To my Father and Mother.
The Carboniferous (Courceyan-Chadian) Sedimentary Facies Mosaic of the Keel-Ardagh Area of County Longford, Eire

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Submission in Partial Requirement for Award of PhD

THE POLYTECHNIC OF WALES
POLITECHNIG CYMRU
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RioFinEx NORTH LTD

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# LIST OF CONTENTS.

| Abstract. | iii |
| Figures. | iv |
| Photographs. | x |
| Introduction. | xvi |

<table>
<thead>
<tr>
<th>CHAPTER.</th>
<th>SUBJECT.</th>
<th>PAGE NUMBER.</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Methodology</td>
<td>1</td>
</tr>
<tr>
<td>1.1</td>
<td>Setting</td>
<td></td>
</tr>
<tr>
<td>1.2</td>
<td>Data points</td>
<td></td>
</tr>
<tr>
<td>1.3</td>
<td>Previous research</td>
<td></td>
</tr>
<tr>
<td>1.4</td>
<td>Correlation</td>
<td></td>
</tr>
<tr>
<td>1.5</td>
<td>Structure and sedimentation</td>
<td></td>
</tr>
<tr>
<td>1.6</td>
<td>Core techniques</td>
<td></td>
</tr>
<tr>
<td>1.7</td>
<td>Petrographic techniques</td>
<td></td>
</tr>
<tr>
<td>1.8</td>
<td>Interpretation of petrographic data</td>
<td></td>
</tr>
<tr>
<td>1.9</td>
<td>Interpretation terminology</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Silurian</td>
<td>24</td>
</tr>
<tr>
<td>3</td>
<td>Devonian (Microconglomerate)</td>
<td>29</td>
</tr>
<tr>
<td>4</td>
<td>Basal Clastics</td>
<td>57</td>
</tr>
<tr>
<td>5</td>
<td>Lower Quartz Sandstone</td>
<td>58</td>
</tr>
<tr>
<td>6</td>
<td>Quartz Pebble Conglomerate</td>
<td>81</td>
</tr>
<tr>
<td>7</td>
<td>Upper Quartz Sandstone</td>
<td>123</td>
</tr>
<tr>
<td>8</td>
<td>Provenance of the Lower Quartz Sandstone, Quartz</td>
<td>138</td>
</tr>
</tbody>
</table>
Pebble Conglomerate and Upper Quartz Sandstone.

<table>
<thead>
<tr>
<th>Page</th>
<th>Section</th>
</tr>
</thead>
<tbody>
<tr>
<td>9</td>
<td>Basal Transition Beds</td>
</tr>
<tr>
<td>10</td>
<td>Lower Mixed Beds</td>
</tr>
<tr>
<td>11</td>
<td>Navan Micrite</td>
</tr>
<tr>
<td>12</td>
<td>Upper Mixed Beds</td>
</tr>
<tr>
<td>13</td>
<td>Shaley Pales</td>
</tr>
<tr>
<td>14</td>
<td>Bioclastic Limestone Unit</td>
</tr>
<tr>
<td>15</td>
<td>Waulsortian Mound Complex</td>
</tr>
<tr>
<td>16</td>
<td>Post Waulsortian lithologies</td>
</tr>
<tr>
<td>17</td>
<td>Conclusions</td>
</tr>
</tbody>
</table>

References. 295
Acknowledgements. 316
ABSTRACT
The stratigraphic relationships and sedimentary environments of the Silurian, Devonian and Lower Carboniferous lithologies in the Keel area are deduced and described. This was accomplished by the analysis of over 10,000M. of diamond drilled core from the area.

The oldest rocks are Silurian shales with turbidites. Deposition was in forearc basins in an active subduction zone on the north margin of Iapetus.

The area was one of net erosion in the Lower Devonian following the continental collision which closed Iapetus.

The Upper Devonian of the area is represented by the Microconglomerate lying unconformably on the Silurian. This lithology was produced by braided streams draining the immediate Keel area.

The earliest Carboniferous sediments are an assemblage of sandstones and conglomerates.

The Lower Quartz Sandstone was laid down in sandy braided streams. The drainage basin of the Keel rivers was considerably larger by this time and sediment was derived from western Ireland.

Uplift in western Ireland made quartz pebbles and higher stream velocities to transport them available. These pebbles formed the Quartz Pebble Conglomerate deposited in pebbly braided streams.

Lowering of the source area by erosion resulted in reduced sediment grain size. Sandy braided streams again dominated the Keel area and deposited the Upper Quartz Sandstone.

Owing to the sea transgressing from the south marginal marine sediments were then deposited in the area, these being represented by the Lower Mixed Beds.

During the advancing transgression facies belts were moving northwards. Lagoonal (Navan Micrite) deposits were laid down behind a barrier complex now represented by the Upper Mixed Beds.

To seaward of the barrier shallow marine calcareous sediments were deposited on a homoclinal ramp. These are represented in the Keel area by the Shaley Pales and the Bioclastic Limestone Unit.

Further to seaward Waulsortian type mudmounds were deposited on the deeper ramp.

Owing to the transgressive regime existing in the Lower Carboniferous each facies belt moved northwards over the Keel area.

At the close of the Courceyan partial subsidence of the ramp took place and basinal/slope shales with turbidites (Calp) were deposited as lateral equivalents of the shallower Oakport Limestone on the surrounding shelf remnant.
FIGURES
## FIGURES.

<table>
<thead>
<tr>
<th>Number</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.</td>
<td>Irish Midlands location map.</td>
<td>xvii</td>
</tr>
<tr>
<td>2.*</td>
<td>Geological map of the Keel area.</td>
<td></td>
</tr>
<tr>
<td>3.</td>
<td>Distribution of Carboniferous rocks in Ireland.</td>
<td>2</td>
</tr>
<tr>
<td>4.*</td>
<td>Location map of the Keel area.</td>
<td></td>
</tr>
<tr>
<td>5.*</td>
<td>Distribution of boreholes in the immediate Keel area.</td>
<td></td>
</tr>
<tr>
<td>6.*</td>
<td>Distribution of bore holes around the Keel and Northern Inliers.</td>
<td></td>
</tr>
<tr>
<td>7.</td>
<td>A correlation of British and Irish Lower Carboniferous successions.</td>
<td>7</td>
</tr>
<tr>
<td>9.</td>
<td>Correlation of the lithological units with conodont zones.</td>
<td>10</td>
</tr>
<tr>
<td>10.</td>
<td>The approximate extent and palaeolatitudes of the Old Red Sandstone continental landmass in Upper Devonian times.</td>
<td>30</td>
</tr>
<tr>
<td>11.</td>
<td>Isopachyte map of the Devonian Microconglomerate.</td>
<td>32</td>
</tr>
<tr>
<td>12.</td>
<td>A clast supported and matrix</td>
<td>46</td>
</tr>
</tbody>
</table>
supported conglomerate couplet in the Microconglomerate. Probable debris flow deposit.

12 A. Key to symbols used on graphic logs throughout this study.

13. Deposition on braid bar tops in the Microconglomerate.


15. Source areas for the basal clastic units.

16. Possible granitic sediment sources in the local Keel area.

17. A possible palaeogeography of Devonian Ireland.

18 (A and B).* Visher plots of the Lower Quartz Sandstone.

19. Markov Chain Analysis of the Lower Quartz Sandstone.

20. Compound bar sequence. South Saskatchewan River.

20 A. Key to Figure 20.


21 A. Key to figure 21.

22. Deposition on a vegetated island. Lower Quartz Sandstone.
23. Sub-environments in a braided stream.

24. Hierarchical organisation of channels and bars in a braided stream.

25.* Diagramatic lithostratigraphic correlations in the Irish Midlands.

26. Water levels and current directions in active and upstream cut-off secondary (level 4) channels of braided rivers.

27. Sequence showing the filling of a level 4 channel. Quartz Pebble Conglomerate.

28. Size distribution of the clasts in the Quartz Pebble Conglomerate.

29. Isopachyte map of the Quartz Pebble Conglomerate showing bars and channels.

30. Quartz Pebble Conglomerate. Deposition on Bars.

31. Distribution of the Ballyvergin Shale and of the Quartz Pebble Conglomerate in the Irish Midlands.

32. Course of the "Keel" river extrapolated seaward.

33. Depositional sequence of the lower
34. Markov Chain Analysis of the Upper Quartz Sandstone.
35. Upper Quartz Sandstone. Deposition in a channel.
36. Upper Quartz Sandstone. Deposition on a sand flat.
37. Postulated palaeohigh in the Keel area in Basal Transition Beds times
38. Rose diagram of pebble orientation in the Quartz Pebble Conglomerate.
39. Thicknesses of the lower clastic units in the Irish Midlands.
40. Locations of Quartz Pebble Conglomerate type lithologies in the Irish Midlands.
41. Possible courses of the "Keel" river.
42. Lower Mixed Beds. Channel sequence.
43. Barrier island and tidal flat complex.
44. Lower Mixed Beds. Channel sequence.
45. Lower Mixed Beds. Beach sequence.
46. Lower Mixed Beds. Mechanism for the transgressive lime mudstone bed.
47. Lower Mixed Beds. Infill of tidal channel meander cut-off.

vii
| 48.* | Strike section along the south of the Keel Inlier. |
| 49.* | "Fence" diagram around the Keel Inlier. |
| 49 A.* | Key to figures 48 and 49. |
| 50. | Generalised over-all sequence of the Navan Micrite. |
| 51. | Navan Micrite. Accretion of carbonate from subtidal source areas onto tidal flats. |
| 52. | Main morphological components of the Barrier model. |
| 53. | Upper Mixed Beds. Broad subdivision. |
| 54. | Thickness variations of the Upper Sandstone in the Irish Midlands. |
| 55. | Upper Mixed Beds. Tidal channel. |
| 56. | Upper Mixed Beds. Shore-face sequence. |
| 57. | Upper Mixed Beds. Washover sequences. |
| 58. | Insoluble residue and carbonate analysis. Shaley Pales and Bioclastic Limestone Unit. |
| 59. | Bioclastic Limestone Unit. |
| 60. | Depositional mechanism for the |
Waulsortian Mudmound complexes.

