*

Parnell (1985) and Rogers (1987) both provide evidence for the former presence of gypsum and halite nodules and horizons. Rogers (1987) has suggested that these sabkha deposits formed on a flat, sandy, evaporitic plain with a water table within 1-2m of the surface. The dominant processes in operation on these sabkhas is growth of salt crusts - probably both gypsum and halite. Other features such as adhesion ripples and desiccation of thin mud layers indicate that damp and even wet conditions prevailed occasionally.

### 2.2.ix Lacustrine deposits

The lacustrine deposits can be separated into two different types. The most abundant deposits are red, green and grey structureless mudstones (Fig. 2.14). These range in thickness from a few cm's to over a metre and are interbedded with fluvial sandstones. Less common, but locally abundant are grey, organic rich, varved, carbonate-silt laminites (Fig. 2.15). The proportion of carbonate to teriginous clay in the laminites is variable and the carbonate has frequently been remobilised during early diagenesis into nodules or bedding parallel sheets. Some of these laminites contain algal laminae and are brecciated as a result of desiccation.

Subordinate clastic facies are found associated with the carbonate silt laminites. These include thick (<1m) grey massive mudstones; massive to flat laminated, occasionally ripple cross-laminated fine sandstones and siltstones; and wave ripple cross-laminated sandstones.

These lacustrine deposits represent different types of lake environment. The first type originated from terrigenous muds which were deposited in temporary playa lakes with water depths of a few metres and a duration of up to a few tens of years (Rogers, 1987). The carbonate-clastic laminites formed in deeper permanent lakes with little clastic input apart from along the margins, as evidenced by the subordinate marginal clastic facies. Suggested depths for these lakes vary from 5 to 60m (Allen,

***Fig. 2.14.*** *Thin lacustrine mudstones which formed in short lived ephemeral lakes. Middle ORS of Easter Ross. (Compass length = 8cm).*

***Fig. 2.15.*** *Grey, varved, organic rich carbonate silt laminites which formed in long lived permanent lakes. Bedding parallel sheets of carbonate (arrowed) formed during diagenesis as a result of remobilisation of carbonate. Middle ORS of Easter Ross. (Pencil = 10cm).*

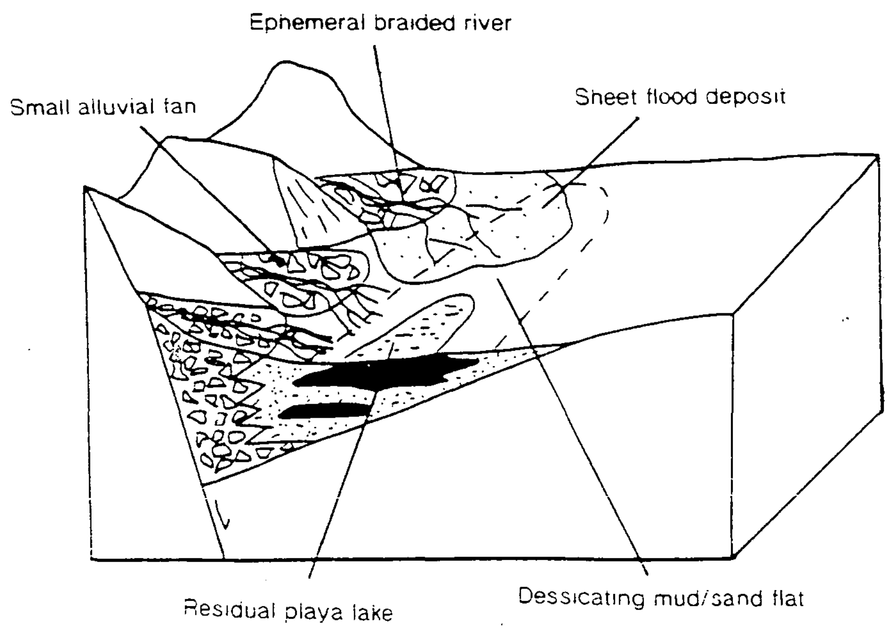



*Fig. 2.A. Block diagram showing distribution of Lower ORS depositional environments.*

1981a,b; Astin, 1985; Trewin, 1986), with a maximum suggested duration of about 10,000 years (Rayner, 1963; Trewin, 1985).

## 2.3 Depositional environments

Descriptions of the depositional environments present in the ORS are given below for the Lower, Middle and Upper ORS. Most detail is given for the environments seen around the Moray Firth, as this is the main thesis area. However, comparisons will be made with other areas to highlight any significant differences.

### 2.3.i Lower ORS

This section is based on summaries of the work of Mykura (1983), Mykura & Owens (1983), Rogers (1987) and on observations made during this study. The Lower ORS consists of three intercalated depositional environments (Fig. 2A):

1) Alluvial fan deposition as evidenced by lenticular matrix supported conglomerates and breccias. The clasts are dominantly sub-angular to angular pebbles, cobbles and boulders, derived locally from Moine or granite basement, set in a matrix that can be either sandy or muddy.

2) Alluvial flood plain deposits forming distally to the alluvial fans which deposited purple and grey, micaceous, medium to fine grained sandstones. These are planar bedded (Sh) with frequent rippled bedding surfaces (Sr) and mudstone intercalations (Fl). Larger scale cross-bedding (St) as a result of channel formation, with sets generally less than 50cm thick is rare.

3) Evaporating playa lakes which deposited grey and grey-green, organic rich siltstones, mudstones and dolomicrites, which are flat or irregularly laminated.

The Lower ORS is dominated by fine grained sediments deposited in localised basins, and as such there is much lateral variation in facies over relatively short distances. The overall pattern is of alluvial fans passing distally into either flood plain

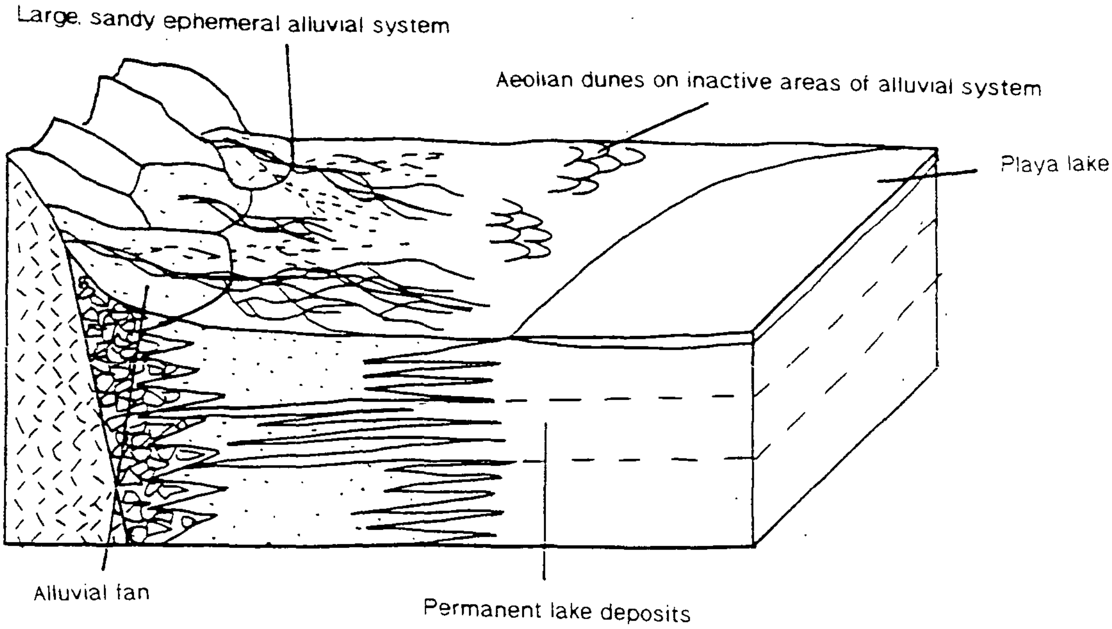



*Fig. 2.B. Block diagram showing distribution of Middle ORS depositional environments.*

deposits or lacustrine deposits, depending on local rates of subsidence and the amount of fluvial input.

## 2.3.ii Middle ORS

There are four depositional environments present in the Middle ORS around the Moray Firth: alluvial fans, sandy fluvial, lacustrine and aeolian, each of which will be discussed in turn. Overall the fluvial deposits dominate with lacustrine intercalations due to the southward expansion of lakes in the Caithness and Orkney area as well as aeolian intercalations (Fig 7.3).

Alluvial fan conglomerates and breccias are common around the margins of the Orcadian Basin where they formed along active synsedimentary faults. The fans are dominated by stream flow deposits with lesser amounts of debris flow deposits, predominantly on the upper reaches of the fans. The alluvial fans intercalated distally with the fluvial deposits below. The size of the fans varied from a few km up to approximately 12km extent from fault scarps (Rogers, 1987) and compare with modern analogues such as those in Death Valley, California, which normally extend for not more than 20km from active fault scarps into the adjacent basin.

The sandy fluvial deposits can be divided into three categories (Rogers, 1987). 1) Medium to coarse intraformational conglomerate bearing sandstones. These are dominated by trough cross-bedding (St) and upper flow regime plane bed horizontal laminations (Shu) with minor amounts of planar cross-bedding (Sp) and ripple cross-lamination (Sr). Thin mudstone intervals and convolute bedding are common. These sediments were deposited in a braided fluvial environment.

2) Fine to medium grained sandstones dominated by upper flow regime plane bed horizontal laminations (Shu) with subordinate low angle laminations (Sl) and ripple cross lamination (Sr). They occur in erosively based beds, <2m thick which can be

isolated or stacked and intercalated with playa mudstones. These are the deposits of unconfined sheet floods.

3) Fine to medium grained channel fill sandstones, dominated by upper flow regime plane bed horizontal laminations (Shu) with subordinate trough cross-bedding (St) and low angle laminations (Sl). These channels are erosively based, commonly overlain by mudstone intraclast conglomerates; convolute bedding is also common. The channels occur within the playa mudstones and sheet flood deposits, either individually or erosively stacked.

The three categories of fluvial deposit represent a proximal to distal variation from a fully developed braided river system to sheet flooding and incised channels which formed on playa surfaces during flood events.

The most common lacustrine deposits are the playa mudstones intercalated with fluvial sandstones and siltstones that formed in temporary lakes. These have a higher proportion of clastic interbeds than the equivalent deposits further north because they are nearer to the basin margin and fluvial input. The permanent lake deposits are less common and are intercalated with fluvial and playa sandstones. They consist of carbonate/silt laminites with subordinate grey mudstones and flat laminated to ripple cross-laminated fine sandstones and siltstones resulting from fluvial input and wave ripple cross-laminated sandstones that formed in shallow water.

The lacustrine deposits of Caithness and Orkney which are from nearer the centre of the Orcadian Basin are quite different. In addition to having the carbonate/silt laminites there are thick sequences of flagstones comprising thinly bedded wave rippled or flat laminated, calcareous or dolomitic, fine sandstones and siltstones, with thin mudstones. These flagstones will have formed in both permanent and temporary lakes. There are also thick sequences (often tens of metres thick) of sandstone which contain cross-bedding, flat laminations, current ripples, wind ripples,

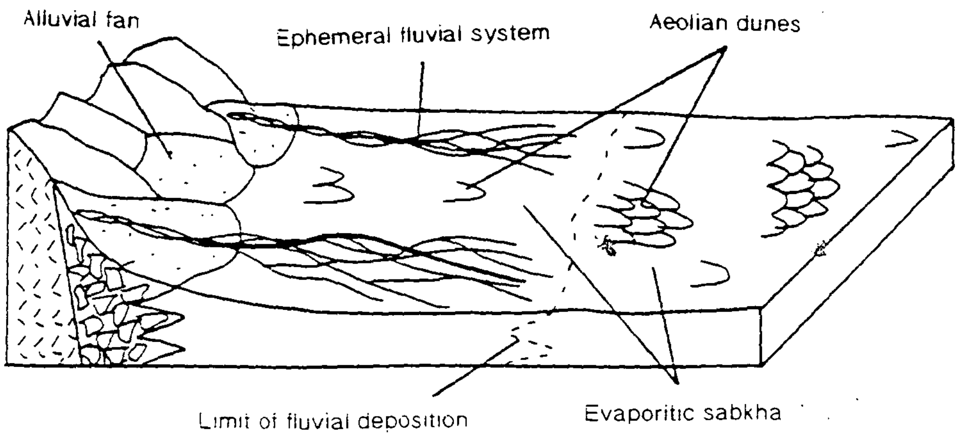



*Fig. 2.C. Block diagram showing distribution of Upper ORS depositional environments.*

and convolute laminations. The sandstones are interbedded with thin mudstones. These represent aeolian fluvial and marginal lacustrine deposits.

Aeolian deposits consist of cross-bedded dune sandstones (Spe) and horizontally laminated grain fall and wind ripple laminae (She). The dunes were probably compound barchanoid dunes - ridges transverse to the palaeowind direction with cuspate slip faces (Rogers, 1987). Dune sandstones within the Middle ORS are separated by flat bedded interdune deposits of sand and mud which show wind ripple flat lamination, granule rich horizons and dissipation structures (See Fig. 2.12). These interdune deposits have a maximum thickness of at least 2m.

## 2.3.iii Upper ORS

The three main Upper ORS depositional environments are fluvial, aeolian and evaporitic sabkha. (Fig 2.C) These have been grouped into four associations in Easter Ross: aeolian; mixed sabkha and aeolian; fluvial; and mixed fluvial/sabkha (Rogers, 1987). Elsewhere around the Moray Firth, only fluvial and aeolian environments have been recognised (Armstrong, 1964, 1977; Peacock *et al*, 1968; Mykura, 1983).

The Tarbat Ness unit, the older of the two units comprising the Upper ORS, is almost totally composed of fluvial deposits. Over 95% of this facies is composed of medium to coarse, often pebbly sandstones. The most dominant lithofacies is trough cross-bedding (St) with subsidiary planar cross-bedding (Sp), low angle and horizontal laminations (Sl, Sh), and ripple cross-laminations (Sr). The majority of the succession is composed of cycles 2 to 40m thick containing St passing up into Sh. Sometimes erosion surfaces can be seen at the base of these cycles but usually neither the erosion surfaces nor the cycles are easily recognised as a result of weathering and erosion. Thin mudstones and siltstones representing overbank deposits and the final stages of channel fills occur but are rare due to erosion by succeeding fluvial cycles as evidenced by the fairly common occurrence of intraformational mud chip conglomerates.

These fluvial deposits share characteristics in common with many modern and ancient braided river deposits (Cant & Walker, 1976, 1978; Miall, 1977, 1978; Bluck, 1979, 1980). They resemble the South Saskatchewan type vertical profile of Miall (1978) which is common as a distal facies in sand dominated braided river deposits. The presence of braided alluvium suggests a high rate of discharge and a high variability in discharge (Miall, 1977). Fluctuating discharge and groundwater levels are also indicated by the presence of mudstone intraclasts and the prevalent red colour of the sediments (Rust, 1984).

***Fig. 2.16.*** *Thick sequence of small scale aeolian dune cross bedding from the Upper ORS of Easter Ross. Note the absence of muddy interdune deposits, in contrast to the Middle ORS aeolian deposits. (Cliff height is approximately 6.5m).*

The thick aeolian dune association only occurs in the Gaza unit towards the top of the Upper ORS (Fig. 2.16). There are two sequences of uninterrupted aeolian dune cross-bedding. The sets are <6.5m thick and a dune height of approximately 8m has been estimated by Rogers (1987) based on partial dune width measurements of 30-40m. Interdune deposits are not preserved along set boundaries and sabkha deposits only occur at the toes of cosets. As in the Middle ORS, the dunes were probably compound barchanoid type. Some large scale soft sediment deformation structures are visible which would have probably resulted from differential loading by subequent dunes or earthquake induced shocks.

The remainder of the Gaza unit is dominated by intercalated aeolian dune and sabkha deposits (Fig. 2.17). Overall the sabkha deposits dominate with the proportion being about 60:40, although locally they reach parity. The dune sets vary from 0.5-2m thick with an estimated dune height of 2-3m (Rogers, 1987). The thickness of the dune intercalations ranges from one set (0.5m) to a maximum of 5m, while the sabkha intervals are <5m thick.

The remaining facies association, interbedded fluvial and sabkha deposits, occurs in both the Gaza and Tarbat Ness units in minor amounts (Fig. 2.18). The fluvial deposits are very much as described above except that lithofacies St is almost entirely dominant with only minor amounts of Sh. The alternating intervals of sabkha sandstones are usually 1-2m thick and fluvial sandstones are 1-5m thick, although there are occasional thicker intervals of more than 10m. The base of the fluvial deposits is


***Fig. 2.17.** Intercalated aeolian dune (D) and sabkha deposits (S) with planar erosional (deflation) surfaces (arrowed) at the base of the aeolian units. Upper ORS from Easter Ross. (Hammer length = 40cm).*


*Fig. 2.18. Intercalated fluvial cross-bedding (F) and sabkha deposits (S). Note the curved, downcutting erosional surfaces at the base of the fluvial units. Upper ORS of Easter Ross. (Hammer length = 40cm).*

erosive into the sabkha sandstones, while the upper surface shows an abrupt change back into sabkha deposits. This planar upper surface probably formed as a result of deflation following periods of fluvial deposition. This facies association represents a more distal environment to that of the fluvial deposits described above.

## 2.4 Controls on sedimentation

The three lithostratigraphic subdivisions, Lower, Middle and Upper ORS, have different facies associations as a result of variations in tectonic and climatic controls. The major difference between the Lower ORS and the Middle and Upper ORS reflects the scale of tectonism - small isolated basins during the Lower ORS coalescing into one large basin during the Middle ORS leading to basin wide sedimentation, which continued through the Upper ORS. Increased tectonism at the start of the Upper ORS caused coarser fluvial deposits to prograde further into the basin. However this tectonism slowly waned as shown by the overall fining-up trend of the

Upper ORS sediments. On a smaller more local scale, active syn-sedimentary faults and avulsion resulted in abandonment of fluvial systems allowing the formation of aeolian dune fields. A modern analogue can be seen in the Lake Eyre Basin, Central Australia (Rust, 1981).

Overprinting the tectonism is a climatic control which is most noticeable in the Middle ORS. Although the overall climate was arid, the presence of varved permanent lake deposits suggests wetter (semi-arid) periods with marked seasonality. The presence/absence of the playa lakes indicates more rapid fluctuations between wetter and drier times within a generally arid climatic setting. In addition there is an overall increase in aridity throughout the ORS, which can be linked to the northward drift of the Devonian continent across the equator.

## 2.5 Summary

A wide spectrum of depositional environments exists within the ORS; aeolian, fluvial, lacustrine and evaporitic sabkha. Braided fluvial deposits are the most dominant, with lesser amounts of alluvial fan, aeolian, permanent and temporary lacustrine (Lower and Middle ORS only) and sabkha (Upper ORS only) deposits. The distribution of the different facies is controlled primarily by tectonism and climate.

A knowledge of the facies, the depositional environments and their distribution, and details about the geometries and scale of the constituent lithofacies is essential when assessing the quality of a potential reservoir rock. When used in combination with diagenetic studies and poroperm characteristics it provides a means of subdividing the reservoir into effective and non effective zones according to their fluid flow properties.

# Chapter 3
# Petrology and Diagenesis

## 3.1 Introduction

For the purposes of this study the Upper and Middle ORS have been largely treated as one because of the many similarities between them. Examples from both units will be used to illustrate the discussion, although any differences will be highlighted.

Over seventy thin sections have been studied with a standard petrological microscope. In addition, twenty sections have also been examined under cathodoluminescence; ten samples with a scanning electron microscope and eleven (eight sandstones and three mudstones) by X-ray diffraction in order to determine information about grain size, sorting, composition, packing, porosity and diagenesis.

## 3.2 Techniques used

### 3.2.i Petrographic techniques

Prior to sectioning, all samples were impregnated with a blue epoxy resin to facilitate the recognition of porosity during petrographic examination. All thin sections that were examined with a petrographic microscope were point counted (300 counts/section) to determine the mineralogical composition of the sample, the proportions of cements and framework grains, and the amount of porosity present. Also, the mean grain size and sorting for each section were obtained from grain measurements.

Compaction estimates were made by randomly selecting 100 grains from each section and counting the number of contacts for each grain and the proportions of the different types of contact present. The terminology for contact types used is that proposed by Taylor (1950): point; planar; concavo-convex; and sutured, which indicates increasing degrees of compaction. The amount of compaction that the sandstones have undergone can be expressed in two ways. First is the compaction index (CI), as suggested by Taylor (1950), and the second is the tight packing index (TPI) introduced by Wilson and McBride (1988). Compaction index is simply the average number of grains in contact with any other grain, while TPI is the average number of planar, concavo-convex, and sutured contacts which any grain has.

Both CI and TPI are dependant on sorting and, to a lesser extent, grain size (Harrell, 1981). For example, in a poorly sorted sand a large grain will have more contacts than a smaller grain thus giving an abnormally high value for CI and, to a lesser extent, TPI. As a result, only sands that were well sorted or better were used. Although several workers have measured CI for uncompacted Holocene sands (Harrell, 1983; Wilson and McBride, 1988; Atkins and McBride, unpublished data, in McBride *et al.,* 1990), the values obtained varied from 0.27 to 1.1.

Both CI and TPI increase with burial as a result of increases in the closeness of packing with burial (Taylor, 1950; Moore, 1975; Wilson and McBride, 1988; McBride *et al.,* !990). The study by McBride *et al.* (1990) of the Palaeocene Carrizo Sandstones from the Texas Gulf Coast suggests that maximum values of CI and TPI were obtained at a depth of 3000m. This, however, is not typical of other examples although they do all have different rates of compaction, which McBride *et al.* (1990) attribute to differing thermal maturities of the sandstones studied.

Some sections were examined with a cold cathodoluminesence microscope to obtain more information on the nature of the carbonate cement present.

### 3.2.ii SEM examination

Samples were selected for SEM examination so that a closer study could be made of grain features and the relationships between grains and cements, and between the various cements themselves. Samples were mounted on 10mm diameter stubs using a conductive silver paint. They were then gold coated to a thickness of approximately 2$\mu$m.

Two scanning electron microscopes were used: a Cambridge 600 in the School of Engineering and Applied Science, University of Durham, and a Cambridge S240 from the Medical School, University of Newcastle Upon Tyne. Operating conditions were 15Kv accelerating voltage at a vacuum pressure of 5 x $10^{-5}$ Torr. All the photographs were taken on the latter microscope, using black and white film at magnifications of up to 8500x.

### 3.2.iii X-ray diffraction

Eight sandstones and three mudstones were analysed semi-quantitatively for their clay content using X-ray diffraction. The samples were coarsely crushed before being powdered mechanically. Next 5mg of powder were added to 100ml of distilled water and left to settle for 30 minutes to separate out the bulk of the denser siliciclastic material. Finally, glass slides were gently placed into the beakers which were then left to desiccate so that the clay minerals settled out onto the slides in an orientated fashion. For comparison some bulk sample slides were also made up by mixing a small amount of the powder with acetone and then spreading on a glass slide and leaving to dry.

The samples were run on a Philips 3kw PW1130 X-ray generator diffractometer assembly using a filament that produced cobalt radiation. The results were plotted graphically to ease interpretation. All samples were glycolated for three hours at 60°C and then re-run to give an estimate of the amount of mixed layer clays present, before finally being heated to 500°C for one hour and run for a third time to ascertain the amounts of chlorite present.

## 3.3 Compositional analysis

For the purposes of point counting the framework grains were taken as monocrystalline quartz (Qm); feldspar (F); and total lithic fragments (Lt). This then enabled their proportions to be plotted on ternary diagrams after the manner of Dott (1964) and Dickinson and Suczek (1979). The feldspar is predominantly K-feldspar, which suggests a granitic source rock, with lesser amounts of plagioclase and microcline. Feldspars are rarely fresh, due to their instability, and show signs of dissolution and/or replacement by calcite. Microcline is the least altered of the feldspars. The lithic fragments consist of polycrystalline quartz fragments and chert (sedimentary and metasedimentary); quartz/feldspar fragments (igneous); and ductile grains of metamorphic origin and shale rip-up clasts.

Other grains present include biotite and muscovite mica (1-3%), but only in fine-grained sandstones and mudstones, both clastic and carbonate. Micas (especially muscovite) are frequently bent or broken. Both types of mica show signs of alteration (see below). Also there are very rare trace amounts of heavy minerals, including rutile, zircon and ?tourmaline. The paucity of heavy minerals could be a result of scarcity in the source area or later dissolution. Although there is no direct evidence of the pre-existence of other heavy minerals, their dissolution could provide material for the clay/Fe-oxide rims observed; Pettijohn (1941) states that a dominance of rutile and zircon could be due to them being more stable than the other heavy minerals. Also Burley (1984) notes that chloritic clays such as those found in the ORS often form as a result of the dissolution of ferromagnesian minerals.

The clay fraction has been studied by XRD. In every sandstone sample studied, except one, kaolinite was the dominant clay present. The other clays identified, in lesser amounts, are chlorite, interstratified chlorite-montmorillonite, illite and muscovite (Fig. 3.1). The mud rocks yielded predominantly a mixture of illite and

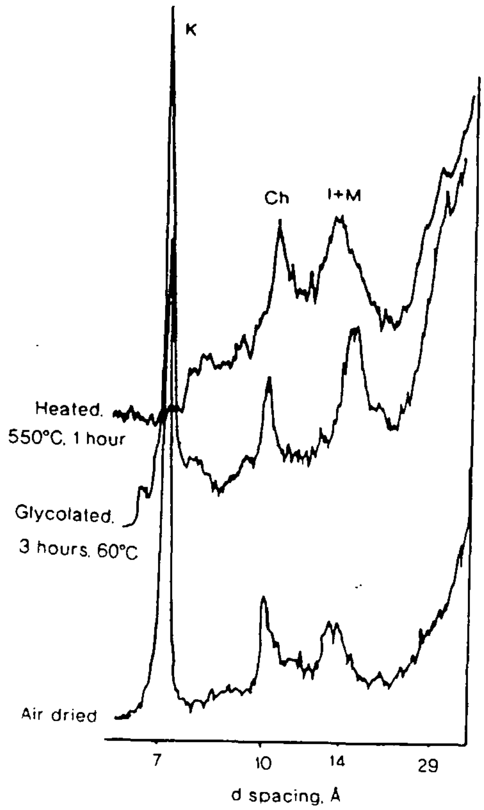



*Fig. 3.1. Typical XRD trace of the clay fraction from ORS sandstones. Kaolinite (K) dominates the assemblage with lesser amounts of chlorite (CH), illite (I) and muscovite (M).*

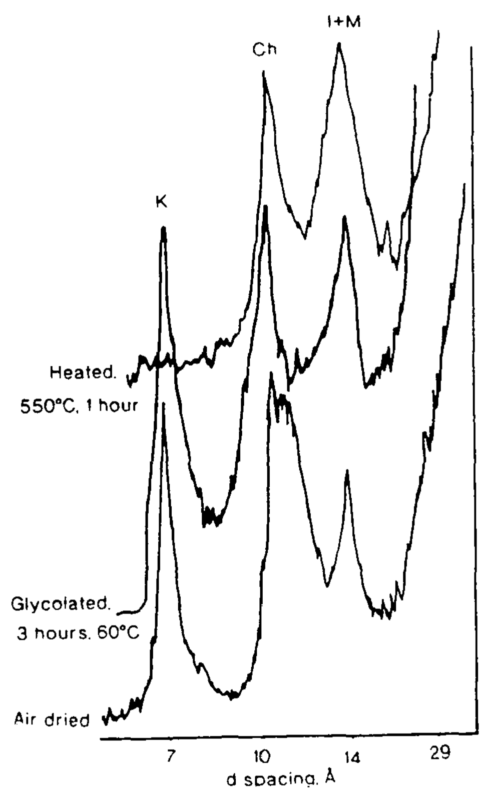



*Fig. 3.2. Typical XRD trace of the clay fraction from an ORS mudstone. The assemblage is dominated by a mixture of illite and muscovite (I+M) with lesser amounts of kaolinite (K) and chlorite (CH).*

muscovite with lesser amounts of kaolinite, and chlorite (Fig. 3.2). This assemblage is similar to those found by Burrolet *et al.,* (1969) and Wilson (1971) in Middle ORS rocks from Caithness and the Moray Firth which contained illite, chlorite, kaolinite montmorillonite and illite/smectite.

The illite, muscovite and chlorite are probably detrital (Wilson, 1971), while the kaolinite and chlorite-montmorillonite are authigenic. Other mineral phases present in the sandstones are all diagenetic and include calcite, haematite and pyrite. These, and the clay minerals, will be discussed in more detail later in this chapter.

The average values for the framework grains are Qm,75%; F,6.5%; Lt,18.5% for the Upper ORS and Qm, 67%; F, 8.7%; Lt,12.3% for the Middle ORS. As can be seen from the ternary plots (Fig. 3.3) the majority of the points fall within the sublitharenite field and there is no change in composition with varying lithofacies. However, there is a slight increase in compositional maturity of the Upper ORS over the Middle ORS, probably due to a certain amount of recycling of sediment which will cause further removal of unstable feldspars and lithic fragments.

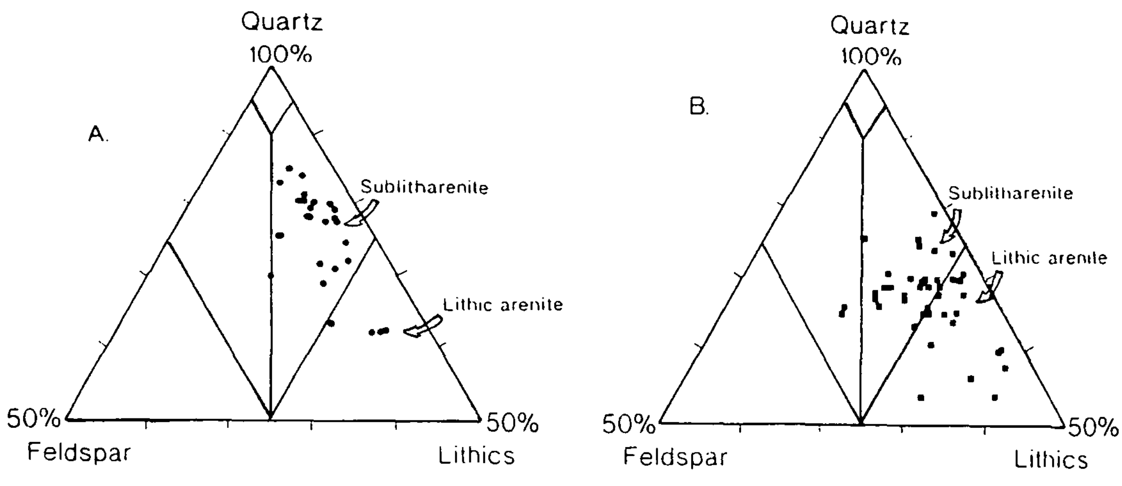



*Fig. 3.3. Compositional plot for ORS sandstones. a) Upper ORS; b) Middle ORS.*

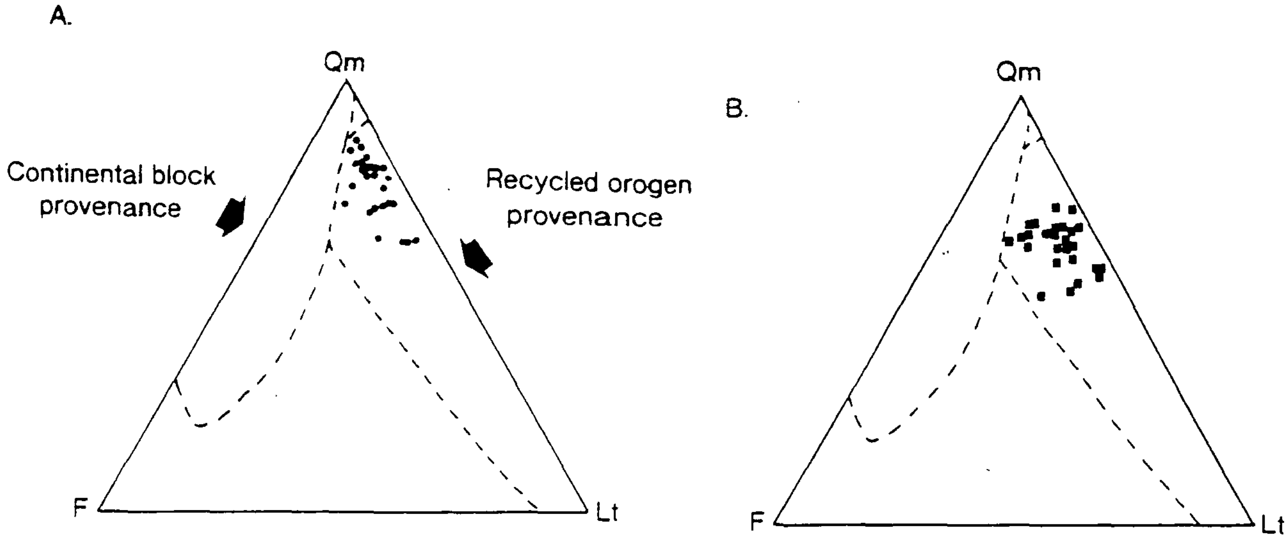



*Fig. 3.4. Compositional plot overlaid with provenance fields of Dickinson and Suckzec, (1979). Qm = quartz grains; F = feldspar grains; Lt = lithic grains. a) Upper ORS; b) Middle ORS.*

When the provenance fields of Dickinson and Suczek (1979) are overlaid on the compositional plots, the data falls within the recycled orogen field (Fig 3.4). Shanmugam (1985) points out that dissolution of unstable framework grains (feldspar and lithic fragments) causing a relative enrichment of quartz can lead to a misinterpretation of the provenance. Also, Mack (1978) suggests that as much as 16% of original feldspars are destroyed during the first 70km of transport. As a result of this the ternary plots should be used with care. However, the recycled orogen field indicated by Fig. 3.4 is still a fairly accurate assessment, as the source area was Dalradian and Moinian metamorphic uplands of the Caledonian orogenic belt (see Chapter 1).

## 3.4 Grain size

The long axes of 100 randomly selected grains from each sample were measured in order to calculate the mean grain size and sorting. The arithmetic mean has been used for the grain size, while the sample standard deviation (in phi units) has been used to obtain the sorting. Computational methods were employed instead of graphic measures because they are far quicker to use. Although such methods are not perfect (*eg.* McCammon, 1962b), graphical methods are not necessarily any better (Folk, 1966). Folk (1966) states that:

> "...each method has its advantages and its drawbacks, and each is equally valid for comparing a suite of samples."

The grain size data have been plotted on histograms because there is no need to extract statistical information from the graphic presentation of the data and because they visually portray the grain size spread adequately. The histograms (Fig. 3.5) show a near normal distribution with only a slight negative skewness. The sands are very clean (they contain virtually no terrigenous clays) and well sorted, with sorting coefficients ranging from 0.39 (well sorted) to 0.73 (moderately sorted) with an average of 0.54 (moderately well sorted). The mean grain size for the Upper ORS (0.19mm) is slightly greater than that for the Middle ORS (0.14mm). This is due to the fact that the Upper ORS deposits are of a more proximal nature, that is, they were deposited by a higher energy system carrying coarser material (although not necessarily physically closer to the source area).

This analysis was not carried out to help environmental interpretation, which is well constrained from field evidence, but rather to aid in assessing the controls on the poroperm characteristics which are discussed in more detail in Chapter 4. However, as a matter of academic interest, some of the data have been plotted on arithmetic probability paper, in the manner of Visher (1969), who carried out extensive work on

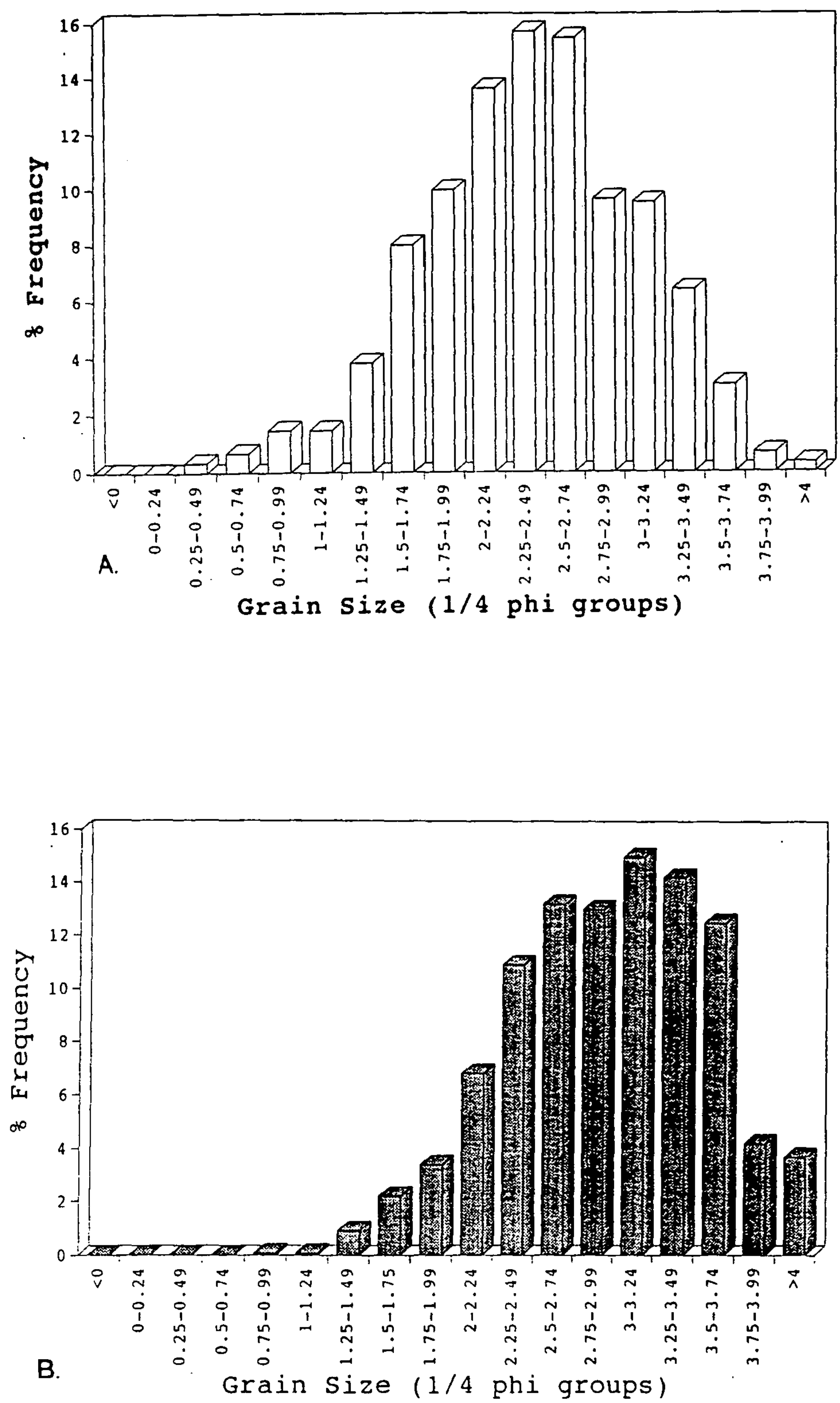



*Fig. 3.5. Grainsize histograms. a) Middle ORS, b) Upper ORS.*

the relationship between the depositional environment and the shape and make-up of the cumulative curve in order to interpret the depositional environment of ancient sandstones. Visher (1969) divided the plot into segments which correspond to the processes of suspension, saltation, and traction (Fig. 3.6).

The fluvial environments of Visher (1969) are dominated by a well developed suspension population and the absence of a traction population. As can be seen from Fig. 3.7 all samples show a well developed saltation population with a suspension population, while the traction population, although generally poorly developed is dominant in some cases.

## 3.5 Diagenesis

Diagenesis is the chemical and physical changes that affect sediments following deposition. These changes can occur anywhere from near surface to deep subsurface environments. In a study of carbonate diagenesis Choquette and Pray (1970) divided the diagenetic regime into three: (1) eogenetic - before effective burial in the environment of deposition; (2) mesogenetic - at any depth of burial above the zone of metamorphism; (3) telogenetic - during exposure following a period of burial. Schmidt and McDonald (1979a) applied this terminology to sandstone diagenesis and the same subdivisions have been used in this study (Fig. 3.8). Early changes include the formation of calcretes and similar changes to those reported by Walker (1976) and Walker *et al.* (1978) from recent desert sediments. The important changes as regards the reservoir characteristics, are those which cause the destruction or creation of porosity. These include compaction, dissolution of unstable grains and cements, and the precipitation of calcite and kaolinite cements.