61. Waulsortian Mudmounds. Isopachyte map. 270

62. Dimensions of the Calp Basin. 282

63. Deep structure of the Keel area. 283

64. Depositional environment of the Microconglomerate. 284

65. Depositional environment of the Lower Quartz Sandstone. 285

66. Depositional environment of the Quartz Pebble Conglomerate. 286

67. Depositional environment of the Upper Quartz Sandstone. 286

68. Depositional environment of the Lower Mixed Beds. 287

69. Depositional environment of the Navan Micrite. 288

70. Upper Mixed Beds. Transgressive sandstone and traces of former barrier islands. 289

71. Depositional environment of the Shaley Pales and Bioclastic Limestone Unit. 290

72. Depositional environment of the Waulsortian Mudmound Complexes. 291

73. Extent of Waulsortian Mudmounds in Ireland. 293

ix
74. Depositional environment of the Calp and the Oakport Limestone.

75.* Overall facies mosaic for the Carboniferous rocks of the Keel area, Central Eire.

* Figure in pocket at rear of volume.
PHOTOGRAPHS
## PHOTOGRAPHS

<table>
<thead>
<tr>
<th>Number</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.</td>
<td>Coarse grained sandstone of the Microconglomerate unconformably overlying the cleaved Silurian. The plane of unconformity is marked (U-U). The sequence youngs to the right.</td>
<td>35</td>
</tr>
<tr>
<td>2.</td>
<td>Coarse grained Silurian sandstone. Note the markedly better rounding of the chert clast (Ch) as compared to the angular quartz (Q). Stained. Crossed Polars. Mag. x 50.</td>
<td>35</td>
</tr>
<tr>
<td>3.</td>
<td>Silurian grit/sandstone. Note the abundant mica. Crossed polars. Mag. x 50.</td>
<td>36</td>
</tr>
<tr>
<td>4.</td>
<td>Calcrete in the siltstones of the Microconglomerate. Contact to the underlying Silurian is under the hammerhead (Silurian to left). Hammerhead is 11.5 cms. long.</td>
<td>36</td>
</tr>
<tr>
<td>5.</td>
<td>Imbrication of well rounded quartzite, vein quartz and jasper clasts in the Microconglomerate</td>
<td>37</td>
</tr>
<tr>
<td>6.</td>
<td>Fining-up cycles in the Microconglomerate. The cycle is between the coins. The sequence fines up from quartz clasts of 1.5 cms. diameter (C) to coarse sandstone (S) to medium sandstone (M). At the base is an erosional contact to an underlying mudstone. The sequence youngs to the right.</td>
<td>37</td>
</tr>
</tbody>
</table>
7. **Microconglomerate.** Small fining-up cycles of coarse to fine grained sandstone in siltstones. The sandstones show sharp basal and top contacts. Deposition was probably into subsidiary channels at times of high discharge. The sequence youngs to the right. The small reduction spot (R) may indicate the presence of vegetation in the channel. The hand lens is 4 cms. long.

8. **Quartz Pebble Conglomerate.** Small exposure on the Keel Inlier. Note the pronounced fining-up and the imbrication of the pebbles. The hammer shaft is 30 cms. long.

9. Fining-up character of the Quartz Pebble Conglomerate as seen in the core. The sequence lies above an erosion surface.

10. **Quartz Pebble Conglomerate.** Shows the high proportion of quartzite clasts (Q) as compared to vein quartz (V). The imbrication indicates transport from the right (south). Note the erosion surface (E-E). The hammer shaft is 30 cms. long.

11. **Quartz Pebble Conglomerate.** Typically sized exposure as seen on the Keel Inlier. Hammer shaft is 30 cms. long.

12. **Quartz Pebble Conglomerate.** Dolomitisation. Note the mudstone clast outlining the shape of a former quartz clast.
13. Quartz Pebble Conglomerate. Dolomitisation. Note the remnants of a mudstone bed (M) confirming the replacement nature of the dolomitisation. The sample is No. 432.

14. Quartz Pebble Conglomerate. Thin section of sample 432. Note the quartz grains "floating in a sea of dolomite. Quartz (Q) . Photomicrograph. Stained. Magnification x50. A grain of microcline is present. Both it and the quartz grains show signs of corrosion. The thin section has been prepared very thick to better show the quartz under crossed polars.

15. Contact of the Upper Quartz Sandstone and the Lower Mixed Beds. Shows the abrupt change in sedimentary "style". Sudden change in grain size and colour. Inch rule as scale.

16. The slightly erosional nature of the contact between the Upper Quartz Sandstone and the Lower Mixed Beds. Note the lack of transitional lithologies. The lines indicate the position of this sequence in the portion shown in P. 15.

18. Navan Micrite. Preferential dolomitisation of burrows in a lime mudstone. The dolomitising fluids may have gained entry to the sediment via the large burrow to the right. Stained. Plain light. Mag x 50. Burrow denoted by B.

19. Navan Micrite. Serpulid worm tubes showing an early isopachous cement and a later equant cement. The "matrix" has been dolomitised. Stained. Plain light. Mag. x 50.

20. Navan Micrite. Serpulid worm tubes. Note the lack of shells etc. for the worms to encrust on. This may provide tentative evidence for the former existence of a hard ground. The lack of worm tubes in the core sections to the left and right may indicate that there were reduced sedimentation rates in the serpulid worm section thus giving rise to the hard ground.


23. Upper Mixed Beds. Oolitic grainstone overlain by
sandstone. Probably marks the site of a tidal channel. Stained. Plain light. Mag x 50.

24. Shaly Pales. Storm lags. Packstone/siltstone couplets. Packstones have erosional bases and some degree of fining-up is present. Length of section 70 cms.


27. Bioclastic Limestone Unit. Sponge spines replaced by ferroan calcite. Note the fragment of silicified bryozoan (B) in the top right. Unreplaced spines are present elsewhere in this thin section Stained. Plain light. Mag. x 50.


29. Shell fragment deforming sediment. Bioclastic Limestone Unit.

30. Storm lags in the Bioclastic Limestone Unit.


32. Contact of Mound and Flank Facies as seen in the core. Waulsortian Mudmound Complex.
33. Waulsortian Mudmound Complex at Carrickbuoy Quarry. Parts of two mounds separated by the Flank Beds (F) are seen. Mound facies (M).

34. Stromotactids in the Mound facies of the Waulsortian Mound complex. Note the flat bases and the high depositional dip of many of the stromotactids. Mud filled stylolite to the left.
INTRODUCTION
INTRODUCTION.

This thesis describes and interprets the sedimentary environments of the (probable) Late Devonian and Lower Carboniferous rocks within a 160 sq. KM. section of the Keel area, County Longford, Eire (Fig.1). In the interests of continuity mention is also made of the Silurian rocks which unconformably underlie them.

In common with much of the Irish Midlands the precise geology of the area is not well known. This is due to a paucity of exposure caused by a thick cover of glacial drift.

Previous studies in the area (Griffiths 1837, Kineham and Symes 1871 and Thompson 1954) laboured under this handicap. Somewhat later studies (Patterson 1970, Slowey 1986) utilised the diamond drilled core produced in the drilling programme for mineralisation in the area. However these studies were carried out with mineralisation in view, only cursory attempts were made to deduce the depositional environments of the rocks.

This study, utilising some of the same core, attempts to correct this omission. Essentially the succession is of a fining upward type that resulted from the marine transgression which effected the area in Lower Carboniferous times (Phillips and Sevastopolou 1986). Unconformably overlying the Silurian are sub-arielly deposited sandstones and conglomerates, these are succeeded by a marginal marine series. Deposition of marine limestones, sandstones and shales of a deeper water character followed culminating in the deposition of Waulsortian type mud mounds. The sequence is completed by the deposition of basinal calcareous shales.
FIG. I. LOCATION MAP OF THE IRISH MIDLANDS.
CHAPTER 1
CHAPTER 1.

Methodology.

1.1 Geographical and geological setting.

The Keel and Lisduff (Northern) Inliers lie approximately 8.6 and 1.6 Kms. respectively south of Longford Town, County Longford, Eire (Fig.4). The inliers, especially the Keel Inlier, dominate the surrounding countryside which has the characteristic flatness of the Irish Midlands. To the north-east lies the Lower Palaeozoic Longford-Down Massif (Fig.1). The two inliers are more properly called Ardagh Hill (Keel Inlier) and Lisduff Hill (Northern Inlier) but in the interests of continuity with previous work the informal names are retained throughout this volume.

Geologically the area forms part of the major expanse of Lower Carboniferous (Dinantian) rocks extending from Ladys Well (County Cork) to Lough Foyle in the north of Ireland (Fig.3). This succession of conglomerates, shales and carbonates lies unconformably on rocks of a predominantly Ordovician and Silurian age. They occasionally rest on red beds of a Devonian or lowest Carboniferous age.

1.2 Location of bore-hole data points and rock exposures.

Rock exposure in the area is very limited owing to the thick cover of glacial drift. This drift blankets the countryside over much of the Irish Midlands and may be up to 30 metres thick. Outcrop, where it exists in the area, is on a small scale and tends to be concentrated on the top of the Keel Inlier where the Quartz Pebble Conglomerate outcrops. Previous researchers made use of the numerous small quarries in the area, but these are now largely abandoned and infilled e.g. the Calp
Fig. 3. Distribution of Carboniferous rocks in Ireland.
quarries on the Ardagh to Longford and Ardagh to Edgeworthstown roads (Fig.4). Only one of these quarries now remains, that at Carrickboy 4.5 Km. south of Keel where the Waulsortian Mound (Reef) Complex is exposed. Even here however the quarry faces are degraded and overgrown.

Bore hole density is highly variable. It is very dense near the area of the Main Keel Fault whilst in other areas it is non existant. The distribution of bore-holes is shown in Figs. 5 and 6.

Figs. 5 and 6 indicate which bore-holes were studied. The bore-holes indicated by a cross mark those holes whose cores were studied in detail, circles indicate less detailed study whilst dots indicate cores not studied.

The studied bore-holes shown in Figs. 5 and 6 also show letter prefixes to indicate which company drilled the hole.

K.A. American Smelting and Refining Corporation (ASARCO).
K. RioFinEx (Keel).
L.F. RioFinEx (Longford).
M.R. RioFinEx (Mine Rock).
K.D. Dressler Minerals.
B.P.H. Charter Consolidated.
D.D.B. Central Mining and Finance.

Six sizes of diamond studded bit were used (drill diameters in brackets); AX (30mm.), BQ (37mm.), BX (41mm.), NQ (47.6mm.), HQ (63.5mm.) and PQ (85mm.). Drill sizes would be stepped down as the hole deepened e.g. PQ to 15M. (thickness of the drift), BX to 300m. and AX thereafter. Different combinations of drill sizes were used by the above
companies. No single company appears to have used all six sizes. The bore-holes were either drilled vertically or at angles of dip varying from 45 to 60 degrees. Acid bottle tests indicated that the amount of drill "wander" was minimal.

1.3 History of previous research.

(a) Eire.

Considering that rocks of a Lower Carboniferous age occur at the surface or beneath the Quaternary cover for over 50% of the area of Ireland (Fig.3) the amount of geological literature is limited. The discovery of major sulphide ore bodies in rocks of this age however stimulated research and the amount of geological information has increased substantially.

The first geologist to inquire into the Lower Carboniferous was Kirwan (1794) who described the distinctive Calp limestones of the Dublin area. Irish geology in the 19th century was dominated by the activities of the Irish Geological Survey who actually mapped in the Keel area. Up to the 1950s there was little geological investigation. Very little exploration was carried out by mining companies despite the fact of there being a long history of metal ore extraction in Ireland (lead had been intermittently worked at Silvermines since the 9th century). A well known geological textbook of the 1950s stated that "there are no mineral deposits in Ireland" (IAEG 1984).

In the 1950s exploration by mining companies became increasingly important. Stream and soil sampling revealed promising anomalies throughout the Lower Carboniferous. Drilling programmes followed and an increased academic interest was generated.
Over the last 25-30 years a large amount of literature has been produced, especially by Sevastopolou and his fellow workers. However owing to the lack of exposure and the fact that most of the bore-hole information relates to those areas around Lower Palaeozoic inliers the geology of a large area of Ireland is essentially unknown. The geology of areas around the inliers and areas of coastal exposure is known to a greater or lesser degree (Graham and Reilly 1976, MacCarthy 1974 etc.).

b) Keel.

The area was surveyed between 1867 and 1872 (Jukes 1867, Kinehan and Symes 1871, Cruse 1872) and memoirs written to accompany the maps produced.

Thompson (1954) made a study of the fossil fauna in the area and put the stratigraphy on a more formal basis. She also made important contributions to a better understanding of the structure of the area.

Patterson (1970) and Slowey (1986) also comment on the stratigraphy of the area in works primarily concerned with the lead and zinc mineralisation which has attracted the attentions of various (1.2) mining companies over the last 25 years.

1.4 Classification and correlation.

(a). Mainland Britain and Ireland.

Fig.7 illustrates a correlation between the Courceyan and Chadian successions of the Central Province (Anderton et.al. 1979) of England with those at locations in Ireland and that at Keel. The Irish Dinantian successions bear close resemblances to those in the Central and Northern Provinces of England reflecting similar sedimentation patterns.

These resemblances are brought out by a comparison of the
succession at Ravenstonedale (Durham-Cumbrian border) and that at Keel.

The basal fluvio-marine Pinskey Gill Beds at Ravenstonedale, resting upon the Lower Palaeozoic, may be equivalent to the lower clastic units at Keel.

The overlying Stone Gill and Coldbeck Beds contain micritic (often dolomitised) limestones, stromatolites and a restricted fauna. These may bear comparison with the Navan Micrite. An open marine bioclastic limestone, the Scandal Beck Limestone, succeeds these beds. This limestone may be compared with the Bioclastic Limestone Unit at Keel.

Although the Longford-Down Massif may have been equatable with the Alston Block etc. in controlling sedimentation it is thought to have exerted far less influence on sedimentation patterns around it than did the blocks of the British Northern Province (Anderton et. al. 1979).

(b). Irish successions.

Philcox (1984) has correlated the stratigraphic succession at Keel with that of Ireland in general. Whilst considered to be very servicable and useful in many respects it can be found wanting in some. Fig.8 shows the local nomenclature at Keel (Patterson 1970, Slowey 1986) and the terms used on a general basis throughout the Irish Midlands (Philcox 1984). Terms used in this text may be from Slowey (1986) or from Philcox (1984). The reasons for using any particular term are explained in the relevant chapter.

1.5 Structure and sedimentation in the Keel area.

The structure of the area consists basically of the two Lower Palaeozoic inliers (Keel and Northern) and three sets of faults surrounding and cutting them (Fig.2). This faulting is apparently much
**Fig. 7. A CORRELATION OF THE COURCEYAN AND CHADIAN ROCKS OF THE BRITISH ISLES AND IRELAND, FROM ANDERTON ET AL. (1979) AND GEORGE ET AL. (1976)**

1. Chadian
2. COURCEYAN

BVB See over page

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**Map showing**

- Scale: 100 Km
- Palaeo land surface

- Locations:
  - Clitheroe
  - Carrick
  - Keel
  - Kingscourt
  - N. Derbys.
<table>
<thead>
<tr>
<th>Location</th>
<th>Code</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Skipton</td>
<td>EL</td>
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</tr>
<tr>
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<td>Haw Bank Limestone</td>
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<td>SHR</td>
<td>Salthill Reef</td>
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</tr>
<tr>
<td></td>
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<td>Coplow Reef</td>
</tr>
<tr>
<td></td>
<td>BVB</td>
<td>Bold Venture Beds</td>
</tr>
<tr>
<td></td>
<td>BEB</td>
<td>Bankfield East Beds</td>
</tr>
<tr>
<td></td>
<td>HB</td>
<td>Horrocksfield Beds</td>
</tr>
<tr>
<td>Kingscourt</td>
<td>KL</td>
<td>Kilbride Sandstone</td>
</tr>
<tr>
<td></td>
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</tr>
<tr>
<td></td>
<td>RS</td>
<td>Rockfield Sandstone</td>
</tr>
<tr>
<td></td>
<td>PL</td>
<td>Porcellanous Limestone</td>
</tr>
<tr>
<td></td>
<td>BG</td>
<td>Basal Grit</td>
</tr>
<tr>
<td>Keel</td>
<td>BL</td>
<td>Bioclastic Limestone</td>
</tr>
<tr>
<td></td>
<td>R</td>
<td>Waulsortian Reef</td>
</tr>
<tr>
<td></td>
<td>BL</td>
<td>Bioclastic Limestone</td>
</tr>
<tr>
<td></td>
<td>MB</td>
<td>Mixed Beds</td>
</tr>
<tr>
<td></td>
<td>BC</td>
<td>Basal Clastics</td>
</tr>
<tr>
<td>Carrick</td>
<td>OG</td>
<td>Oakport Group</td>
</tr>
<tr>
<td></td>
<td>KG</td>
<td>Kilbryan Group</td>
</tr>
<tr>
<td></td>
<td>BSG</td>
<td>Boyle Sandstone Group</td>
</tr>
</tbody>
</table>