### 3.5.i Calcretes

Calcretes have been widely identified in Devonian sediments in Britain (Maufe, 1910; Burgess, 1961; Allen, 1965, 1973, 1974a,b; Leeder, 1976a). They occur

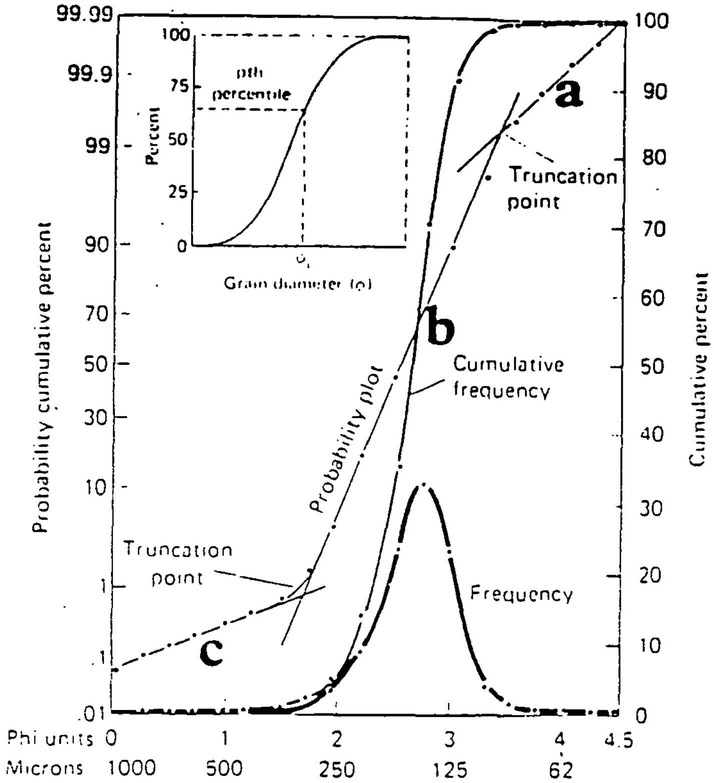



*Fig. 3.6. Segmented cumulative curve plotted on lognormal probability paper showing (a) the suspension, (b) the saltation and (c) the traction populations (adapted from Visher, 1969).*

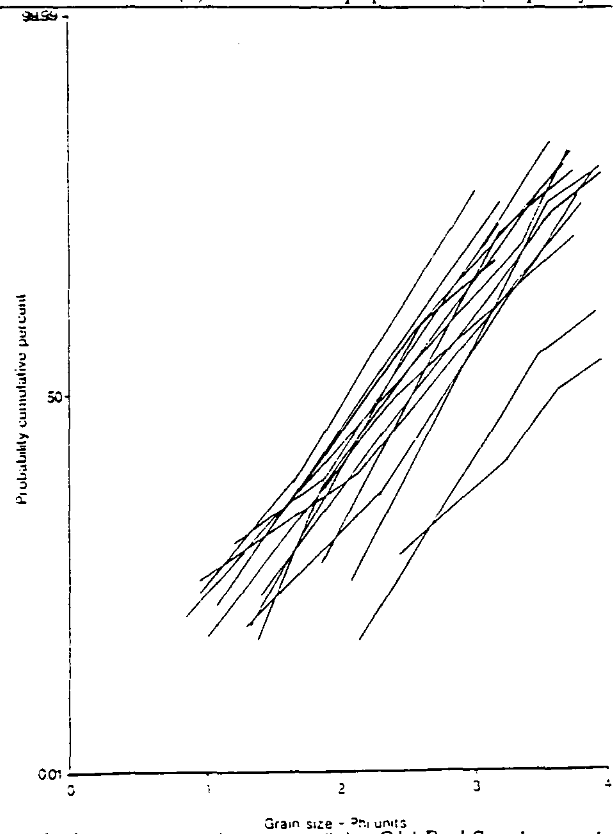



*Fig. 3.7. Typical cumulative segmented curves of the Old Red Sandstone demonstrating a well developed saltation population but poorly developed suspension and traction populations.*

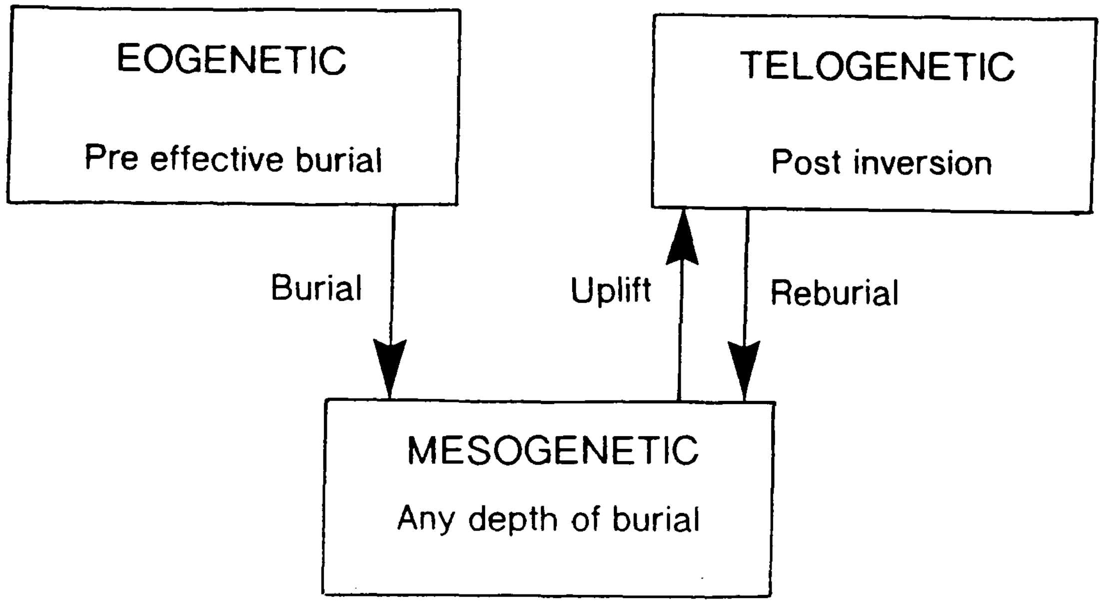



*Fig. 3.8. Diagenetic regimes of development of secondary sandstone porosity. Redrawn from Schmidt & McDonald, (1979).*

throughout the Orcadian Basin succession (although they are more common in the Upper ORS), including around the Inner Moray Firth (Donovan, 1982; Parnell, 1983). The calcretes occur both in-situ in mudstones and sandstones as well as being reworked into channel deposits. They represent periods of non-deposition and pedogenesis (indicating increased tectonic stability).

Calcretes have been classified in a variety of ways. Firstly there is a morphological classification which relates to stages of development of the calcrete profile (Netterberg, 1980). This type of classification has been used by many workers for both Quaternary and pre-Quaternary calcretes (Gile *et al.*, 1966; Allen, 1974a,b; Steel, 1974; Machette, 1975) with the classification presented by Machette (1985) being probably the most comprehensive (Table 3.1). In this classification six stages have been recognised, representing a progressive increase in maturity, from thin grain coatings to complex fabrics of laminae, commonly brecciated and recemented.

| Stage | Detrital content | Diagnostic features | $CaCO_3$ distribution | Maximum $CaCO_3$ content |
|---|---|---|---|---|
| 1 | High | Thin discontinuous coatings on pebbles, usually on undersides. | Coatings sparse to common. | Trace to 2% |
| | Low | Few filaments in soil or faint coatings on ped surfaces. | Filaments sparse to common. | Trace to 4% |
| 2 | High | Continuous, thin to thick coatings on tops and undersides of pebbles. | Coatings common, some carbonate in matrix. | 2-10% |
| | Low | Nodules, soft 5-40mm in diameter. | Nodules common, generally non-calcareous to slightly calcareous. | 4-20% |
| 3 | High | Massive accumulations between clasts, fully cemented in advanced forms | Continuous in matrix to form K (alpha) fabric. | 10-25% |
| | Low | Mainly coalesced nodules, matrix is firmly to moderately cemented. | Continuous in matrix to form K (alpha) fabric. | 20-60% |
| 4 | Any | Thin (<2mm) to thick (10mm) laminae capping hard pan. | Cemented, platy to tabular structure. 0.5-1m thick | >25%in high detrital content. >60 in low detrital content |
| 5 | Any | Thick laminae (>10mm) with thin to thick pisoliths above. Laminated carbonate may coat fracture surfaces. | Indurated, dense, strong, platy to tabular. 1-2m thick | >50%in high detrital content. >75 in low detrital content |
| 6 | Any | Complex fabric of multiple generations of laminae, brecciated and recemented, pisolitic. Typically with abundant peloids and pisoliths in fractures. | Indurated, dense, thick, strong, tabular structure. >2m thick | >75% in all detrital contents. |

***Table 3.1.*** *Classification of pedogenic calcretes . Redrawn from Machette (1985). High detrital content refers to >50%; low is <20%. Non pedogenic carbonate must be ignored when assessing* $CaCO_3$ *distribution.*

Secondly, calcretes can be classified according to their hydrological setting (Carlisle, 1980, 1983). In addition to near surface conditions, calcretes can also form within or just below the capillary fringe (Arakel & McConchie, 1982) or with or without the influence of phreatophytic plants (Semeniuck & Meagher, 1981) at depths of up to tens of metres below the surface.

Netterberg (1980) suggested a mineralogical division of carbonate duricrusts based on their dolomite content. It would also be possible to devise classifications according to whether the duricrusts were gypsiferous, siliceous or ferroan but there seems little point to this (Wright & Tucker, 1991).

Finally, microstructure can be used as a method of classification. Wright (1990) has recognised two end member fabrics. Alpha fabrics, which correspond to K-fabrics of Gile *et al.* (1965) and Bal (1975) which are predominantly physio-chemical precipitates, and beta fabrics which are mainly biogenic in origin. Alpha fabrics are dominated by a dense micritic microfabric with nodules, complex cracks, rhombic calcite crystals and floating sediment grains. Beta fabrics typically exhibit needle fibre calcite, microbial tubes, alveolar-septal fabric and the microfossil *Microcodium.*

***Fig. 3.9.*** *Nodular calcrete (stage II) developed in Upper ORS fluvial mudstones. (Pencil length = 14cm).*

Morphological classifications are quick and easy to use, and for the purposes of this study Machette's (1985) classification has been adopted, although comparisons with the microstructural classification of Wright (1990) will also be made.

The calcretes occur in both mudstones and sandstones and have a variety of forms. In the mudstone they occur as white, vertical or horizontal nodules about 5-15cm x 10-15cm (Fig. 3.9). In the sandstones, calcretes occur in two forms: as red nodules, either <1mm or about 5-20cm in size (Fig. 3.10), or as white and red nodules in a chalky white matrix (Fig. 3.11).

In thin section a variety of pedomicromorphological features can been seen. In some samples there are microspar nodules (0.1-0.2mm in diameter) which grew either replacively or displacively in the phreatic zone, as indicated by the lack of fractured grains. If calcite growth occurred in the vadose zone as meniscus cements, stresses caused by the calcite growth would concentrate at grain/grain contacts and lead to fracturing of the grains (Buczynski and Chafetz, 1987). The sandstone is otherwise unaltered (Fig.3.12). These are equivalent to stage 2 calcretes (Table 3.1).

***Fig. 3.10.*** *Red calcareous nodules in Upper ORS fluvial sandstone (stage II calcrete). (Hammer length = 40 cm).*

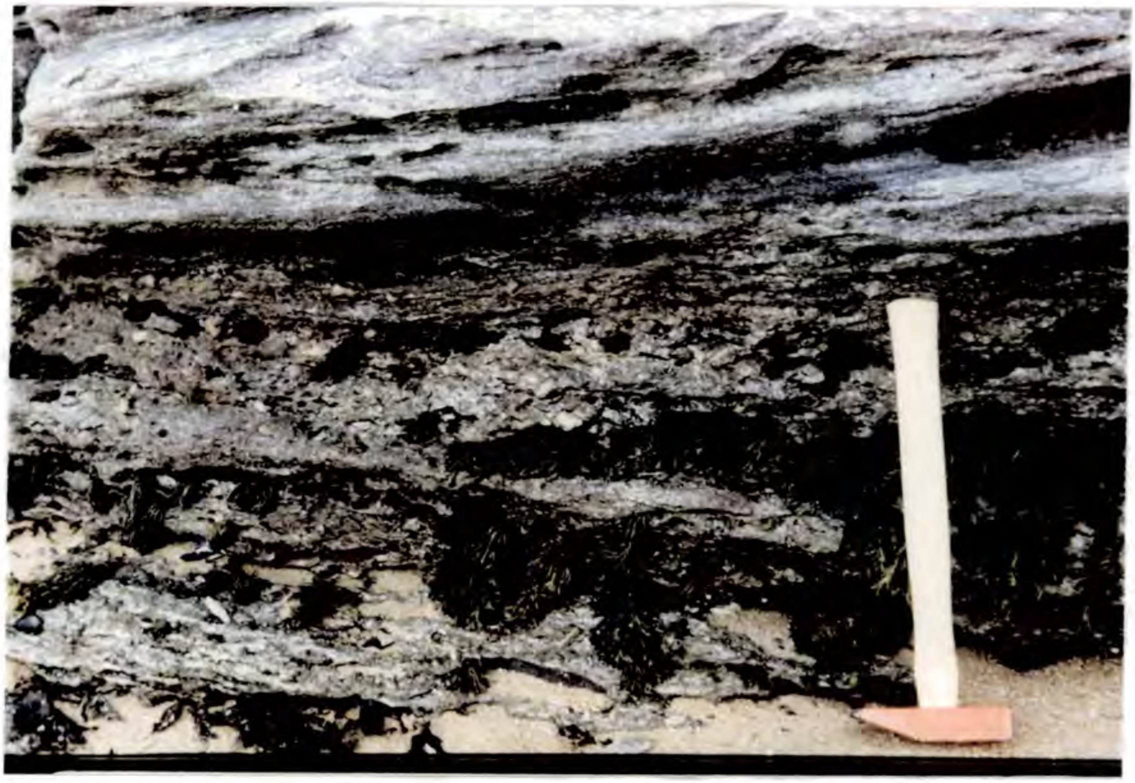


*Fig. 3.11. Fluvial cross-bedded sandstone with white nodules in a chalky martix: stage III calcrete from the Upper ORS. (Hammer length = 40cm).*

Other examples consist of micritic nodules (0.2-3mm) in a matrix of microsparry calcite with occasional patches of coarse calcite spar and infrequent framework grains (approx. 30%) which are predominantly quartz (Fig. 3.13). The micritic nodules have sharp boundaries which suggests that they did not form *in situ*. The microspar has good crystal shape, often showing rhombic faces, and has a varied grain size leading to a mottled fabric as a result of inclusion of earlier nodules into the groundmass (Wright, 1990). Under cathodoluminescence the calcite exhibits alternating colour bands which suggests slow, perhaps episodic growth. The overall lack of framework grains, and the predominance of quartz amongst them, suggests that the calcite has grown largely replacively, at the expense of the less stable lithic fragments and feldspars. This fabric closely resembles the alpha fabric of Wright (1990), and the calcretes belong to the more mature stage 3 calcretes according to Machette's classification.

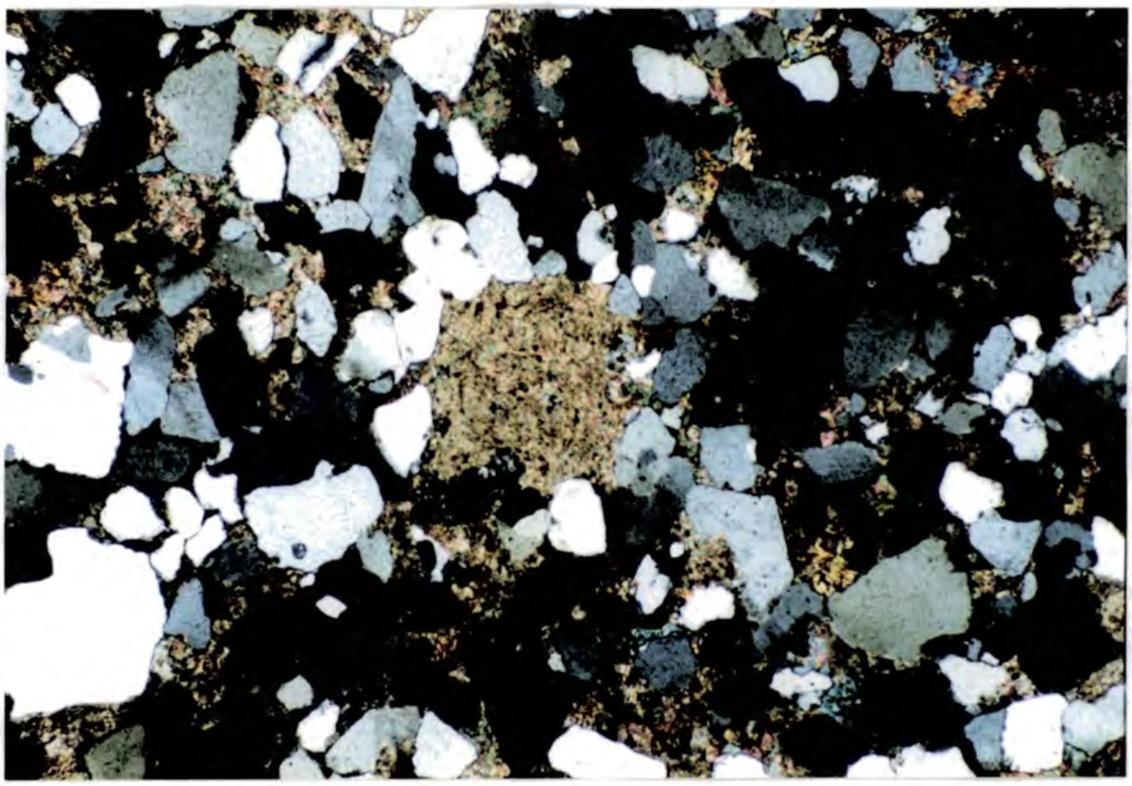


*Fig. 3.12. Photomicrgraph of a similar sample to that seen in Fig. 3.10 showing a nodule in an otherwise unaltered sandstone. (XPL, FOV=3mm).*

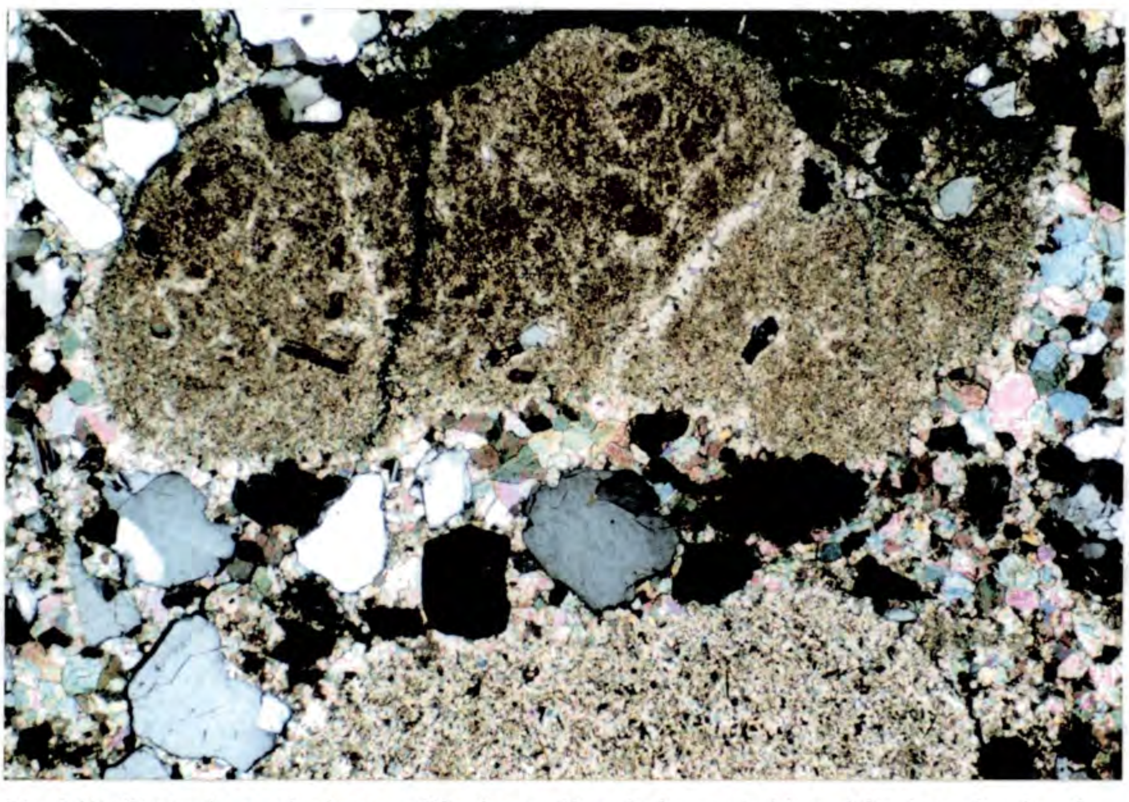


*Fig. 3.13. Photomicrgraph of a stage III calcrete. Note the large micritic nodules, irregular size of calcite crystals in matrix and sparse framework grains. (XPL, FOV = 3mm).*

The presence of Wright's (1990) alpha fabric in the more mature calcretes shows that they formed as result of physio-chemical rather than biogenic processes. The vertical nodules in the mudstone consist of structureless micrite. They lack any of the features diagnostic of rhizoliths described by Klappa (1980) and are therefore unlikely to be biogenic. The only evidence of any biogenic affect on the ORS is a solitary root mold from a lacustrine limestone that formed close to the margin of a Middle ORS lake (Fig 3.14).

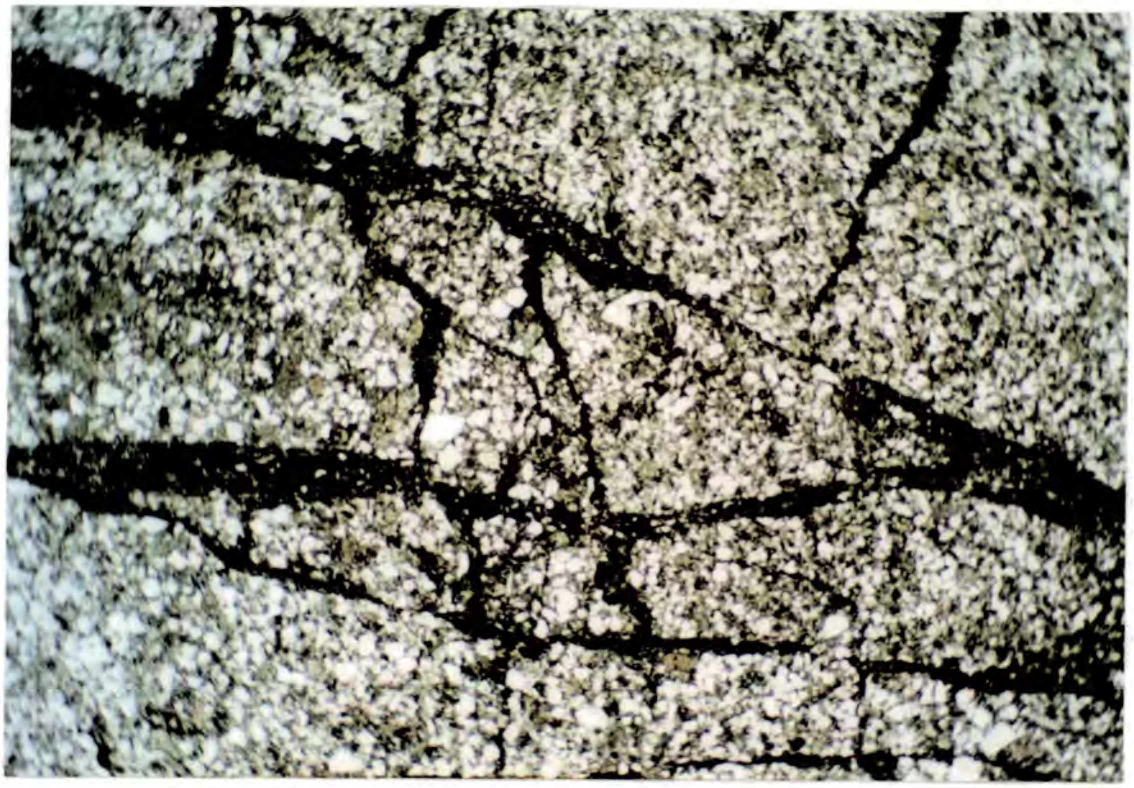


*Fig. 3.14. Photomicrograph of a root mold from a Middle ORS calcareous mudstone. (PPL, FOV = 12mm).*

There are two major controls on calcrete formation: climate and sedimentation rate. Calcretes form in a variety of environments. However, if it is too arid only superficial deposits will form, and if it is too humid there will be excessive leaching of soil solubles (Reeves, 1970). Goudie (1973, and references therein) suggests that 500mm is the maximum level of precipitation which will allow the formation of calcretes, although he does cite examples forming in areas with rainfall as high as 1200mm per annum. A lower limit for rainfall is less clear cut, although a value of around 50mm seems reasonable (Goudie, 1973). At low precipitation rates gypcretes and evaporites will form instead of calcretes (Goudie, 1983).

Apart from precipitation rate, the other important climatic factor is seasonality (Arkley, in Reeves, 1970; Goudie, 1973; Semeniuck and Searle, 1985). Calcrete formation is greatest when there is a mild wet winter followed by a hot dry summer. In regions where the precipitation is concentrated in the hot season (*ie.* seasonally peaked) the rainfall can exceed the rate of evapo-transpiration and leaching of the carbonate will occur (Goudie, 1973).

The other important factor is sedimentation rate: the slower the sedimentation rate the greater the chance of a calcrete developing. Williams & Polach (1971) and Williams (1973) assembled radiocarbon datings on late Quaternary soils which show that episodes of pedogenesis of the order of 100,000 years were required to form calcrete profiles of stages 2/3 of Machette (1985) (Table 3.1) although this can vary with climate (Wright and Tucker, 1991). Periods of non-deposition of similar duration are likely to have been required during ORS times also. Rogers (1987) has suggested that the greater occurrence of calcretes in the Upper ORS is due to a period of tectonic quiescence and reduced rates of subsidence.

### 3.5.ii Eodiagenesis

Walker (1976) and Walker *et al.* (1978) have studied both recent and ancient red beds and have identified a series of early authigenic changes that occur in arid oxidising conditions. These are mechanical infiltration of clay, removal of framework grains of feldspar and ferromagnesian silicates, and precipitation of authigenic phases including quartz, haematite and calcite. These diagenetic alterations result in changes to texture (*eg.* creation of dissolution voids and infiltration of interstitial clays) and chemical composition (*eg.* removal by ground water of ions released by dissolution) of the sediment.

Mechanical clay infiltration and dissolution of unstable grains has occurred in the ORS and the result of this can be seen in red, mixed iron oxide/clay rims on detrital grains that are clearly observed in thin section (Fig. 3.15). The fact that these rims are present at grain contacts is a clear indication of their early timing. Formation of clay rims continued after early compaction as shown by their development as pore linings.

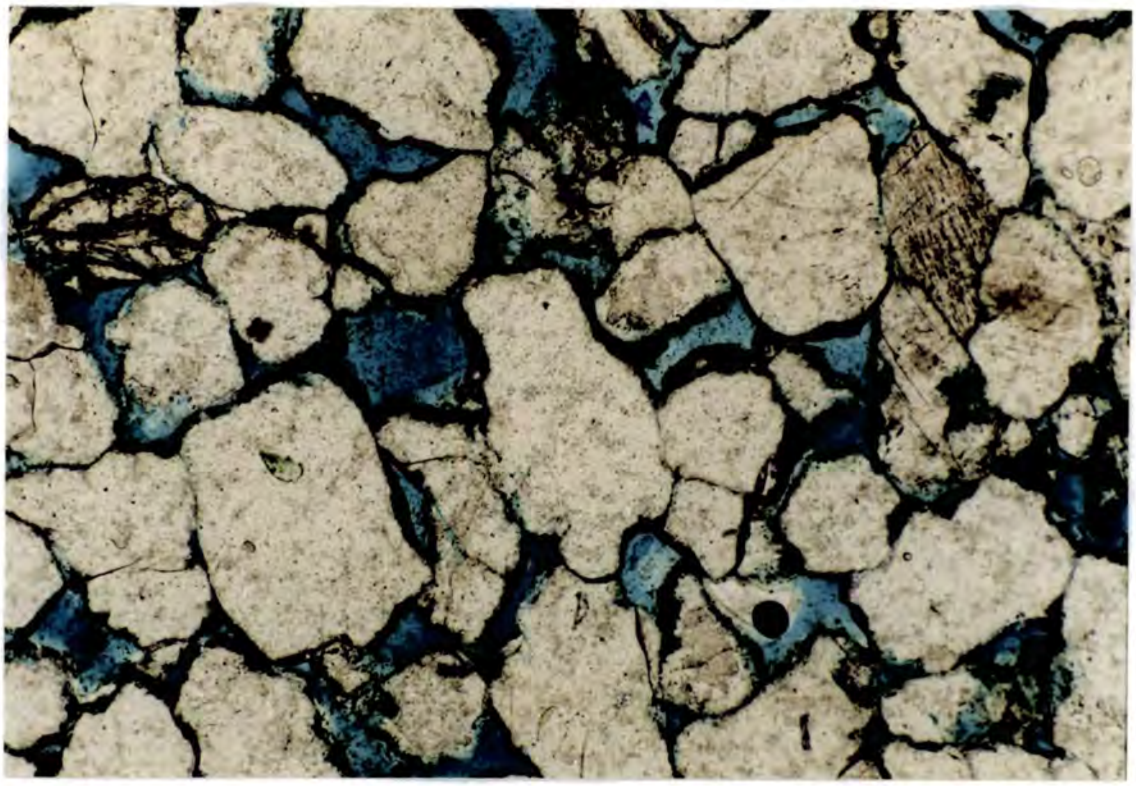


***Fig. 3.15.*** *Dark brown mixed clay/Fe-oxide rims around detrital grains. (PPL, FOV = 1.3mm).*

Grains most susceptible to early dissolution are the unstable grains such as the heavy minerals (hence their paucity in the rocks now), some lithic fragments and K feldspars. Early dissolution released many cations into solution, principally K, Al, Si, Mg and Fe (Walker *et al.*, 1978; Turner, 1980). The Fe would have been initially precipitated as a precursor mineral *ie.* amorphous FeOOH which later aged to haematite (Turner, 1980). This relationship is indicative of arid oxidising conditions.

Both biotite and muscovite micas show signs of alteration. Biotite is more susceptible to weathering than muscovite due to differences in crystal structure (Bassett, 1960; Deer *et al.,* 1966; Weaver and Pollard, 1973). Theoretically biotite should weather to vermiculite under acid conditions and to vermiculite plus montmorillonite under neutral and alkaline conditions. However, this transformation is only rarely reported from red bed sequences (Turner, 1980). Walker (1949), Seddoh and Pedro (1974) and Burley (1984) have all noted that biotite can oxidise almost totally with little or no vermiculite forming despite substantial potassium loss.

Farmer *et al.* (1971) proposed three alternative mechanisms for the oxidation of biotite not requiring the formation of vermiculite: a) by loss of interlayer cations, b) by loss of hydroxyl protons, and c) by the loss of octahedral iron. Turner (1980) summarises the evidence for and against all three methods and concludes that the third is most likely. The loss of octahedral iron occurs as follows:

$$(Si_3AlO_{10})Fe_3^{2+}(OH)_2K + {}^{1}\!/\!{}_{4}O_2 + {}^{1}\!/\!{}_{2}H_2O \text{ -----> } (Si_3AlO_{10})Fe_2^{3+}(OH)_2K + FeOOH$$

A problem with the above relationship is that oxidised biotites do not necessarily have lower iron contents (Rimsaite, 1970) and that interlayer iron oxides have been shown to exist (Farmer *et al.*, 1971).

A possible solution to this problem is the development of an intermediate step whereby iron is expelled through holes in the silicate sheet to an interlayer space. This interlayer oxide is probably amorphous iron hydroxide or a crystalline phase of ß-FeOOH (Turner, 1980). There are two possible destinations for this interlayer iron oxide. Either it can be expelled to form haloes around the biotite (Walker, 1967a, 1976; Walker *et al.,* 1978; McBride, 1974) or it can remain in the interlayer space and convert to haematite and replace the original biotite (Turner and Archer, 1977). It is the interstitial conditions that dictate what actually happens to the iron oxide: alkaline conditions will result in retention of the iron and acidic conditions will lead to expulsion of the iron (Turner, 1980).

Biotite micas observed in ORS samples from this study are undergoing psuedomorphous replacement by Fe oxide. This is visible in thin section as dark brown opaque sheets of haematite parallel to the basal cleavage and results in partial to total


***Fig. 3.16.** Biotite mica undergoing psuedomorphousreplacement by haematite (H). See text for discussion. (PPL, FOV=0.66mm).*

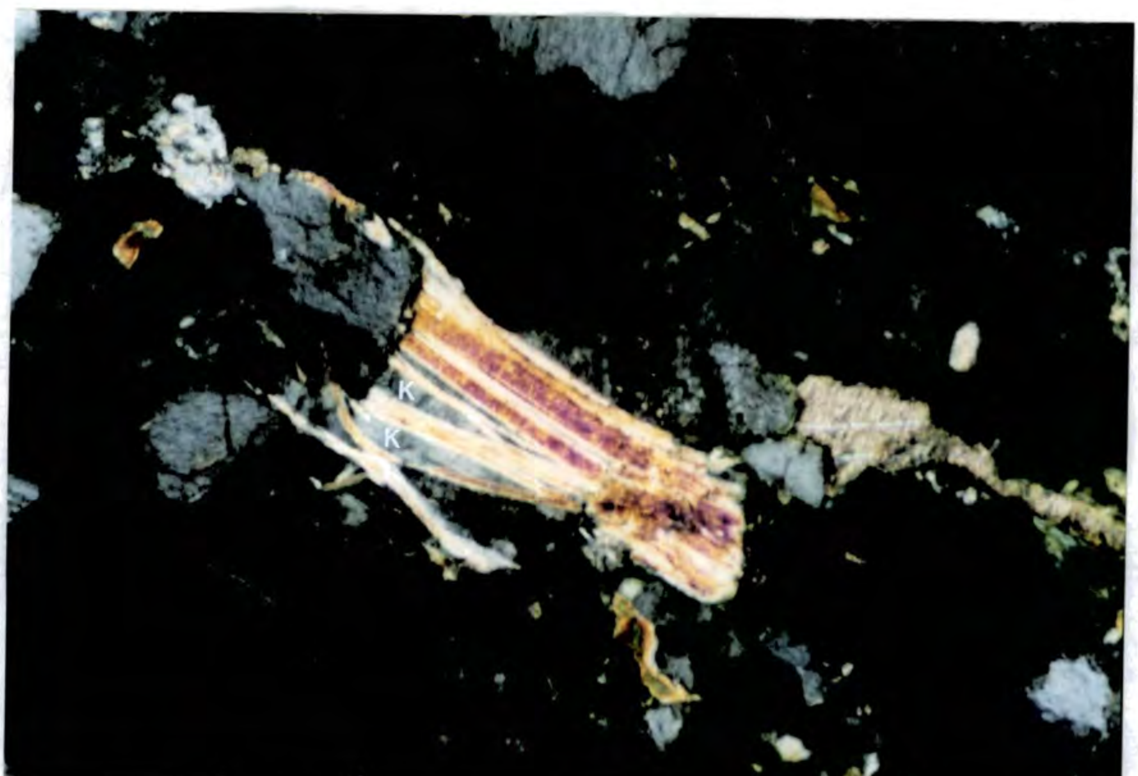



***Fig. 3.17.** Muscovite mica undergoing alteration to kaolinite (K) along cleavage planes. (PPL; FOV=0.66mm).*

loss of pleochroism depending on the extent of the replacement (Fig. 3.16). This therefore indicates alkaline conditions, consistent with the general climatic setting and oxidising, alkaline nature of the early pore solutions.

In thin section muscovites are seen to be severely splayed, parallel to the basal cleavage, towards the ends before passing into kaolinite (Fig. 3.17). This is a common feature (Blanche and Whitaker, 1978; Bjørlykke *et al.,* 1979; Burley, 1984) and occurs as a result of cation loss, principally K, according to the following equation (Burley, 1984):

$$\underset{\text{Muscovite}}{K_2Al_4[Si_6Al_2O_{20}](OH,F)_4} \xrightarrow[K_2\text{ loss}]{} \underset{\text{Kaolin}}{Al_4[Si_4O_{10}](OH)_8}$$

### 3.5.iii Compaction

Taylor (1950) and Moore (1975) both examined variations in grain packing, that is types and numbers of grain contacts, during studies into burial compaction. In an uncemented sand, initial grain rearrangement will cause an increase in point and planar contacts. This will continue until the intergranular volume is reduced to 26%, which represents rhombohedral packing, the tightest possible by grain rearrangement alone (Graton and Fraser, 1935; Füchtbauer, 1967). As compaction proceeds further, grain interpenetration will commence. This will result in the formation of concavo-convex and sutured contacts. Bell (1978) claims that in a well sorted sand, three or more contacts per grain suggests that the porosity has been reduced by burial compaction.

In cemented sandstones, however, the story is different. Cements such as quartz and calcite, if present in sufficient quantity, are strong enough to retard compaction. Precise amounts required for this are uncertain (McBride *et al.,* 1990), although Mack (1984) suggests that 5% by volume of quartz cement could be sufficient (see also Chapter 4). If a cement was introduced in large volumes early on before significant compaction had occurred, it could result in an abnormally high amount of point and plānar contacts due to its inhibiting effect on grain interpenetration.

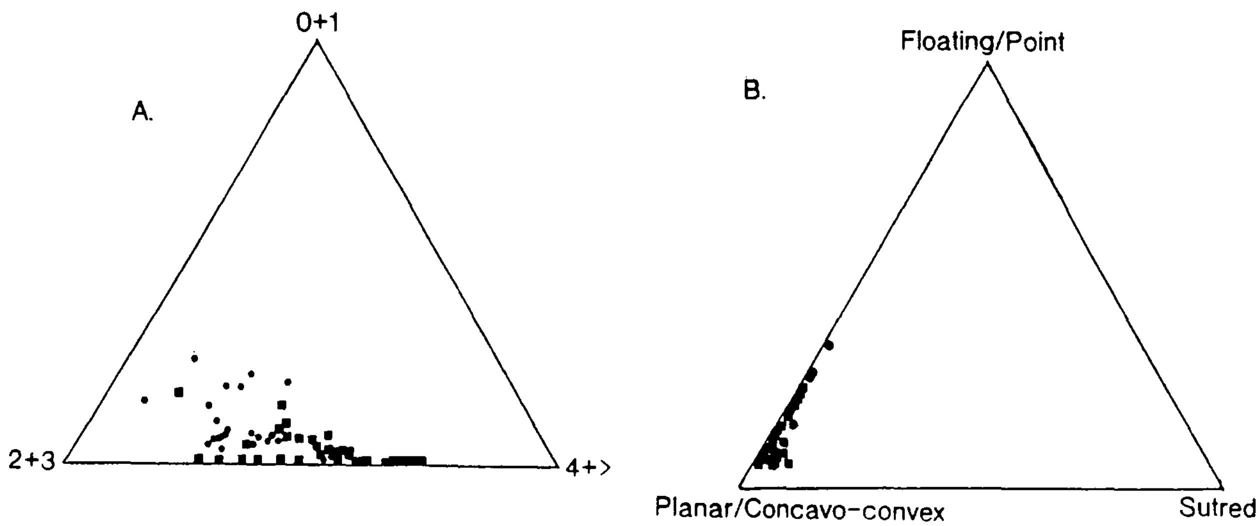



*Fig. 3.18. Triangular diagrams showing: a) the number of contacts per grain, and b) the type of contacts present. Circles = Upper ORS; Squares = Middle ORS.*

Once the data pertaining to compaction for the ORS samples had been obtained (see Appendix 2), it was then plotted on ternary diagrams. Planar contacts dominate with lesser amounts of concavo-convex and point contacts, while sutured and floating contacts are negligible (Fig. 3.18b). The number of grain contacts falls between the 2+3 and the 4 and above poles on Fig. 3.18a, with the average number of contacts being just over 3 (3.04 for the Upper ORS and 3.43 for the Middle ORS). These relationships suggest that while a moderate amount of compaction has occurred, only grain rearrangement has occurred in significant quantities, with minimal grain interpenetration. This is probably due to the early precipitation of a calcite cement (see also Chapter 4).