8
**Fig. 8. Lower Carboniferous Stratigraphy at Keel, Comparison with the nomenclature of Slowey (1986) and Philcox (1984).**

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Calp Limestone</td>
<td>Calp Limestone</td>
<td>Calp Argillaceous LST</td>
<td>Calp Limestone</td>
</tr>
<tr>
<td>Visean Bioclastic Limestone</td>
<td>Visean Bioclastic Limestone</td>
<td>Calp Argillaceous LST</td>
<td>Calp Limestone</td>
</tr>
<tr>
<td>Massive (Reef) Limestone</td>
<td>Bioclastic Limestone</td>
<td>Reef Limestone</td>
<td>Waulsortian Mudmound Complex</td>
</tr>
<tr>
<td>Bioclastic Limestone</td>
<td>Argillaceous Bioclastic Limestone</td>
<td>Bioclastic Limestone</td>
<td>Bioclastic Limestone Unit</td>
</tr>
<tr>
<td>Upper Mixed Beds</td>
<td>Upper Shaly Pales</td>
<td>Upper Shaly Pales</td>
<td>Upper Shaly Pales</td>
</tr>
<tr>
<td>Upper Mixed Beds</td>
<td>Middle Shaly Pales</td>
<td>Middle Shaly Pales</td>
<td>Middle Shaly Pales</td>
</tr>
<tr>
<td>Upper Mixed Beds</td>
<td>Upper Sandstone Pales</td>
<td>Upper Sandstone Pales</td>
<td>Lower Shaly Pales</td>
</tr>
<tr>
<td>Bioclastic Dolomite</td>
<td>Lower Mixed Beds</td>
<td>Navan Micrite</td>
<td>Lower Mixed Beds</td>
</tr>
<tr>
<td>Lower Mixed Beds</td>
<td>Lower Quartz SST</td>
<td>Quartz Pebble Cong.</td>
<td>Upper Quartz SST</td>
</tr>
<tr>
<td>Upper Quartz SST</td>
<td>Lower Quartz SST</td>
<td>Lower Quartz SST</td>
<td>Quartz Pebble Cong.</td>
</tr>
<tr>
<td>Microconglomerate</td>
<td>Sandstone</td>
<td>Microconglomerate</td>
<td>Lower Quartz SST</td>
</tr>
<tr>
<td>Lower Palaeozoic</td>
<td>Lower Palaeozoic</td>
<td>Lower Palaeozoic</td>
<td>Lower Palaeozoic</td>
</tr>
</tbody>
</table>
more in evidence around the Keel Inlier but this may be a function of the much greater drill hole density around Keel and the consequent increased geological knowledge as opposed to that around the Northern Inlier.

The inliers show the east/north-east Caledonian trend typical of Irish Lower Palaeozoic inliers. As is noted below (Chap. 3f) it is believed that the Keel Inlier was not a significant structure during the Devonian and Lower Carboniferous but uplift has certainly occurred at some stage since the Carboniferous sedimentaries now dip off the Keel Inlier at angles ranging from 10-30 degrees (Slowey 1986). Hercynian tectonics have been postulated as an agent of this uplift (Slowey opp. cit.).

The faults may be divided into three groups;

(1). Main Keel Fault.

The fault follows the Caledonian structural trend and can be detected for some 8 Km. as it trends along the south of the Keel Inlier (Fig.2). The fault is really a series of discrete faults rather than a single one. Each of these faults dips at 55-60 degrees to the south-east with a throw of about 210 metres (Slowey opp.cit.). Underground work in the Keel shaft gave no indication of movement on the faults during deposition of the Quartz Pebble Conglomerate and Upper and Lower Quartz Sandstones but their local fracturing gives evidence that movement took place after their lithification. This movement may have had profound effects on the later carbonate lithologies.

(2). Minor faults parallel to the Keel Fault.

A series of minor faults striking parallel to the main Keel Fault
are present (Fig. 2). These dip at about 75 degrees to the north-west. Throws are small averaging about 0.5 metres.

(3). North-west faults.

A third set of steeply dipping faults striking roughly north-west postdates the above faults. Mineralisation at Garrycam, 1 Km. to the east of Keel, associated with a north-west trending fault appears to indicate that this set of faults was active during the Carboniferous (Slowey opp.cit.).

Deep structure.

Recent gravity and magnetic investigations in the Irish Midlands have revealed something of the underlying deep structure. In essence it appears that the deeper rocks comprise Pre-Cambrian crystalline basement at a depth of 3-4 Km. (Brown and Williams 1985) overlain by thick north-west trending Ordovician volcanic blocks separated by sedimentary troughs. These blocks and troughs are covered by varying thicknesses of Silurian, possible Devonian and Carboniferous sedimentary rocks.

With regard to the Keel area it has been suggested (Jacobs et.al. 1985) that a large granite body exists at depth to the south of Keel whilst large linear blocks of Ordovician volcanics, of the type alluded to above, occur to the north-west and south-east of Keel (Brown and Williams 1985). It would seem possible that this granitic body may have been an important control on sedimentation during deposition of the Quartz Pebble Conglomerate and the Upper and Lower Quartz Sandstones.

Sedimentology.

The pattern of sedimentology in the area is of an overall fining upward character. Coarse sandstones and conglomerates of a fluvial
origin at the base are succeeded by finer grained, shallow marine sandstones, shales and limestones which are in turn followed by deeper water limestones. Deep water shales are found at the top of the succession. The strata bear witness to deposition of sediment over some 10 million years (Clayton and Higgs 1979) largely inspired by a marine transgression from the south which ultimately drowned the Lower Palaeozoic inliers which had probably stood up as slight eminances above a peneplain. The transgressing sea, probably an arm of the Rheic Ocean, entered the south of Ireland in Late Devonian (Tnl) times and had reached the area of Clew Bay (Fig.15) by early Visean times (Clayton and Higgs opp.cit.) by which time deep water (Calp) shales and mudstones were being deposited in the Keel area.

The activating agent for the marine transgression has been variously postulated as the result of downwarping of the earths crust (Johnson 1982), North Atlantic rifting (Haszeldine 1984) as well as more generalised eustatic causes (Ramsbottom 1973, George 1978).

1.6 Techniques employed in the investigation.
(a). Bore-hole logging.

The cores selected for logging were chosen so as to express the maximum vertical and lateral extent of the rock facies. This was undertaken to facilitate the construction of measured sections, isopachyte maps etc.

Twenty four weeks were spent at the Keel mine site and over 10,400 metres of core were logged in detail. An additional 3,500 metres was also examined in a less rigorous manner to yield information on formation thicknesses etc.
(b). Methods employed.

The following procedure was carried out in the preliminary examination of the core (Siemers and Roderick 1981). A checklist was made up incorporating the following points.

1. Primary (i.e. depositional) composition.
2. Secondary (i.e. diagenetic) composition.
3. Grain texture (i.e. mean size and sorting).
4. Rock texture (i.e. fabric and orientation).
5. Primary physical sedimentary structures.
7. Secondary (i.e. deformational) sedimentary structures.

Close attention was paid to the vertical distribution and variation of these features and contact relationships (e.g. erosional, grading etc.) between vertically adjacent lithological units.

Water was freely applied to the core so as to facilitate logging.

Alizarin Red "S", Potassium Ferricyanide and 10% Hydrochloric acid were used as necessary. Due to the commonly weathered state of the core it was frequently slabbed on site, using a rock saw, in order to reveal a fresh surface.

Grain size was determined using standard sand grains in small clear plastic tubes fixed to a strip of wood. The strip was placed over the sample of interest and what appeared to be the average sized grain matched to grains in one of the tubes. Care was taken to include in each tube grains at both ends of the particular range for that tube e.g. in the case of a medium grain sandstone care would be taken to include grains of size 0.25 mm. and of 0.5 mm. in the tube. The amount of matrix
was determined visually. Although colour in the rock cores was noted, recourse was not made to colour charts for the determination of colour since this is not considered a very reliable indicator of depositional environments (Flugel 1982). Situations where colour changes were considered important e.g. in the mudstones associated with the Quartz Pebble Conglomerate are brought out in the text.

After the preparation, staining and examination of thin sections, slabs and peels the information gained, along with that from the check list, was plotted onto lithological graphic logs on a scale of 1:76 (4 cm. to 10 feet). These logs were designed to bring out the following features;

(1) An indication of the vertical grain size profile.
(2) The composition of the lithologies.
(3) An indication of the contact relationships between lithological units.
(4) Physical and biogenetic sedimentary structures.

Lithological units were recognised by features which originated by relatively constant processes or that varied in a uniform manner. The practical difficulties of achieving this (Flugel opp.cit.) are however realised. Features used to identify the tops and bases of these units are described in the relevant chapters.

Some consideration was given to designing a check list suitable for storing in a computer for further analysis (Melton and Ferm 1978, Lehman 1978) but on reflection it was thought that the amount of simplification of rock types involved would have substantially blunted recognition of the depositional environments. This was borne out subsequently when it
was realised how many rock types were present in the Keel cores. Whilst some of the differences were quite subtle and probably incapable of incorporation in a computer scheme e.g. increasing amounts of silt in the Navan Micrite they pointed to quite marked changes in environment e.g. proximity to a shoreline or barrier island.

(c). Sampling.

Due to the large amount of core involved it was felt that statistical sampling of the core (Flugel 1982) was impractical as this would either result in an impossibly large number of thin sections or, if numbers were kept down to a more realistic number, a situation where possibly vital areas of the succession would be missed. Accordingly the core was logged as above and the core sampled only in the following circumstances:

(1). Slabs taken to examine sedimentary structures under laboratory conditions.
(2). Samples for thin sections to point count sandstones.
(3). Slabs for acetate peels of limestones.
(4). Samples for chemical and heavy mineral analysis.
(5). Samples for thin sections for microfacies analysis, cements etc.

Samples taken would enable the true nature of each lithological unit to be ascertained. Some 700 samples were taken of which 386 were further processed to thin sections and 53 to acetate peels. Samples were taken as half cores after the core had been sliced on the rock saw. Whole core was never taken.

Substantial numbers of photographs were taken at Keel, both of the
core and of the limited outcrop.

(d). Problems encountered.

Whilst the procedure for measuring and describing a core sequence is similar to measuring and describing a stratigraphic section in outcrop, problems can arise. Some of the following points apply specifically to Keel core, others refer to core analysis in general.

(1). Due to the limited amount of material available in a core it is not possible to examine lateral lithological variations and lateral aspects of sediment body geology e.g. channels, bars etc. (Cant 1984).

(2). Whilst identity had been lost on some core boxes and ink on some footage blocks had faded, occasionally so as to become indecipherable, retention of core identity was, in the main, good. However confusion can arise from the different methods of packing the core boxes practised by different companies. Dressler Minerals (DDB holes), for example, placed their core in the box so that each 1.6 metre length started from the bottom of the box with increasing depth going from right to left. ASARCO core (KA) started at the bottom left with the next 1.6 metre starting at the top of the next slot in a manner of packing almost the complete opposite of Dressler core. This packing problem is one of those encountered where more than one company has been drilling in the same license area.

(3). The Keel core was not grooved for orientation. Thus no directional work was possible on structures or fracture patterns. Orientation marks would also greatly assist in correctly fitting the core back together where breaks had occurred.

(4). It is quite often the case that shaley material will be
dropped out of the core barrel (Siemers and Roderick 1981). It was noted at Keel, particularly in the lower clastic units (Lower Quartz Sandstone-Quartz Pebble Conglomerate-Upper Quartz Sandstone), that two lengths of sandstone were separated by an apparently erosive and dirty basal surface on the "up-hole" side. It must be presumed that the shale parting which had previously separated the two beds had been lost.

(5). Grinding of the core at breakage points, probably due to lithological contacts, possibly resulted in lowered core recoveries particularly of shales and mudstones.

(6). Disintegration of the core, particularly shales and mudstones, due to dehydration over time is a marked feature of the Keel core.

(7). Breakage of the core into 1.6 metre lengths for storage, breaking along contacts, mishandling etc. makes the recognition of changes in the direction of cross-bedding very difficult. Types of cross-bedding are very difficult to recognise even under the best conditions. It is usually described in this volume as undifferantuated cross-bedding.

(8). The core, particularly the older material, tended to become oxidised and weathered. In these cases recourse was made to slabbing on site to reveal a fresh surface.

1.7 Petrographic techniques.

(a). Thin sections and peels.

Thin sections were prepared on normal 75 x 25 mm glass slides. Each slide was stained for calcite and dolomite, including ferroan varieties (Dickson 1965). This was to facilitate the observation of cements in both the limestones and sandstones where a carbonate cement was
frequently encountered.

Whilst the sandstones were prepared to the normal thickness of <30 microns the limestones were prepared to a thickness of approximately 50 microns since this is known to facilitate the observation of textures in limestones (Flugel 1982).

Slides were not cover slipped or lacquered since this would prevent restaining of the slide at a future date without the removal of the cover slip or lacquer.

Peels were prepared (Adams et. al. 1984) from stained slabs. Peels were found particularly useful in the study of moderately large samples such as those of crinoidal turbidites in the Calp lithology.

(b). Point counting.

Objectives.

1. The correct terminology of a rock type e.g. sub-litharenite (Pettijohn 1975) of a particular formation can only be made after careful study of a number of thin sections using point counting techniques.

2. Identification of the types of quartz present in a sandstone and the percentage of these types will give valuable indications of the provenance of that sandstone e.g. stretched quartz showing undulate extinction would indicate that at least some of the quartz was derived from a high grade metamorphic terrain.

Similar considerations apply to any lithic fragments or feldspars present in the sandstone e.g. a significant percentage of the orthoclase feldspar microcline would indicate a granitic provenance. The degree of weathering of such feldspars may give valuable indications as to the
transport history of the sediment and to the prevailing palaeoclimate.

(3). Point counting of the lower clastic units and the Upper Mixed Beds might indicate if the sandstones of the two formations were derived from a common source i.e if the sandstones of the Upper Mixed Beds were first deposited in the sea via a fluvial mechanism similar to the lower clastic units and then remobilised into a barrier island.

Point counting was carried out using a Swift Model F automatic instrument. Stepping distances of the stage were tailored to the particle size so as to cover the whole slide, without double counting any grain, in 300 counts. The stepping distance was usually 0.5 mm, the distance found most appropriate for the medium grained sandstones (Wentworth 1922) commonly found.

A minimum of 300 grains were counted. This gives an absolute error of between 5 and 6% on 50% of any one particle type (Plas and Tobias 1965). The absolute error declines for percentages lower than 50%.

Where point counting results were used to construct graphs for the establishment of depositional environments e.g. Visher (1969) the point counting data was converted to sieve data by the use of the method of Friedman (1958).

With regards to point counting sandstones straight extinction quartz includes the slightly undulate quartz (1-5 degrees rotation of the stage) of Folk (1968). Undulate extinction quartz involves a 5-20 degree rotation of the stage.

(c). Cathodelumininescance.

Sandstones were examined under cathodeluminescance to assist with provenance studies.
In accordance with accepted practice (Marshall 1977) details of the instrument and of the working parameters are given below;

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Specification</th>
</tr>
</thead>
<tbody>
<tr>
<td>Beam energy</td>
<td>15 KV for sandstones</td>
</tr>
<tr>
<td>Beam energy</td>
<td>10 KV for limestones</td>
</tr>
<tr>
<td>Beam current</td>
<td>4-500 microamps</td>
</tr>
<tr>
<td>Spot diameter</td>
<td>6 mm</td>
</tr>
<tr>
<td>Gun type</td>
<td>Cold cathode</td>
</tr>
<tr>
<td>Ambient gas</td>
<td>Residual air. 0.5 Torr.</td>
</tr>
<tr>
<td>Instrument</td>
<td>Technosyn 8200 Mk.2</td>
</tr>
</tbody>
</table>

Samples were examined under a magnification of x 200. All thin sections for cathodeluminescence work were double polished to a thickness of <30 microns.

As with all the thin sections and peels the microscope used was a Nikon type 104. Photomicrographs were produced using a microscope mounted Nikon FX 35A.

(d). Chemical analysis.

Insoluble residue analysis (Ellingboe and Wilson 1964) and carbonate analysis (Bruce and Harper 1955) were carried out on the calcareous units to illustrate the declining influence of terrigenous sediment sources as the transgression proceeded.

(e). Heavy mineral analysis.

A small programme of heavy mineral analysis utilising 21 samples was carried out as a part of provenance studies on the Upper and Lower Quartz Sandstones and the Quartz Pebble Conglomerate. After the rock had been crushed and sieved bromoform was used as the separating agent (Allman and Lawrence 1972).
1.8. Interpretation of petrographic data.

a). Grain size after Wentworth (1922).


Pettijohns classification was selected since it includes chert among the quartz clasts and not among the rock fragments as does the classification of Folk (1968). Due to the hardness of chert, as compared to many of the fine grained rock fragments, and its significant presence in many of the lithologies it is thought that its inclusion in the quartz fraction gives a truer impression of a depositional environment.


The classification was selected since;

1). It usually gives a good impression of the depositional environment and

2). It lends itself to the rapid identification of limestones in the field.


e). Maturity of sandstones after Folk (1951).


g). Flaser and lenticular bedding after Reineck and Wunderlich (1968).

1.9. Facies interpretation terminology.

a). Carbonate ramp.

Huge carbonate bodies built away from positive areas and down gentle regional palaeoslopes. No striking break in slope exists and facies patterns are apt to be wide and irregular belts with the highest
energy zone relatively close to the shore (Wilson 1975).

b). Carbonate platforms.

Huge carbonate bodies built up with a more or less horizontal top and abrupt shelf margins where "high energy" sediments occur (Wilson opp.cit.).

c). Shelf.

An area on the top of a ramp or platform (Wilson opp.cit.).

d). Shelf margin.

The edge of the shelf on a platform (Wilson opp.cit.).

e). Mound.