### 3.5.iv Mesodiagenesis

In thin section silica overgrowths are present in trace amounts (<1%) only, predominantly in the more deeply buried Middle ORS. These overgrowths, however, can be more clearly seen under the SEM, where they are only a few tens of microns thick on average (Fig. 3.19). Although overgrowths are a common early diagenetic feature (Franks and Forester, 1984), strongly alkaline pore fluids (see below) inhibit precipitation of silica because its solubility decreases with increasing pH (Blatt *et al.,* 1980).

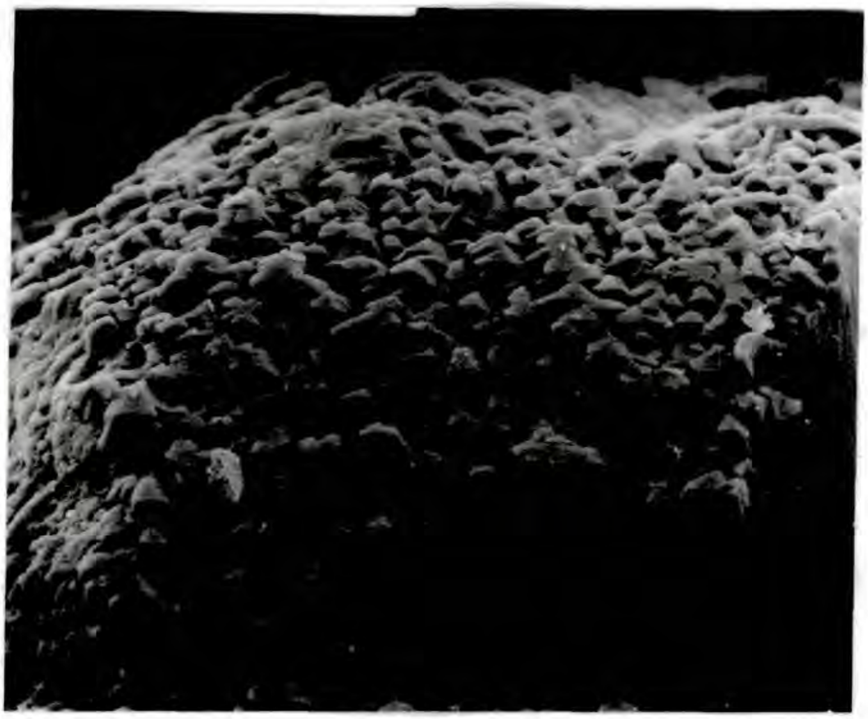


***Fig. 3.19.*** *Scanning electron micrograph of thin silica overgrowth. (Scale bar = 20μ m).*

Another possible reason for the lack of overgrowths is the presence of small amounts of the clay mineral chlorite (Heald, 1965; Cecil and Heald, 1971), identifiable under the SEM (Fig. 3.20) and by XRD (Fig. 3.1) as result of the increase in intensity of the 14Å peak on heating (Carroll, 1970). Wilson (1971) found that Middle ORS sandstones from around Caithness contained chlorite derived from the surrounding metamorphic terranes. However, the delicate rosette morphology of the chlorite in

Fig. 3.20 implies an authigenic origin as it is unlikely to have survived transportation in this form. Also the clay/Fe oxide rims formed during eodiagenesis are very likely precursors of chlorite (Dixon *et al.,* 1989). The chlorite, is therefore likely to be a mixture of detrital and authigenic grains.

***Fig. 3.20.*** *Scanning electron micrograph of an authigenic chlorite rosette. (Scale bar = 5μ m).*

The major cementation event to occur was the precipitation of a blocky, poikilotopic calcite cement (Fig. 3.21). This is patchy and constitutes between 0 and 26% of the rock volume. That this occurred early on is shown by the low levels of compaction (see above and Chapter 4). Apart from supporting the framework grains the calcite has also corroded them along their margins and occasionally replaces them either partially or totally. Under cathodoluminesence the calcite appears a uniform dull red colour with no evidence of zonation (Fig. 3.22). This lack of zonation shows that uniform conditions of pore water geochemistry and growth rate lasted for the duration of the cementation event (Ten Have and Heijnen, 1985).

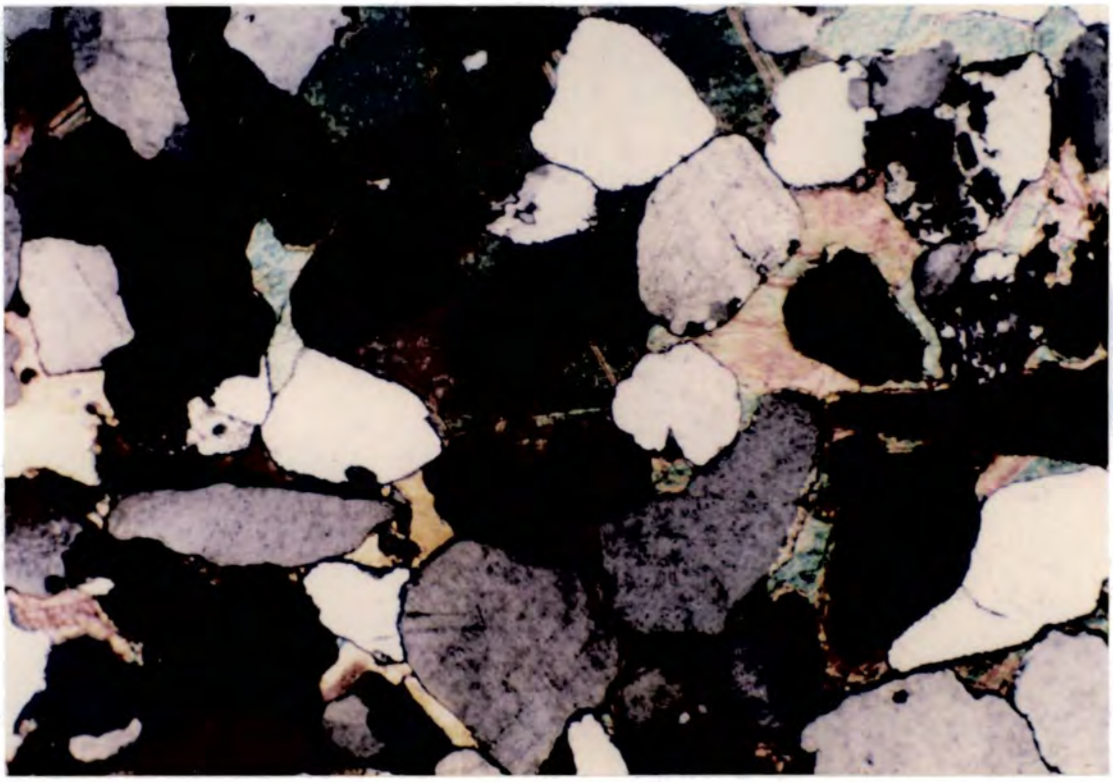


***Fig. 3.21.** Typical example of poikilotopic calcite cement occluding much of the porosity. (PPL; FOV = 1.3mm).*

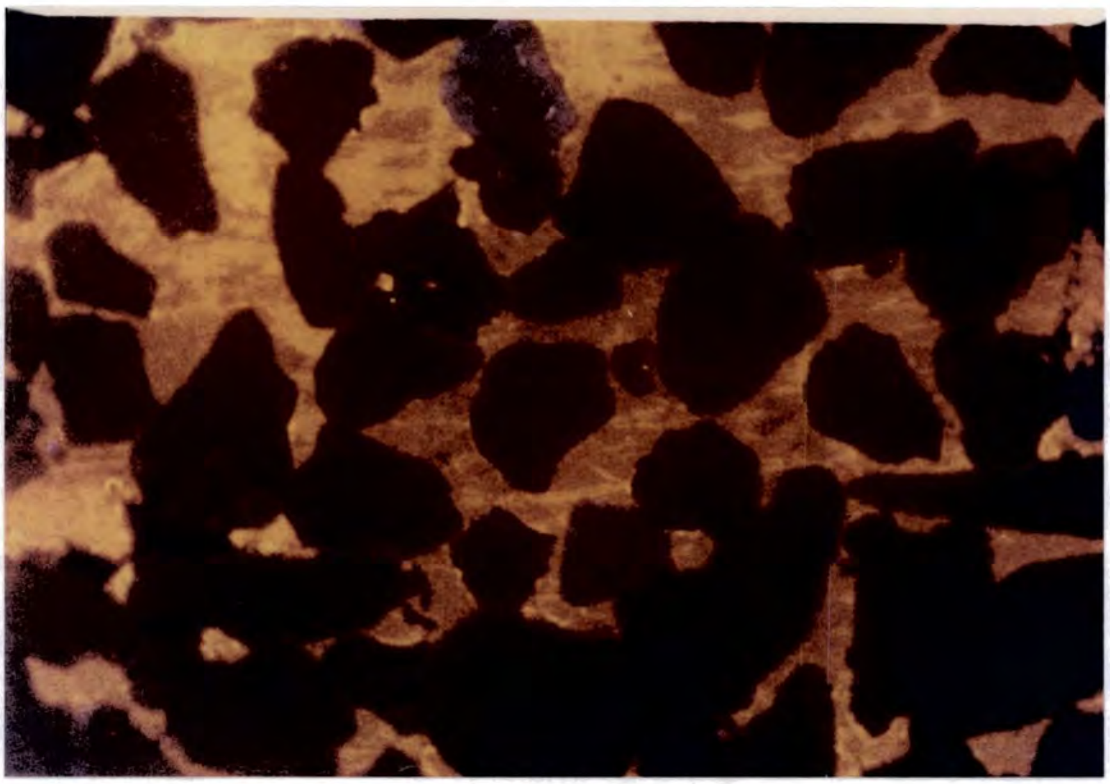


***Fig. 3.22.** Same view as above under luminesence. Note uniform luminesence of the calcite cement. See text for details. (FOV = 1.33mm).*

Two of the major controls on calcite solubility are temperature and pH. Solubility decreases with an increase in either temperature or pH but is more sensitive to the latter (Blatt, 1979). Calcite precipitates at a pH of 7 and above, but the fact that quartz grains have undergone dissolution and replacement along their margins suggests that the pH might have been 9 or above (Fig 3.23). Another requirement is a low $P_{CO2}$, which is linked to the pH, and a low Mg/Ca ratio. This ratio needs to be in the range of about 1:1 to 1:10 for the precipitation of calcite, rather than dolomite, to occur. The blocky anhedral morphology of the calcite described above requires a ratio below 1:2 (Folk and Land, 1975). The actual cut off will vary according to factors such as salinity and temperature.

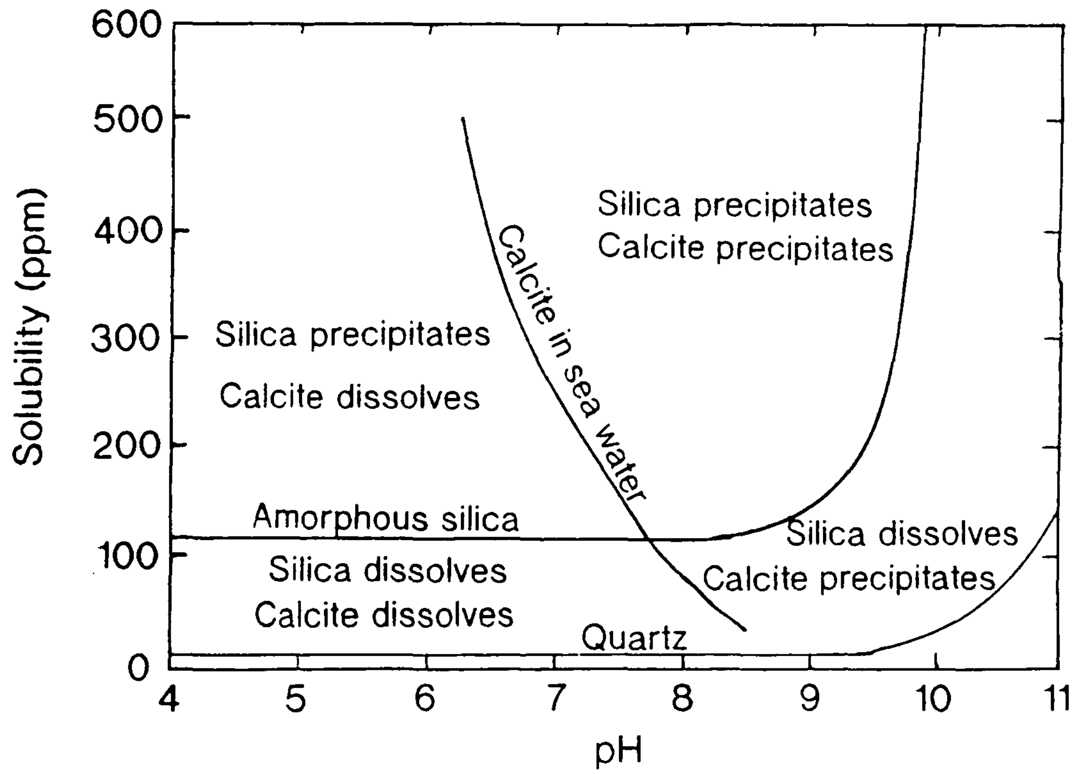



*Fig. 3.23. Relationship between pH and the solubilities for calcite, quartz, and amorphous silica. (Redrawn from Blatt et al., 1980)*

Biotite micas can be seen that have been split by the growth of calcite (Fig. 3.24). The micas may have been slightly split initially by early compaction which created new mineral faces. These faces then absorb hydrogen creating strongly alkaline conditions around the mica which encourages calcite precipitation (Boles, 1984), thereby further displacing the mica.


*Fig. 3.24.* *Biotite mica split by the growth of calcite (C) along cleavage planes. See text for discussion. (XPL; FOV = 1.3mm).*

The source of the calcite is uncertain. Franks and Forester (1984) suggested two end member sources :

> "...early (shallow) marine carbonate (including recycled shell material) and late (deep) carbonate from organic reactions in shales."

There is no marine influence in the ORS except perhaps in the Givetian (late Upper ORS) (Marshall *et al.*, 1990). However, the Orcadian permanent lake deposits are limestones and carbonate/clastic/organic laminites. This shows that the lacustrine waters were carbonate saturated and if they were buried along with the sediment and expelled into the adjacent sandstones they would form a likely carbonate cement source.

Most of the limestones have undergone neomorphism (from micrite to microspar) and some show stylolites (Fig. 3.25), both of which could act as another source for calcite cement. There is no petrographic evidence to support the presence of a late carbonate cement such as that suggested by Franks and Forester (1984) from reactions in the shales.

***Fig. 3.25.*** *Corse grained calcite spar formed following neomorpism of a micritic lacustrine limestone. (XPL; FOV=3mm).*

Pyrite is present in trace amounts, locally reaching an abundance of 1-2% (Fig. 3.26). It is present as anhedral masses or as a pore lining, and is replacive of the calcite. The presence of pyrite indicates that the pore waters became locally reducing in nature with increasing burial.

## 3.5.v Telodiagenesis

Many framework grains have corroded margins and pores often have irregular outlines. Also there are dispersed patches of calcite that have uniform extinction and corroded margins (Fig. 3.27). All these were considered by Schmidt and McDonald

(1979b) to be petrographic evidence of dissolution processes. Burley and Kantorowicz (1986) and Maliva and Siever (1990) have recognised that sand grains and their overgrowths often are covered in pits (1 - 2$\mu$m), notches (<20$\mu$m), and embayments (>20$\mu$m) which formed as a result of silicate dissolution during precipitation of carbonate cements (calcite, dolomite, and siderite). Therefore, if these features can be recognised in a porous sandstone they can be used as indicators of the previous presence of a carbonate cement. However, when the ORS samples were studied under the SEM these features were not readily visible. This is probably due to the fact that the cement is poikilotopic with a crystal size that is often equal to or greater than the grain size which causes the corrosion textures to be less distinctive if present at all (Burley and Kantorowicz, 1986).

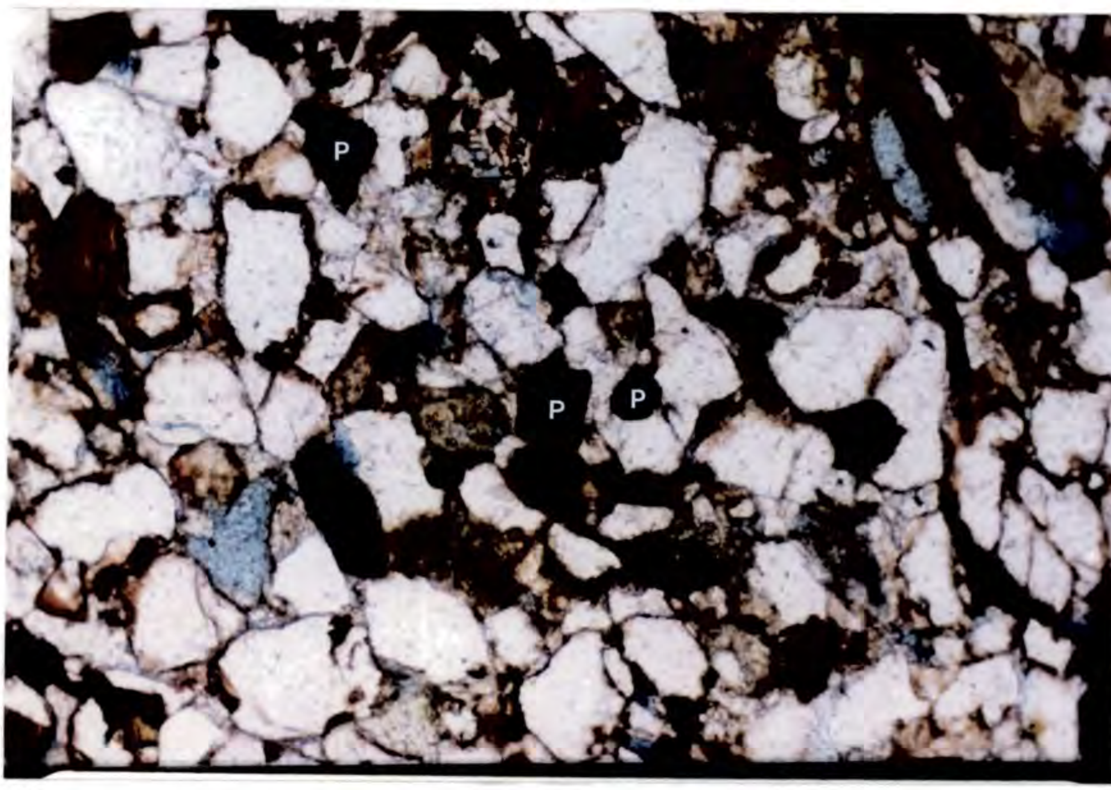



***Fig. 3.26.*** *Diagenetic pyrite (P) in a sandstone. (PPL; FOV = 3mm).*

Using the criteria outlined in the discussion on the precipitation of calcite, one of the requirements for dissolution to proceed is a low pH of less than 7 (Blatt *et al.*, 1980). The other requirement is a large volume of acidic fluid and sufficient porosity to allow the fluid to circulate. The question is, what causes the increase in acidity.

There are three main causes: a) meteoric waters containing atmospheric $CO_2$; b) formation waters containing organic acid anions (carboxylic) from the decarboxylation of organic matter (Surdam *et al.*, 1984); and c) formation waters containing $CO_2$ from decarboxylation of organic matter (Schmidt and McDonald, 1979b). In this study, meteoric waters are the most likely cause, as described below. The theory behind the latter two models is that during thermal maturation, kerogen will break down releasing $CO_2$ and carboxylic acid anions prior to oil generation. These then go into solution and migrate into the potential reservoir rocks.


***Fig. 3.27.*** *Dispersed calcite fragments (C) in optical continuity that were orig-inally part of one crystal: evidence for a period of calcite dissolution. (XPL; FOV = 1.3mm).*

Several authors have provided mass balance arguments against the method involving $CO_2$ from decarboxylation of organic matter (Bjørlykke, 1981 and 1984; Lundegard *et al.*, 1984; Giles and Marshall, 1986). Bjørlykke (1984) and Giles and Marshall (1986) also put forward the problem of the mineralogical content of the source rocks. If the source rocks were rich in carbonate or feldspar then the $CO_2$ would react with these before entering the reservoir rock. Using values for $CO_2$

generation from Tissot et al (1974) it is obvious that algal and mixed source rocks are incapable of producing sufficient $CO_2$ and that only humic source rocks might be able to do so (Giles and Marshall, 1986). The Middle ORS source rocks are lacustrine laminites that are composed of organic/carbonate/clastic triplets (Trewin,1986) with the organic content dominated by mixed and algal derived material (Marshall *et al.*, 1985) and as such decarboxylation of organic matter is untenable.

With regard to organic acid anions, Carothers and Kharaka (1978) showed that they can be present in oil-field waters throughout the temperature range 80 to 200°C, with their peak concentrations between 80 and 100°C. They also proposed that these anions control alkalinity. Surdam *et al.* (1989) describe the organic diagenetic reactions that lead to the formation of the carboxylic acid anions and their affect on mineral stability. Hillier and Marshall (1992) show that palaeotemperatures resulting from burial around the Moray Firth are in the region of 100-120°C, and therefore the rocks studied will have passed through the zone of peak concentrations of acid anions.

However, Lundegard *et al.* (1984) and Giles and Marshall (1986) have their doubts about this method also. In particular, Giles and Marshall (1986) again cite the problem of source rock mineralogy, as for the method involving the production of $CO_2$.

This leaves meteoric flushing as the most likely cause of calcite dissolution in the samples studied. The waters must have a high $CO_2$ content and be undersaturated with respect to calcite. Meteoric water is unlikely to be able to cause dissolution at depth because a) it will reach equilibrium with reactive minerals early on; and b) higher temperatures at depth will cause faster reaction rates therefore leading to faster attainment of water/rock equilibrium (Giles and Marshall, 1986). The problem of the large volumes of water required for this dissolution is overcome by the length of time that these rocks have been at the surface. Since inversion during the late Carboniferous the rocks have been at or near the surface until the present day. Astin

(1990) further suggested that the topography has not changed significantly since the Triassic. All this supports the idea that the dissolution event occurred after tectonic uplift.

Kaolinite is clearly visible in most samples and can also be identified with XRD by the disappearance of the 7Å peak on heating to 550°C (Carroll, 1970) (Fig 3.2). It occurs as a pore lining, pore filling, or occasionally as a grain replacing cement. Petrographic evidence for the timing of kaolinite formation includes fragments of calcite cement surrounded by kaolinite and the corroded nature of calcite in contact with kaolinite (Fig. 3.28). This evidence indicates that the kaolinite formed after calcite dissolution, and that it was probably formed as a result of the same acidic solutions that caused the dissolution event itself.

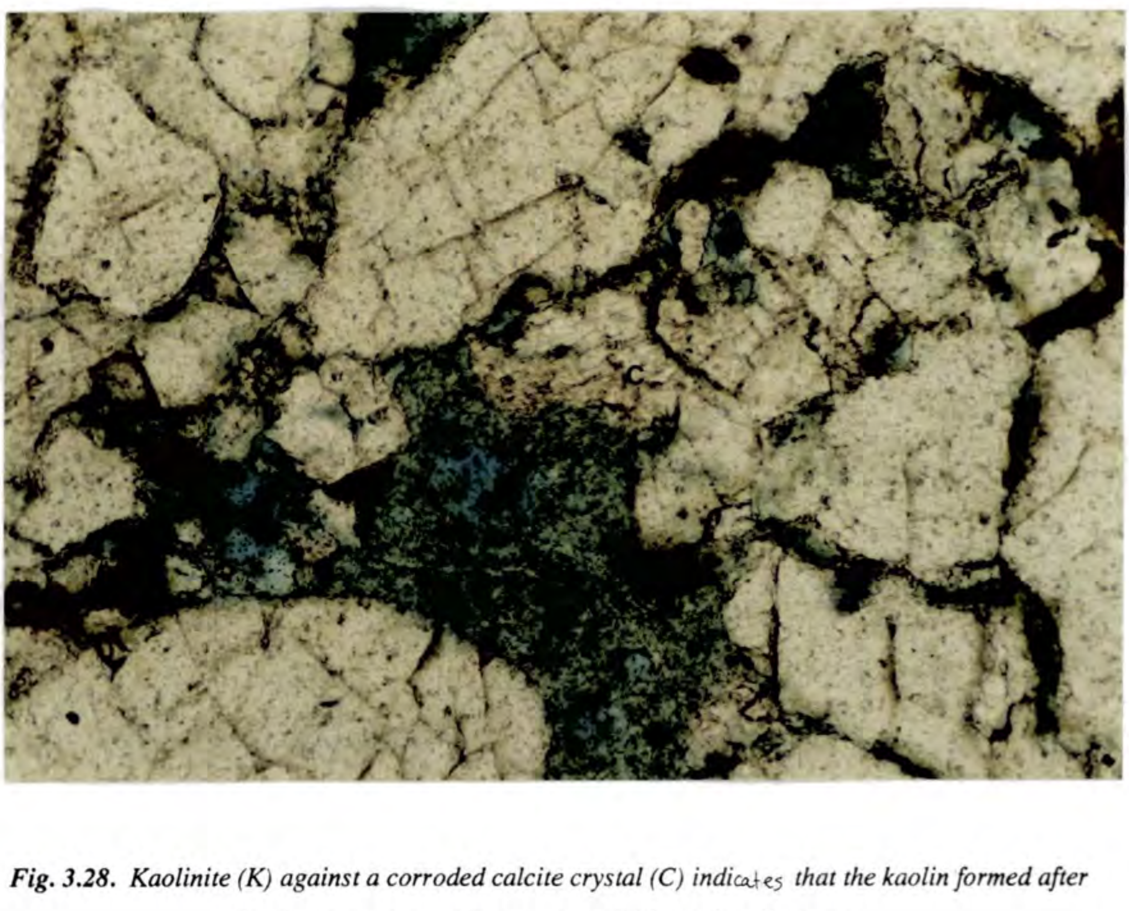



***Fig. 3.28.*** *Kaolinite (K) against a corroded calcite crystal (C) indicates that the kaolin formed after dissolution of the calcite occured. (PPL; FOV = 0.66mm).*

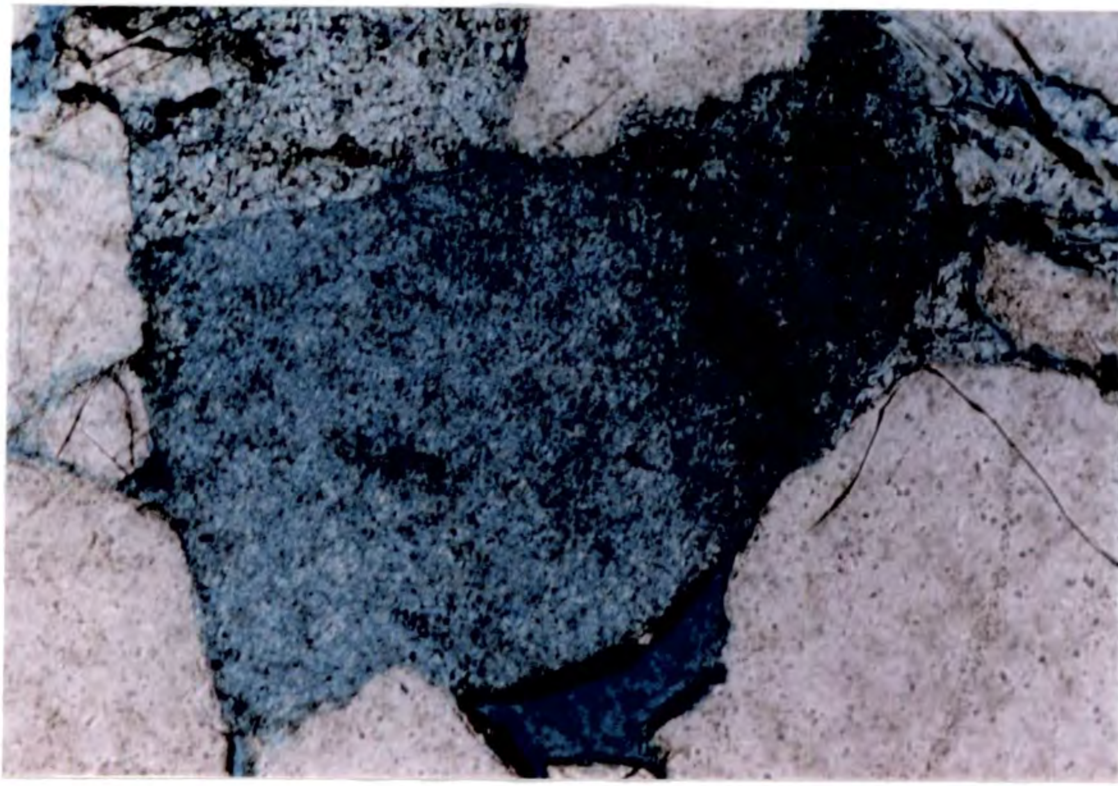


***Fig. 3.29.*** *Fine grain kaolinite replacing a detrital grain, probably feldspar. (PPL; FOV = 1.33mm).*

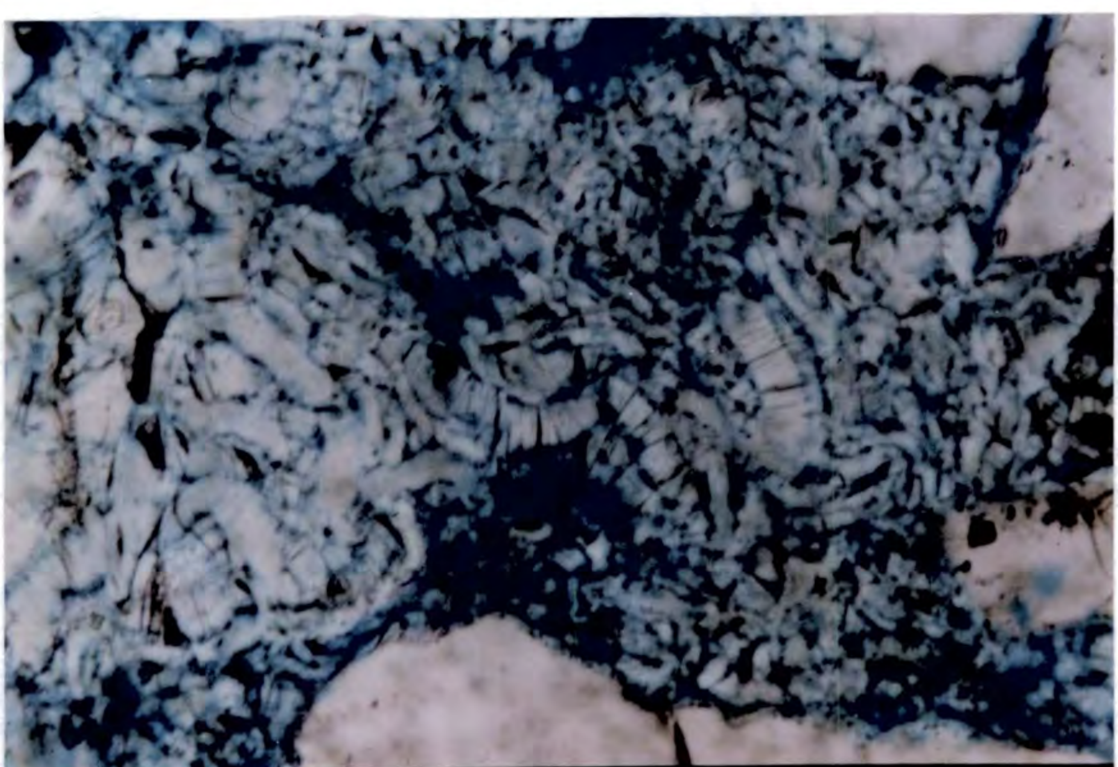


***Fig. 3.30.*** *Coarse grain pore filling kaolinite, showing vermiform structures. (PPL; FOV = 0.325mm).*

Petrographic observations also indicate that there have been two stages of kaolinite precipitation. The first is a fine grain, blocky kaolinite which is pore lining and grain replacive (probably of K feldspar) (Fig. 3.29). This was followed by a later, coarser, blocky to vermiform, pore filling kaolinite (Fig. 3.30). This morphology of euhedral, flat plates is typical of residual (authigenic) kaolin according to the textural classification of Keller (1978) (Fig. 3.31).

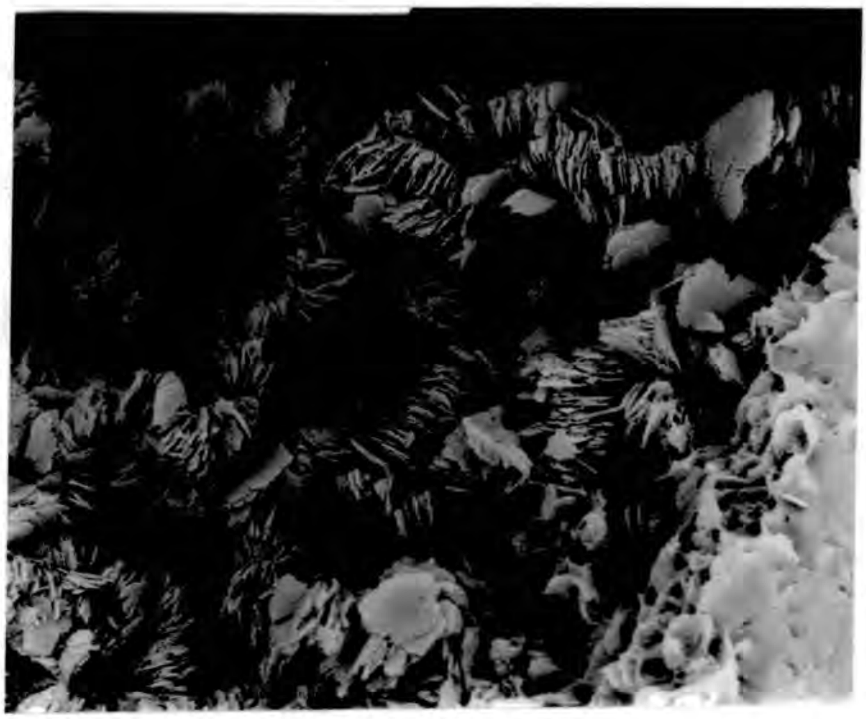


*Fig. 3.31. Scanning electron micrograph of kaolinite showing stacked sub to euhedral plates. (Scale bar = 20μ m).*

Franks and Forester (1984) state that kaolinite is commonly associated with dissolution events and give several examples, although in all cases the precipitation of kaolinite is claimed to have occurred at depth. Isotope data is given for one example which shows the temperature of formation to be 70 to 100°C. They give mean pH ranges for kaolinite forming reactions of between 5.3 and 6.0 (calculated at approximately 10,000' burial and 100°C - not the conditions under which the ORS samples studied formed). However, Hem (1985) states that the pore water composition of many modern meteoric aquifers falls within the stability field of kaolinite which supports the idea of meteoric flushing causing the dissolution of the calcite and the precipitation of the kaolinite.

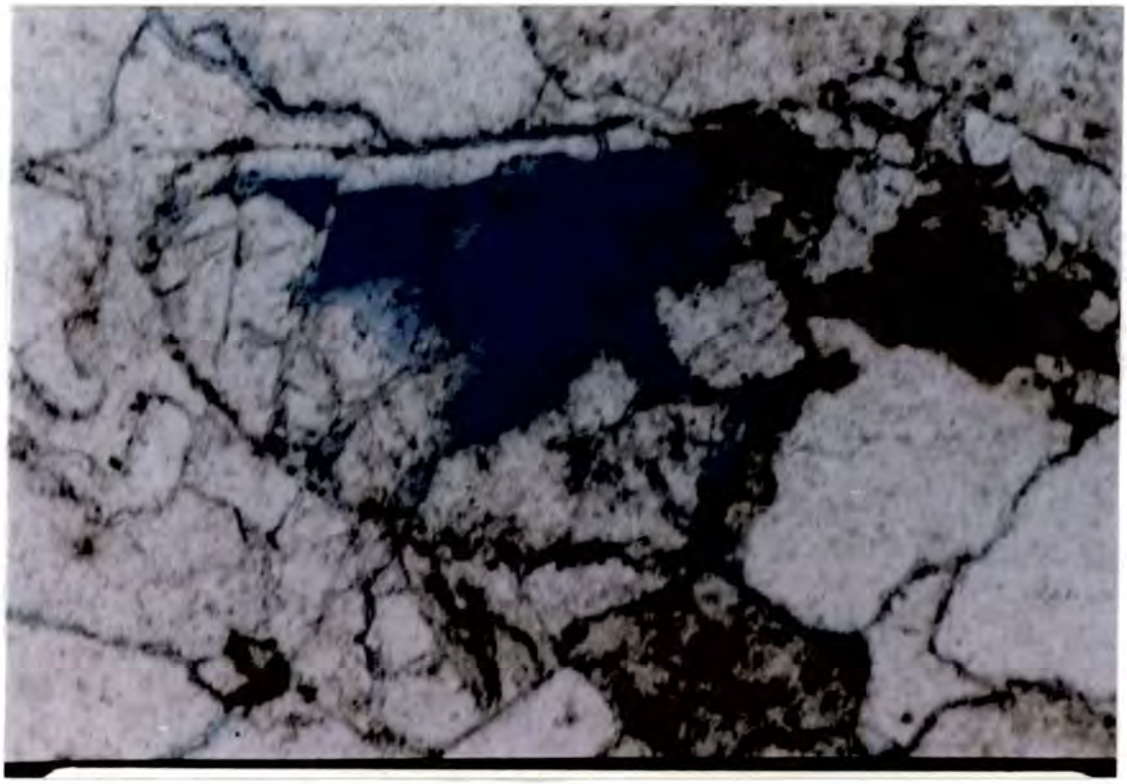


***Fig. 3.32.*** *Feldspar dissolution is the most likely source of Alumuinium for the formation of kaolinite. (PPL; FOV = 1.33mm).*

A source of Al is required for the formation of kaolinite. Also low Log$[Na^+/H^+]^2$ and Log$[K^+/H^+]^2$ ratios (<13 and<20 respectively) are required (Blatt, 1979). In river waters these ratios average about 8 and 5 respectively. Given the presence of skeletal feldspars (Fig. 3.32) it would seem reasonable to suggest feldspar leaching as a source of Al and silica. This would also release Na and K into the pore waters, although meteoric waters could remove these elements (Bjørlykke, 1984). Although kaolinite rarely forms within leached feldspars it is always present near to them, suggesting preservation of Al on a local scale, it simply having transferred from the feldspar to the kaolinite.

Surdam *et al.* (1989) state that the pH of a fluid phase is the primary control of the distance of mass transfer. Therefore if acidic waters leached a feldspar and then attacked nearby calcite cement, the pH would start to rise as the calcite was dissolved thus allowing the precipitation of kaolinite. This would explain the commonly observed occurrence of kaolinite adjacent to corroded calcite.

## 3.6 Summary

During burial, first compaction and then cementation have affected the sediments. Early diagenetic changes are similar to those of modern arid environments followed by the precipitation of a pore filling calcite cement which inhibited further compaction during mesodiagenesis. The calcite has also corroded and replaced detrital grains. Following inversion, meteoric waters flushed the rocks causing dissolution of the calcite and unstable grains such as feldspars which led to the precipitation of kaolinite. This diagenetic sequence of events is summarised in Fig. 3.33.

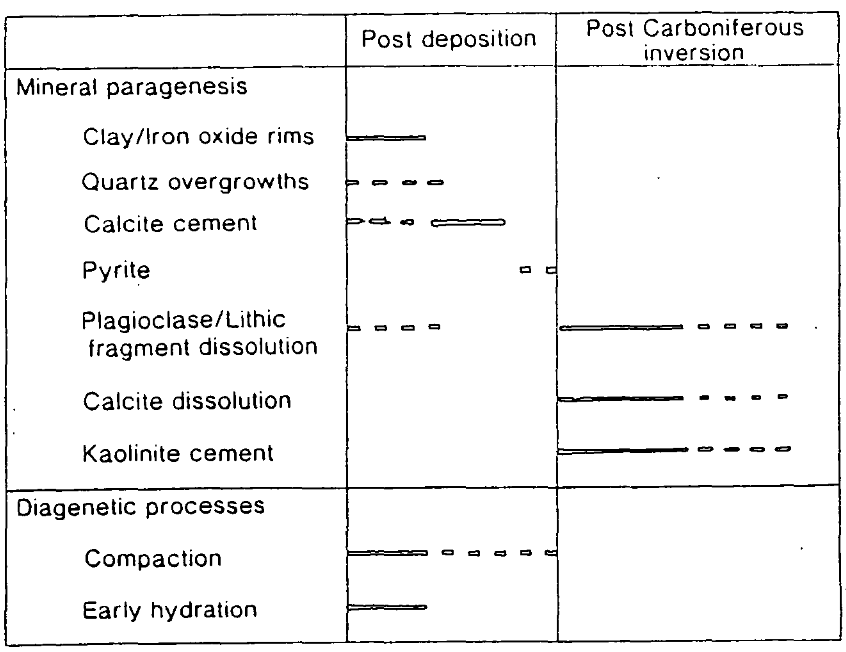



*Fig. 3.33. Summary of the diagenetic sequence of the Old Red Sandstone around the Moray Firth.*

With regard to the timing of the formation/destruction of porosity, early compaction resulted in porosity losses that would have outweighed any gains due to dissolution during eodiagenesis. The major period of porosity loss however, was during mesodiagenesis with the precipitation of the calcite cement which occluded a large proportion of the porosity. It was only after uplift and telodiagenetic changes that there was an increase in porosity following dissolution of the calcite and other unstable grains. The major period of hydrocarbon generation was probably during late mesodiagenesis after deposition of the Upper ORS. However, significant migration of the hydrocarbons would have been inhibited by the low porosity and permeability of the sandstones until the formation of secondary porosity and faulting during late Carboniferous and Permian times (Marshall, *et al.,* 1985; Parnell, 1985; Trewin, 1985).