Equidimensional or ellipsoidal buildup (Wilson opp.cit.).
CHAPTER 2
CHAPTER 2.

SILURIAN.

A. Introduction.

Throughout the study area the basal beds of the Upper Devonian/Lower Carboniferous lie unconformably upon dark grey, green or red mudrocks, siltstones and coarse sandstones of a Silurian age (Patterson 1970). The unconformity is inferred from the pronounced angular nature of the contact as seen in the core (Pl), the marked change in sedimentary petrology and inferred depositional environment, the marked cleavage in the Silurian strata which is not seen in the overlying lithologies and the traces of bedding seen in the Silurian (approximately 80 degree dip) as compared to that in the overlying lithologies at 10-30 degrees (Slowey 1986).

Silurian rocks form outcrops (Fig.2) on both the Keel and Northern Inliers and thus presumably also form their cores.

The thickness of the Silurian in the area is unknown.

B. Lithology.

Essentially the Silurian rocks in the Keel area can be sub-divided into 2 lithofacies.

Lithofacies a) Mudrocks and siltstones.

The mudrocks can contain up to 50% mica. This is predominantly muscovite with subsidiary biotite. Small quartz clasts, occasionally making up 30% of the lithology, are frequently present. The quartz and mica occur in a very fine grained quartzitic groundmass.

The marked cleavage may indicate that the mudstones have been subjected to low grade regional metamorphism, possibly to a slate grade.
Occasional bedding (1-2cm. thick) is present. The silts are generally seen to lie between the mudstones and sandstones.

**Lithofacies b) Coarse sandstones (P2 & 3).**

The clasts of these sandstones are very poorly sorted and mainly angular. They may reach 8mm. long. The sandstones are immature, highly arkosic lithic arenites, as deduced from point counting, and frequently show a well marked upward grading.

An average of the point counting results is given below.

<table>
<thead>
<tr>
<th>Clast</th>
<th>Average %</th>
<th>% Std. Dev.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Monocrystalline straight</td>
<td>9.1</td>
<td>3.4</td>
</tr>
<tr>
<td>extinction quartz.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Polycrystalline straight</td>
<td>2.4</td>
<td>3.2</td>
</tr>
<tr>
<td>extinction quartz.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Monocrystalline undulate</td>
<td>36.2</td>
<td>15.8</td>
</tr>
<tr>
<td>extinction quartz.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Polycrystalline undulate</td>
<td>2.5</td>
<td>2.1</td>
</tr>
<tr>
<td>extinction quartz.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Feldspar.</td>
<td>14.9</td>
<td>7.8</td>
</tr>
<tr>
<td>Rock fragments.</td>
<td>18.0</td>
<td>8.2</td>
</tr>
<tr>
<td>Matrix.</td>
<td>8.0</td>
<td>6.0</td>
</tr>
<tr>
<td>Chert.</td>
<td>8.9</td>
<td>12.1</td>
</tr>
</tbody>
</table>

The high standard deviation figures illustrate the great
variability of the sandstones. The average is a recalculated one after the exclusion of carbonate cements. These cements are composed of non-ferroan calcite occurring in tectonic veins. Occasionally quartz veining is also seen. The chert clasts tend to be the largest present and are well rounded. A distinct orientation of the longer clasts is noted in many of the sandstones. Much of the feldspar present is strongly weathered and forms a pseudo-matrix (Nockolds et.al. 1978). Occasionally mica inclusions are seen in both undulate and straight extinction variety quartz clasts. The rock fragments are fine grained and of a supra crustal origin. Very fine grained material, thought to be chlorite, is common. Where chlorite occurs micas tend to absent.

C. Interpretation.

The fine grained nature and the lack of wave formed sedimentary structures suggest that the mudstones were deposited in quiet waters below storm wave base, taken as 200M. (Ewing 1973).

The coarse sandstones testify to periodic high energy events whereby coarse grained terrestrial sediments were emplaced into the normally low energy regime.

Morris (1983) has suggested that the Silurian rocks at the western end of the Longford-Down Massif (Fig.3) were deposited in 200M. of water and states that the depth may have been as great as 4000m. in some cases.

The occurrence of high and low energy lithologies deposited in deep water argues that at least some of the sediment was deposited by turbidity currents. The graded sandstones mark the division A of Bouma (1962) with the mudstones and siltstones representing divisions B to D.
Unfortunately, the cleavage in the siltstones and mudstones has disrupted any of the diagnostic sedimentary features of divisions B to D originally present. The bulk of the mudstones (division E) represent background deposition from suspension. The presence of Division A indicates that the deposits were of a proximal turbidite origin (Blatt et al. 1980). A pronounced orientation of clasts, as is seen in many of the of the lithic arenites of lithofacies b), has been observed in turbidite deposits (Colton 1967).

The extensive sequences of mudstones, lacking sandstone and siltstone interbeds, indicate deposition from suspension only, with no turbidite influence.

Central Ireland at this time was positioned on the northern margin of the closing Iapetus Ocean, above a subduction zone with oceanic crust moving to the north-west (Phillips and Sevastopolou 1986). Most of the sediment would have been trapped in fore-arc basins lying over the trench (Dickinson and Yarborough 1976). Here the turbidites would probably have been deposited on the outer portions of submarine fans which accumulated at the foot of submarine canyons. These canyons crossed the continental slope from the north-west and deposited their sediment load in the fore-arc basins at the foot of the slope where the gradient eased.

The "background mudstones" of lithofacies a) would have been deposited away from the fans.

Deposition of the turbidites would probably have been "triggered" by seismic shock waves produced as the oceanic plate was subducted.

The presence of a cleavage and of chlorite indicates low-grade
metamorphism of the sediments. Chloritisation of the micas is indicated by the lack of mica in samples containing chlorite.

D. Provenance.

The high percentage of undulate extinction quartz and the presence of mica inclusions within some quartz indicate derivation from a metamorphic terrain. The abundant chert found within the lithic arenites of lithofacies b) may have been reworked from chert bearing sandstones occurring within the Pre-Cambrian rocks of North-West Ireland. However such rocks appear to be absent in the area today. Chert is however abundant in the Ordovician strata of Western Ireland e.g. the Lough Nafovey Group of South Mayo and North Galway (Phillips 1981). This may indicate that Ordovician rocks were undergoing uplift and erosion in Silurian time. The absence of microcline is thought worthy of comment since it tends to be abundant in many overlying units. Its absence indicates that granite was not being eroded at this time. The granitic plutons emplaced during the Caledonian Orogeny, and their associated veins, were therefore not available for erosion. This lack of microcline suggests that the straight extinction quartz may have been derived from recrystallised metaquartzites and gneisses, the presence of mica inclusions in the quartz may confirm this view (Folk 1968).

Thus to summarise the Silurian strata in the Keel area were probably derived from the Pre-Cambrian and Ordovician rocks of the North-West of Ireland being deposited by turbidity currents in fore-arc basins over an active subduction zone.
CHAPTER 3.

DEVONIAN.

Following the closure of the Iapetus Ocean, extensive highlands were created throughout the newly formed Old Red Sandstone Continent (Anderton et al. 1979).

Ireland was located on the southern margin of this continent (Fig.10) and by Upper Devonian shallow marine sediments were being deposited in the area south of the Cork-Kenmare line i.e. South Munster Basin. (Phillips and Sevastopolou 1986). This was in marked contrast to the area north of this line where more terrestrial fluvial Old Red Sandstone conditions prevailed. A parallel situation existed in mainland Britain with the marine Devonian of South-West Britain and the Old Red Sandstone of the remainder of the country.

The newly uplifted continental masses were subject to erosion by the Late Silurian and throughout the Devonian (Graham 1983). By Upper Devonian times more extensive deposition prevailed. This was the result of the easing of river gradients due to deposition on the fluvial plain to the south and to the wearing down of the source areas.

MICROCONglomerATE.

A. Introduction.

The only deposits of a possible Devonian age in the Keel area are those belonging to the rock unit termed the Microconglomerate (Fig.8).

A Devonian age is assigned to the lithology since it unconformably overlies the Silurian and is succeeded by the Lower Quartz Sandstone or by the Quartz Pebble Conglomerate which are both of a Lower Carboniferous age (Phillips and Sevastopolou 1986). Support for a
FIG. 10, APPROXIMATE EXTENT AND PALAEOCLATITUDES OF THE OLD RED SANDSTONE CONTINENTAL LANDMASS IN UPPER DEVONIAN TIMES. (PARTLY AFTER WOODROW ET. AL. 1973)
Carboniferous rather than a Devonian age might be found in the fact that the possibly equivilant Red Beds at Navan (which also unconformably overlie Lower Palaeozoic rocks) have been given a Lower Carboniferous age by MacDermot and Sevastopolou (1972).

Philcox (1984) has placed the Microconglomerate in his Basal Sandstone. In the local Keel nomenclature (Slowey 1986) it is placed in the Basal Clastics. However, in common with the succeeding terrestrial units it is assigned a separate identity in this report.

B. Lithology.

Fine sandstones and siltstones dominate the succession but conglomerates and coarser sandstones are locally very important.

The average thickness in the cores examined, where the lithology is present, is 2.8M. The thickness ranges from 1-15M. and changes rapidly between holes as close together as 70M. (Fig. 11). In 6 holes out of the 34 where the Silurian was reached no Microconglomerate was found.

The Microconglomerate may be divided into 4 lithofacies;

a) Fine grained sandstones and siltstones.

b) Medium to coarse grained sandstones.

c) Clast supported conglomerates.

d) Matrix supported conglomerates.

Lithofacies a) Fine sandstones and siltstones.

The fine grained rocks of this lithofacies are predominantly red in colour although green siltstones are moderately common. Occasionally discrete horizons of fine grained sandstones are associated with the siltstones. Instances of brecciation, associated with concentrically
Fig. II. Microconglomerate, thickness variations and sparse distribution of the lithology in the Keel Area.
laminated growths (vadose pisoliths) and vadose silt, are quite commonly developed in the red siltstones (P4) especially where this lithofacies directly overlies the Silurian. These are thought to represent calcrete horizons. Between the breccia "clasts" and along rootlet traces through the red siltstones the red colour has been altered to green.

Lithofacies a) makes up 47% of the Microconglomerate in the cores examined.

Lithofacies b) Medium to coarse grained sandstones.

Medium to coarse grained sandstones comprise 21% of the succession. The sandstone is frequently conglomeritic. Petrologically the sandstones are poorly sorted, immature lithic arenites. The clasts are angular and of low sphericity. The lithology is occasionally extremely coarse grained.

Point counting results are given below as an average. The high standard deviation figures illustrate the great variability of the sandstones in the Microconglomerate. The average is a recalculated one after the exclusion of the carbonate cement which may be composed of ferroan calcite or of dolomite, either ferroan or non-ferroan.

Point counting conditions are as Chap.1.7b.

<table>
<thead>
<tr>
<th>Clast</th>
<th>Average%</th>
<th>% Std.Dev.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Polycrystalline undulate extinction quartz</td>
<td>9.3</td>
<td>5.4</td>
</tr>
<tr>
<td>Monocrystalline undulate extinction quartz</td>
<td>36.5</td>
<td>20.4</td>
</tr>
<tr>
<td>Monocrystalline straight</td>
<td>14.1</td>
<td>9.2</td>
</tr>
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</table>

33
### Table of Composition

<table>
<thead>
<tr>
<th>Component</th>
<th>Abundance 1</th>
<th>Abundance 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>extinction quartz</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Polycrystalline straight</td>
<td>0.9</td>
<td>0.8</td>
</tr>
<tr>
<td>extinction quartz</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Chert</td>
<td>6.5</td>
<td>5.0</td>
</tr>
<tr>
<td>Rock fragments</td>
<td>20.4</td>
<td>17.4</td>
</tr>
<tr>
<td>Microcline</td>
<td>0.3</td>
<td>0.2</td>
</tr>
<tr>
<td>Plagioclase feldspar</td>
<td>0.5</td>
<td>0.6</td>
</tr>
<tr>
<td>Biotite</td>
<td>0.2</td>
<td>0.2</td>
</tr>
<tr>
<td>Matrix</td>
<td>11.3</td>
<td>11.9</td>
</tr>
<tr>
<td>Muscovite</td>
<td>0.1</td>
<td>0.1</td>
</tr>
</tbody>
</table>

n=8

The rock fragments fall into two categories, angular to well rounded, dark, fine grained supra crustal rock fragments and angular mudstone and siltstone fragments. Many of the supra crustal fragments show a well marked cleavage. The mudstone and siltstone clasts, which comprise the bulk of the rock fragments, are frequently deformed by the former type of clasts implying a greater hardness of the former. It is considered that deformation of the mudstone and siltstone clasts may be responsible for the creation of some of the matrix (pseudomatrix, Nockolds et al. 1978). The clasts are frequently mica rich. The chert clasts frequently show a pre-depositional quartz veining and are thus not authigenic. The plagioclase feldspar is occasionally very heavily kaolinised. The microcline is much "fresher" looking. Jasper clasts are present. Very rarely well rounded 1-2cm. clasts of porphyritic andesite are seen.
P. I. Coarse grained sandstone of the Microconglomerate unconformably overlying the cleaved Silurian. The plane of unconformity is marked (U-U). The sequence youngs to the right.

P. 2. Coarse grained Silurian sandstone. Note the markedly better rounding of the chert clast (Ch) as compared to the angular quartz (Q). Stained. Crossed Polars. Mag. x 50.
P. I. Coarse grained sandstone of the Microconglomerate unconformably overlying the cleaved Silurian. The plane of unconformity is marked (U-U). The sequence youngs to the right.

P. 2. Coarse grained Silurian sandstone. Note the markedly better rounding of the chert clast (Ch) as compared to the angular quartz (Q). Stained. Crossed Polars. Mag. x 50.
*Mag. x 50.*

P.4. *Calorete in the siltstone of the Microconglomerate. Contact to the underlying Silurian is under the hammerhead (Silurian to left). Hammerhead is 11.5 cms. long.*
   Mag. x 50.

P.4. Calcrete in the siltstone of the Microconglomerate. Contact to the
   underlying Silurian is under the hammerhead (Silurian to left).
   Hammerhead is 11.5 cms. long.
P. 5. Imbrication of well rounded quartzite, vein quartz and jasper clasts in the Microconglomerate.

P. 6. Fining-up cycle in the Microconglomerate. The cycle is between the coins. The sequence fines up from quartz clasts of 1.5 cms. diameter (C) to coarse sandstone (S) to medium sandstone (M). At the base is an erosional contact to an underlying mudstone. The sequence youngs to the right.
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P. 7. Microconglomerate. Small fining-up cycles of coarse to fine grained sandstone in siltstones. The sandstones show sharp top and basal contacts. Deposition was probably into subsidiary channels at times of high discharge. The sequence youngs to the right. The small reduction spot (R) may indicate the presence of vegetation in the channel. The hand lens is 4 cms. long.

P. 8. Quartz Pebble Conglomerate. Small exposure on the Keel Inlier. Note the pronounced fining-up and the imbrication of the pebbles. The hammer shaft is 30 cms. long.
P. 7. Microconglomerate. Small fining-up cycles of coarse to fine grained sandstone in siltstones. The sandstones show sharp top and basal contacts. Deposition was probably into subsidiary channels at times of high discharge. The sequence youngs to the right. The small reduction spot (R) may indicate the presence of vegetation in the channel. The hand lens is 4 cms. long.

P. 8. Quartz Pebble Conglomerate. Small exposure on the Keel Inlier. Note the pronounced fining-up and the imbrication of the pebbles. The hammer shaft is 30 cms. long.
Lithofacies c) Clast supported conglomerate.

The conglomerate is a clast supported, moderately well to poorly sorted polymict conglomerate of both an intraformational and extraformational provenance. It makes up 28% of the total Microconglomerate in the cores examined.

Owing to the large clast size of the conglomerate (clasts over 6cm. long are common) point counting of thin sections was considered irrelevant.

Clasts in this lithofacies consist of moderately rounded white quartz and well rounded quartzite frequently green or red in colour. Quartzite clasts can make up 40% of the total clast number. Angular Silurian rock fragments showing a good cleavage are present where the conglomerate directly overlies the Silurian. Well rounded chert and jasper clasts are locally very abundant. Variably sized clasts of mudstone and siltstone are present, often making up 30% of the total clast number. The colour of the conglomerate varies from white to green to red. It is the presence or absence and the colour of the mud and siltstone clasts which is largely responsible for this varying colour.

The matrix of the conglomerate is a coarse grained sandstone similar to that in lithofacies b).

Lithofacies d) Matrix supported conglomerate.

This lithofacies has a very patchy distribution within the Microconglomerate. In the 3 cores in which it appears it makes up 10% of the total. In others it does not appear at all. In total 4% of the Microconglomerate examined is of this lithofacies. It usually consists of subangular black and green mudstone clasts within a mudstone matrix.
The colour is probably a consequence of subaqueous deposition when iron in the sediment would be reduced to the ferrous form. One clast of very carbonate rich material was found, presumably derived from a calcrete horizon of lithofacies a). Calcretised mudstone fragments have been recognised in similar conglomerates (Allen 1983). Very occasionally the matrix supported conglomerate is seen to form the top 7.5cm. of a matrix support (lithofacies d) / clast supported conglomerate (lithofacies c) couplet (Fig.13).

C. Carbonate cementation.

Medium crystalline equant calcite is occasionally present infilling tectonic veins. Rarely it is slightly ferroan in character.

The occurrence of vadose silt in the calcretes has been commented upon.

D. Sedimentary structures.

Lithofacies a)

No structures are visible in the lithofacies apart from a pronounced fining-up from fine grained sandstone to siltstone where interbedded sandstones are present.

Lithofacies b)

Poorly developed cross-bedding is occasionally seen. Where the sandstones are conglomeritic a low angle imbrication is sometimes seen.

Lithofacies c)

A low angle imbrication (P5) is common where the clasts are disc shaped. Rare but well developed fining-upward cycles are developed (P6) with erosive basal contacts.