This sequence of events closely resembles the situation in the Buchan field sandstones (Upper Devonian/?Lower Carboniferous) from the Outer Moray Firth except that these rocks have undergone deeper, more prolonged burial. As a result late stage calcite and dolomite cements formed with illite forming after kaolinite and smectite (Benzagouta, 1991). Unfortunately, Benzagouta (1991) has paid scant attention to the causes/timing of the calcite dissolution that he recognises.

Chapter 4

# Poroperm Characteristics

## 4.1 Introduction

For a sandstone to act as a suitable reservoir rock it requires sufficient porosity and permeability to permit the storage and passage of hydrocarbons at commercial rates. As a result the study of the controls on porosity and permeability has increased dramatically over the past two decades, due mainly to oil company interest in more efficient reservoir modelling and extraction. Some of the more important studies include Pittman (1979), Schmidt and McDonald (1979a,b), Shanmugan (1984) and Kaiser (1984). Schmidt and McDonald (1979a) and Shanmugan (1983) proposed porosity classifications based on the physical and genetic characteristics of pore geometry. Studies like these have recognised many factors which usually result in the partial or total destruction of porosity. For example depositional porosities can range from 17% to 56%, while permeabilities have values in the region of several darcies (Pryor, 1973). However in the ORS samples analysed in this study, porosities range from 0 to 14%, and permeabilities from 0 to 270md. This is due to changes to the pore network which occur during diagenesis. The following discussion considers the various controls on the porosity and permeability in these ORS samples.

## 4.2 Porosity

### 4.2.i Introduction

Pryor (1973) in his study of porosity and permeability characteristics of Holocene sandbodies found that point bar sands had porosities with a range of 17 to 52% and a mean of 41%; other environments had values ranging from 39 to 56% with means of around 49%. Other workers have reported porosity values in the range of 35 to 45% from a variety of modern environments (Fraser, 1935; Hamilton and Menard, 1956). Beard and Weyl (1973) made artificial grain packs and found porosities ranging from 27.9 to 42.4%. These porosities were dependant on the sorting of the samples as follows: very poorly sorted, 27.9%; poorly sorted 30.75; moderately sorted, 34%; well sorted, 39%; very well sorted, 40.8%; and extremely well sorted, 42.4%. Thus, depositional porosities can vary considerably, with an overall range of 17 to 56%.

Since the ORS samples are mainly well sorted (see Chapter 3) 40% has been chosen as an average value for the original depositional porosity of the rocks, although this is probably a conservative estimate. However, the porosities measured from the ORS samples range from 0 to 14.3% with a mean of only 6.2%, suggesting that a large proportion of the original primary porosity has been destroyed during burial and diagenesis.

### 4.2.ii Classification

Porosity can be classified in two ways, either genetically or texturally. A classification based on texture is more useful than a genetic one when considering

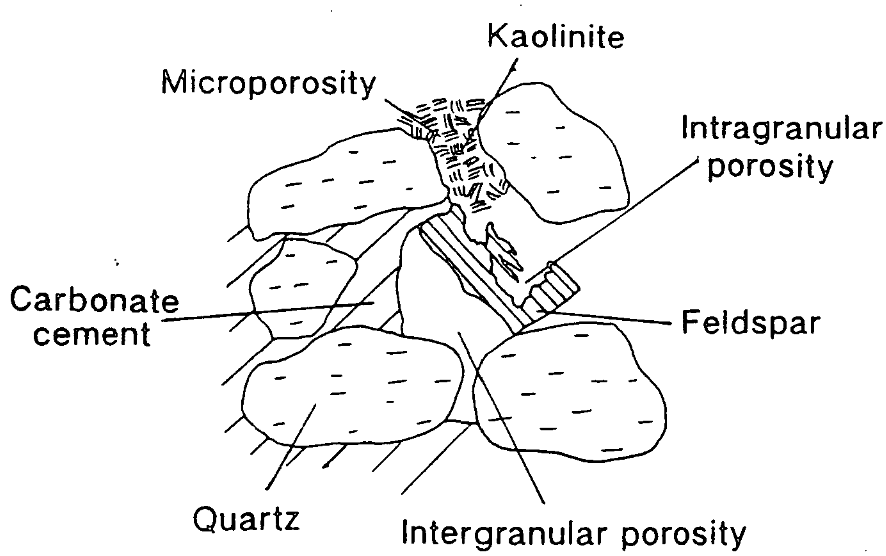



*Fig. 4.1. Diagram illustrating the textural classification of porosity.*

reservoir characteristics because it is the physical appearance of the pores (*ie.* micro- or macroporosity) and the resultant degree of connectivity that relates to permeability, and as such it is more important than how the pores formed. However, textural classes can be further described according to their genetic origin thereby providing insight into the relative timing of porosity formation. The textural classification has three subdivisions: 1) intergranular porosity; 2) intragranular porosity; and 3) microporosity (Fig. 4.1). Intergranular porosity is all the pore space between detrital grains and includes remnant primary (depositional) porosity, secondary porosity, produced by the dissolution of either grains or cements, and hybrid porosity which is a combination of both. Intragranular porosity is secondary porosity within grains formed by the partial dissolution of the grains. Microporosity includes any pores smaller than 0.5$\mu$m in size, which mainly includes pores within clay minerals and the smaller intragranular pores.

### 4.2.iii Porosity measurement

Samples were collected by hand from the field (with a hammer) and then brought back to the department where any obviously weathered and discoloured surfaces were removed with a rock cutter. The samples were then impregnated with a blue resin to facilitate the recognition of porosity, before finally being made into thin sections. Porosity values were then obtained by point counting and therefore only represent macroporosity (*ie.* inter- and intragranular porosity) and not microporosity. Some workers however, are of the opinion that the porosity of authigenic kaolinite can be estimated at *c.* 50% (Ehrenberg, 1990; A. Hurst and P. Nadeau in Bjϕrkum *et al.*, 1990). A more accurate method of assessing the porosities (*eg.* helium or water saturation) could not be used on samples collected by hand because these require the use of core plugs which could not be cut from the samples owing to their size.

Most of the porosity present within the ORS is intergranular, with lesser amounts of both intragranular and microporosity. The intergranular porosity is a mixture of remnant primary porosity, secondary porosity after dissolution of calcite cement and unstable grains such as lithic fragments and feldspars, and hybrid porosity. However, it is impossible to accurately distinguish between remnant primary porosity and that which formed after dissolution of cement, because of their often similar appearance (Fig. 4.2), hence the usefulness of a textural, rather than a genetic, classification. The intragranular porosity is mainly due to dissolution of feldspar (Fig. 4.3), although some lithic fragments are probably involved as well. The majority of the microporosity is associated with clay minerals as shown in Fig. 4.3.

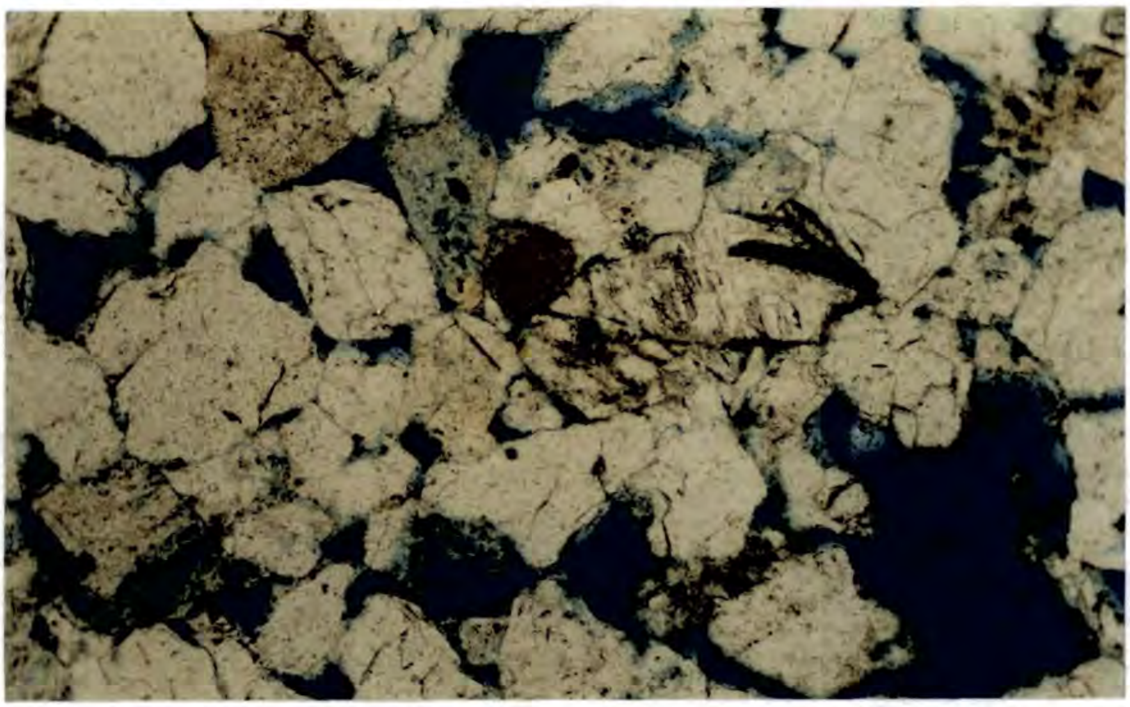


***Fig. 4.2.*** *Photomicrograph showing intergranular porosity (IR), highlighting the difficulty in distinguishing dissolution and primary porosity. (PPL, FOV = 1.3mm).*

## 4.2.iv Porosity evolution

Porosity in recent sandbodies is clearly related to grain size and sorting. In meandering river point bar sands, Pryor (1973) found that an increase in either grain size or sorting leads to an increase in porosity. However, in the ORS samples there is no correlation between porosity and grain size (Fig. 4.4) or sorting (Fig. 4.5). This is because diagenetic events in some cases totally overprint textural controls and can reduce/enhance porosity (Pittman, 1979; Brown *et al.,* 1989; Ehrenberg, 1990).

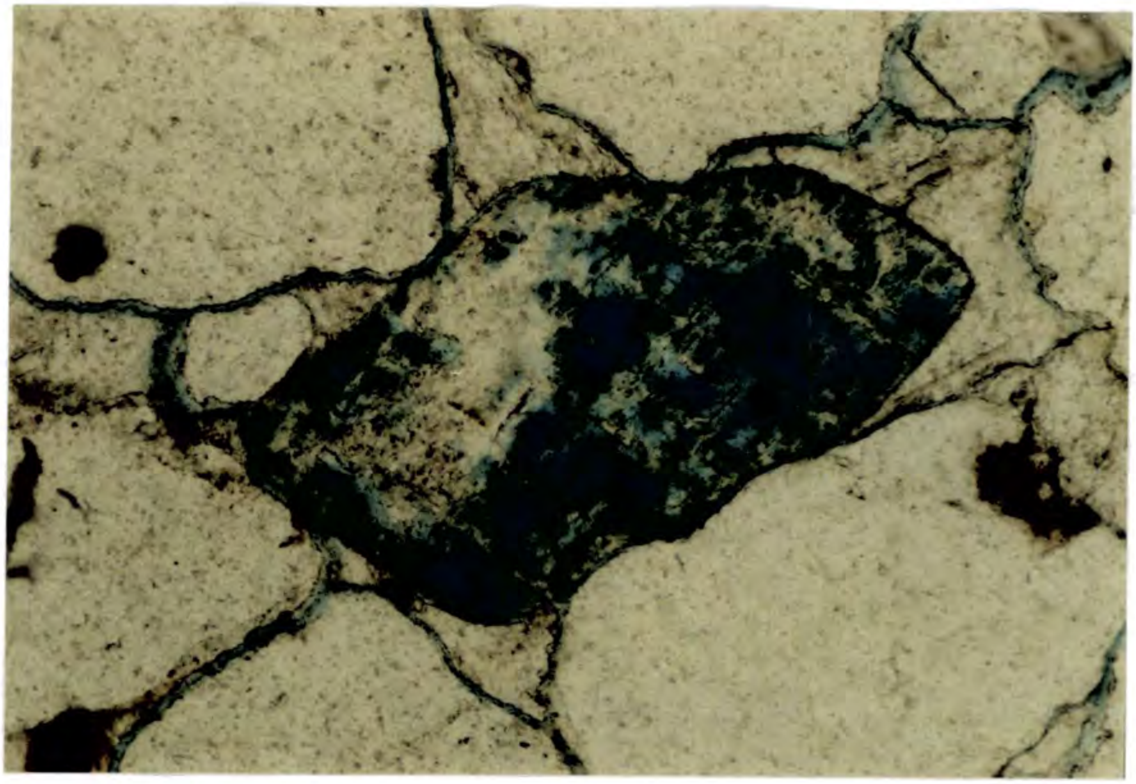


*Fig. 4.3. Photomicrograph showing intragranular porosity (IA) (here after feldspar) and microporosity (M) associated with kaolinite. (PPL, FOV = 0.66mm).*

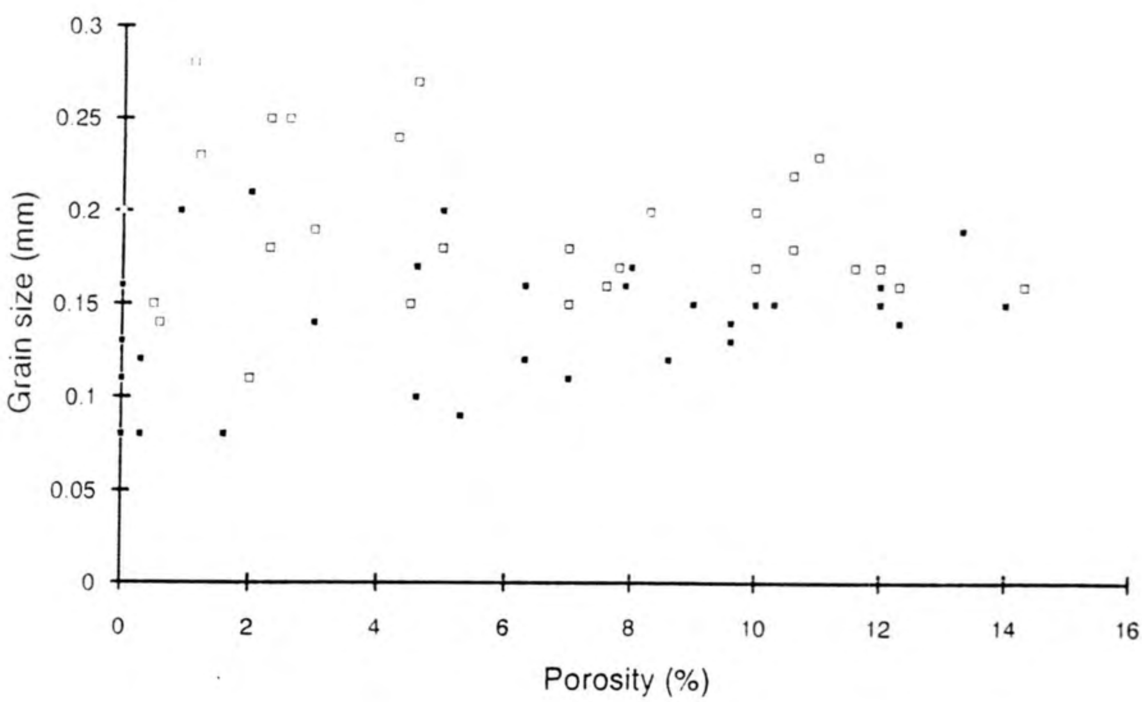



*Fig. 4.4. Plot of porosity against grain size. . . Key (for all following crossplots): Black squares = Middle ORS; white squares = Upper ORS.*

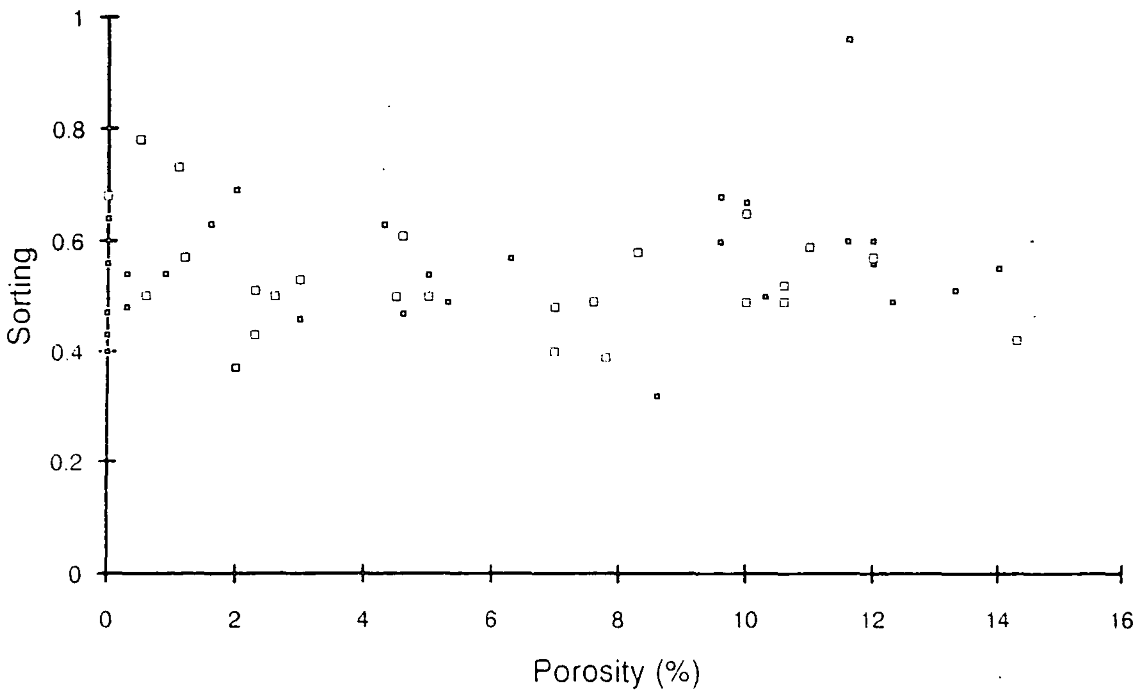



*Fig. 4.5. Plot of porosity against sorting.*

## A. Porosity reduction

The established view, ever since Sorby's pioneering work (1863, 1908), is that there are three processes which reduce porosity: a) mechanical compaction; b) chemical compaction; and c) cementation. The following definitions have been taken from Houseknecht (1987):

*Mechanical compaction:* This is bulk volume reduction resulting from processes other than grain dissolution and is characterised by the repacking and reorientation of competent grains, or local fracturing of competent grains and deformation of ductile grains. This leads to the reduction of intergranular volume and therefore porosity.

*Chemical compaction:* This is the bulk volume reduction caused by the dissolution of framework grains at points of contact (see Füchtbauer, 1967). This reduces the volume of framework grains, and the intergranular volume, and therefore porosity.

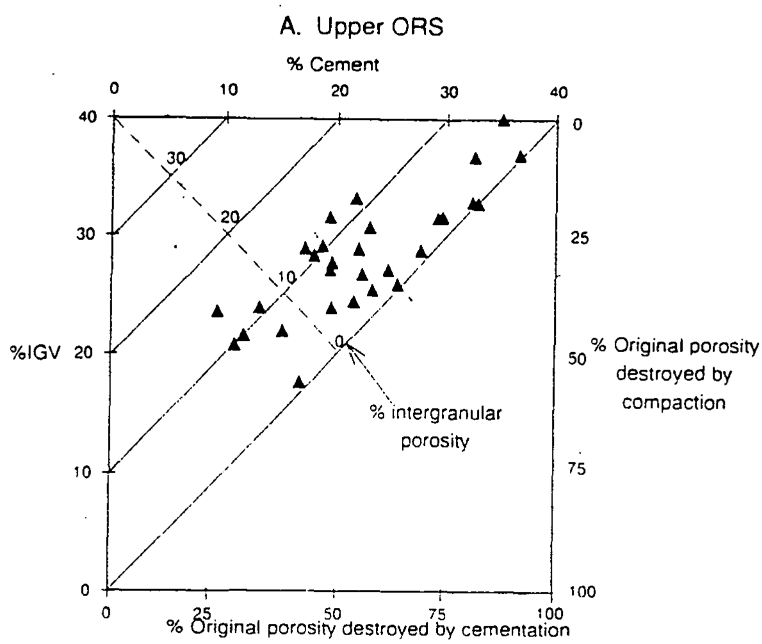
*Cementation:* This is the occlusion of intergranular volume (and therefore porosity) as a result of the precipitation of authigenic minerals without a reduction in bulk volume.

Intergranular volume (IGV) is defined as the sum of intergranular porosity plus all cements that occupy intergranular space (Houseknecht, 1987). It was apparently first used by Weller (1959), and is synonymous with minus-cement porosity (Rosenfeld, 1949; Heald 1956), and pre-cement porosity (Wilson and McBride, 1988).

However, Tada and Siever (1989) state that there are many discrepancies between theoretical, experimental and field evidence as regards chemical compaction. Stephenson *et al.* (1992) provide a model for sandstone compaction by grain interpenetration that is an entirely mechanical process, and as a result suggest that the term mechanical compaction be replaced with *grain rearrangement* and that chemical compaction be replaced with *grain interpenetration.* The terms of Stephenson *et al.* (1992) have been used hereafter because they describe the processes that are in operation, whether grain interpenetration is a chemical or a mechanical phenomena.

As a sediment is buried it will undergo grain rearrangement and then grain interpenetration, both of which serve to reduce the intergranular volume from its original 40%. The relative importance of these two processes is very difficult to assess but assuming spherical grains, the tightest packing possible by grain rearrangement alone is rhombohedral which has an intergranular porosity of 26%. Therefore any rock with an intergranular volume less than this value will have undergone grain interpenetration. Cementational porosity loss is more straight forward and can be quantified from point counting.

Houseknecht (1987) introduced a method to rapidly assess the relative importance of compaction and cementation as methods of destroying intergranular porosity from point count data. To do this a plot of intergranular volume (vertical



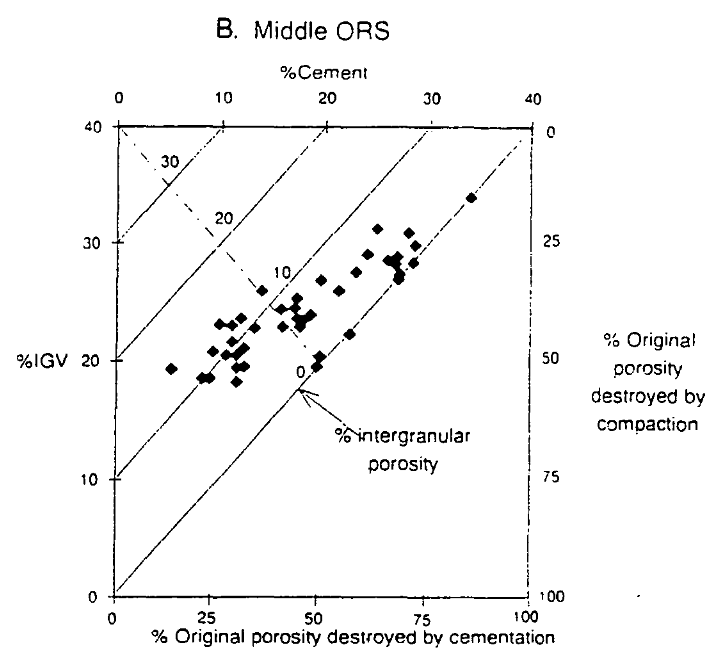




*Fig. 4.6. Plot of cement against intergranular volume. A) Upper ORS; B) Middle ORS. Dashed diagonal line separates those samples in which cementational porosity loss dominates (upper right) from those in which compactional porosity loss dominates. Note the dominance of cementational porosity loss.*

axis) against the percentage of cement present (horizontal axis) is constructed (Fig. 4.6). The vertical axis of Fig. 4.6 can also be used to quantify the percentage of original porosity destroyed by compaction according to the following equation:

$$\text{\% original porosity destroyed by compaction} = \frac{40 - \text{intergranular volume}}{40} \times 100 \qquad \text{(Eq.4.1)}$$

where intergranular volume is expressed as a percentage of whole rock and an original porosity of 40% has been assumed.

Equally, the horizontal axis of Fig. 4.6 can be used to ascertain the amount of original porosity destroyed by cementation using the following equation:

$$\text{\% original porosity destroyed by cementation} = \frac{\text{\% cement}}{40} \times 100 \qquad \text{(Eq.4.2)}$$

where the volume of cement is expressed as a percentage of whole rock.

Intergranular porosity is represented on this diagram by straight diagonal lines and can be estimated directly from the diagram or calculated from the following equation:

intergranular porosity =% intergranular volume-% cement. (Eq.4.3)

The dashed diagonal on Fig. 4.6 divides those samples in which cementation is more important (upper right) from those in which compaction has been more important in destroying intergranular porosity (lower left). This allows rapid assessment to be made of the importance of compaction and cementation.

When the data for the ORS samples are plotted, the majority of the points clearly fall in the upper right hand half of the diagram, which shows that overall cementation has destroyed more porosity than compaction. Average figures for the Upper and Middle ORS are:

| | % original porosity destroyed by cementation. | % original porosity destroyed by compaction. |
|---|---|---|
| Upper ORS | 54 | 30 |
| Middle ORS | 46.2 | 38.7 |

(The complete data set is in Appendix 3.)

The increase in importance of compaction in the Middle ORS, although only slight, reflects its greater burial depth. Average intergranular volumes for the Upper and Middle ORS are 28% and 24.5% respectively, and bearing in mind that the probable minimum as a result of grain rearrangement alone is 30%, this suggests that grain interpenetration has played only a minor role. As a result there has been virtually no grain volume reduction except by deformation of ductile grains. This in line with the findings of Wilson and McBride (1988) from their study of sparsely cemented sandstones from the Ventura Basin, California, where they showed that grain rearrangement has been the greatest cause of porosity loss (up to 53% of original porosity in some samples).

Both Ehrenberg (1989) and Pate (1989) have suggested modifications to this type of diagram that would have the overall effect of increasing the importance of compaction relative to cementation by altering the equation used to calculate the amount of porosity destroyed by compaction (Eq. 4.1). Pate's (1989) modification, however, involves filtering the data through a set of equations before plotting and thus, the data does not represent the physical characteristics of the rocks analysed (Houseknecht, 1987). Ehrenberg's (1989) technique is to be preferred because it keeps the original axes of the diagram allowing direct plotting of the petrographic data (Houseknecht, 1987). However, the main reason for using Houseknecht's (1987) original equations in this study, is because when these and Ehrenberg's (1989) equations were used to calculate the porosity of the samples, Houseknecht's (1987) equations were consistently more accurate when compared to the point counted values.

There are, however, limitations to this method of assessing porosity destruction. For example, if a sandstone contains abundant ductile grains or intragranular porosity it will not work. Despite the high proportion of lithic grains within the ORS (between 10 and 39%) they are mainly non-ductile so their affect is minimal. Likewise the intragranular porosity is only a minor fraction of the total porosity. As a result this method allows a rapid, objective assessment of porosity destruction.

When intergranular volume is plotted against compaction (represented by the compaction index) there is a clear trend of decreasing intergranular volume with increasing compaction (Fig. 4.7), as is to be expected. Also when cement is plotted against intergranular volume (Fig. 4.8) an increase in cement causes an increase in intergranular volume (*ie.* intergranular volume is preserved), again as is to be expected. This confirms that compaction results in porosity loss due to bulk volume reduction, while cementation preserves bulk volume. Plots of cement and compaction against porosity show a far more interesting relationship. A plot of total cement against porosity shows a clear trend of decreasing porosity with increasing cement (Fig. 4.9) as expected, whereas when compaction is plotted against porosity there is an increase in porosity with increasing compaction (Fig. 4.10). It is also worth noticing the inverse relationship between cement and compaction (Fig. 4.11). These latter relationships are due to the early timing of the cement halting compaction and occluding the majority of the porosity. However, where there was less cement, compaction could proceed to a greater extent, although nowhere does it occlude as much porosity as the cement. The fact that the higher porosities are associated with higher degrees of compaction and the lower amounts of cement adds further support to the dominance of cementation over compaction as a means of porosity destruction. Thus, there has only been a minimal reduction of bulk volume.

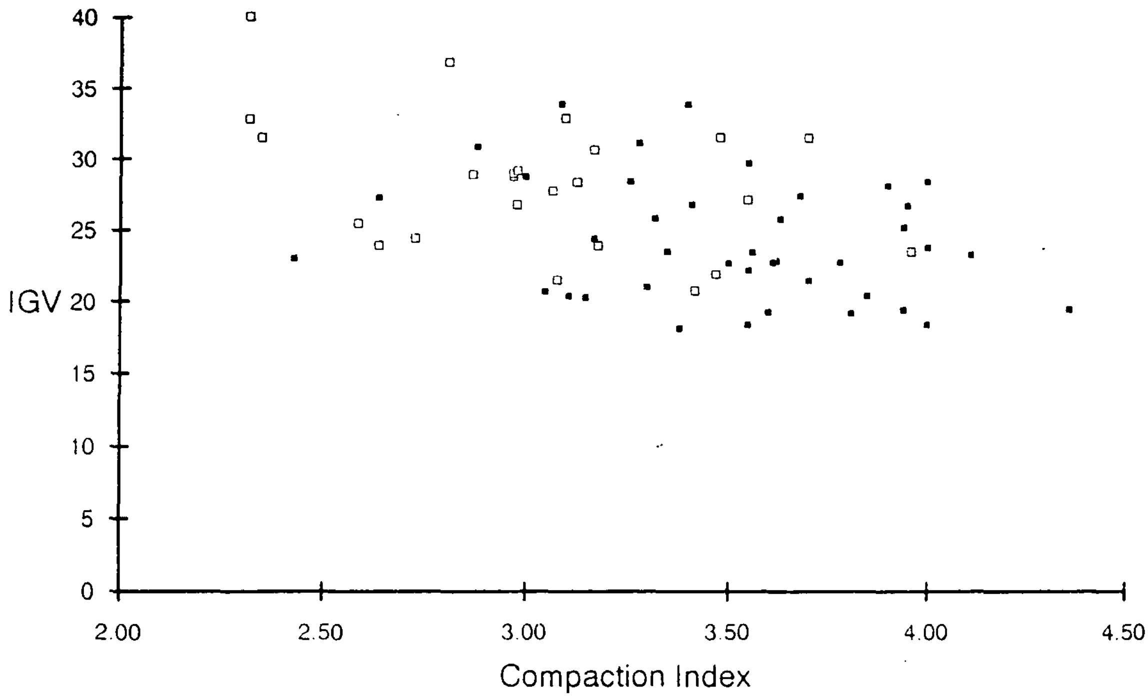



*Fig. 4.7. Plot of intergranular volume against compaction.*

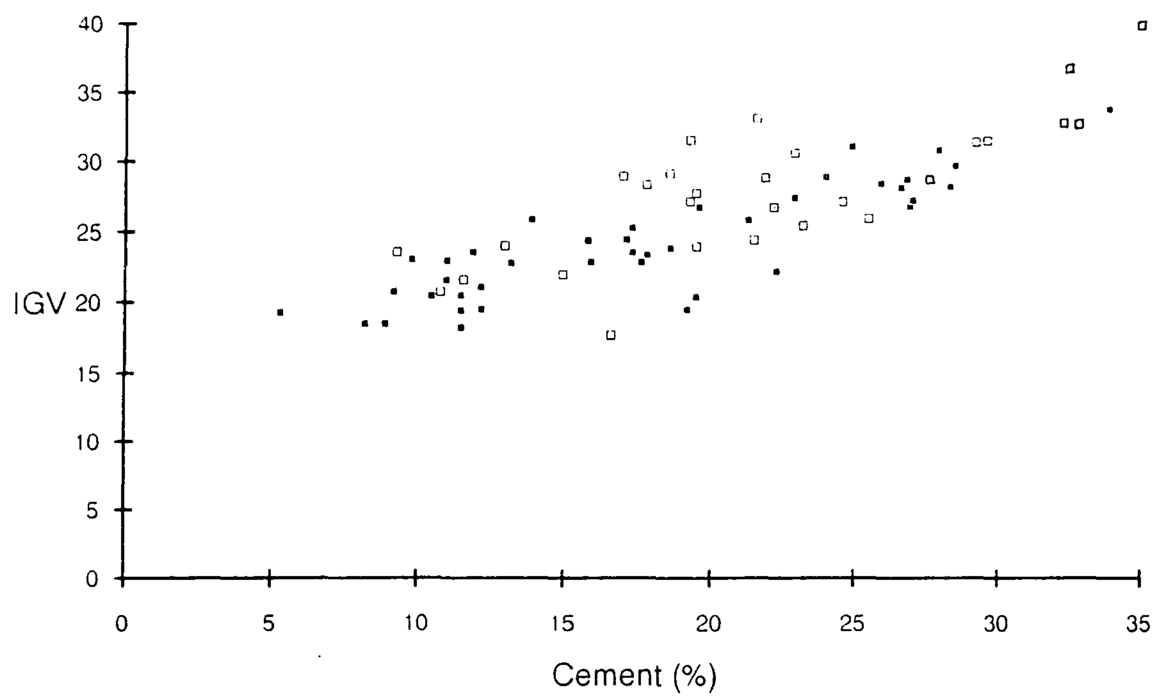



*Fig. 4.8. Plot of intergranular volume against cement.*

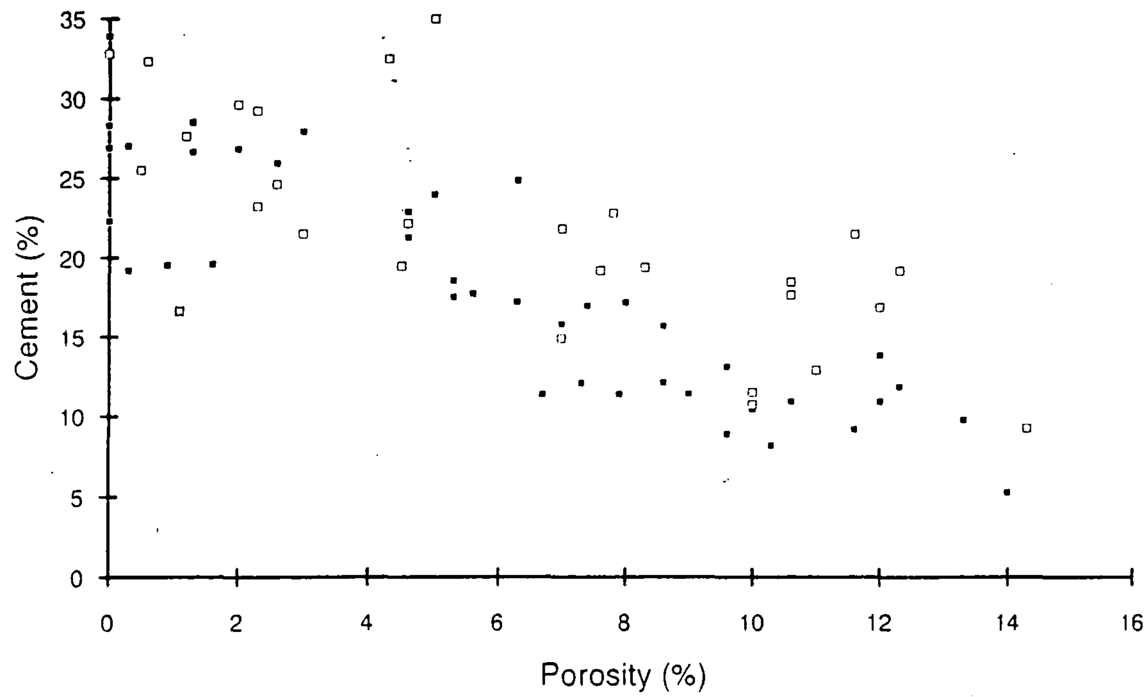



*Fig. 4.9. Plot of porosity against cement clearly showing the control of cement over porosity.*

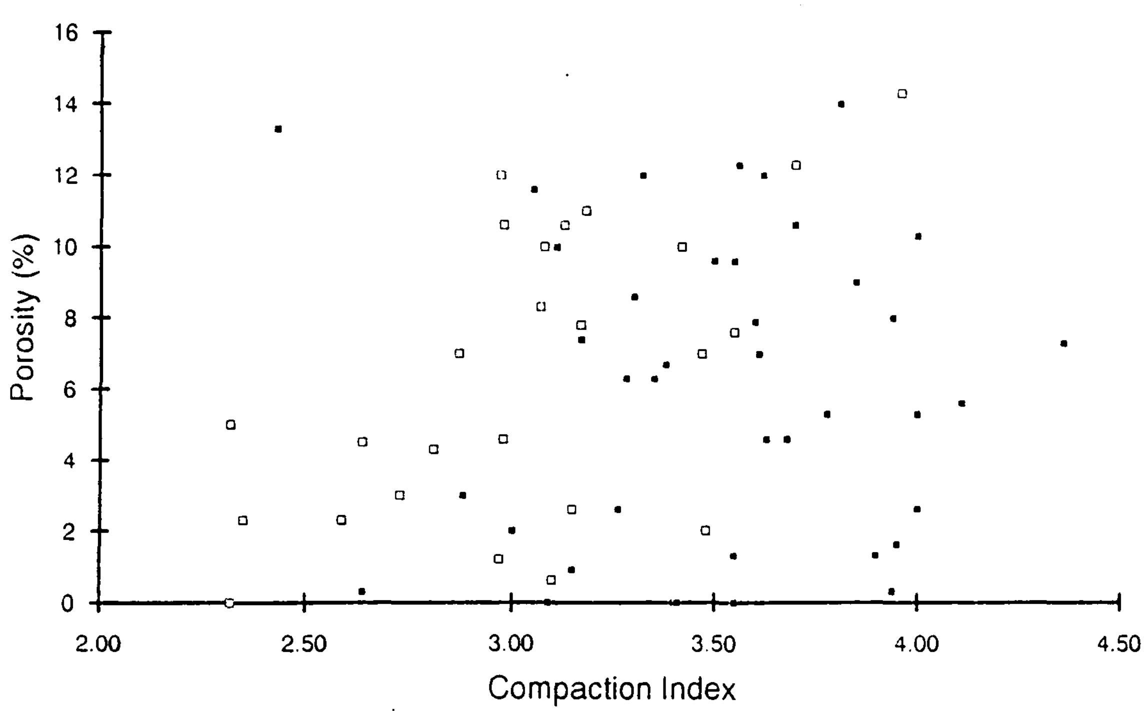



*Fig. 4.10. Plot of porosity against compaction showing an unexpected inverse relationship. See text for discussion.*

Further evidence comes from a plot of porosity against depth (Fig. 4.12). Although the depths on Fig .4.12 are only relative and not spaced regularly one would still expect a reasonable correlation between depth and porosity. However, there is no correlation because of the dominance of cementation (which is not depth dependant) over compaction. The only cement present in these samples with sufficient strength to retard compaction is calcite. However, there is little data available on the precise amounts of calcite required to achieve this (McBride *et. al.*, 1990). Mack (1984) suggests that 5% by volume of quartz cement is needed to retard compaction while Stephenson *et al.* (1992) suggest a wide range of values. They divide cement into two types: supporting (load-bearing) cement; and passive (non load-bearing) cement. Supporting cement is that which is deposited around intergrain contacts such that it increases the area of contact and thus acts to increase the resistance to further grain interpenetration (compaction). Passive cement on the other hand, is any pore filling material that does not affect the rate of interpenetration (Fig. 4.13). An example from Stephenson *et al.* (1992) which uses data from Houseknecht (1987) indicates that as little as 1% supporting quartz cement will act to inhibit compaction a little, while 15% supporting cement could be sufficient to halt compaction totally. As the average values of calcite present in the ORS are 11-12% this is probably sufficient to retard compaction.

Scherer (1987) in a study of the factors influencing porosity in sandstones from all around the world found that the dominant controls were age, depth of burial, sorting, and detrital quartz content in samples that had little or no cement. However, in the ORS samples these factors have no effect on porosity, which is due to the high cement proportions. Scherer (1987) and McBride *et. al.* (1990) have both suggested that a high thermal gradient could cause increased compaction by increasing the solubility of quartz and thus increasing grain interpenetration. Scherer (1987) is more specific in that he considers a thermal gradient would have to be in excess of 40°/km to be of any major importance. A probable thermal gradient for the Inner Moray

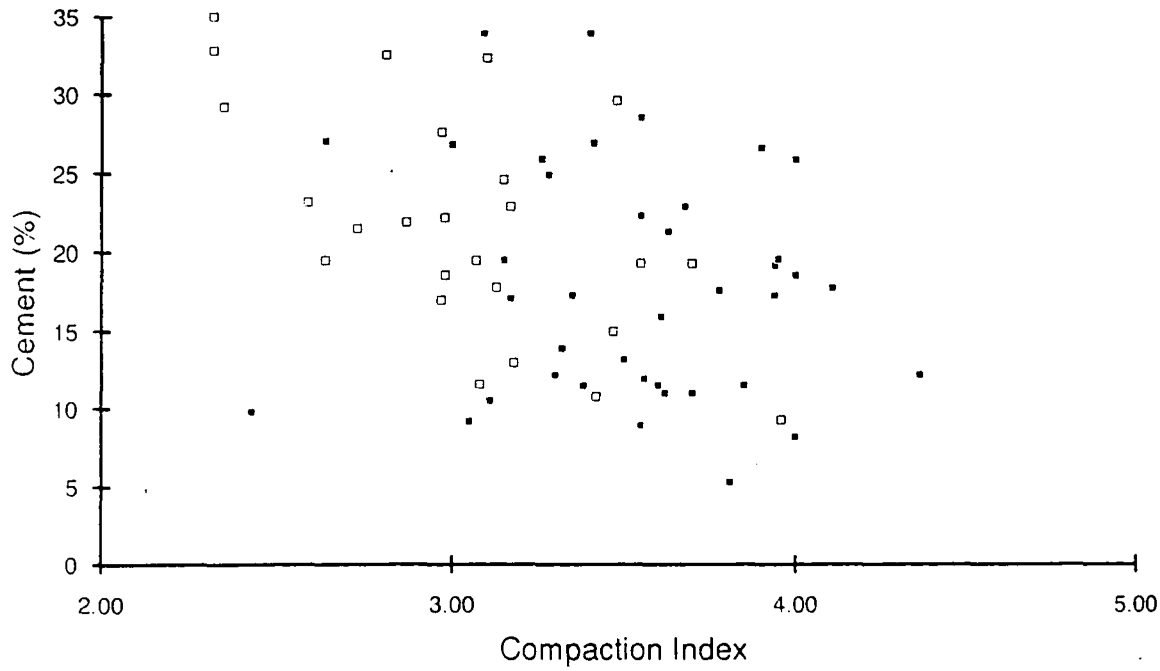



*Fig. 4.11. Plot of cement against compaction,*

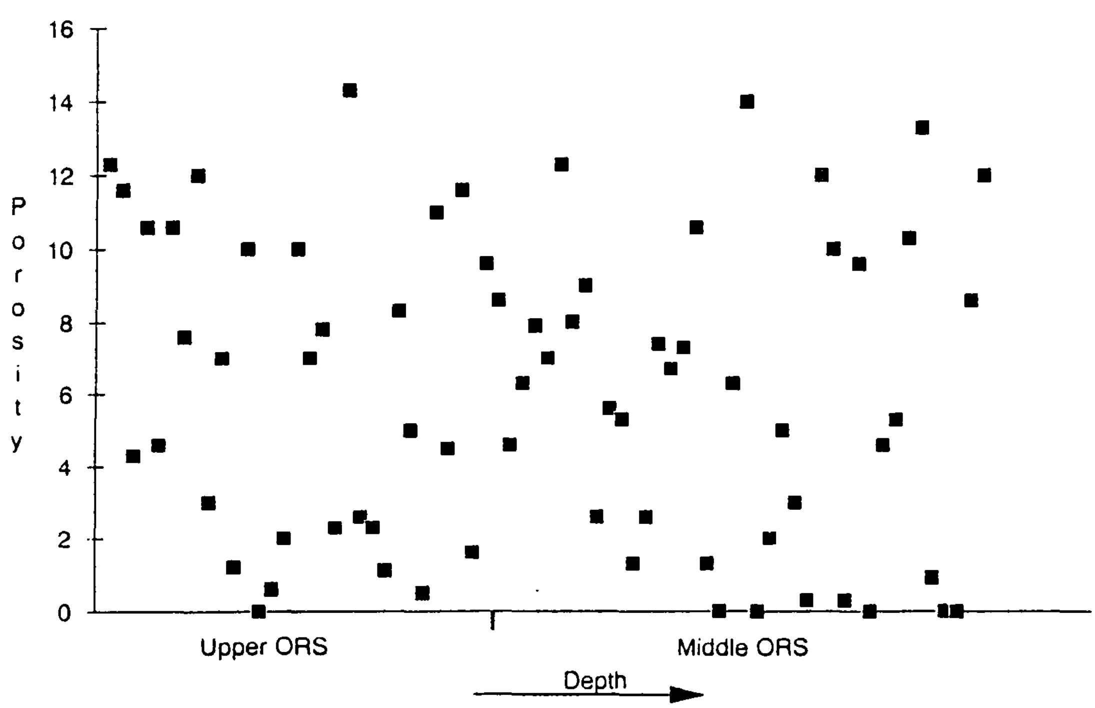



*Fig. 4.12. Plot of porosity against depth. Note that porosity is not depth dependant.*

Firth during the Devonian is 35°/km (T.R. Astin, *pers. comm.*) which is unlikely to have greatly affected compaction according to the value determined by Scherer (1987). Stephenson (1977) and Stephenson *et al.* (1992) suggest that variations in geothermal gradient will only have an effect on compaction at depths of burial greater than 2.5-3km.

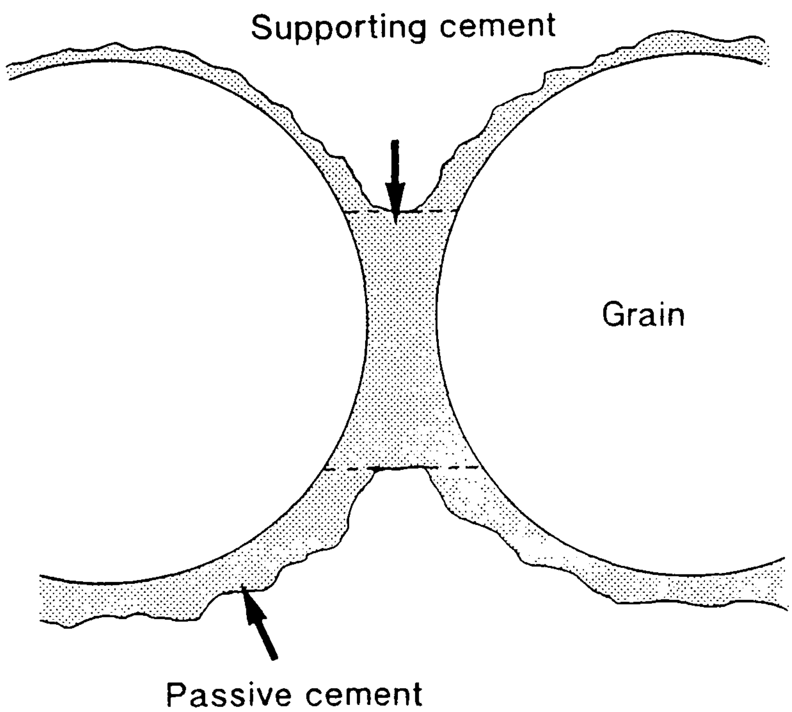



*Fig. 4.13. Diagram illustrating passive and supporting cement. (After Stephenson et. al., 1991)*

## B. Porosity Enhancement

Secondary porosity due to dissolution of cements or detrital grains is an important form of porosity in many reservoir sandstones (*eg.* Blackbourn, 1984; Knox *et al.*, 1984; Burley, 1986). In the ORS the importance of cementation in porosity destruction provides a high intergranular volume that can potentially be returned to porosity. Hayes (1979) and Schmidt and McDonald (1979b) have listed several criteria for the recognition of secondary porosity in thin section including such features as corroded grain margins, irregularly shaped pores and patches of authigenic cement in optical continuity, all of which have been observed within these samples (See Figs.

3.22, 3.23). The secondary porosity has formed mainly as a result of dissolution of the calcite cement (see Chapter 3). The fact that dissolution has occurred at all indicates that there must have been some residual primary porosity in the sandstones to allow the leaching fluids to pass through, and therefore the samples with the highest primary porosity have the highest potential for leaching (Bjørlykke, 1984). Unfortunately, it is impossible to differentiate between intergranular porosity due to dissolution and that which is remnant primary porosity (Fig. 4.2).

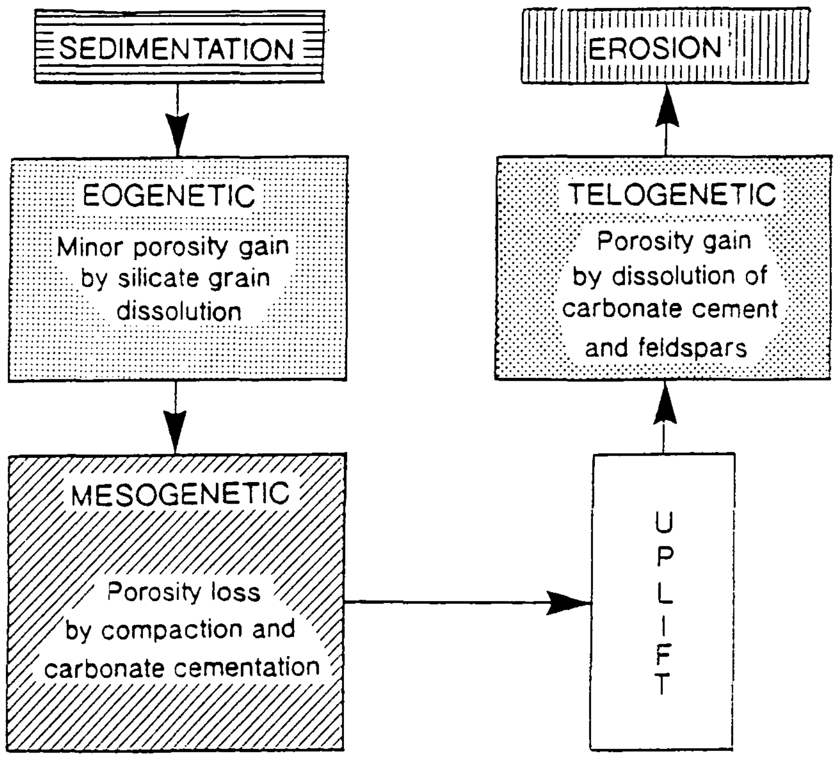



*Fig. 4.14. Diagram demonstrating the relative timing of porosity reduction/enhancement.*

Although the formation of kaolinite associated with the leaching of the calcite, fills intergranular pore space it does not totally occlude it, because of the microporosity associated with the kaolinite (Fig. 4.3) Nevertheless, this reduction in pore diameter is detrimental to the permeability (see below). Fig. 4.14 shows the relative timing of these porosity changes and the diagenetic regimes in which they occurred. Another possible method of increasing the porosity is by fractures, both microfractures and macrofractures, as a result of overpressure or compaction of the reservoir section. Several thin sections contain heavily fractured grains. However, these were almost certainly formed during section preparation, a view which is supported by the variability of fracturing across individual sections and an absence of fractured grains during SEM examination. Larger-scale fractures (although still on the micro-scale:0.1-0.3mm wide) were only observed in very few samples (Fig. 4.15) and this fracturing has had only a slight effect on the volume of total porosity. However, its effect on permeability is somewhat more important, as in the ORS of the Buchan Field (see below).

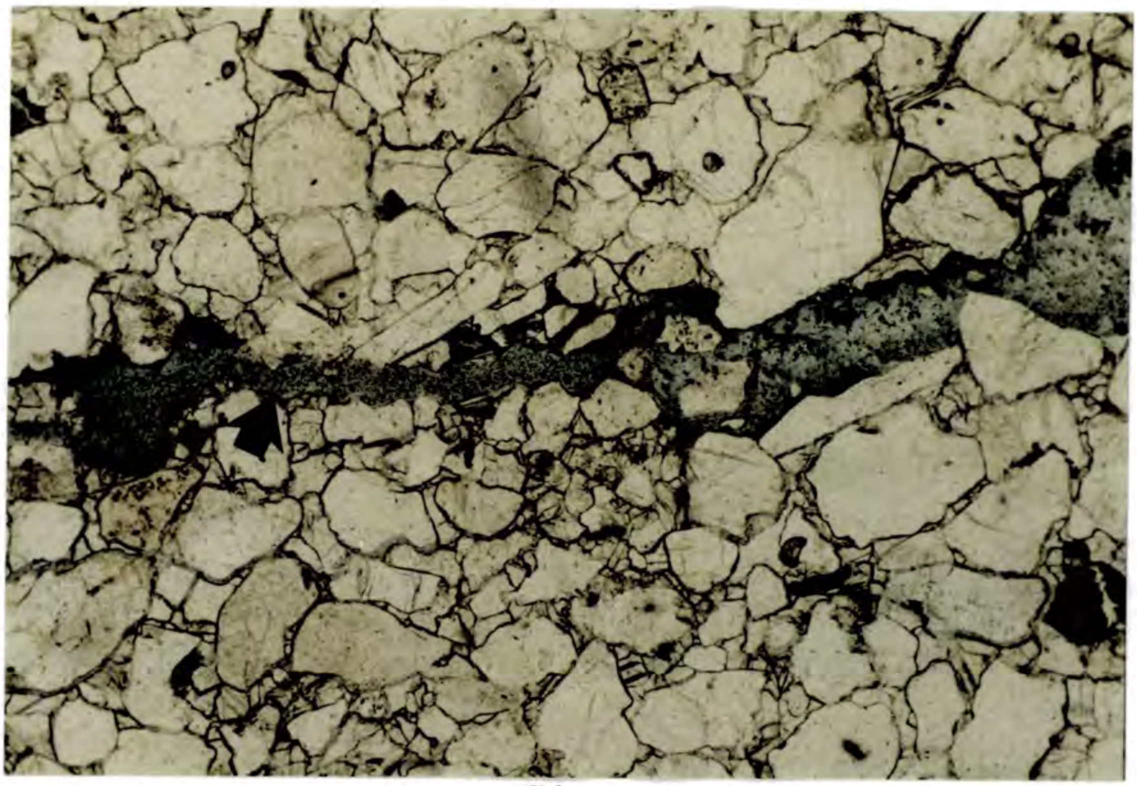


*Fig. 4.15. Photomicrograph showing open fracture (arrowed). (PPL, FOV = 3mm).*

## 4.3 Permeability

### 4.3.i Introduction

The overall permeability values for the ORS are very poor. They range from 14.5 to 270md (arithmetic mean of 50.9md, geometric mean of 34.7md) for the Upper ORS, and 0 to 48md (arithmetic mean of 16.7md, geometric mean of 15.9md) for the Middle ORS. The fact that these values are very low compared to the recent sandbodies studied by Pryor (1973, 4md to 195 darcys for point bar sands and 46 to

104 darcys for aeolian dune sands) is probably due to diagenetic effects, as for the porosity.

### <u>4.3.ii Permeability measurement</u>

A minipermeameter was used to obtain the permeability data (Fig 4.16). The instrument was loaned by R. M. Slatt of ARCO International Oil and Gas Company, Ltd., Plano, Texas, USA. It was constructed according to the design of Eijpe and Weber (1971) and consists of a probe, a series of flow meters and an air pressure gauge housed in a wooden box, and an air supply. The probe is a metal tube about 15cm long with an internal diameter of 6mm. There is a rubber ring at the tip of the probe to prevent air leakage between the tip and the rock surface.

*Fig. 4.16. Photograph showing the minipermeameter in the field.*

The main box consists of five flow meter mounts which could house any of the eight flow meters provided, although only one meter was 'on line' at any one time. The flow meters cover a wide range of flow rates, with the smallest being 0 to 10 cc/min, and the largest being 0 to 8 l/min. There is also a pressure gauge which reads in psi and a valve to control the flow of air through the permeameter. The air supply is contained in a metal cylinder which holds about 200lb of air.

Calibration of the minipermeameter was done by taking measurements from a set of seven core plugs which had known permeabilities in the range of 3 to 563 md. During calibration, and at all times thereafter, a constant air pressure through the probe of 1 psi was maintained whilst taking readings. For each plug a set of six measurements (three from each end) were made with each flow meter. The highest and lowest values were discounted, and the remaining four readings were averaged to give a single value for each plug with each flow meter (see also Eijpe & Weber, 1971). These flow rate values were then plotted against the known permeabilities of the plugs to construct a correlation curve (Fig. 4.17). Flow meter A (0 to 10 cc/min) gave consistently low readings and was thus discounted, also flow meter D (0 to 250 cc/min) gave high readings for the plugs with permeabilities less than 40md, and consequentially these readings were also discounted. A regression line was then fitted to the remaining data points (Fig. 4.18) used to convert all measured flow readings to permeabilities.

The original plan was to use the minipermeameter to obtain some detailed information from outcrops about permeability anisotropy associated with cross bedding, similar to the work done by Weber *et al.* (1972) and Weber (1982) as well compiling a more general data set. However, when the minipermeameter was used in the field it was impossible to get a good enough seal between the probe tip and the rock surface to get any sort of reading, due to the roughness of the rock and the inflexibility of the rubber ring around the tip. As a result samples had to be collected by hand and brought back to the department where smooth faces were cut to take measurements from. As a result the detailed field measurements which had originally been included in the research proposal could not be made.

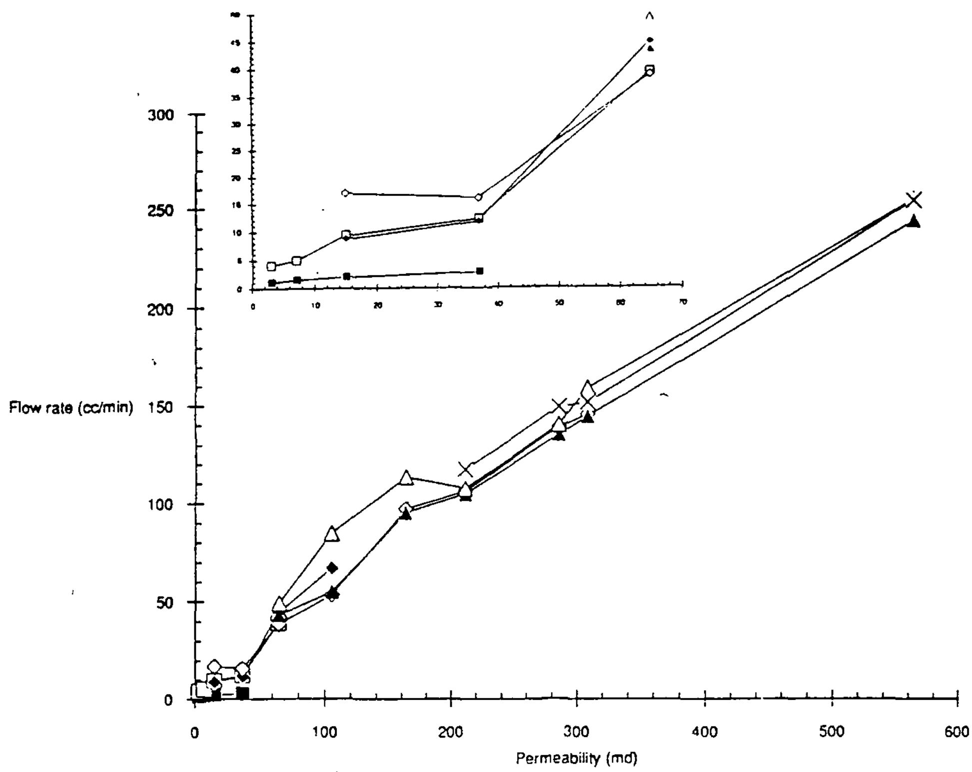



*Fig. 4.17. Plot of flow rate against permeability for the different flow meters.*

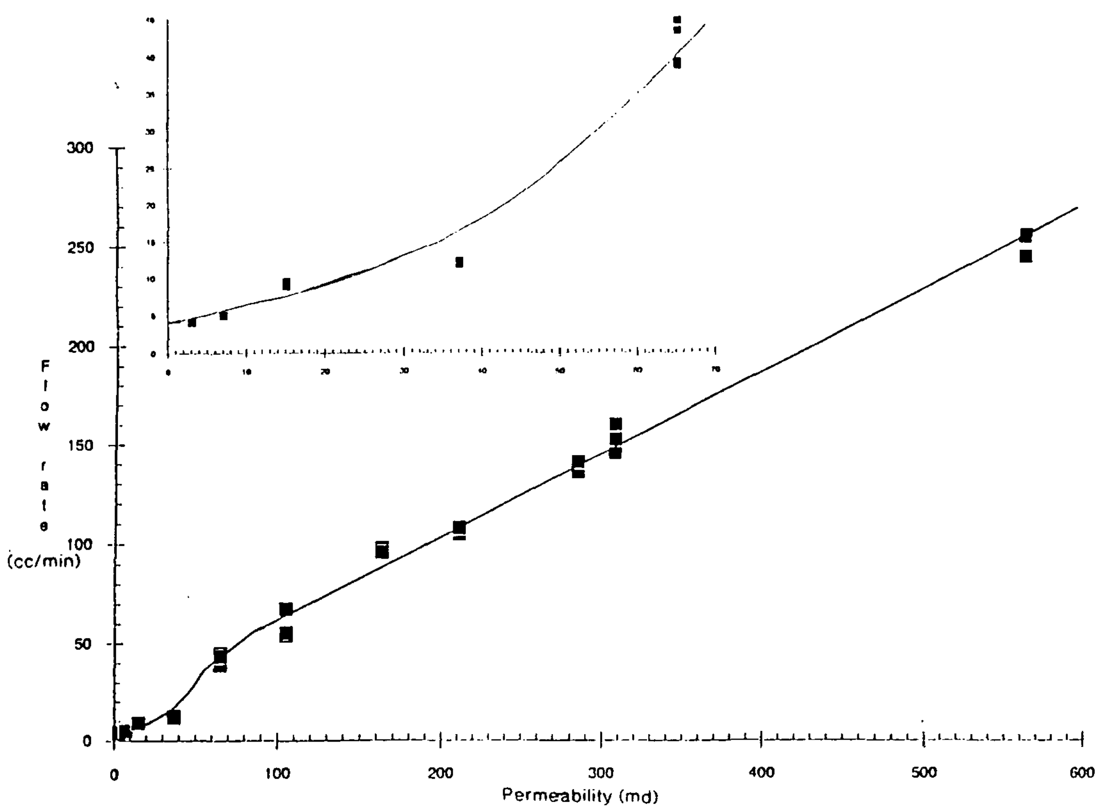



*Fig. 4.18. Correlation curve used to convert flow reading into permeability.*

#### 4.3.iii Permeability controls

Pryor's (1973) study showed that in recent fluvial and aeolian sandbodies permeability is controlled by grain size and sorting, which are directly related to the depositional environment. The relationship between texture and environment results from the unique set of physical conditions associated with any environment or bedform. Under these circumstances different lithofacies should intuitively have quite distinct permeability characteristics. However, when the lithofacies are plotted against permeability (Fig. 4.19), there is a marked overlap in permeability values for the different lithofacies and there is no correlation between permeability and grain size (Fig. 4.20) or sorting (Fig. 4.21). It is also worth noting that when porosity is plotted against permeability for different lithofacies, there is again an overlap (Fig. 4.22).

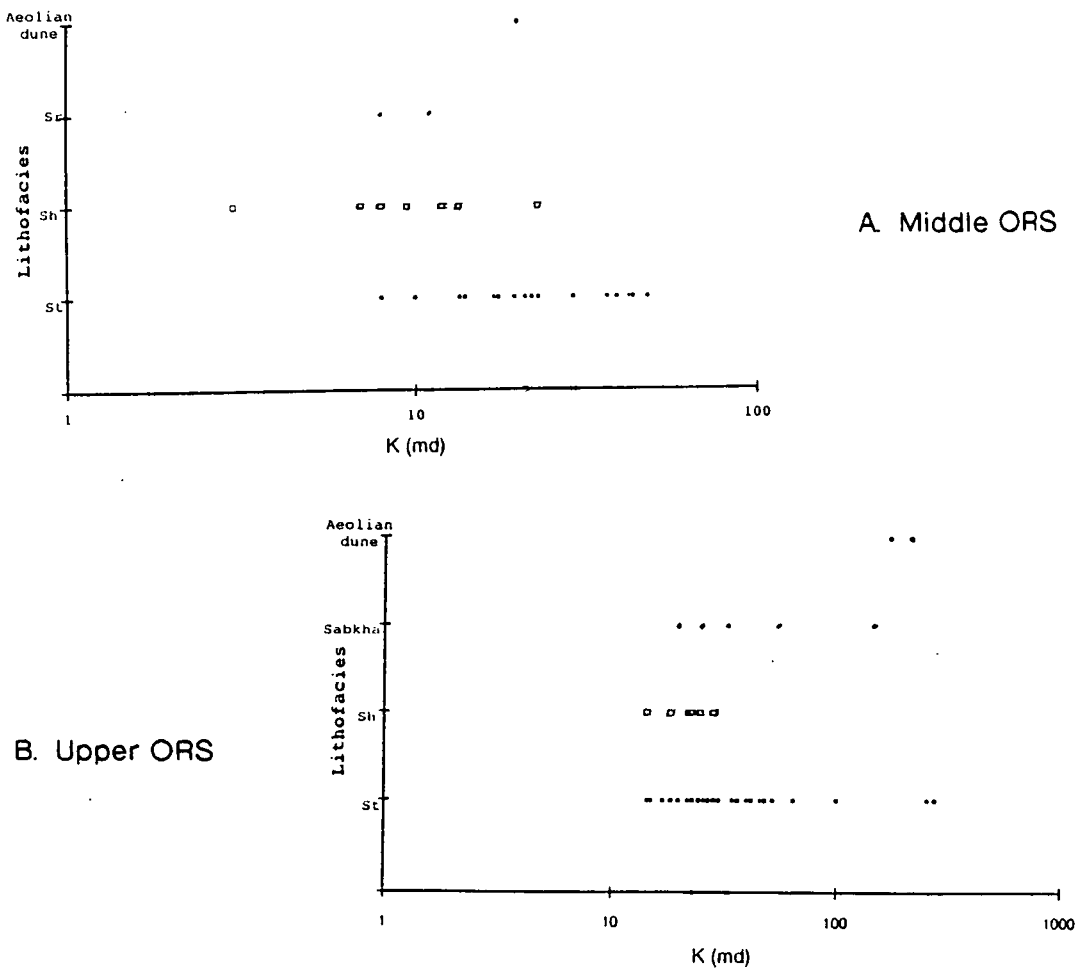



*Fig. 4.19. Plot of lithofacies against permeability. Note the large overlap in values for different lithofacies. St = trough crossbeds; Sh = horizontal laminations; Sr = ripple cross lamination.*

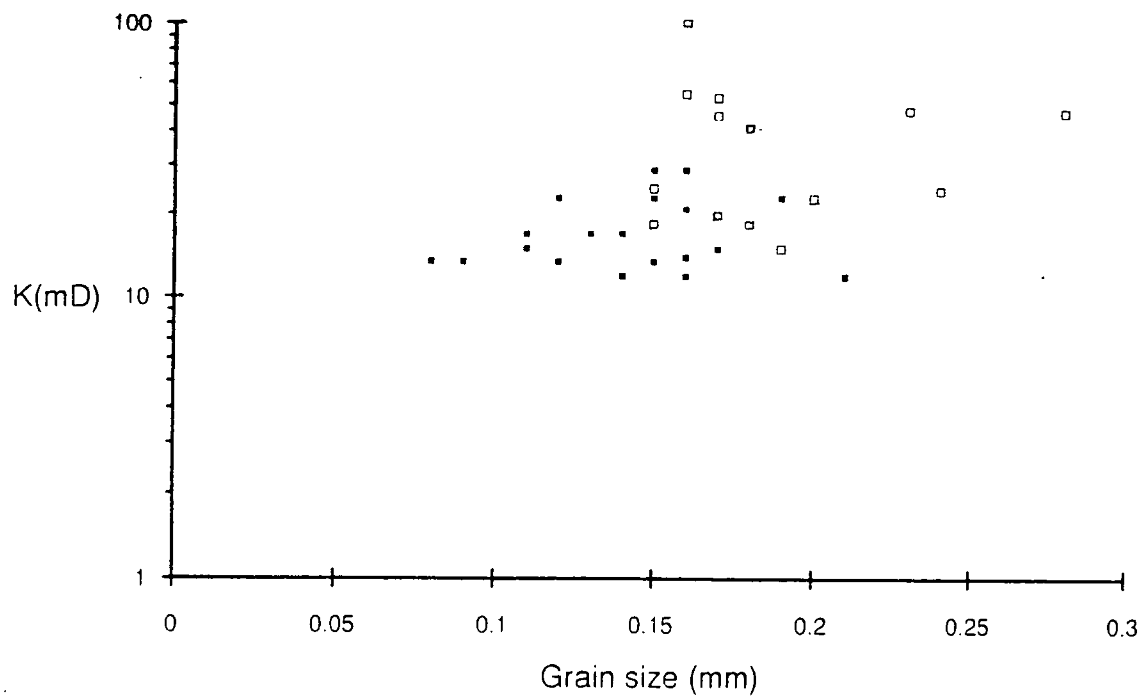



*Fig. 4.20. Plot of permeability against grain size, showing no correlation at all.*

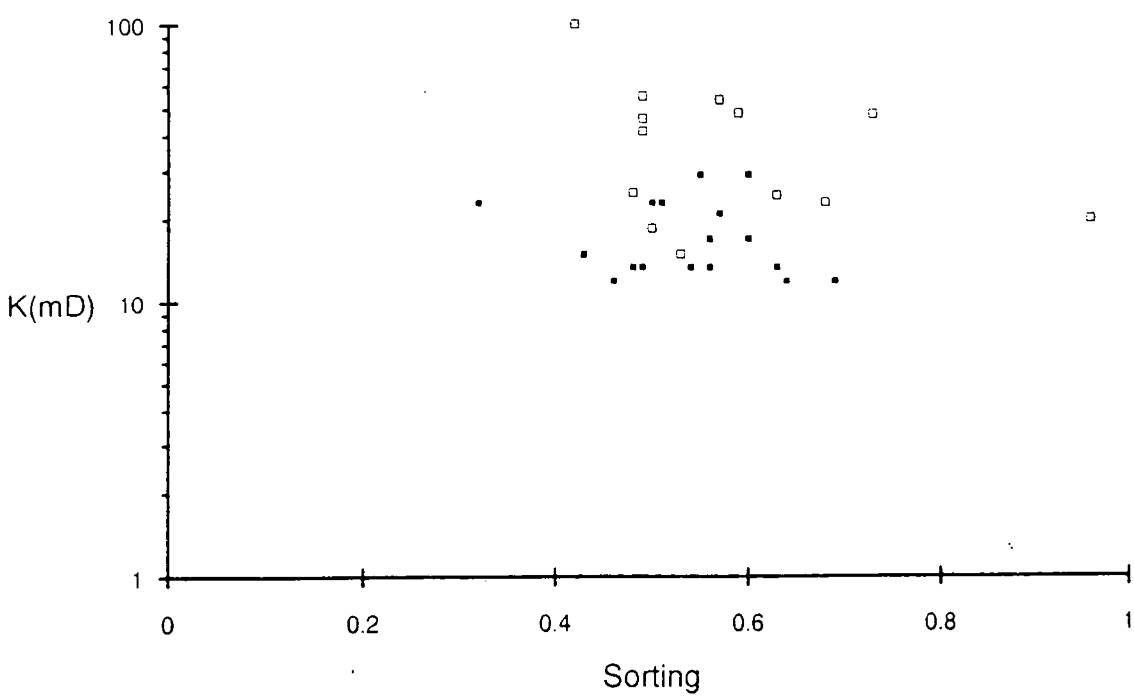



*Fig. 4.21. Plot of permeability against sorting, again showing no correlation.*

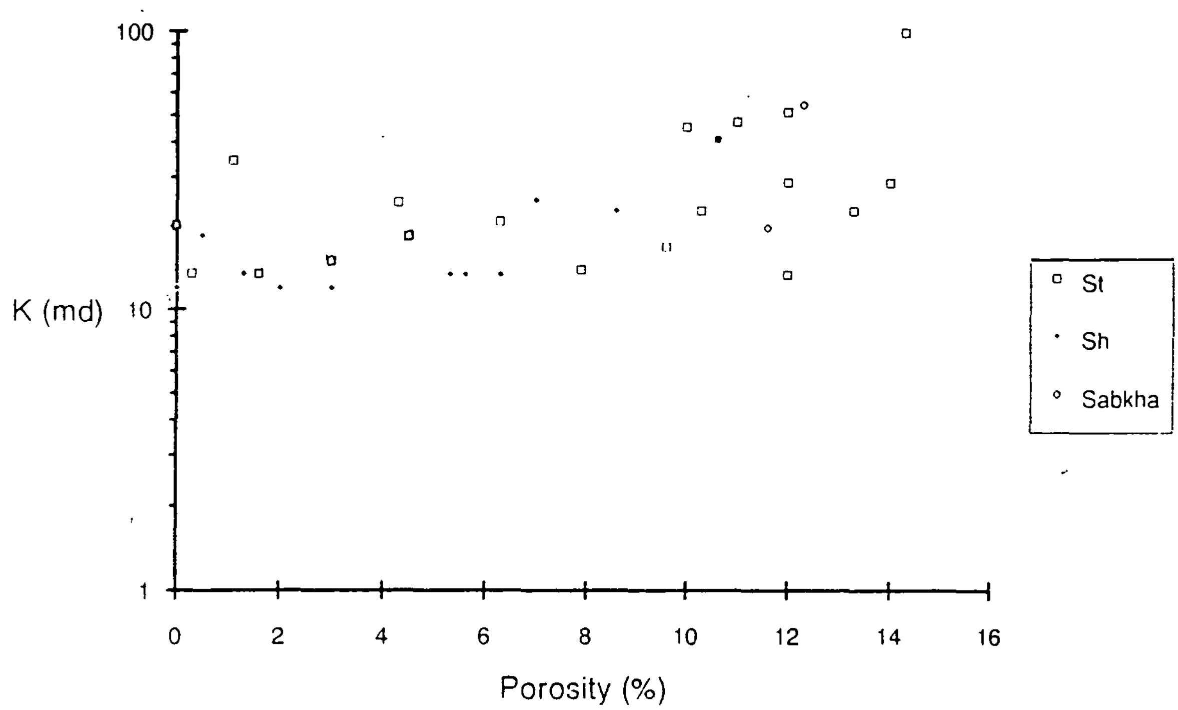



*Fig. 4.22. Plot of porosity against permeability for the different lithofacies, showing clear overlap in values for the different lithofacies. St = truogh crossbeds; Sh = horizontal laminations.*

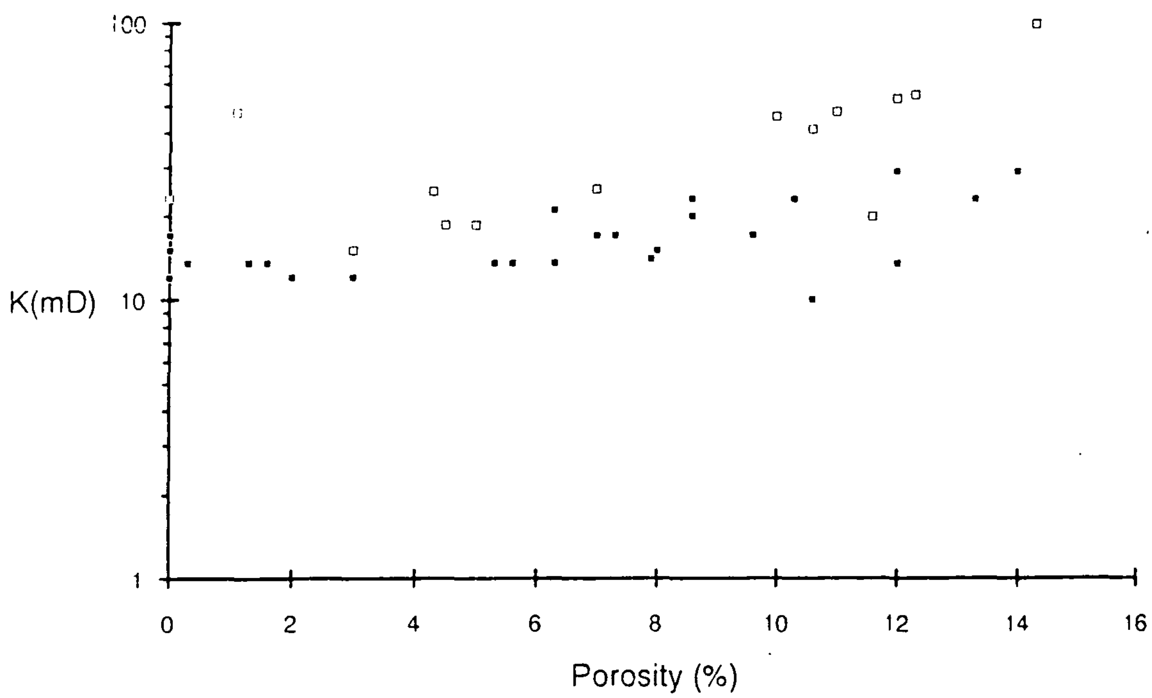



*Fig 4.23. Plot of porosity against permeability showing a clear relationship despite the narrow range of permeabilities.*

However, there is a good correlation between porosity and permeability (Fig. 4.23), despite the narrow range of permeabilities. Earlier in this Chapter it has been shown that the calcite cement was the major control on the amount of porosity, which in turn exerts a control on permeability. Thus, the cement must also affect the permeability, as can be seen in Fig. 4.24. The plots of permeability against intergranular volume (Fig. 4.25) and compaction (Fig. 4.26) show inverse relationships similar to those shown by the respective plots against porosity, which is to be expected in view of the dependence of permeability on porosity. It follows therefore, that because the porosities are low, so too are the permeabilities. However, there is still the problem of the much lower permeability values in the Middle ORS relative to the Upper ORS, even for similar lithofacies. This appears to be due to a smaller average pore diameter in the Middle ORS samples, resulting from a small crystal size of the calcite cement. This in turn leads to smaller pores after dissolution, thereby restricting the free flow of fluids (*ie.* decreasing the permeability).

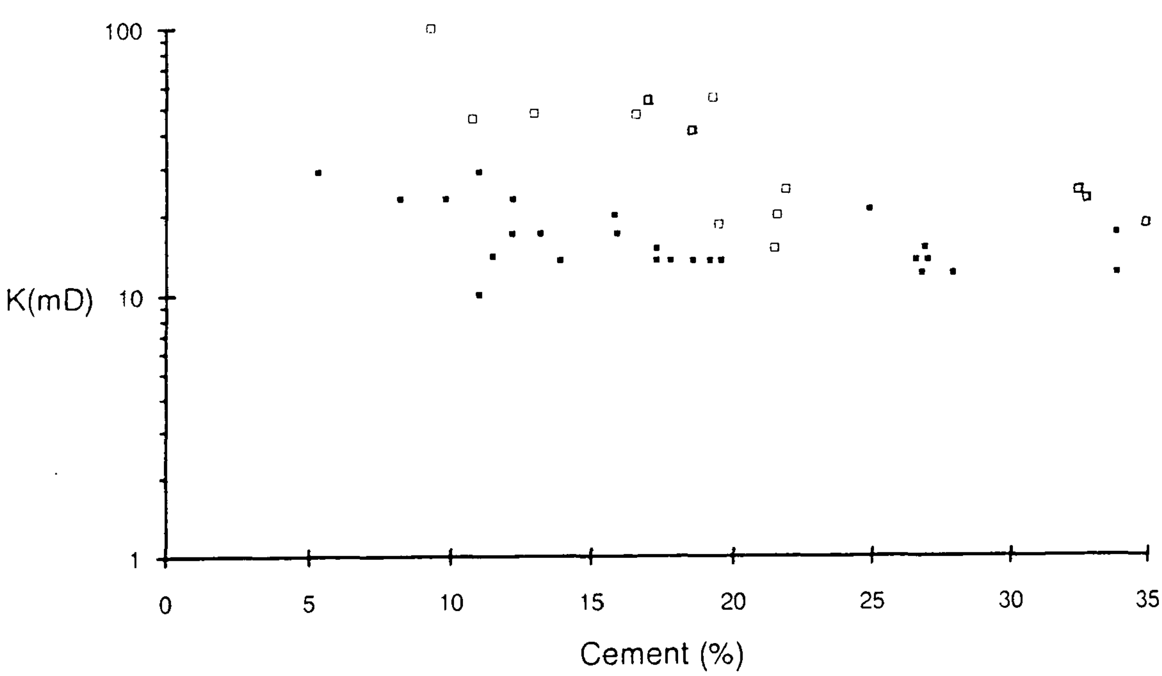



*Fig. 4.24. Plot of permeability against cement indicating the control exerted by the cement on the permeability.*

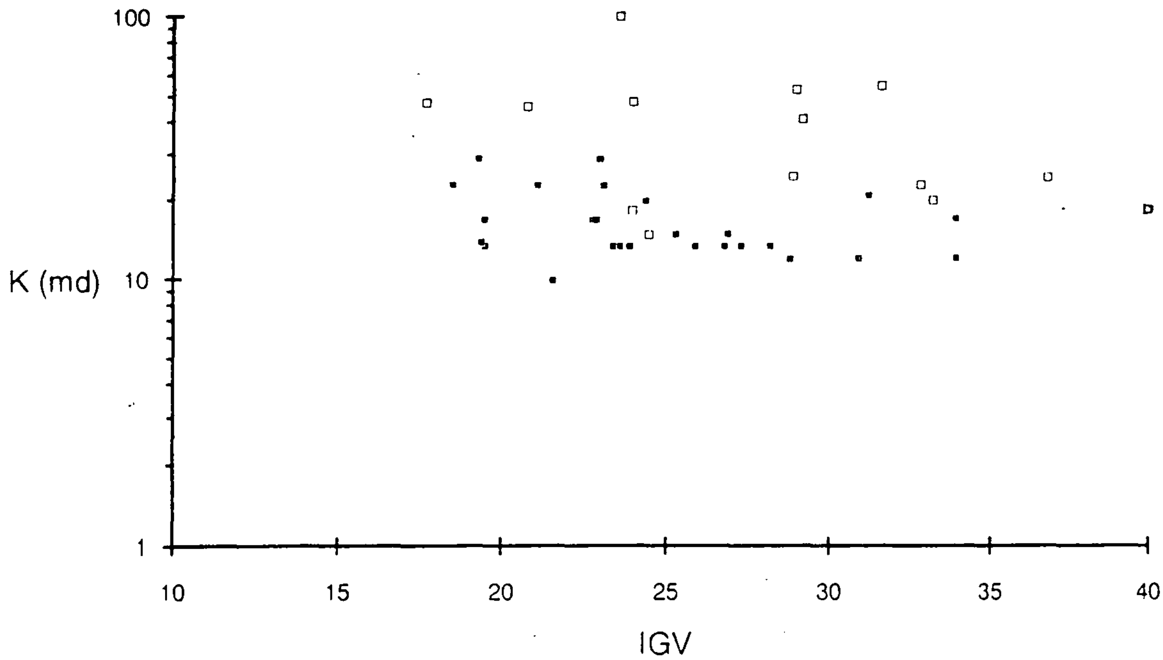



*Fig. 4.25. Plot of permeability against intergranular volume*

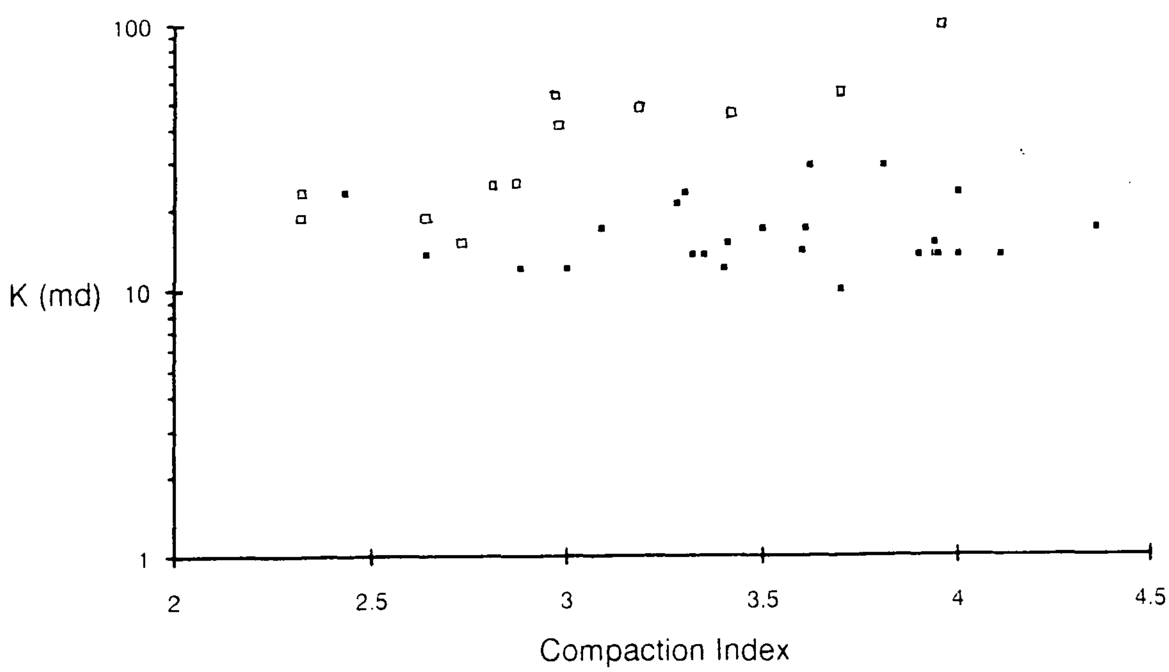



*Fig. 4.26. Plot of permeability against compaction,*

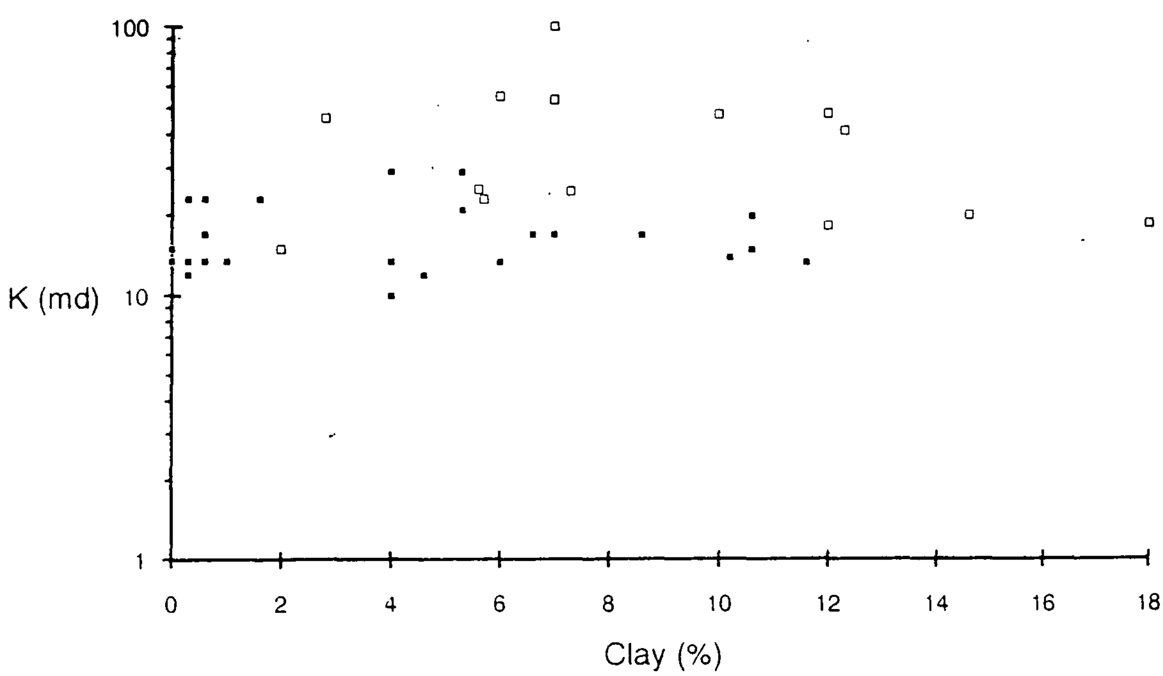



*Fig. 4.27. Plot of permeability against clay. Note that the clay has no effect on permeability until it is present in quantities greater than c. 8% by volume.*

Clay minerals can have a very detrimental effect on permeability by occluding intergranular porosity, which is the type of porosity most relevant to permeability. Fibrous clays like illite, which have high surface areas and bridge pore throats, have the worst effect, while platy clays like kaolinite, are less detrimental (Stadler, 1973). Kaolinite is the most common clay in the ORS of the study area, and even then it is usually present in only small amounts, with the result it probably has only a minor effect on permeability. Ehrenberg (1990) in a study of cores from the Jurassic Garn Formation in the North Sea showed that illite only has a significant effect on porosity when it occupies a major percentage of the total porosity (*ie.* there is a significant reduction in the average pore diameter). A similar relationship may apply to the ORS, in that the plot of clay against permeability (Fig. 4.27) appears to show no correlation until the clay minerals are present in amounts in excess of about 8% rock volume, after which any further increase causes a decrease in permeability. However, the effect will be less marked than that observed by Ehrenberg (1990) because the dominant clay is kaolinite not illite.

Selley (1985) states that fractures are important in enhancing low permeability, brittle and semi-brittle rocks into economic reservoir rocks. There are several examples of fields producing from fractured rocks to support this view: a) Devonian-Carboniferous ORS, Buchan Field, North Sea (Butler *et al.,* 1976); b) Jurassic age Franciscan sediments, Long Beach and other fields, California (Truex, 1972); and c) the Augila Field (fractured granite) of the Sirte Basin, Libya (Williams, 1972).

It is not the total number of fractures that is important when considering permeability, but the number of fractures that are open, and not filled by later cements. The rare fractures that are seen in thin section at the micro scale, are all open, and the few samples tested for permeability that were fractured had very high values (550-1400md) compared to host rock permeabilities. However, many fractures seen in the field around the Moray Firth were obviously closed, while with many others it was impossible to tell from their surface expression. Thus, although fractures do improve permeability it is not certain as to whether there are a sufficient number of open fractures to make an economic reservoir. It is also worth noting that the majority of the fracturing occurs in the Upper ORS which has a very high sand/shale ratio (c. 9) thus making it more homogeneous and more brittle than the Middle ORS.

## 4.4 Summary

Porosity and permeability are closely linked, with permeability depending on the presence of porosity. In recent sandbodies the values for porosity and permeability are very high and are controlled by textural characteristics (*ie.* grain size and sorting) which vary according to depositional environment. In the ORS samples however, permeabilities are reduced by several orders of magnitude and porosities by over 80%, on average, resulting in very poor values. The lack of correlation between textual parameters and porosity suggests that it is primarily controlled by diagenetic effects. The relative contributions of compaction and cementation to porosity loss vary, but on average cementation has been more important.

Intergranular macroporosity is the most important pore type present. This is a mixture of primary porosity and dissolution porosity after calcite cement, but it is impossible to quantify precise amounts of dissolution porosity. There are minor amounts of intragranular and microporosity also, but these are not as significant. A schematic summary of the porosity evolution path is shown in Fig. 4.28.

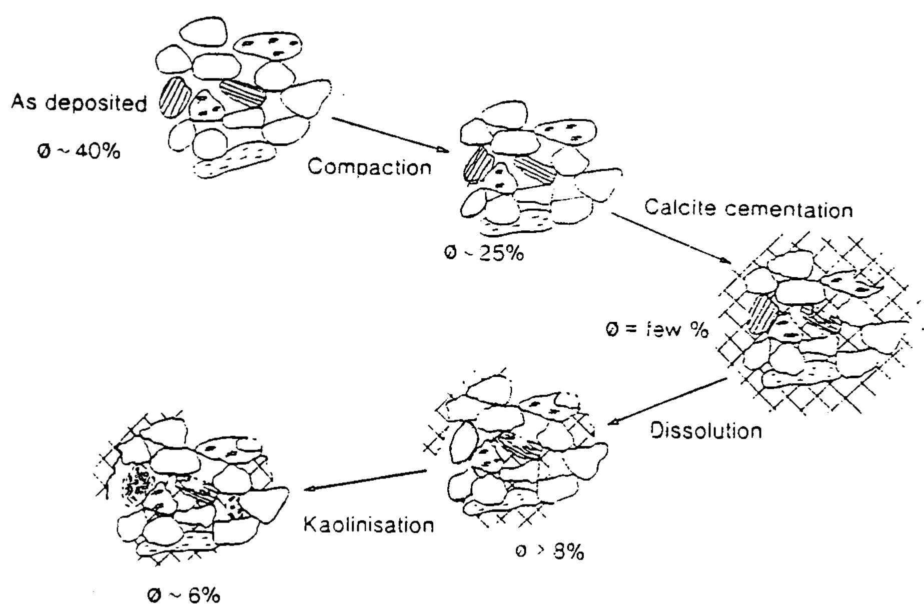



*Fig. 4.28. Schematic representation of the porosity evolution path.*

Permeability is directly related to macroporosity, which is controlled by the above mentioned factors. When clay minerals occlude a large proportion of the macroporosity, the resultant decrease in pore size causes a reduction in permeability. However, only a few samples contain sufficient amounts of clay to drastically affect the permeability. The overall reduction in permeability values in the Middle ORS relative to the Upper ORS is due to a decrease in average pore diameter of the macroporosity. Fractures, despite locally increasing permeability, might not be present in sufficient quantity to make a significant difference to the reservoir characteristics, as they do in the Buchan Field.

## Chapter 5

# Reservoir Heterogeneities

### 5.1 Introduction

No rock body, carbonate or siliciclastic is entirely homogeneous. The presence of internal heterogeneities in rock bodies such as physical barriers to fluid flow or gross variations in permeability, can greatly reduce the effective recovery of oil from a reservoir. Problems such as these have been recognised for many years, as evidenced by the early works of Mills (in Ambrose, 1921) on the effects of permeability variations in trapping oil during waterflood. Thus, analysis of the types of heterogeneity present, and their distribution, within a reservoir is an essential part of any study into the fluid flow and reservoir characteristics of a rock body. In most reservoirs the wide spacing of wells means that there is insufficient data to model heterogeneities adequately, therefore analogues have to be derived from outcrop studies. The presence and probable distribution of heterogeneities is generally predictable due to their close relationship with depositional environments, diagenetic history and tectonic setting (Weber, 1986).

### 5.2 Classification of heterogeneities

Heterogeneities occur on a variety of scales. Pettijohn *et al.* (1973) proposed a hierarchical four fold sequence of controls on permeability. Haldorsen and Lake (1984) proposed a modification to this model as follows: 1) microscopic, or pore scale, 2) macroscopic,based on the scale of individual core plugs, 3) megascopic,based on the size of grid blocks in a field model, and 4) gigascopic, which is the total formation or reservoir scale. Weber (1986) however, subdivided the scales slightly differently and arrived at a fivefold breakdown. Dreyer *et al.* (1990) have combined both of these models into a single diagram (Fig. 5.1).

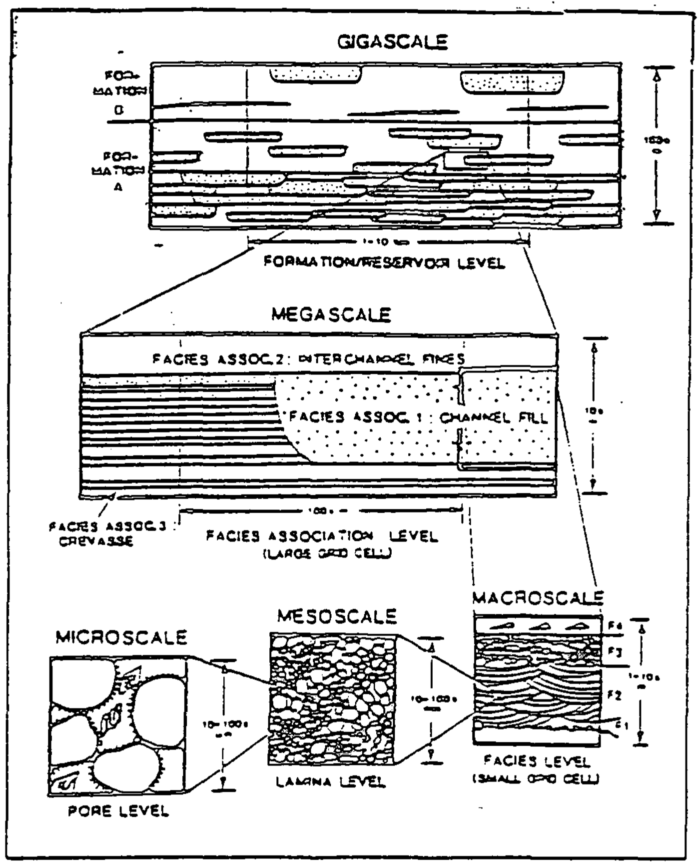


*Fig. 5.1. Levels of reservoir heterogeneity in relation to facies-related permeability classes. Example from a fluvial reservoir. (Redrawn from Dreyer et al., 1990).*

According to Weber (1986) size is very significant when assessing the importance of heterogeneities, and large tectonic features, *eg.* faults, have a greater effect on overall fluid flow characteristics than small scale features, *eg.* sedimentological characteristics. As a result, Weber (1986) suggested a classification of heterogeneities based on their size and genetic origin which can be used in the quantification of heterogeneities (Fig 5.2).

## 5.3 Description and quantification of heterogeneities from the ORS

The description of heterogeneities is a problem of size. Faults may be visible on seismic sections, whilst most other heterogeneities have an extent less than average well spacing. In particular, the extent of shale intercalations is very difficult to ascertain from logs and correlations. In general, with the exception of faulting and fracturing, the key to the analysis of heterogeneities is the correct identification of environments of deposition and diagenetic history (Weber, 1986). Comparison with modern day analogues and outcrop studies is essential, hence cores of the reservoir are necessary. In this section, each class of heterogeneity, from large to small scale (Fig. 5.2), will be described as they occur in the ORS, and their affect on the reservoir potential of the ORS will be discussed.

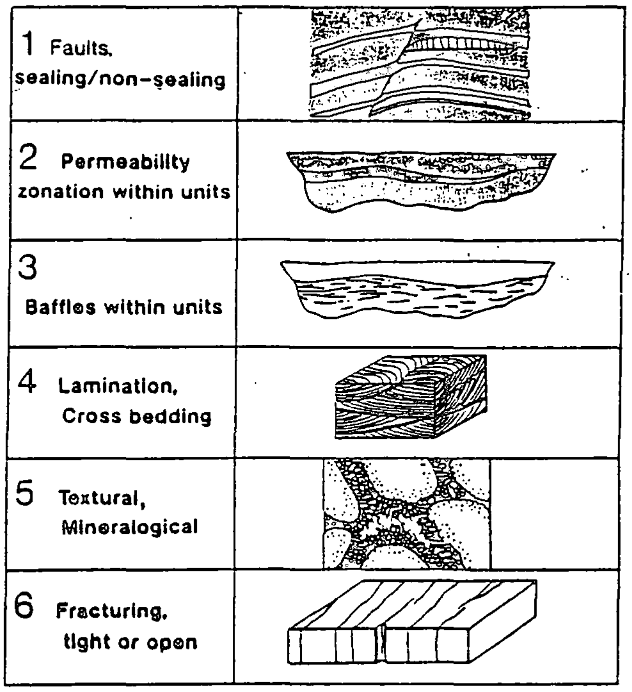



*Fig. 5.2. Classes of heterogeneity. (After Weber, 1986).*

## 5.3.i Faults

Faults can have either beneficial or detrimental affects on a reservoir depending on whether they are sealed or not. Weber (1986) maintains that although faults are often clearly visible on seismics or from log correlations their sealing capacity is difficult to ascertain unless fluid contacts or pressures can be observed across the fault. As a result it also virtually impossible to adequately quantify the effects of faulting from field observations, especially if the actual fault planes are only rarely visible as is often the case in the Orcadian basin.

There are three main mechanisms by which fault sealing may occur: clay smearing along the fault plane, cataclasis of fault gouge material and diagenetic sealing of faults. Clay smearing is a process particularly important during synsedimentary faulting of clay/sand sequences, as for example in deltaic environments (Weber, 1986). This process has been described from Niger delta oil-fields (Weber *et al.,* 1978) and from Louisiana Gulf Coast oil-fields (Smith, 1980). The resulting sealing capacity of the fault is mainly a function of the percentage of clay in the section that passed over the fault (Weber *et al.,* 1978).

Cataclasis can strongly reduce fault zone permeability, and additionally, the fine grain gouge is more susceptible to diagenetic alteration. It is more common in wrench and reverse faulting (Weber, 1986). Diagenetic sealing of otherwise permeable faults has been described by Smith (1980) and is very difficult to predict or quantify.

Clay smearing is unlikely to be a major feature of the Devonian faults, despite the fact that many of them were active synsedimentary faults. This is because there is relatively little mud in the ORS succession, and because the synsedimentary faults were at the basin margin where the coarsest deposits (sand and gravel) are to be found. Equally, the majority of the faults are normal extensional faults, and therefore unlikely to have undergone much cataclasis. There are however, transfer faults linking many normal faults which have a wrench motion, and as such could be susceptible to cataclasis. Diagenetic sealing of faults is the most likely mechanism to have occurred, given the preponderance of calcite cement throughout the ORS. Nevertheless continuous movement along synsedimentary faults would keep reopening them.

### 5.3.ii Fractures

Fractures, in terms of their effect on reservoir characteristics, can be classified as open or closed, as with faults. Some differences between faulting and fracturing are that fractures tend to be on a smaller scale than faults, and there is only one method of sealing fractures, that is diagenetically. Open fractures have a beneficial effect in that they can greatly increase porosity and especially permeability, thereby enhancing the potential of what might have been an otherwise poor reservoir. Closed fractures on the other hand are very detrimental in that they can divide a perfectly good potential reservoir into unconnected units.

In the ORS fractures occur on two scales: micro fractures, and macro fractures, or joints. Micro fractures are best observed in thin section, being 0.1-0.3 mm wide and extending for a few cm's (Fig. 5.3). All observed examples of these small fractures are uncemented and open. Larger scale fractures or joints are visible in the field (Fig 5.4). These can extend for several tens of metres and have on average, a spacing of 1m. They are found almost exclusively in the Upper ORS, probably as a result of the very high sand/shale ratio (approx. 9) which makes it far more homogeneous and therefore more likely to undergo brittle deformation. The jointing also appears to be more abundant in aeolian sandstones in the succession than in the fluvial sandstones.

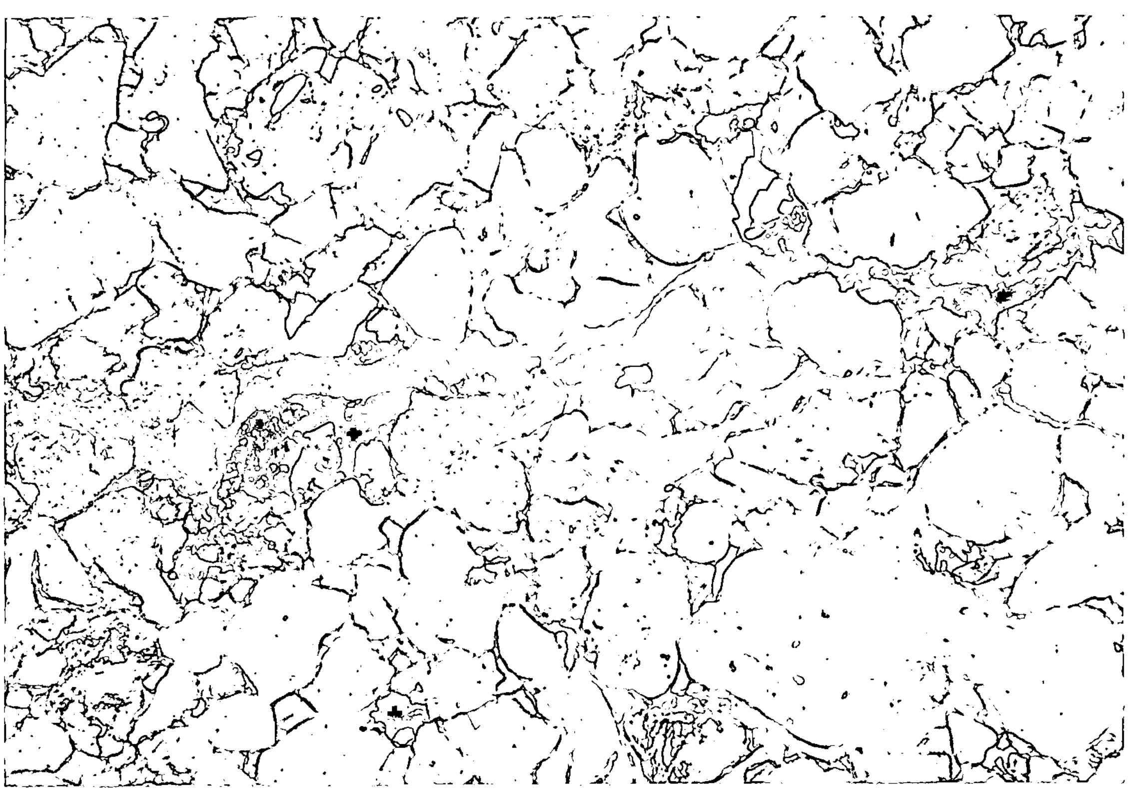


*Fig. 5.3. Open microfracture in a fluvial sandstone enhancing porosity and permeability. (PPL; FOV = 3mm).*

In the field the joints are easily recognised because they are typically seen to be standing proud of the surrounding sandstone (Fig. 5.4). This is a weathering effect due to the fact that the joints are sealed by a calcite cement (*ie.* they are closed) and are more resistant than the surrounding sandstone which is less heavily cemented. Petrographically (including under cathodoluminescence) this cement has the same features as the calcite that occurs as a cement throughout the ORS. The orientation of the joints (approx. NE-SW) is consistent with them having formed during Devonian extensional faulting. This would imply that the joints formed soon after burial and early cementation.

***Fig. 5.4.*** *Calcite cemented joints standing proud of the host sandstone. (Hammer length = 40cm).*

Supporting evidence for early formation of the joints comes from the presence of the infilling calcite cement. If it is assumed that the calcite cement along the joints originated from the same source as the rest of the calcite cement, it is possible to ascertain a relative timing for the formation of the joints. After burial and the onset of the cementation by calcite, the rocks would rapidly become partially lithified, and brittle enough to fracture with continued syn-sedimentary faulting. This fracturing would lead to the formation of conduits of high porosity and permeability when compared to the surrounding rock, along which any further carbonate saturated fluids would preferentially pass. As a result the joints would become preferentially sealed with calcite cement leaving the surrounding sandstone less well cemented.

The micro fractures, which are all open, have a very positive effect on the permeability. Fractured samples gave permeabilities of 550-1400md, which is significantly higher than matrix permeabilities. Despite this the overall effect on the reservoir potential is going to be negligible because of two factors. Firstly, the fractures are very small and only extend for a few cms with the result that increased permeability is going to be very localised. Secondly, fracture density is low with less than 1% of the samples on which permeability measurements were made containing such fractures.

*Fig. 5.5. Outcrop of aeolian dune cross bedding with tightly cemented subvertical to vertical joints* compartmentalising it *. (Cliff height = approx. 3m).*

Because the joints, which are of much greater extent and connectivity than the micro fractures, are mainly sealed by calcite cement they do not enhance permeability.

Indeed, the sealed joints form permeability barriers and effectively partition the host sandstone into unconnected compartments (Fig. 5.5). In the case of the ORS the main host for the fractures is the aeolian sandstone which is the only unit in the succession with reasonable matrix permeabilities.

### 5.3.iii Permeability zonation

In a sandbody unaffected by diagenetic alteration textural variations coincide with lithofacies variations, and therefore, with permeability variations. Thus, it follows that variations in gross permeability would occur vertically and laterally, as for example through an upwards fining channel fill sequence. However, as shown in chapter 4 diagenetic events have exerted a strong control on poroperm characteristics and largely overprint the textural control. As a result permeabilities are very similar for the different lithofacies with differences between lithofacies of only a few tens of millidarcies; so small as to be almost insignificant.

However, as mentioned in Chapter 4, there is a marked difference in permeabilities for similar lithofacies from the Upper and Middle ORS. Values for the Middle ORS are consistently lower than those in the Upper ORS despite similar porosities. A visual examination indicates that this is due to the average pore size in the Middle ORS being less than in the Upper ORS. This in turn appears to be a function of the smaller average size of the calcite crystals in the Middle ORS. The cause of this is probably the finer grain size in the Middle ORS and therefore the closer spacing of nucleation centres for calcite in the finer sediment resulting in smaller crystals.

It is very interesting to note that the aeolian sandstones from both the Middle and Upper ORS have very different poroperm characteristics despite being from similar depositional environments. The Upper ORS aeolian deposits have permeabilities significantly higher (around 200md) than any other facies while the Middle ORS aeolian sandstones have lower permeabilities (only about 20md). Aeolian deposits in general tend to be very clean, well sorted and texturally mature, and as such they have high initial porosities and permeabilities. This will result in cementing fluids readily entering them and occluding large proportions of the porosity, as has happened in the Middle ORS. With the Upper ORS however, the presence of early fractures (see above) has had a beneficial effect in that the cementing fluids have preferentially passed along the fractures instead of the host rock, thus leaving it less well cemented, and with higher porosities/permeabilities, although more compartmentalised.

## 5.3.iv Permeability baffles

Low permeability silts and impermeable mudstones are present to varying degrees in all depositional environments. These cause anisotropies in reservoir sands by greatly inhibiting fluid flow in the vertical direction, therefore knowledge of their lateral continuity and how they terminate within the sandbody is essential in describing the reservoir characteristics. Zeito (1965) examined rocks from several different environments (marine, deltaic and channel sands) and measured the length of the shale breaks and their frequency in an attempt to draw up some confining limits. Other workers have since added further data (Verrien *et al.,* 1967; Weber, 1982; Geehan *et al.,* 1986) and Fig. 5.6 combines all the data onto a single graph that relates the probability of a shale break extending over a certain length within a particular depositional environment. Shale breaks from different environments have markedly different characteristics and continuous layers, for example, are far more prevalent in marine sands than in channel sands. Even within channel sands the nature and discharge characteristics of the channel effect the thickness, abundance and lateral continuity of shale breaks.

The Middle and Upper ORS have very different distributions and frequencies of shale breaks as a result of the different depositional environments present. In the Middle ORS lacustrine deposits are common throughout, particularly in the lower part. These consist of mudstones, a few tens of cm's thick, that formed in short-lived

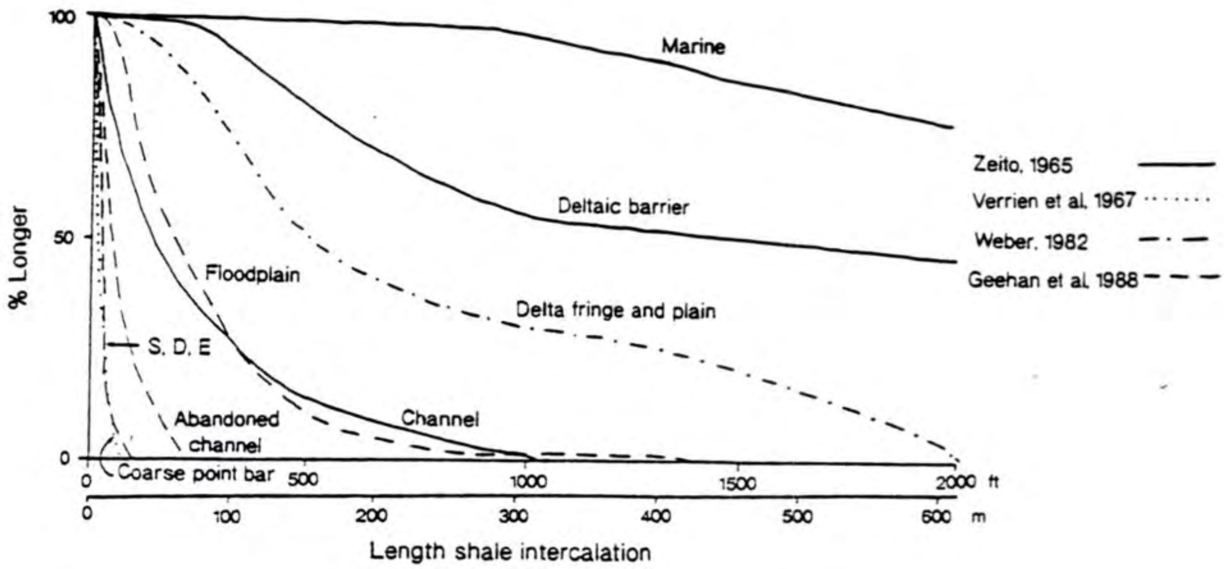



*Fig. 5.6. Observed shale length as a function of depositional environment. (S,D,E, = slough/drape/eddy facies of Geehan et al., 1988).*

*Fig. 5.7. Thin laterally continuous lacustrine muds from the Middle ORS. (Compass length = 8cm)*

ephemeral lakes (Fig. 5.7), and thicker (up to 1m around the inner Moray Firth) carbonate/clastic deposits that represent times of lake highstand (Fig. 5.8). These latter deposits are often associated with thick deposits of siltstone, especially nearer the centre of the basin (*eg.* the Caithness Flagstones). The lacustrine deposits are laterally very extensive with the highstand carbonates probably being deposited across the whole basin concurrently (the most extensive of these, the Achanarras Horizon, has been correlated across most of the basin). The continuity of these horizons means that the Middle ORS is effectively separated into a series of layers with little or no vertical connectivity which would have an enormous effect on reservoir modelling and any recovery methods used.

***Fig. 5.8.*** *Thick laterally extensive lacustrine laminites from the Middle ORS. (Hammer length = 40cm).*

The rest of the Middle ORS is dominated by sandy fluvial deposits with occasional thin mudstones which are frequently eroded by subsequent fluvial deposits to form mud chip conglomerates, similar to those described below. Additionally the interdune deposits from the Middle ORS aeolian sequences consist of interbedded

***Fig. 5.9.*** *Aeolian cross-bedding with muddy interdune deposits. From the Middle ORS of Easter Ross. (Hammer length = 40cm).*

***Fig. 5.10.*** *Thin laterally discontinuos fluvial mudstone from the Upper ORS of Easter Ross. (Hammer length = 40cm).*

siltstones and mudstones in packets up to 2m thickness (Fig. 5.9), with a lateral continuity of around several tens of metres.

The situation in the Upper ORS is notably different. Here lacustrine deposits are absent and fluvial and aeolian sands dominate. As can be seen in Fig. 5.6 shale breaks are much less continuous in fluvial environments than in most others. The Upper ORS is approximately 95% sand and the only shale breaks present are within the fluvial sandstones. These are overbank deposits and consist of red muds, <50cm thick with a lateral continuity of only 15-20m (Fig. 5.10). They all terminate laterally as a result of erosion by superseding sandbodies, and it is this erosion and reworking that has caused the paucity of mudstones, as is shown by the common presence of mud chip conglomerates. The effect of these shale horizons on poroperm characteristics is going to be very minimal because of their lateral discontinuity and the lack of terminal convergence. The aeolian sands are very clean and associated muddy interdune deposits have not been observed (Fig. 5.11).

***Fig. 5.11.*** *Upper ORS aeolian sands consisting of dune cross-bedding and sandy horizontal laminations. There are no muddy interdune deposits, unlike the Middle ORS in Fig. 5.9. (Rucksac = 80cm).*

Allen (1978) used a quantitative model to examine the architecture of avulsion controlled alluvial suites. During this study he found that in alluvial suites containing more than 50% fines that the sandbodies were virtually unconnected, whereas the degree of connectivity grew rapidly as the proportion of sand increased above 50%. The Upper ORS contains about 95% sand and as a result has a very high degree of connectivity Although the Middle ORS contains about 70% sand, the fines are concentrated into the extensive lacustrine horizons (a phenomena not accounted for in Allen's model) which means that the degree of vertical connectivity is zero, while the horizontal connectivity will be reasonable. Thus, there is strong gross anisotropy present in the Middle ORS. Overall the Upper ORS is much more homogeneous than the Middle ORS and the sandbodies have far higher connectivity.

## 5.3.v Lamination and cross-bedding

Cross-bedding imparts a major heterogeneity in sandstones and the ORS is no exception to this, given the large proportion of cross-bedded sands that it contains. Heterogeneities associated with sedimentological features form as a direct result of depositional processes, although diagenesis can greatly enhance the effects of these heterogeneities (Weber *et al.,* 1972). In general, the effects of cross-bedding on permeability in both recent and ancient sediments are fairly well known, largely as a result of the work done by Weber *et. al.* (1972), Weber (1982), Pryor (1973) and Pryor & Fulton (1978).

Weber *et al.* (1972) concluded that the maximum permeability direction is along the direction with the least bottom-sets per unit length (*ie.* parallel to the channel axis) because these are usually of low permeability due to their association with high concentrations of mud and silt. In fluvial systems the sets tend to be preferentially aligned and elongate parallel and subparallel to the flow direction (Fig. 5.12). Pryor (1973) also found that there was a directional permeability parallel to the length of sets and increasing down the dip of the foresets (Fig. 5.13).

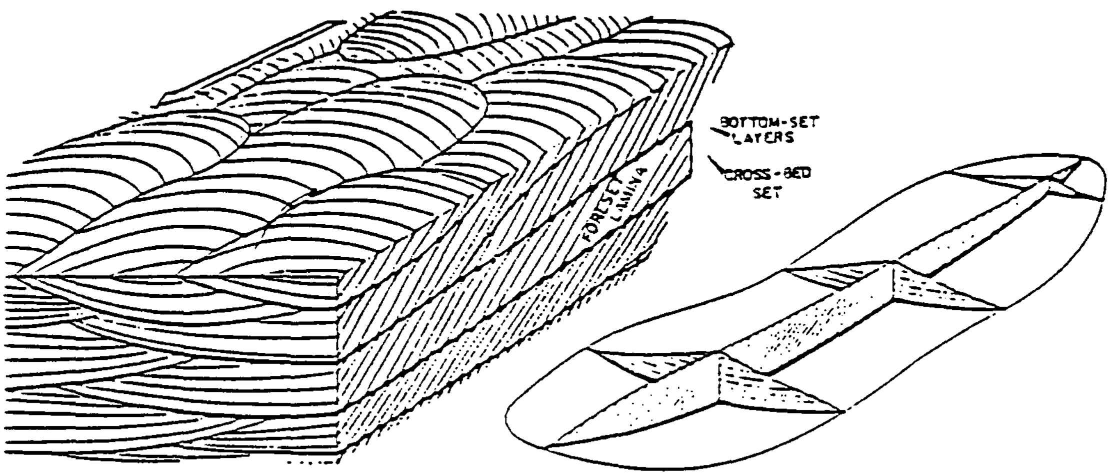



*Fig. 5.12. Block diagram of a trough cross-bedded sand body showing typical internal geometry and the principle fluid flow direction (arrow). (From Weber et al., 1972).*

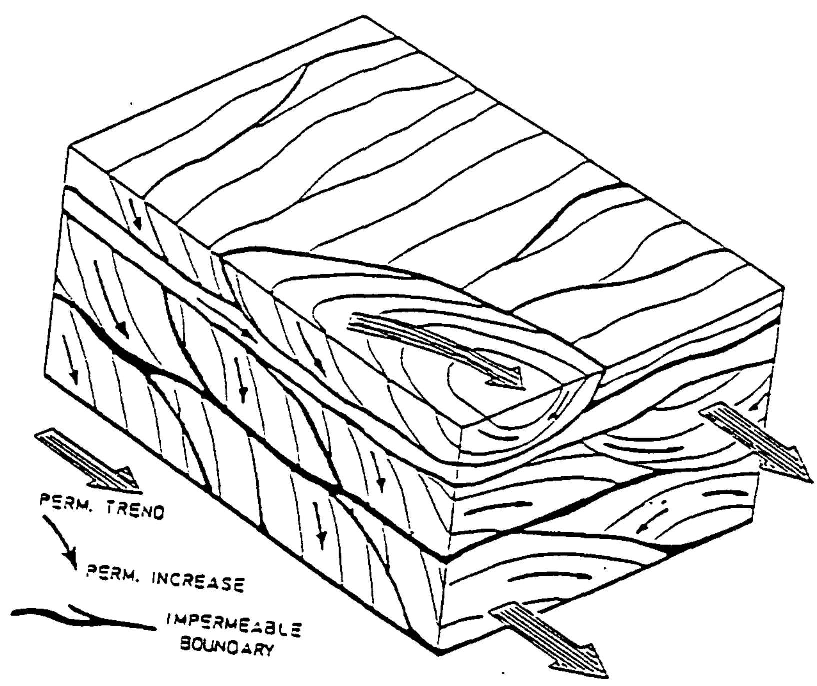



*Fig. 5.13. Block diagram showing typical permeability trends within a river bar sand body (From Pryor, 1973).*

Weber (1982) shows a method whereby the permeability anisotropy of a cross-bed set can be calculated using measurements obtained with a minipermeameter. Initially the intention was to carry out a similar study on cross-bed sets throughout the ORS around the inner Moray Firth, but problems arose in using the minipermeameter in the field (see Chapter 4) and the idea had to be abandoned. However, some useful general points can still be made based on field studies. The proportion of mud associated with the sands is very low and clay drapes are not present along foresets (Fig. 5.14) thus reducing any permeability anisotropy. The only observed exception to this was in the Middle ORS where some rippled fine sandstones have mud infills to their troughs. Also bottom-sets appear to be very thin and insignificant which further reduces their effect on permeability. These poorly developed bottom-sets suggest that the suspension load was poorly developed relative to the bedload.

***Fig. 5.14.*** *Fluvial trough cross-bedding from the Upper ORS of Easter Ross. Note the lack of muddy bottom sets along set boundaries. (Hammer length = 40cm).*

Variations in set thickness resulting from the effects of variables such as erosion and water depth also effect heterogeneity in that a decrease in set thickness will bring the bottom-sets closer together thereby increasing the permeability anisotropy. For the fluvial cross-bedded sandstones the average set thicknesses for the Upper ORS are 50-60cm, whilst in the Middle ORS they are only half this thickness suggesting an even greater anisotropy.

The aeolian cross-bedded sands act as a much more homogeneous unit because they are cleaner sands and do not have muddy bottom-sets, although grain fall deposits are more common along the base of foresets, and because changes in grain size across set boundaries will probably have a slight effect. In particular the aeolian sands present in the Upper ORS have no muddy interdune deposits associated with them either, leading to the development of thicker bodies of sand. In contrast, Middle ORS dune sands are interspersed with mudstones and siltstones (see above).

Illing (1939) stated that the most difficult zone to flush with water is the coarse/fine interface. Thus, the more textural changes within a rock body the harder it will be to extract any oil. As textural changes occur with changes in lithofacies, this will have more serious consequences for the Upper ORS where fluvial/sabkha and aeolian/sabkha interbedding occurs on a scale of usually less than 2m, leading to a rapid interbedding of lithofacies, thus significantly enhancing the heterogeneity (Fig. 5.15).

Horizontally laminated and rippled sands will also have strong permeability anisotropies associated with them. Laminated sands will have a much greater horizontal permeability than vertical permeability as a result of the textural changes resulting from the stacking of several thin beds. Additionally the long axes of the grains are parallel to the palaeo flow direction which will act to increase this horizontal anisotropy. Rippled sands will have a similar anisotropic effect to that of cross-bedded sands except on a smaller scale, and thus will have a more detrimental effect. Both rippled and laminated sands and silts are fairly common in the Middle ORS, whereas they are much less frequent in the Upper ORS.

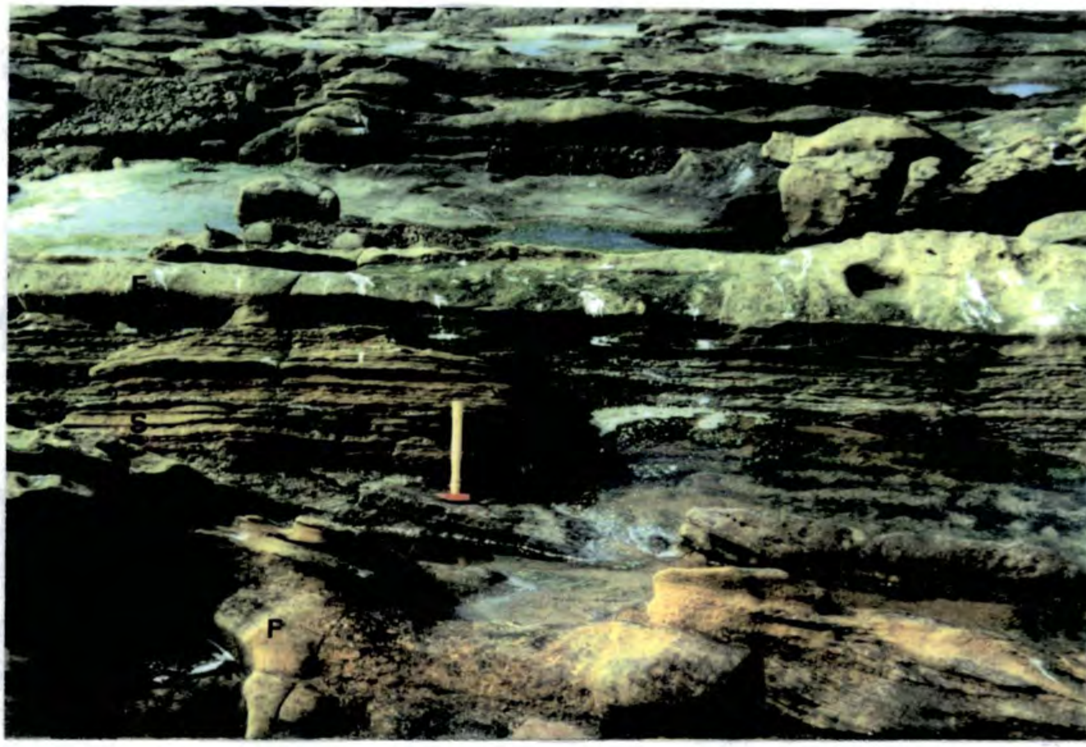



*Fig. 5.15. Interbedded sabkha (S) and fluvial cross-bedded (F) sands leading to frequent occurence of bounding surfaces. (Hammer length = 40cm).*

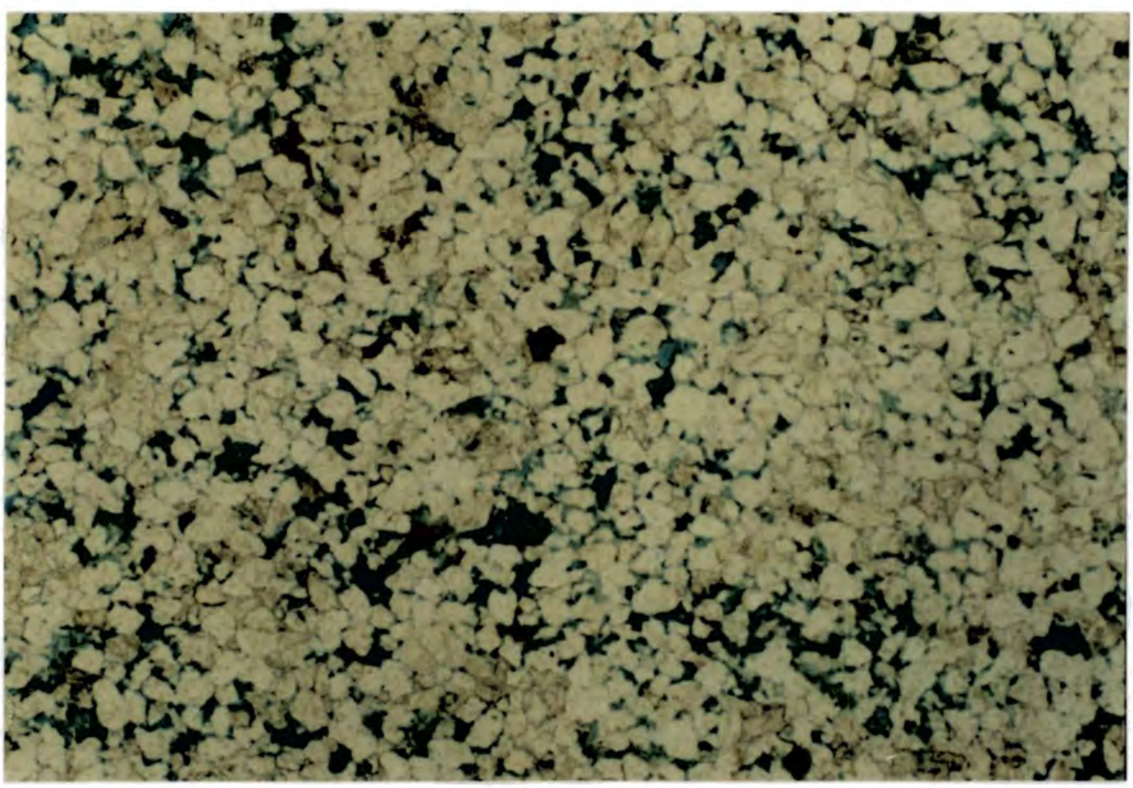


*Fig. 5.16. Uneven distribution of calcite cement causes patchy poorly connected porosity. (PPL; FOV = 6mm).*

## 5.3.vi Microscopic heterogeneities

Microscopic heterogeneities are those associated with rock texture and diagenetic events. In the previous chapter it was shown that porosity and, in particular, permeability were controlled by the diagenesis and not by texture, therefore any heterogeneities will be associated with the distribution of diagenetic minerals and variations in compaction.

There are two causes of heterogeneity at the microscopic scale. The first cause is the dominant calcite cement. This is patchy in distribution (Fig. 5.16) and therefore the porosity is also patchy, which will result in poorly or unconnected areas of porosity forming cul-de-sacs which will inhibit oil recovery. The second cause is the blocky, pore filling kaolinite (Fig. 5.17). Although this has microporosity associated with it, the small size of this porosity will effectively seal off the macropores containing the kaolinite from the main pore network. Despite this the kaolinite is of minor importance since in most samples studied it is present in amounts of less than 8% by volume, below which it has no effect on permeability (see Chapter 4).

***Fig. 5.17.*** *Backscattered scanning electron micrograph of a kaolinite filled pore illustrating microporosity. (Scale bar = 50μ m).*

The patchy nature of the calcite is also evident on a slightly larger scale. As can be seen in Fig 5.18 there are cul-de-sacs similar to those seen in thin section, except that they are a few metres in scale Heterogeneities such as these are typically chaotic and random and as such are very difficult to characterise quantitatively in reservoir modelling (Lasseter *et al.,* 1986).

***Fig. 5.18.*** *Fluvial sands with irregular distribution of intensive calcite cementation (light areas) forming cul-de-sacs which would act to trap oil. (Hammer length = 40cm).*

## 5.4 Summary

Knowledge of the type, distribution, and influence of heterogeneities within a reservoir is critically important to its development. Modelling of heterogeneities requires a combination of cores, logs, production tests, and outcrop analogues, and because of the large amount of data input computer aided modelling is now common (*eg.* Haldorsen & Lake, 1982; Martin and Cooper, 1984; Johnson & Krol, 1984; Lasseter *et al.,* 1986).

The effect of different heterogeneities on fluid flow characteristics and oil recovery is varied, but is almost always detrimental. This is certainly the case in the ORS. At the microscale diagenetic heterogeneities have greatly reduced porosity and permeability to virtually insignificant levels. The effect of cross-bedding has been to exert a directional anisotropy parallel to the length of the sets, along the flow direction. Additionally, the Middle ORS is separated into discrete segments from 10-100m thick by laterally extensive lacustrine deposits. This has resulted in there being virtually no vertical connectivity between sandbodies.

The Upper ORS has a much higher sandbody connectivity as a result of its greater sand content (about 95%, well above the 50% cut off suggested by Allen, 1978, above which sandbody connectivity rapidly increases). However, it suffers from presence of a large number of sealed joints which compartmentalise the sandbodies. The only heterogeneity to enhance the reservoir characteristics are the micro joints, although these are so sparse that the detrimental effects of the other heterogeneities far outweighs their positive influence. Overall the presence of the heterogeneities is very detrimental to the reservoir potential of the ORS.

## Chapter 6

# Conclusions and Discussion

### 6.1 The Old Red Sandstone lithology and sedimentology

Old Red Sandstone deposition in North East Scotland occurred during Devonian times within the Orcadian Basin, a northeast-southwest orientated structure. The basin formed in an extensional tectonic regime that resulted from gravitational collapse of over-thickened Caledonian crust. Pure shear in the lower crust caused an overall thinning of the crust with a surface expression of listric extensional faulting and basin development controlled by a Caledonian anisotropy in the middle and upper crust.

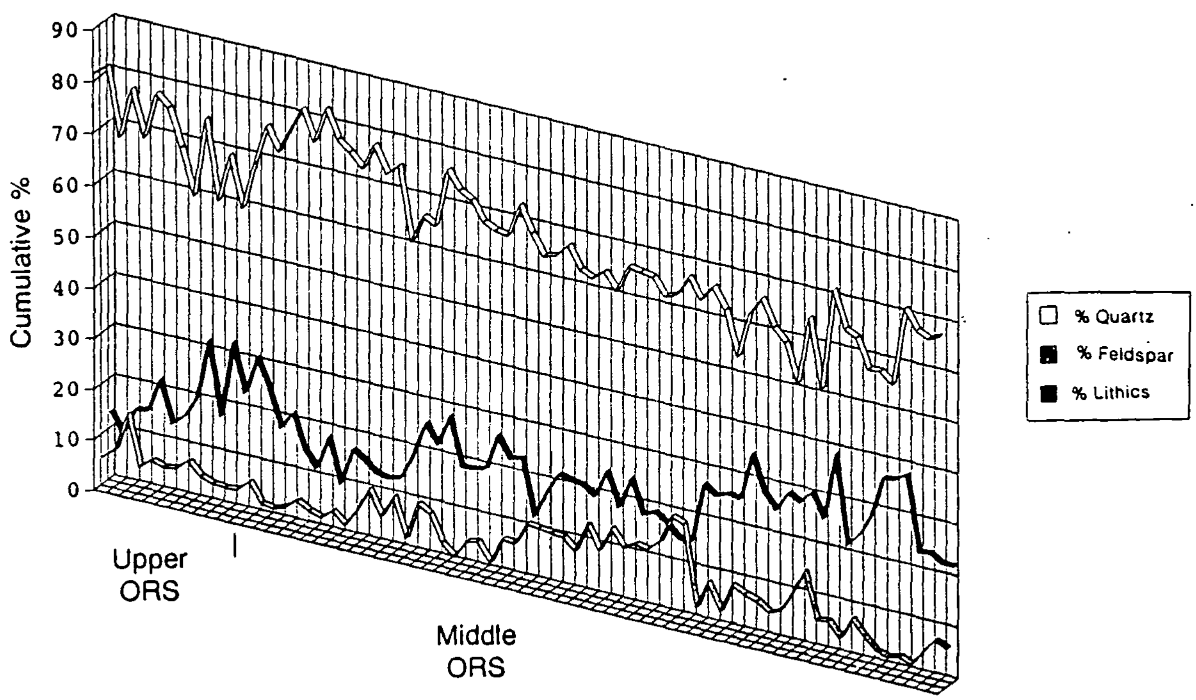



*Fig. 6.1. Diagram illustrating the relationship between mineralogical maturity of the Middle and Upper ORS sandstones and stratigraphy, from the Inner Moray Firth.*

The ORS sandstones are texturally and mineralogically sub-mature to mature sublitharenites. They are sourced from Dalradian and Moinian metamorphic highlands intruded by granites, which occur over much of northern Scotland, to the south and west of the Orcadian Basin. Compositional maturity increases up the succession, probably as a result of sediment recycling (Fig. 6.1). Sorting is very consistent throughout the ORS with the sediments being on average well sorted, while the grain size shows a slight increase in the Upper ORS compared to the Middle ORS (Fig. 6.2).

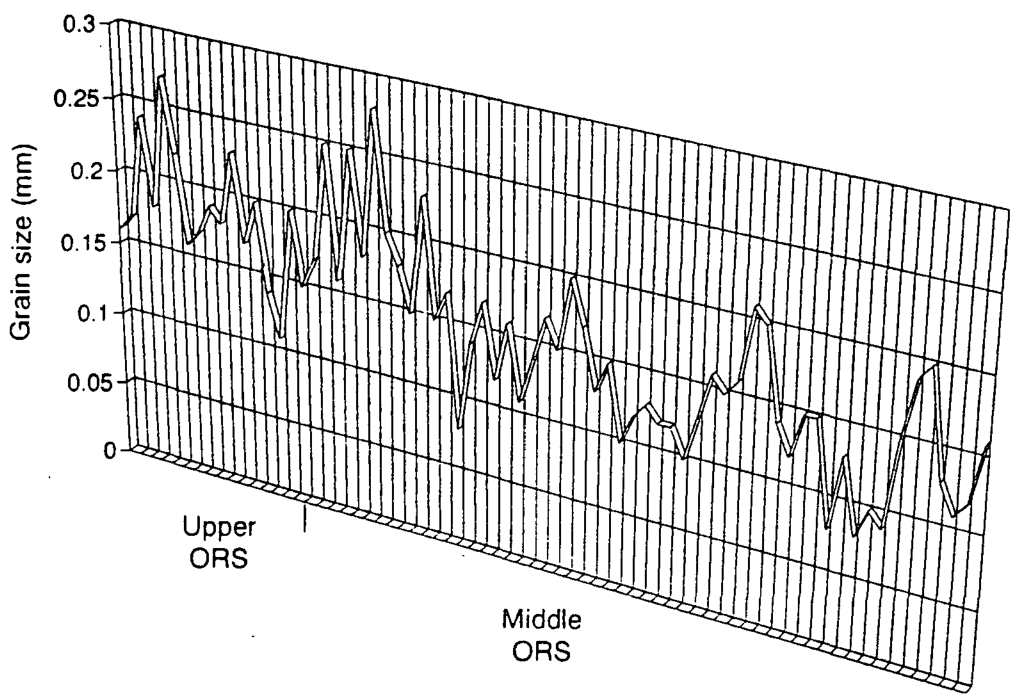



*Fig. 6.2. Diagram illustrating the relationship between grain size and stratigraphy for the Middle and Upper ORS around the Inner Moray Firth.*

Grain size distribution plots show that there is a shift from a fine grain skewness in the Middle ORS to a more symmetrical distribution with a coarse tail in the Upper ORS. This change in grain size distribution and the increase in average grain size at the Middle/Upper ORS boundary are as a result of increased tectonic activity, which caused uplift of the source area and progradation of coarser clastics further into the basin.

The ORS sediments comprise five main depositional environments: alluvial fan; sandy fluvial; lacustrine; aeolian; and sabkha. The Lower ORS deposits are dominated by planar and current rippled medium to fine grain sandstones, organic rich siltstones and mudstones and dolomicrites. These represent floodplain and lacustrine deposits. Alluvial fan conglomerates occur along the lines of active synsedimentary faults.

Following the initiation of half graben faulting Lower ORS deposition occured in small, unconnected, locally sourced, fault bounded basins.

The beginning of Middle ORS times saw the onset of basinwide deposition for the first time, as opposed to the small, unconnected basins present during the Lower ORS. The dominant environments represented are sandy braided fluvial and lacustrine with marginal conglomerates. The fluvial deposits consist of approximately equal proportions of trough cross-bedded and horizontally laminated sands with minor amounts of ripple cross-laminated sands and silts. These represent part of a braided fluvial system and sheet floods. Lacustrine deposits occur in two forms, clastic muds or organic-rich carbonate/clastic laminites. The former represent short lived, shallow temporary lakes and the latter long lived, deeper water, more permanent lakes. Aeolian deposits are present in the form of dune cross-bedding, which represent the remnants of barchanoid dunes, and interdune deposits.

The Upper ORS consists predominantly of trough cross-bedded and planar bedded, often pebbly sandstones with soft sediment deformation, deposited within a braided fluvial system. Aeolian deposits are present as either compound barchanoid dunes, as in the Middle ORS, or as smaller isolated barchan dunes which frequently occur interbedded with fluvial sands or horizontally bedded evaporitic sabkha sands. Sabkha deposits are also interbedded with fluvial cross-bedded sands. The Upper ORS differs from the Middle ORS in that the lacustrine environment is replaced by the evaporitic siliciclastic sabkha which occurs distal to the fluvial deposits, and the fluvial sandstones are of a more proximal nature.

## 6.2 Controls on the reservoir potential of the ORS

### 6.2.i Poroperm controls

Porosities from the ORS are much lower than those from similar Holocene sandbodies and artificial grain packs, ranging from 0-14.3% with a mean of 6.2%.

Since textural parameters, grain size and sorting show no correlation with porosity, diagenetic changes, principally compaction and cementation must be the dominant controls on porosity. Both compaction and cementation have acted to reduce the porosity to levels well below its primary depositional value of about 40%. Compaction (predominantly grain rearrangement as opposed to grain interpenetration) has destroyed 30-39% of the original porosity, while cementation has destroyed 46-50%. The cement responsible for this is a poikilotopic framework supporting calcite cement which was precipitated during early diagenesis as indicated by the low levels of compaction.

During inversion in the late Carboniferous meteoric flushing caused dissolution of the calcite cement which lead to a slight increase in macroporosity. However, kaolinite formed at the same time and occluded much of this porosity. Micro fratures have also contributed a small amount of porosity, but this is too small to be significant.

Permeability is also much lower in the ORS than in recent sandbodies, sometimes by several orders of magnitude (14-270 md as opposed to <200 darcies). Since permeability is closely related to porosity the controlling parameters are essentially the same. Thus, permeability shows no relationship with grain size or sorting, but has a clear correlation with the proportion of calcite cement. The presence of authigenic kaolinite only has a detrimental effect on permeability when present in amounts greater than 8% by volume.

## 6.2.ii Heterogeneities within the ORS

Heterogeneities within a rock body play an important part in controlling its fluid flow characteristics, and thus its suitability as a reservoir rock. Heterogeneities within the ORS include fractures, shale breaks and diagenetic changes. These occur on a variety of scales from the microscale to the megascale.

Fractures occur on two scales: micro and macro. Micro fractures are 0.1-0.3mm wide by a few cm's long and are usually open, thereby increasing porosity and permeability. However, fractures are relatively scarce and their overall effect on poroperm values is negligible. Macrofractures or joints are far more common especially in the more homogeneous sand dominated Upper ORS. However, the majority of these appear to be cemented with calcite and, rather than acting to increase porosity and permeability, they tend to compartmentalise the sandbodies thereby rendering them almost useless as reservoirs.

Another major heterogeneity within the ORS are shale breaks. These are most important in the Middle ORS (Fig. 6.3), where there are frequent, laterally extensive lacustrine deposits, some of which cover the entire Orcadian Basin. These horizons mean that there is virtually no vertical connectivity between sandbodies which will inhibit any vertical movement of fluids. The Upper ORS has a much higher degree of connectivity, with the only shale breaks being thin laterally discontinuous overbank mudstones (Fig. 6.4) which impart only a slight anisotropy to the sandbodies.

On a smaller scale heterogeneities occur associated with sedimentary structures. Trough cross-bedding, which dominates the Upper ORS in particular, has a maximum permeability parallel to the palaeocurrent direction, which in general is to the north-east in the Orcadian Basin. The Middle ORS has a higher proportion of horizontally laminated and rippled sands which will have strong permeability anisotropies associated with them, with horizontal permeabilities being much greater than vertical permeabilities.

As mentioned above, microscopic heterogeneities are a result of diagenetic changes rather than primary depositional textural characteristics. The widespread calcite cement has not only greatly reduced total porosity levels, but because of its patchy distribution, the porosity is only poorly interconnected, thereby reducing the effective porosity and the permeability even further. Pore filling kaolinite cement

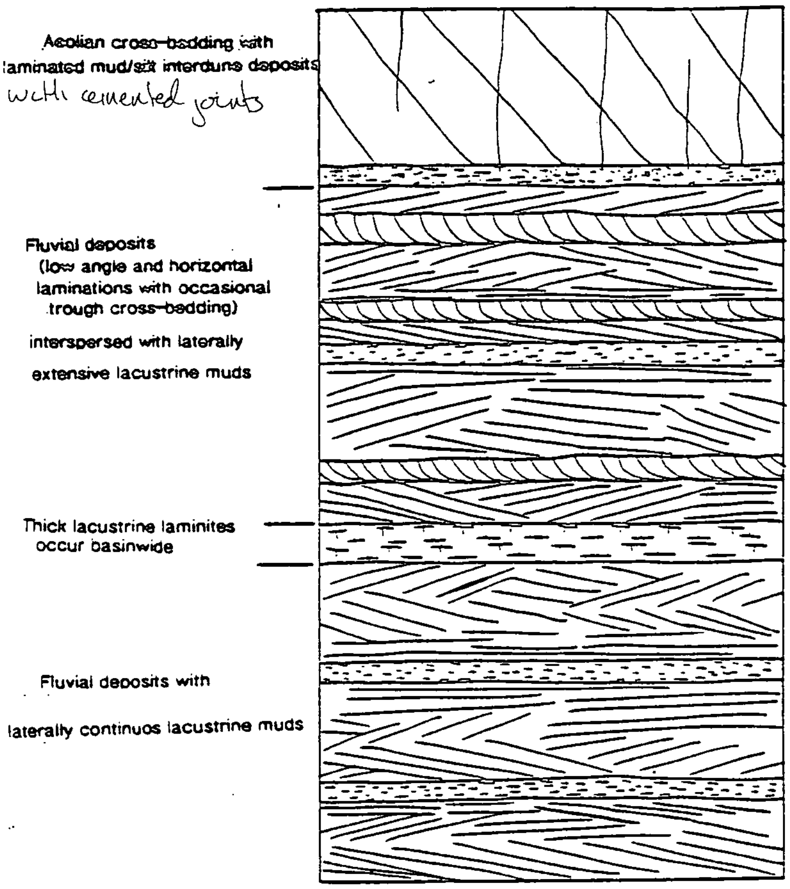



*Fig. 6.3. Cartoon summarising the heterogenities from the Middle ORS.*

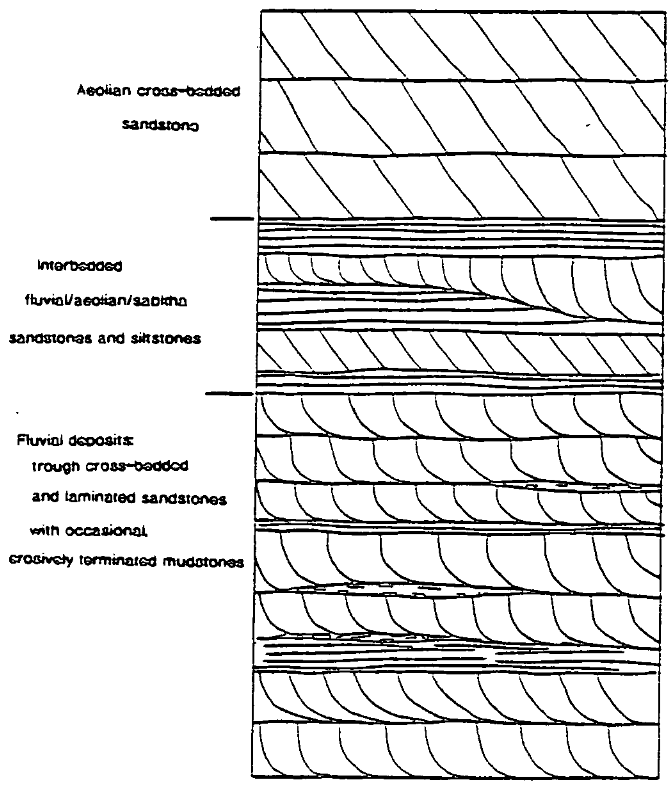



*Fig. 6.4. Cartoon summarising the heterogeneities from the Upper ORS.*

which has also reduced porosity and pore size has a detrimental effect on permeability.

## 6.3 Source rock potential and hydrocarbon generation in the Orcadian Basin

Residual hydrocarbons have been recognised in joints and veins from around the Moray Firth for well over one hundred years (Witham, 1825; Anderson & McKenzie, 1863; Barron, 1883: Morrison, 1883, 1888b, 1889). It was also recognised that Devonian organic-rich argillites were the probable source of these hydrocarbons (Murchison, 1859; Morrison, 1888a). Organic rich Lower ORS shales were described by Horne and Hinxman (1914) from around the southwestern Orcadian Basin and Hall & Douglas (1983) showed these to have significant organic contents dominated by oil prone amorphous organic matter (probably of algal origin). Middle ORS laminites have also been shown to contain oil prone organic matter derived from an algal source (Donovan, 1978; Hall & Douglas, 1983; Douglas *et al.*, 1983).

Several workers have recently published work on the organic matter content and thermal maturation levels of potential hydrocarbon source rocks from the Orcadian Basin (Hamilton & Trewin 1985; Parnell, 1985; Marshall *et al.,* 1985; Irwin & Meyer, 1990; Hillier & Marshall, 1992). These studies all indicate that total organic contents of the lacustrine facies are commonly around 0.5% although values of between 1% and 5% are not unusual, especially in Caithness, where Marshall *et al.* (1985) consider some 20% of the flagstone succession to have in excess of 1% organic matter, and around the Moray Firth. Marshall *et al* (1985) found that the organic matter is mainly Type II and Type IIIA kerogen (classification of Tissot and Welte, 1978) which would produce high wax oils and gas (Fig. 6.5).

The maturation level has been measured in a variety of ways with the most detailed study being that of Hillier and Marshall (1992). Maturation levels measured at outcrop throughout the Orcadian Basin show great variation. In the Moray Firth area the highest maturities (vitrinite reflectances ($R_0$) <10%) are along the line of the Great Glen fault. Elsewhere in the region $R_0$ values range from 0.5% to 2.4%, with an average of about 1%. Southern and central Caithness have high maturities (>3% $R_0$), while in northern Caithness maturities are much lower (0.6-1.9% $R_0$).

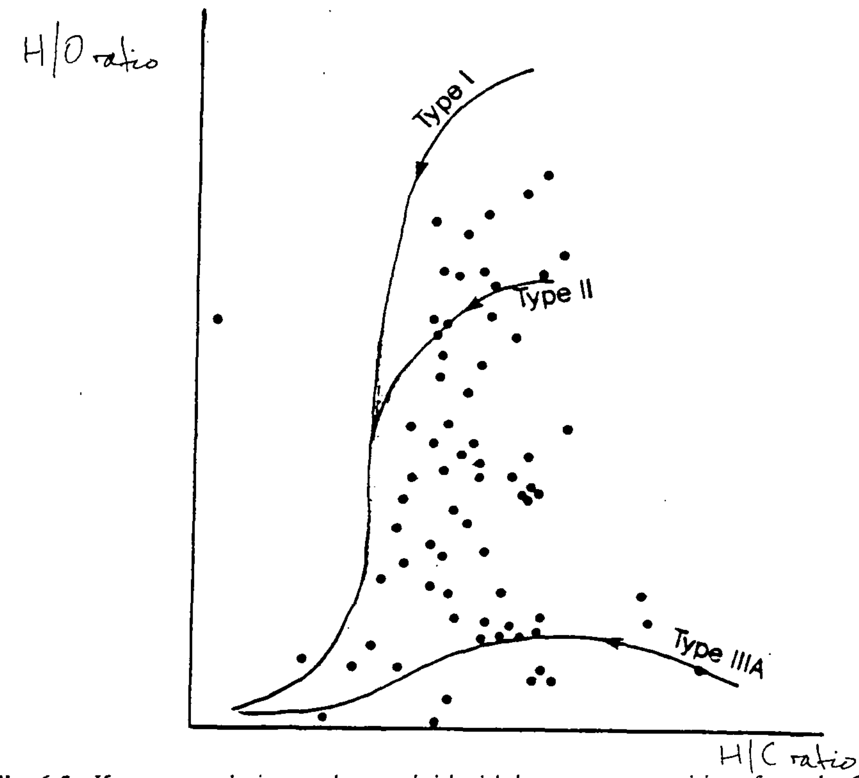



*Fig. 6.5. Kerogen evolution paths overlaid with kerogen compositions from the Orcadian area. Kerogens range from oil prone Type II's to gas prone Type IIIA's (After Marshall, et al., 1985).*

Vitrinite reflectance values from Orkney show little variation (1-1.5%) indicating a more uniform level of maturity. It is important to note that there is no link between thermal maturity and stratigraphy in Caithness and Orkney (Hillier & Marshall, 1992). The maturity values from Shetland fall into two groups on either side of the Melby Fault. To the west of the Melby Fault vitrinite reflectance values are between 0.7 and 1.7% $R_0$, while to the east of the Melby Fault values range from 3.2 to 8.9% $R_0$.

In terms of oil generation scales the wide range of vitrinite reflectances are equivalent to near the base of the oil window (approx. 0.5% $R_0$) to beyond the dry gas zone (3-4% $R_0$) (Heroux *et al.*, 1979). Areas with high $R_0$ values in excess of about 4% must have reached temperatures in the region of 250-350°C. In the Walls Basin on Shetland the Sandsting Granite (a Late Devonian intrusion) is an obvious cause of these high temperatures. Magnetic anomalies under southeast Shetland (interpreted as basic intrusions by Wilson, 1965) and an exposed metamorphic aureole on Fair Isle to the south of the mainland (Mykura, 1972) suggest that contact metamorphism related to intrusions caused the heating of the sediments.

In Caithness there is no direct evidence of any contact metamorphism. Thus, if the maturation levels were to be explained by burial alone a very deep and complex burial history would be required. However, vitrinite reflectance values similar to those of Shetland, and the presence of a large magnetic anomaly below Wick which has been interpreted as an intrusion at depth (Flinn, 1969), have led to the suggestion of the presence of a 'Caithness Granite' at some 1-2km depth (Hillier & Marshall, 1992). This granite intrusion could then have metamorphosed the adjacent sediments.

The other area of high maturities along the Great Glen fault is very much restricted to the line of the fault zone. Given the presence of highly sheared and metamorphosed rocks, including ORS sediments, between Foyers and Fort William (Eyles & MacGregor, 1952) Hillier & Marshall (1992) have suggested that the zone of high maturities is a northwards extension of the metamorphism and probably related to post Devonian strike slip movements along the Great Glen fault.

The lower levels of maturity in the remainder of the basin will be as a result of burial maturation with the maximum temperatures reached being around 100-120°C (Barker & Pawlewicz, 1986). The thickness of late Devonian to Carboniferous sediments, since removed, that caused the organic rich facies to enter the oil window is uncertain. Trewin (1985) estimates that 1km of sediment is missing based on Middle ORS sediment thicknesses proposed by Donovan *et al.,* (1974). However recent work (*eg.* Astin, 1990; Rogers, *pers. comm.* 1992) indicates that the thickness of the Middle ORS sediment pile should be reduced and Hillier & Marshall (1992) have suggested that as much as 3km of sediment has been removed.

The Orcadian Basin, consisting of a series of half-grabens filled with lacustrine and alluvial sediments can be classed as a continental rift basin (Enfield & Coward, 1987; Rogers, 1987; Astin, 1990). Generally rift basins have high but variable geothermal gradients (Robbins, 1983) and this is shown by the Mid to Late Devonian igneous intrusions. This generally high heatflow and contact metamorphism associated with the intrusions combined with the maximum depth of ORS burial occurring at the end Devonian/Carboniferous would have caused an early maturation event leading to the generation of hydrocarbons in the late Devonian (Parnell, 1985; Trewin, 1985; Hillier & Marshall, 1992). Evidence for this is seen on Orkney where Permian dykes have altered hydrocarbons already reservoired in the sandstone (Astin, 1990). The dykes themselves had little effect on maturation levels (Marshall *et al.,* 1985).

As a result of early cementation by calcite, especially in the sandstones associated with the organic rich laminites, porosities and permeabilities were very low, thereby restricting the migration of any hydrocarbons produced at this time. Another problem is the lack of good reservoir rocks within the Devonian. As this study demonstrates there are no suitable reservoirs around the Inner Moray Firth, and the situation elsewhere within the Orcadian Basin is little better with only some aeolian dune sandstones from Orkney (Yesnaby Sandstone, Lower ORS), Eday Group sandstones (Middle ORS) and the Upper ORS of Hoy and Dunnet Head showing solid hydrocarbons (Astin, 1985; Parnell, 1985a). Aeolian sandstones from Shetland occurring near to laminite horizons on Shetland are the only other possible reservoirs (Mykura, 1976; Allen & Marshall, 1981; Astin in Marshall *et al.,* 1985).

It is interesting to note that not only are the highest maturity areas all near to the Great Glen/Walls Boundary fault zone, but solid hydrocarbon occurrences as fracture and vein fills, and in vugs are commonly associated with fracture systems and igneous intrusions (Parnell, 1983, 1985). This would suggest that the fault systems provided a conduit for both igneous intrusions and hydrocarbons.

The Variscan inversion event would have had two effects. Firstly it would have halted hydrocarbon generation and secondly it would have either redistributed reservoired hydrocarbons or exposed them to the degrading effects of meteoric waters. Permian and Mesozoic sedimentation was only very thin over the areas of onshore outcrop (Andrews *et al.,* 1990; Hall, 1991) and would have been insufficient to cause a further period of hydrocarbon generation. However, further east sedimentation rates increased such that in the offshore regions as much as 4km of sediments accumulated in some of the Mesozoic half grabens, mainly in the late Jurassic to early Cretaceous (Andrews *et al.,* 1991). This period of burial will have produced hydrocarbons from the ORS source rocks which had the lowest maturities at the end of the Palaeozoic (Hillier & Marshall, 1992).

As well as causing a second period of hydrocarbon generation Mesozoic sedimentation provided plenty of potential reservoir rocks. Duncan & Hamilton (1988) suggested that the oil in the Jurassic age Beatrice Field in the Outer Moray Firth was at least partly sourced from the Devonian. Since then these results have been confirmed by Peters *et al.* (1989), who estimated a Devonian component of 60%, and Bailey *et al.* (1990) who suggest that the Beatrice Field is entirely sourced from Devonian lacustrine sediments.

## 6.4 Future exploration areas

It is unlikely that any of the hydrocarbons reservoired during Devonian times will be preserved *in situ* because of losses from fault leakage or degradation by meteoric waters during the Variscan inversion event (Marshall *et al.,* 1985; Trewin, 1985). As a result the most promising sites for future exploration are offshore within the Moray Firth in areas where the three following criteria are met: (1) A thick sequence of oil prone lacustrine sediments; (2) These source rocks must have had low thermal maturities at the end of the Palaeozoic; and (3) Sufficient Mesozoic sedimentation to initiate hydrocarbon generation.

The Lower ORS deposits are restricted in distribution to localised basins and have only been found in two areas offshore (Richards, 1985; Andrews *et al.,* 1991) (Fig 6.5). Norton *et al.* (1987) consider it likely that the centres of Mesozoic deposition were the same as those where Devonian rifting began, therefore it is likely that many of the younger basins will have Lower ORS at great depth in them. Middle ORS lacustrine deposits similar to those found in Caithness and Orkney probably occur widely across the Moray Firth, to the east and north of the onshore outcrop (Mykura, 1983), and have been found in several places offshore (Andrews *et al.,* 1991) (Fig. 6.6).

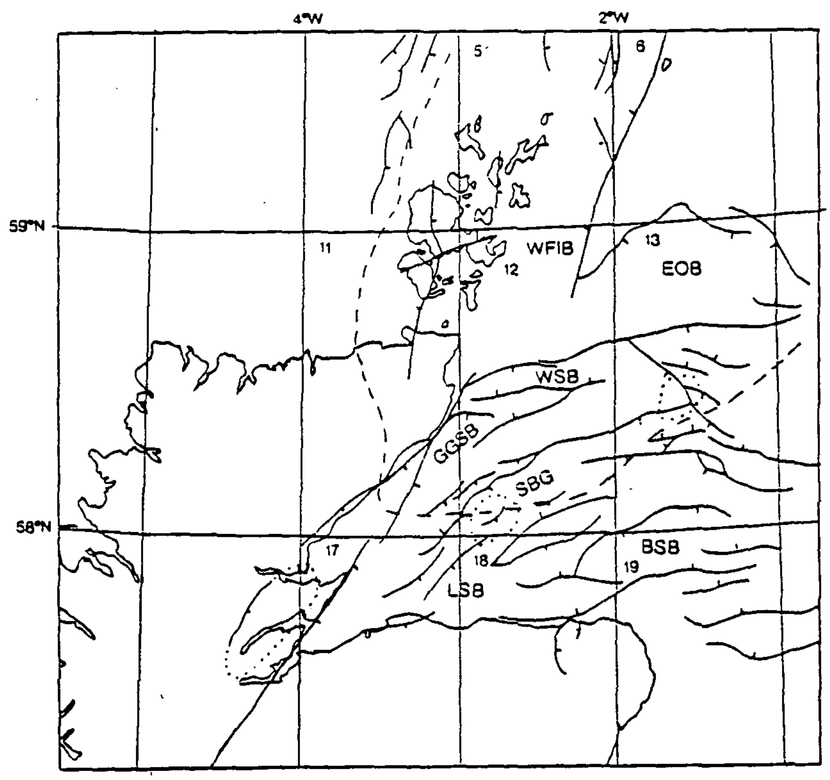



*Fig. 6.6. Map of the Moray Firth showing areas with potential Devonian reservoirs. See text for discussion. Dotted line = proven extent of Lower ORS source rocks, dashed line = possible extent of Middle ORS source rocks; WFIB = West Fair Isle Basin, EOB = East Orkney Basin, WSB = Wick Sub Basin, GGSB = Great Glen Sub Basin, SBG = Smith Bank Graben, LSB = Lossiemouth Sub Basin, BSB = Banff Sub Basin.*

As regards the second requirement, thermal maturities at or near the base of the oil window at the end of the Palaeozoic, the high levels of maturity seen onshore are all associated with magmatism and low levels of metamorphism around fault zones, it is therefore highly likely that large areas of the basin away from the Great Glen/Walls Boundary fault systems will have maturities similar to the lower levels onshore: near to the base of the oil window (Hillier & Marshall, 1992).

To satisfy the third requirement at least 2km of Mesozoic burial are necessary. Three main areas in the Moray Firth had sufficient levels of sedimentation: the Great Glen Sub Basin, the Wick Sub Basin and the Smith Bank Graben. By the end of the Cretaceous the Great Glen Sub Basin had undergone over 4km of burial while the Wick Sub Basin and the Smith Bank Graben had undergone about 3½km. The bulk of this sedimentation occurred during the late Jurassic and early Cretaceous, with sufficient depth of burial to cause hydrocarbon generation during the Middle to Upper Jurassic.

Areas further north with good source rock potential *e.g.* the West Fair Isle Basin and the East Orkney Basin have not undergone sufficient Mesozoic burial to generate hydrocarbons. In the south of the Moray Firth, where Mesozoic sedimentation reached levels of around 2km, the Devonian sediments are more proximal in nature and source rocks are largely absent.

The best reservoir properties are shown by the Upper ORS. This about 1km thick, although this will vary according to the amount of Variscan erosion, which is generally lower to the east, and is about 95% sandstone (Fig. 6.7). Permeabilities are typically 15-30md, (sometimes >200md) with porosities typically <10%. In this more distal area the Upper ORS will consist of interbedded sabkha, fluvial and aeolian sandstones in approximately equal proportions, typically in units 1-5m thick. The aeolian dune crossbedding, which has the highest permeabilities at 100-200md locally occur in sequences 12-15m thick.

The Middle ORS while being thicker at around 2km, has lower porosities (typically 5-10%) and permeabilities (<50md) in the more proximal deposits studied. However, further into the basin, these marginal deposits are not present, and are replaced instead with sequences of flagstones (interbedded mud and silt/fine sand) around 10m thick, permanent lake organic rich carbonate/clastic laminites (the potential source rocks) at 1-2m thick and playa lake clastic mudstones (<1m) (Fig. 6.8).

Potential traps within the ORS will most likely be in sandstone accumulations banked against irregular basal unconformities (*eg.* the Yesnaby sandstone) (Trewin, 1985) and in fold structures *eg.* rollovers adjacent to faults. It is also possible that Devonian sourced hydrocarbons are reservoired in Mesozoic sandstones in fault traps, rollover structures or fault block drapes. The area of the Moray Firth that best fulfils all the above requirements covers the southeastern corner of Quadrant 11, the southern half of Quadrant 12 and the southwestern corner of Quadrant 13 (Fig. 6.6), in particular the Great Glen Sub Basin, the Wick Sub Basin and the Smith Bank Graben.

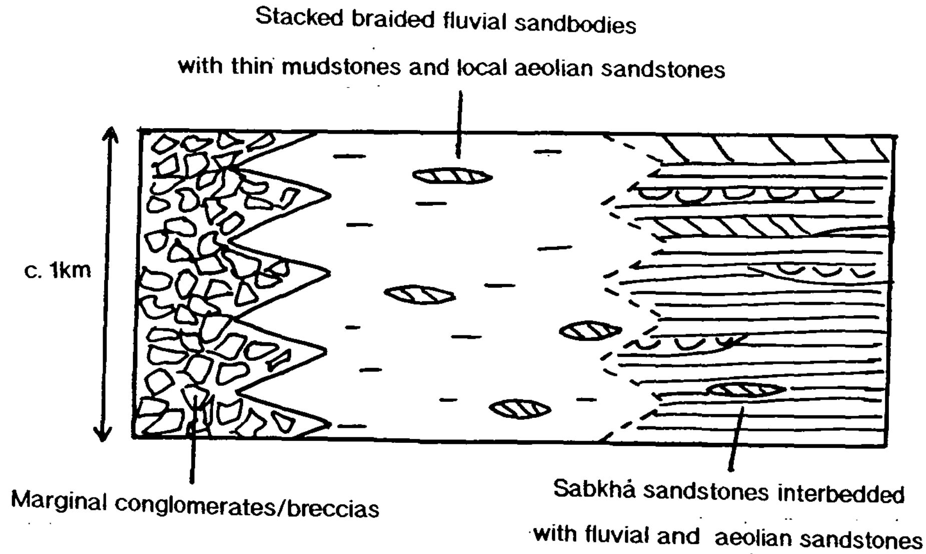



*Fig. 6.7. Schematic representation of the facies distrbution in the Upper ORS.*

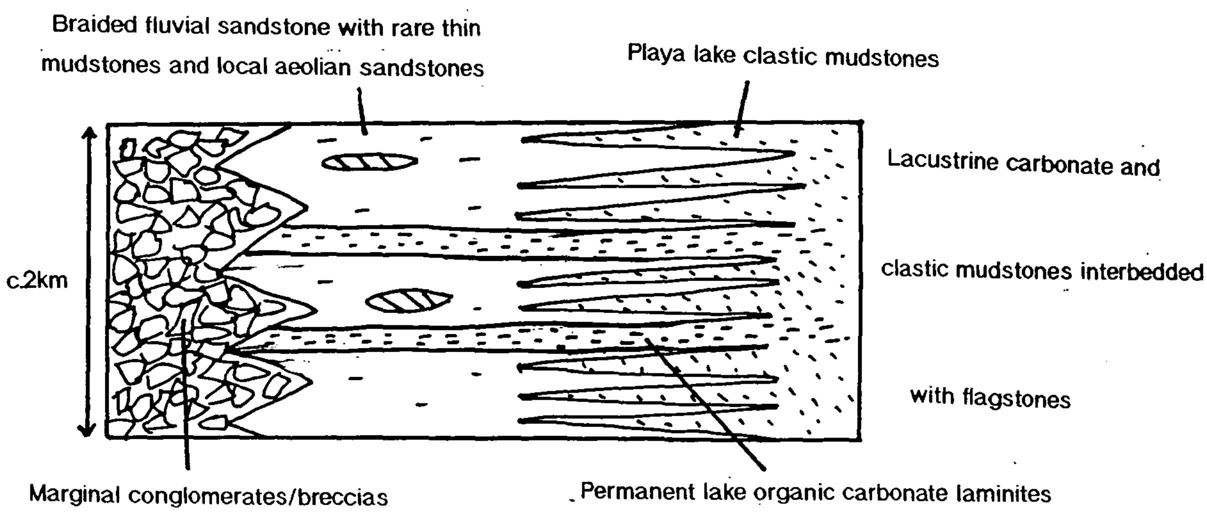



*Fig. 6.8. Schematic representation of the facies distrbution in the Middle ORS.*

# Point count data

| T/S No | %Qtz | %Fsp | %Lt | Grain Sz. | Sorting | Porosity | %IGV | %Clay | %Calcite | %FeO |
|---|---|---|---|---|---|---|---|---|---|---|
| 1/1 | 79.8 | 3.2 | 17 | 0.25 | 0.5 | 2.6 | 27.2 | 4.6 | 20 | 0 |
| 1/2 | 85.1 | 5.3 | 9.7 | 0.16 | 0.42 | 14.3 | 23.6 | 7 | 0 | 2.3 |
| 1/3 | 78.2 | 3.5 | 18 | 0.25 | 0.51 | 2.3 | 25.5 | 1.6 | 21.6 | 0 |
| 1/5 | 84.1 | 4.3 | 12 | 0.17 | 0.39 | 7.8 | 30.7 | 4 | 18.3 | 0.6 |
| 1/6 | 79.7 | 5.7 | 15 | 0.15 | 0.48 | 7 | 28.9 | 5.6 | 16.3 | 0 |
| 1/12 | 74.7 | 3.9 | 21 | 0.2 | 0.65 | 10 | 21.6 | 5 | 5.6 | 1 |
| 1/13 | 79.1 | 3.1 | 18 | 0.11 | 0.37 | 2 | 31.6 | 17.6 | 0 | 12 |
| 1/14 | 71.3 | 3.5 | 25 | 0.14 | 0.5 | 0.6 | 32.9 | 25 | 0 | 7.3 |
| 1/15 | 62.1 | 7.2 | 31 | 0.2 | 0.68 | 0 | 32.8 | 5.7 | 25.7 | 1.4 |
| 1/16 | 71.8 | 5.1 | 23 | 0.17 | 0.49 | 10 | 20.8 | 2.8 | 6 | 2 |
| 1/18 | 62.4 | 5 | 33 | 0.23 | 0.57 | 1.2 | 28.8 | 1.6 | 23.6 | 2.4 |
| 1/21 | 77.6 | 5.1 | 17 | 0.18 | 0.4 | 7 | 22 | 4 | 10.5 | 0.5 |
| 1/23 | 62.3 | 6 | 32 | 0.19 | 0.53 | 3 | 24.5 | 2 | 19.5 | 0 |
| 1/25 | 70.9 | 8.5 | 20 | 0.17 | 0.57 | 12 | 29 | 7 | 10 | 0 |
| 2/2 | 68.9 | 9.3 | 22 | 0.15 | 0.5 | 4.5 | 24 | 18 | 1.5 | 0 |
| 2/3 | 63.5 | 10.8 | 26 | 0.23 | 0.59 | 11 | 24 | 12 | 0 | 1 |
| 2/4 | 77.6 | 3.4 | 19 | 0.15 | 0.78 | 0.5 | 26 | 18 | 7.5 | 0 |
| 2/5 | 75.4 | 11 | 14 | 0.18 | 0.5 | 5 | 40 | 12 | 10.5 | 12.5 |
| 2/6 | 80.5 | 6.6 | 13 | 0.2 | 0.58 | 8.3 | 27.8 | 5.3 | 13.6 | 0.6 |
| 2/7 | 75.3 | 11.3 | 13 | 0.28 | 0.73 | 1.1 | 17.7 | 10 | 6.6 | 0 |
| 2/8 | 77.9 | 6.8 | 15 | 0.18 | 0.43 | 2.3 | 31.5 | 11.6 | 7.3 | 10.3 |
| 3/2 | 78 | 6.4 | 16 | 0.16 | 0.49 | 7.6 | 27.2 | 8 | 11.3 | 0 |
| 3/3 | 80.3 | 6.1 | 14 | 0.22 | 0.52 | 10.6 | 28.4 | 7.3 | 9.9 | 0.6 |
| 3/4 | 71.1 | 6.9 | 22 | 0.27 | 0.61 | 4.6 | 26.8 | 4 | 17.6 | 0.6 |
| 3/5 | 80 | 5 | 15 | 0.18 | 0.49 | 10.6 | 29.2 | 12.3 | 6.3 | 0 |
| 3/8 | 70.2 | 14.9 | 15 | 0.24 | 0.63 | 4.3 | 36.8 | 7.3 | 23.6 | 1.6 |
| 3/11 | 83.2 | 7.3 | 9.5 | 0.17 | 0.96 | 11.6 | 33.2 | 14.6 | 5 | 2 |
| 3/13 | 81.4 | 5.3 | 13 | 0.16 | 0.49 | 12.3 | 31.6 | 6 | 13.3 | 0 |
| 7/1 | 68.9 | 6 | 25 | 0.14 | 0.46 | 3 | 30.9 | 0.3 | 23.3 | 4.3 |
| 7/2 | 65.5 | 7.7 | 28 | 0.2 | 0.54 | 5 | 29 | 8 | 16 | 0 |
| 7/3 | 56.4 | 8.5 | 35 | 0.21 | 0.69 | 2 | 28.8 | 4.6 | 15.6 | 6.6 |
| 7/4 | 64.9 | 9.7 | 26 | 0.16 | 0.64 | 0 | 33.9 | 0.3 | 22.3 | 11.3 |
| 7/9 | 69 | 4 | 26 | 0.15 | 0.55 | 14 | 19.3 | 5.3 | 0 | 0 |
| 7/10 | 65.8 | 9.2 | 25 | 0.16 | 0.57 | 6.3 | 31.2 | 5.3 | 19.6 | 0 |
| 7/11 | 69.7 | 3.6 | 27 | 0.13 | 0.56 | 0 | 33.9 | 0.6 | 23.3 | 10 |
| 8/2 | 70 | 7.5 | 23 | 0.16 | 0.6 | 12 | 23 | 4 | 7 | 0 |
| 8/4 | 68.9 | 8.6 | 23 | 0.12 | 0.32 | 8.6 | 21.1 | 0.3 | 8.6 | 3.3 |
| 8/8 | 70 | 6 | 24 | 0.11 | 0.43 | 0 | 26.9 | 0 | 25.3 | 1.6 |
| 8/9 | 73.7 | 2.6 | 24 | 0.13 | 0.4 | 0 | 22.3 | 0 | 20.3 | 2 |
| 8/13A | 58.2 | 3.7 | 38 | 0.2 | 0.54 | 0.9 | 20.4 | 0 | 19.2 | 0.3 |
| 8/14 | 60.3 | 3.1 | 37 | 0.19 | 0.51 | 13.3 | 23.1 | 0.6 | 1.6 | 7.6 |
| 8/16 | 60 | 3.7 | 37 | 0.15 | 0.5 | 10.3 | 18.5 | 1.6 | 6.3 | 0.3 |
| 8/18 | 65.7 | 6 | 28 | 0.09 | 0.49 | 5.3 | 23.9 | 0.3 | 17 | 1.3 |
| 8/19 | 66.9 | 8.9 | 24 | 0.1 | 0.47 | 4.6 | 27.5 | 0 | 16.3 | 6.6 |
| 8/20 | 74 | 4.4 | 22 | 0.08 | 0.47 | 0 | 28.3 | 0 | 20.3 | 8 |
| 9/1 | 53.9 | 7.2 | 39 | 0.13 | 0.68 | 9.6 | 18.5 | 0.6 | 8 | 0.3 |
| 9/6 | 67.5 | 6.9 | 26 | 0.08 | 0.48 | 0.3 | 19.5 | 0.6 | 18.6 | 0 |
| 9/8 | 53.9 | 15.8 | 30 | 0.15 | 0.67 | 10 | 20.5 | 2.6 | 7.6 | 0.3 |
| 9/10 | 61.3 | 11.2 | 28 | 0.15 | 0.56 | 12 | 25.9 | 1 | 12.6 | 0.3 |

Appendix 1

| | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| 9/11 | 63.8 | 7.2 | 29 | 0.12 | 0.54 | 0.3 | 27.3 | 0 | 27 | 0 |
| 10/1 | 67.6 | 4.2 | 28 | 0.17 | 0.6 | 11.6 | 20.8 | 6.6 | 2.3 | 0.3 |
| 10/2 | 79.1 | 2.2 | 19 | 0.08 | 0.63 | 1.6 | 26.8 | 6 | 7.6 | 11.6 |
| 10/3 | 75.6 | 5.6 | 19 | 0.14 | 0.6 | 9.6 | 22.8 | 8.6 | 4.3 | 0.3 |
| 10/5 | 70.7 | 2.7 | 27 | 0.17 | 0.65 | 4.6 | 25.9 | 8 | 12.3 | 1 |
| 10/8 | 69.6 | 7.9 | 23 | 0.12 | 0.57 | 6.3 | 23.6 | 4 | 11.3 | 2 |
| 10/9 | 69.2 | 7.5 | 23 | 0.16 | 0.61 | 7.9 | 19.4 | 10.2 | 1.3 | 0 |
| 10/10 | 75.7 | 12 | 12 | 0.11 | 0.62 | 7 | 22.9 | 6.6 | 9.3 | 0 |
| 10/11 | 70.7 | 11.7 | 18 | 0.14 | 0.47 | 12.3 | 23.6 | 6.3 | 4 | 1.6 |
| 10/14 | 66.8 | 11.3 | 22 | 0.17 | 0.69 | 8 | 25.3 | 10.6 | 6.7 | 0 |
| 10/15 | 67.7 | 11.1 | 21 | 0.15 | 0.6 | 9 | 20.5 | 5.6 | 5.3 | 0.6 |
| 10/17 | 70.2 | 9 | 21 | 0.2 | 0.63 | 2.6 | 28.5 | 2.6 | 15.3 | 8 |
| 10/21 | 65.9 | 14.8 | 19 | 0.17 | 0.61 | 5.6 | 23.4 | 11.6 | 5.6 | 0.6 |
| 10/22 | 65 | 10.2 | 25 | 0.13 | 0.63 | 5.3 | 22.9 | 10 | 6.6 | 1 |
| 10/23 | 66.8 | 15.1 | 18 | 0.15 | 0.44 | 1.3 | 29.8 | 6.2 | 20.7 | 1.6 |
| 11/4 | 63.5 | 11.9 | 25 | 0.1 | 0.65 | 2.6 | 28.5 | 2 | 15.6 | 8.3 |
| 11/5 | 68.9 | 13.2 | 18 | 0.12 | 0.59 | 7.4 | 24.5 | 5.5 | 11.3 | 0.3 |
| 11/8 | 68.4 | 12.7 | 19 | 0.13 | 0.43 | 6.7 | 18.2 | 3.2 | 7.7 | 0.6 |
| 11/13 | 68 | 14.7 | 17 | 0.12 | 0.43 | 7.3 | 19.5 | 7 | 2.6 | 2.6 |
| 11/16 | 64.8 | 20.3 | 15 | 0.12 | 0.52 | 10.6 | 21.6 | 4 | 7 | 0 |
| 11/17 | 65.8 | 19.3 | 15 | 0.1 | 0.45 | 1.3 | 28.2 | 1 | 20.3 | 5.6 |

# Contact data

| T/S No | CI | TPI | Contact types (%) | | | | Nº contacts (%) | | |
|---|---|---|---|---|---|---|---|---|---|
| | | | Point | Long | C-C | Sutred | 0+1 | 2+3 | 4+> |
| 1/1 | 3.2 | 2.2 | 31.6 | 52 | 13.8 | 2.6 | 4.9 | 58.5 | 36.6 |
| 1/2 | 4 | 3.2 | 18.6 | 65.1 | 15.1 | 1.3 | 0 | 26.3 | 73.7 |
| 1/3 | 2.6 | 2.4 | 10.5 | 59.7 | 28.2 | 1.7 | 21.1 | 55.5 | 23.4 |
| 1/5 | 3.2 | 2.4 | 21.9 | 61 | 16.6 | 0.5 | 6.6 | 54.1 | 39.3 |
| 1/6 | 2.9 | 2.5 | 13.4 | 60.5 | 25.6 | 0.6 | 11.7 | 63.6 | 25 |
| 1/12 | 3.1 | 2.4 | 20.4 | 58.6 | 19.4 | 1.6 | 6.5 | 63 | 29 |
| 1/13 | 3.5 | 2.5 | 27.5 | 62.4 | 10 | 0 | 1.5 | 46.2 | 52.3 |
| 1/14 | 3.1 | 2.4 | 20.8 | 57.5 | 21.3 | 0.5 | 6.3 | 54 | 39.7 |
| 1/15 | 2.3 | 1.9 | 17.5 | 63.8 | 17.5 | 1.2 | 17.6 | 67.6 | 14.7 |
| 1/16 | 3.4 | 3.1 | 13.9 | 59.4 | 25.3 | 1.4 | 6.3 | 54.6 | 48.1 |
| 1/18 | 3 | 2.4 | 21.8 | 49.7 | 26.9 | 1.5 | 9.4 | 62.5 | 28.1 |
| 1/21 | 3.5 | 2.9 | 15.6 | 56.9 | 25.7 | 1.7 | 3 | 44.4 | 52.5 |
| 1/23 | 2.7 | 2.3 | 16.8 | 57 | 24 | 2.2 | 21.2 | 45.5 | 33.3 |
| 1/25 | 3 | 2.7 | 8.8 | 58 | 30.1 | 3.1 | 9.4 | 56.3 | 34.4 |
| 2/2 | 2.6 | 2.3 | 14.1 | 53.8 | 31.4 | 0.6 | 23.7 | 49.2 | 27.1 |
| 2/3 | 3.2 | 3 | 10.5 | 47.9 | 40.5 | 1.1 | 7 | 52.6 | 40.4 |
| 2/4 | 2.6 | 1.9 | 27 | 51.3 | 21.7 | 0 | 16.9 | 62.7 | 20.3 |
| 2/5 | 2.3 | 1.7 | 25.2 | 48.9 | 25.9 | 0 | 26.6 | 60 | 13.3 |
| 2/6 | 3.1 | 2.5 | 15.7 | 49.4 | 34.3 | 0.6 | 5.4 | 64.3 | 30.3 |
| 2/8 | 2.4 | 2 | 18.4 | 52.8 | 27 | 0.2 | 22.1 | 57.4 | 20.5 |
| 3/2 | 3.6 | 3.5 | 9.7 | 48.1 | 35.1 | 1.1 | 4.9 | 39 | 56.1 |
| 3/3 | 3.1 | 2.8 | 9.3 | 51 | 37.6 | 2.1 | 6.5 | 56.5 | 37 |
| 3/4 | 3 | 2.7 | 11.2 | 50 | 38.2 | 0.6 | 6.8 | 64.4 | 28.8 |
| 3/5 | 3 | 2.7 | 11 | 52.3 | 36.7 | 0 | 5.8 | 67.3 | 26.9 |
| 3/8 | 2.8 | 2.7 | 7.7 | 52.3 | 37.4 | 2.6 | 7.4 | 66.6 | 25.9 |
| 3/13 | 3.7 | 3.3 | 11.4 | 48.5 | 40.1 | 0 | 0 | 44.4 | 55.6 |
| 7/1 | 2.9 | 2.7 | 21.3 | 54.3 | 24.4 | 0 | 14 | 52.6 | 33.3 |
| 7/3 | 3 | 2.2 | 25.5 | 49.5 | 24.5 | 0.5 | 11.9 | 49.2 | 39 |
| 7/4 | 3.4 | 2.6 | 22.3 | 61.1 | 16 | 0.6 | 1.9 | 55.8 | 42.3 |
| 7/9 | 3.8 | 3.4 | 9 | 61.5 | 29.5 | 0 | 1.9 | 32.1 | 66 |
| 7/10 | 3.3 | 2.8 | 12.4 | 58.9 | 28.6 | 0 | 5.3 | 43.9 | 50.9 |
| 7/11 | 3.1 | 2.5 | 19.7 | 59.5 | 20.8 | 0 | 10.7 | 50 | 39.3 |
| 8/2 | 3.6 | 3.3 | 12.1 | 53.5 | 33.5 | 0.9 | 3.4 | 43.1 | 53.4 |
| 8/4 | 3.3 | 3 | 8.3 | 53.7 | 37.1 | 0.9 | 4.8 | 41.3 | 54 |
| 8/8 | 3.4 | 2.9 | 15.2 | 54.9 | 28.8 | 1.1 | 5.6 | 46.3 | 48.1 |
| 8/9 | 3.6 | 3.3 | 9.4 | 56.3 | 34.3 | 0 | 3.8 | 45.3 | 50.9 |
| 8/13A | 3.2 | 2.7 | 13.3 | 49.7 | 36.4 | 0.6 | 7.3 | 50.9 | 41.8 |
| 8/14 | 2.4 | 2.1 | 10.5 | 62.9 | 26.6 | 0 | 16.7 | 68.3 | 15 |
| 8/16 | 4 | 3.5 | 12.5 | 54.7 | 29.7 | 0.3 | 0 | 31.3 | 68.7 |
| 8/18 | 4 | 3.6 | 11 | 57 | 29.2 | 0.3 | 0 | 38.9 | 61.1 |
| 8/19 | 3.7 | 3.4 | 8.6 | 61.4 | 25.7 | 0.4 | 0 | 52.6 | 47.4 |
| 9/6 | 3.9 | 3.6 | 8.5 | 60.6 | 29.6 | 0 | 0 | 33.3 | 66.7 |
| 9/8 | 3.1 | 2.8 | 10.2 | 47.5 | 40.7 | 1.7 | 0 | 68.4 | 31.6 |
| 9/10 | 3.3 | 2.9 | 12.7 | 46 | 38.1 | 0.3 | 5.3 | 47.4 | 47.3 |
| 10/1 | 3.1 | 2.9 | 6.3 | 51.6 | 39.1 | 0.3 | 4.8 | 61.9 | 33.7 |
| 10/2 | 4 | 3.8 | 3.6 | 48.2 | 43.4 | 3.6 | 0 | 33.3 | 66.7 |
| 10/3 | 3.5 | 3.2 | 7.9 | 55.6 | 28.6 | 6.3 | 5.6 | 50 | 44.4 |
| 10/5 | 3.6 | 3.3 | 7.1 | 51.8 | 44.6 | 5.4 | 6.3 | 43.7 | 50 |
| 10/8 | 3.4 | 3.2 | 3.9 | 51.9 | 40.3 | 3.9 | 0 | 60.9 | 39.1 |

Appendix 2

| | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| 10/9 | 3.6 | 3.4 | 6.9 | 48.6 | 41.7 | 2.8 | 0 | 40 | 60 |
| 10/10 | 3.6 | 3.3 | 9.2 | 44.6 | 44.6 | 1.5 | 0 | 44.4 | 55.6 |
| 10/11 | 3.6 | 3.4 | 4.7 | 50 | 45.3 | 0 | 5.6 | 50 | 44.4 |
| 10/14 | 3.9 | 3.6 | 8.5 | 40.8 | 45.1 | 2.8 | 0 | 38.9 | 61.1 |
| 10/15 | 3.9 | 3.55 | 7.9 | 53.90 | 34.2 | 3.9 | 5 | 40 | 55 |
| 10/17 | 4 | 3.30 | 16.5 | 50.6 | 32.9 | 0 | 0 | 35 | 65 |
| 10/21 | 4.1 | 3.94 | 3.8 | 62 | 32.9 | 1.3 | 0 | 27.8 | 72.2 |
| 10/22 | 3.8 | 3.50 | 7.4 | 47.1 | 39.7 | 5.9 | 0 | 38.9 | 61.1 |
| 10/23 | 3.6 | 3.40 | 4.4 | 57.4 | 32.4 | 5.9 | 5 | 40 | 55 |
| 11/4 | 3.3 | 2.89 | 11.3 | 58.1 | 25.8 | 4.8 | 0 | 63.2 | 36.8 |
| 11/5 | 3.2 | 2.89 | 8.8 | 52.6 | 33.3 | 5.3 | 0 | 72.2 | 27.8 |
| 11/8 | 3.4 | 3.24 | 4.3 | 53.6 | 34.8 | 7.2 | 4.8 | 38.1 | 57.1 |
| 11/13 | 4.4 | 4.09 | 6.3 | 47.9 | 45.8 | 0 | 0 | 27.3 | 72.2 |
| 11/16 | 3.7 | 3.48 | 6.4 | 55.1 | 34.6 | 3.8 | 0 | 42.9 | 57.1 |
| 11/17 | 3.9 | 3.70 | 5.1 | 60.3 | 33.3 | 1.3 | 0 | 30 | 70 |

# Porosity loss data

| T/S No | Porosity | Cement | IGV | % original porosity desroyed by: compaction | cementation |
|---|---|---|---|---|---|
| 1/1 | 2.6 | 27.2 | 24.6 | 32 | 61.5 |
| 1/2 | 14.3 | 23.6 | 9.3 | 41 | 23.2 |
| 1/3 | 2.3 | 25.5 | 23.2 | 36.25 | 58 |
| 1/5 | 7.8 | 30.7 | 22.9 | 23.25 | 57.25 |
| 1/6 | 7 | 28.9 | 21.9 | 27.75 | 54.75 |
| 1/12 | 10 | 21.6 | 11.6 | 46 | 29 |
| 1/13 | 2 | 31.6 | 29.6 | 21 | 74 |
| 1/14 | 0.6 | 32.9 | 32.3 | 17.75 | 80.75 |
| 1/15 | 0 | 32.8 | 32.8 | 18 | 82 |
| 1/16 | 10 | 20.8 | 10.8 | 48 | 27 |
| 1/18 | 1.2 | 28.8 | 27.6 | 28 | 69 |
| 1/21 | 7 | 22 | 15 | 45 | 37.5 |
| 1/23 | 3 | 24.5 | 21.5 | 38.75 | 53.75 |
| 1/25 | 12 | 29 | 17 | 27.5 | 42.5 |
| 2/2 | 4.5 | 24 | 19.5 | 40 | 48.75 |
| 2/3 | 11 | 24 | 13 | 40 | 32.5 |
| 2/4 | 0.5 | 26 | 25.5 | 35 | 63.75 |
| 2/5 | 5 | 40 | 35 | 0 | 87.5 |
| 2/6 | 8.3 | 27.8 | 19.5 | 30.5 | 48.75 |
| 2/7 | 1.1 | 17.7 | 16.6 | 55.75 | 41.5 |
| 2/8 | 2.3 | 31.5 | 29.2 | 21.25 | 73 |
| 3/2 | 7.6 | 27.2 | 19.3 | 32 | 48.25 |
| 3/3 | 10.6 | 28.4 | 17.8 | 29 | 44.5 |
| 3/4 | 4.6 | 26.8 | 22.2 | 33 | 55.5 |
| 3/5 | 10.6 | 29.2 | 18.6 | 27 | 46.5 |
| 3/8 | 4.3 | 36.8 | 32.5 | 8 | 81.25 |
| 3/11 | 11.6 | 33.2 | 21.6 | 17 | 54 |
| 3/13 | 12.3 | 31.6 | 19.3 | 21 | 48.25 |
| 7/1 | 3 | 27.9 | 30.9 | 22.75 | 69.75 |
| 7/2 | 5 | 24 | 29 | 27.5 | 60 |
| 7/3 | 2 | 26.8 | 28.8 | 28 | 67 |
| 7/4 | 0 | 33.9 | 33.9 | 15.25 | 84.75 |
| 7/9 | 14 | 5.3 | 19.3 | 51.75 | 13.25 |
| 7/10 | 6.3 | 24.9 | 31.2 | 22 | 62.25 |
| 7/11 | 0 | 33.9 | 33.9 | 15.25 | 84.75 |
| 8/2 | 12 | 11 | 23 | 42.5 | 27.5 |
| 8/4 | 8.6 | 12.2 | 21.1 | 47.25 | 30.5 |
| 8/8 | 0 | 26.9 | 26.9 | 32.75 | 67.25 |
| 8/9 | 0 | 22.3 | 22.3 | 44.25 | 55.75 |
| 8/13A | 0.9 | 19.5 | 20.4 | 49 | 48.75 |
| 8/14 | 13.3 | 9.8 | 23.1 | 42.25 | 24.5 |
| 8/16 | 10.3 | 8.2 | 18.5 | 53.75 | 20.5 |
| 8/18 | 5.3 | 18.6 | 23.9 | 40.25 | 46.5 |
| 8/19 | 4.6 | 22.9 | 27.5 | 31.25 | 57.25 |
| 8/20 | 0 | 28.3 | 28.3 | 29.25 | 70.75 |
| 9/1 | 9.6 | 8.9 | 18.5 | 53.75 | 22.25 |
| 9/6 | 0.3 | 19.2 | 19.5 | 51.25 | 48 |
| 9/8 | 10 | 10.5 | 20.5 | 48.75 | 26.25 |

| | | | | | |
|---|---|---|---|---|---|
| 9/10 | 12 | 13.9 | 25.9 | 35.25 | 34.75 |
| 9/11 | 0.3 | 27 | 27.3 | 31.75 | 67.5 |
| 10/1 | 11.6 | 9.2 | 20.8 | 48 | 23 |
| 10/2 | 1.6 | 19.6 | 26.8 | 33 | 49 |
| 10/3 | 9.6 | 13.2 | 22.8 | 43 | 33 |
| 10/4 | 8.6 | 15.8 | 24.4 | 39 | 39.5 |
| 10/5 | 4.6 | 21.3 | 25.9 | 35.25 | 53.25 |
| 10/8 | 6.3 | 17.3 | 23.6 | 41 | 43.25 |
| 10/9 | 7.9 | 11.5 | 19.4 | 51.5 | 28.75 |
| 10/10 | 7 | 15.9 | 22.9 | 42.75 | 39.75 |
| 10/11 | 12.3 | 11.9 | 23.6 | 41 | 29.75 |
| 10/14 | 8 | 17.3 | 25.3 | 36.75 | 43.25 |
| 10/15 | 9 | 11.5 | 20.5 | 48.75 | 28.75 |
| 10/17 | 2.6 | 25.9 | 28.5 | 28.75 | 64.75 |
| 10/21 | 5.6 | 17.8 | 23.4 | 41.5 | 44.5 |
| 10/22 | 5.3 | 17.6 | 22.9 | 42.75 | 44 |
| 10/23 | 1.3 | 28.5 | 29.8 | 25.5 | 71.25 |
| 11/4 | 2.6 | 25.9 | 28.5 | 28.75 | 64.75 |
| 11/5 | 7.4 | 17.1 | 24.5 | 38.75 | 42.75 |
| 11/8 | 6.7 | 11.5 | 18.2 | 54.5 | 28.75 |
| 11/13 | 7.3 | 12.2 | 19.5 | 51.25 | 30.5 |
| 11/16 | 10.6 | 11 | 21.6 | 46 | 27.5 |
| 11/17 | 1.3 | 26.6 | 28.2 | 29.5 | 66.5 |

# Middle ORS poroperm data

| | | Sample No | Perm/md | Porosity | Lithofacies |
|---|---|---|---|---|---|
| **Unit:** | Cromarty | **Sample No** | **Perm/md** | **Porosity** | **Lithofacies** |
| **Age:** | ? Eif/Giv | 15/1-91 | 15 | -- | Gm |
| **Locality:** | Miller's Bay | 15/2-91 | 43.5 | -- | Sh |
| | | 7/1 | 12 | 3 | Sh |
| | | 7/9 | 29 | 14 | St |
| | | 7/10 | 21 | 6.3 | St |
| | | 7/3 | 12 | 2 | Sl |
| | | 7/4 | 12 | 0 | Sl |
| | | 7/11 | 17 | 0 | Sr |
| **Unit:** | Geanies/Cadboll | **Sample No** | **Perm/md** | **Porosity** | **Lithofacies** |
| **Age:** | Ems/Eif | 9/6 | 13.5 | 0.3 | Sm |
| **Locality:** | Hilton | 7/4-91 | 0 | -- | Sm |
| | | 8/2 | 29 | 12 | St |
| | | 8/14 | 23 | 13.3 | St |
| | | 8/16 | 23 | 5.3 | St |
| | | 9/10 | 13.5 | 12 | St |
| | | 9/11 | 13.5 | 0.3 | St |
| | | 6/4-91 | 48 | -- | St |
| | | 6/5-91 | 42.5 | 8 | St |
| | | 6/6-91 | 39 | -- | St |
| | | 6/8-91 | 8 | -- | St |
| | | 7/1-91 | 19.5 | 5 | St |
| | | 7/2-91 | 17.5 | -- | St |
| | | 8/4 | 23 | 8.6 | Sh |
| | | 8/8 | 15 | 0 | Sh |
| | | 8/18 | 13.5 | 5.3 | Sh |
| | | 6/1-91 | 9.5 | -- | Sh |
| | | 6/7-91 | 8 | -- | Sh |
| | | 7/3-91 | 0 | -- | Sh |
| | | 6/2-91 | 8 | -- | Sr |
| | | 6/3-91 | 11 | 4.3 | Sr |
| | | 3/1-91 | 20 | -- | Spe |
| **Unit:** | Rockfield | **Sample No** | **Perm/md** | **Porosity** | **Lithofacies** |
| **Age:** | Eif/Giv | 10/2 | 13.5 | 1.6 | St |
| **Locality:** | Rockfield | 10/3 | 17 | 9.6 | St |
| | | 10/9 | 14 | 7.9 | St |
| | | 10/14 | 15 | 8 | St |
| | | 11/13 | 17 | 7.3 | St |
| | | 8/2-91 | 36.5 | -- | St |
| | | 8/4-91 | 0 | -- | St |
| | | 8/6-91 | 22 | 5.6 | St |
| | | 8/7-91 | 10 | -- | St |
| | | 10/4 | 20 | 8.6 | Sh |
| | | 10/8 | 13.5 | 6.3 | Sh |
| | | 10/10 | 17 | 7 | Sh |
| | | 10/21 | 13.5 | 5.6 | Sh |
| | | 11/16 | 10 | 10.6 | Sh |
| | | 11/17 | 13.5 | 1.3 | Sh |
| | | 8/1-91 | 7 | -- | Sh |
| | | 8/3-91 | 3 | -- | Sh |
| | | 8/5-91 | 0 | -- | Sh |

# Upper ORS poroperm data

| | | Sample No | Perm/md | Porosity | Lithofacies |
|---|---|---|---|---|---|
| **Unit:** | Tarbat Ness | | | | |
| **Age:** | Givetian | 1/2 | 100 | 14.3 | St |
| **Locality:** | Embo | 1/15 | 23 | 0 | St |
| | | 1/16 | 46 | 10 | St |
| | | 1/23 | 15 | 3 | St |
| | | 1/25 | 53.5 | 12 | St |
| | | 1/3-91 | 15 | -- | Sp |
| | | 1/6 | 25 | 7 | Sh |

| | | Sample No | Perm/md | Porosity | Lithofacies |
|---|---|---|---|---|---|
| **Unit:** | Tarbat Ness | | | | |
| **Age:** | Givetian | 2/2 | 18.5 | 4.5 | St |
| **Locality:** | Tarbat Ness | 2/3 | 48 | 11 | St |
| | | 2/7 | 34.5 | 1.1 | St |
| | | 4/2-91 | 42 | 10 | St |
| | | 4/3-91 | 22 | -- | St |
| | | 4/6-91 | 27.5 | 7 | St |
| | | 4/7-91 | 22 | -- | St |
| | | 4/8-91 | 14.5 | 2.5 | St |
| | | 4/9-91 | 14.5 | 3 | St |
| | | 4/10-91 | 30 | -- | St |
| | | 4/11-91 | 17 | -- | St |
| | | 4/12-91 | 23 | 6 | St |
| | | 4/13-91 | 30 | -- | St |
| | | 4/14-91 | 17 | 4 | St |
| | | 4/15-91 | 26 | -- | St |
| | | 4/16-91 | 28.5 | -- | St |
| | | 4/19-91 | 36.5 | 6.3 | St |
| | | 2/5 | 18.5 | 5 | Sl |
| | | 4/1-91 | 28.5 | -- | Sh |
| | | 4/4-91 | 14.5 | 2 | Sh |
| | | 4/5-91 | 23 | -- | Sh |
| | | 4/17-91 | 22 | 5 | Sh |
| | | 4/18-91 | 23 | -- | Sh |

| | | Sample No | Perm/md | Porosity | Lithofacies |
|---|---|---|---|---|---|
| **Unit:** | Nairn Sandstone | | | | |
| **Age:** | Givetian | Nairn-A | 40 | 9 | St |
| **Locality:** | Nairn | Nairn-B | 27 | 7 | St |

| | | Sample No | Perm/md | Porosity | Lithofacies |
|---|---|---|---|---|---|
| **Unit:** | Gaza | | | | |
| **Age:** | Giv/Fras | 4/21-91 | 64 | -- | St |
| **Locality:** | Tarbat Ness | 4/23-91 | 250 | 20 | St |
| | | 9/1-91 | 270 | -- | St |
| | | 9/2-91 | 170 | 16 | Spe |
| | | 9/3-91 | 210 | -- | Spe |
| | | 1/8-91 | 57 | 13.6 | Sabkha |
| | | 4/20-91 | 33 | -- | Sabkha |
| | | 4/22-91 | 145 | -- | Sabkha |
| | | 9/4-91 | 25.5 | 6.2 | Sabkha |

Appendix 4

| | | | | | |
|---|---|---|---|---|---|
| **Unit:** | Gaza | **Sample No** | **Perm/md** | **Porosity** | **Lithofacies** |
| **Age:** | Giv/Fras | 3/5 | 41.5 | 10.6 | Sm |
| **Locality:** | Portmahomack | 3/8 | 24.5 | 4.3 | St |
| | | 3/11 | 20 | 11.6 | Sh |
| | | 3/13 | 55 | 12.3 | Sr |

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