Lithofacies d)
No sedimentary structures are present apart from an erosive basal contact where the conglomerate overlies mudstones of lithofacies a). There is no clast orientation.

E. Palaeontology.

No body fossils have been detected in the unit. Rootlet traces are found in the siltstones. Burrow mottled sandstones are present indicating that the sandstones were not deposited under anoxic conditions.

F. Interpretation.

The coarse nature of the conglomerate clasts (over 6cms.long) indicates that high stream velocities were dominant on occasions (18 KM./Hr., Sundborg 1956). The presence of abundant siltstones shows that lower energy states were probably common. Fluctuating currents are indicative of fluvial processes (Visher 1972). The absence of body fossils could substantiate this (Sykes 1973) as marine life was prolific throughout the Devonian and Carboniferous. Staining has failed to reveal any trace of calcareous debris.

No bi-polar bedding exists in the Microconglomerate core examined.

It is considered that the Microconglomerate was the product of a fluvial environment.

The Microconglomerate does not meet the criteria for alluvial fans as set out by Collinson (1978. A) and Bull (1972). The proposition that the Microconglomerate is a product of streams draining such alluvial fans is however considered tenable.

The deposit is thought to be the result of streams showing a strong braiding influence. Actual sequences from the core indicating specific
areas of the braided stream environment and thus differentiating it from that of a meandering stream, the other possible environment, are described below.

The fining-upward cycles represent deposition on small bars or in channels. It is highly likely that in such a fluctuating regime the silts at the cycle top, both on the bars and in the channels, would become dessicated and cracked at times of low stage flow. Renewed flow in these circumstances would erode the dessication chips and thus furnish the mudstone and siltstone clasts found in the sandstones (lithofacies b) and conglomerates (lithofacies c and d). The predominantly red colouration of the siltstones and mudstones implies a considerable amount of sub-arid exposure. Iron, resulting from the break-down of ferro-magnesium minerals, remained in or was converted to the ferric state by the oxidising conditions and thus stained the sediment red.

The variability of flow competence is illustrated in Fig.13 when, after the initial fining-upward cycle with conglomerate at the base, the overlying second cycle commences with coarse sandstone.

It is thought that more sustained flow took place in some channels as evidenced by the better sorting, cleaner nature and pebble imbrication of some of the conglomerates of lithofacies c) (P.5). However these cleaner conglomerates are rare in relation to those containing abundant poorly sorted and angular mudstone clasts.

Sequences resulting from overbank deposition

The thick (3-4M.) siltstone horizons occasionally seen are thought to result from either deposition in abandoned braid channels or by
overbank flow. Having regard to the fact that the section shown in P. 4 was deposited directly onto the Silurian and well developed calcrete horizons are present it is considered that this particular deposit resulted from repeated overbank flow directly onto the Silurian. Deposition in an abandoned braid channel would involve the deposition of coarser material prior to the deposition of the silt. A deposit of coarse sand commencing a fining-upward cycle is present, this is the product of renewed flow involving a more major flow than previously. Calcrete horizons have been reported in similarly aged Red Bed lithologies in other areas of the Irish Midlands e.g. Sion Hill (Philcox 1984) and in Old Red Sandstone lithologies in mainland Britain (Allen 1974). Rootlet traces indicate that the area became vegetated. Reducing ground waters later used these traces to percolate through the silt and reduce the red ferric iron to the ferrous state in the immediate vicinity of the trace. However it is also known for reduction along rootlet traces to be caused by humic acid production by the plants themselves (Thompson 1970)

Sequences resulting from deposition in subsidiary channels

Deposition in subsidiary channels is inferred (Fig.14). Initial deposition was onto the Silurian and a thin conglomeritic lag deposit consisting of re-worked Silurian material is present at the base of the sequence. Owing to this fact it is not thought that this channel was of the Secondary type of Doeglas (1962) which involves a pre-existing channel cut off from the main stream. Rather it was probably higher ground than the primary channel only becoming flooded at high discharge when silt would be deposited. At very high discharge sand would be
deposited. Note the erosional basal contact in P.7. The channel would undergo sub-ariel exposure between discharges. Calcrete is found in the red siltstone low down in the sequence so initially at least intervals between discharges were probably quite substantial. Deposits in such subsidiary channels are commonly disturbed by bioturbation soon after deposition (Stanley and Fagerstrom 1974). This may account for the lack of ripple structures the existence of which would normally be expected in this environment. The overlying siltstone is green indicating that the sediment was no longer exposed at all times between discharges. This indicates that either the water table was becoming higher or that there was a more consistent flow of water.

The deposition of the overlying conglomerate indicates that the channel was now fully active and was probably flooded at all times. Deposition of the Lower Quartz Sandstone (Chap. 5) then followed.

Sequences resulting from bartop deposition

The sequence shown in Fig.13 exhibits repeated fining-upward cycles inferred to have been developed on a bar. The red colour of the sediment indicates that the sediment has undergone sub-ariel exposure. This fact would mitigate against an origin in an abandoned channel where the sediment would be under water at some time and thus in a reducing environment. An origin in a "high ground" subsidiary channel such as is mentioned above is unlikely since conglomerate would probably not be deposited at such a height in the braid tract.

Deposition commenced with 18cms. of conglomerate fining upwards into 17cms. of coarse sandstone (Fig.13). The abrupt contact to 4cms. of siltstone indicates that the first cycle (A on Fig.13) was eroded by
renewed flow which then deposited the siltstone (B). The overlying coarse sandstone fining up into siltstone marks the base of another cycle (C). Minor sandstone beds (D and E) probably indicate other cycles. The overall fining upward aspect shows that as the bar was being built higher, it became less prone to flooding except by very large discharges.

**Sequences resulting from debris flow deposits.**

The matrix supported conglomerates of lithofacies d) are thought to be the deposits of debris flows resulting from the undercutting of banks and braid bars. Unconsolidated silt along with dessication chips would be deposited into the stream as a consequence. Where the matrix supported conglomerate forms a couplet with clast supported conglomerates of lithofacies c) it is thought that the clasts of quartz, quartzite etc. low on the bar were deposited into the stream first followed very closely by the overlying silt and dessication chips (Fig.12). Deposition would have been very rapid to prevent mixing of the two types. Undercutting of bars and banks is common in modern braided streams (Kelling 1968, Cant 1978, Rust and Jones 1987). In view of this matrix supported conglomerates would be commonly expected in deposits of this type. However the probable poor preservation potential of such material under renewed flow conditions accounts for its comparative rarity.

As commented upon above the lithology shows a very patchy distribution in the Keel area (Fig.11). This pattern tends to be repeated throughout the Irish Midlands both with regard to whole areas and within those areas. In cores from Sion Hill (Fig.2) Red Beds of a
**Figure 12. Microconglomerate, Debris Flow Deposit, Core K 63.**

**Lithology**

- Red mudstone
- Pebbly red mudstone
- Matrix supported conglomerate
- Clast supported conglomerate
- Clast supported conglomerate merging to green mudstone through pebbly mudstone

**Interpretation**

- Products of undercutting of bar-top.
- Products of undercutting of lower-most portion of the bar.
- Channel lag deposit followed by deposition of mudstone under low discharge conditions.
**Figure 12A. Key to Symbols Used on Graphic Log Sections Throughout This Volume.**

- Clast supported conglomerate.
- Matrix supported conglomerate.
- Sandstone.
- Impure micaceous sandstone.
- Siltstone.
- Mudstone.
- Shale.
- Intraformational lime mudstone or mudstone clasts.
- Extraformational clasts.
- Slumping.
- Flaser bedding.
- Lenticular bedding.
- Planar bedding.
- Lime mudstone.
- Ripples.
- Cross bedding.
- Unconformity.
- Erosion surface.
- Fossiliferous bands.
- Ooids.
- Bi-directional cross bedding.
- Bioturbation.
- Burrows.
- Rootlets.
- Imbricated clasts.
- Beach lamination.
**Figure 12 A continued.**

**Grain sizes.**

- **C** Clay
- **M** Mudstone
- **S** Siltstone
- **F** Fine-grained sandstone.
- **D** Medium-grained sandstone.
- **R** Coarse-grained sandstone.
- **N** Conglomerate.
**FIGURE 13. MICROCONGLOMERATE, BAR-TOP DEPOSIT, CORE K63.**

**LITHOLOGY**

<table>
<thead>
<tr>
<th>Depth down hole (m)</th>
<th>Lithology</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>159</td>
<td>Fine grained sandstone (Cycle E)</td>
<td>Bar being built higher resulting in less energy and thus finer grain sized sediments.</td>
</tr>
<tr>
<td></td>
<td>Fine grained sandstone (Cycle D)</td>
<td>Cycle C</td>
</tr>
<tr>
<td></td>
<td>Coarse-grained sublitharenite fining-up into siltstone.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Siltstone (Cycle B)</td>
<td>Deposition after renewed flow</td>
</tr>
<tr>
<td>1583</td>
<td>Coarse grained sublitharenite</td>
<td>Cycle A</td>
</tr>
<tr>
<td></td>
<td>Clast supported conglomerate.</td>
<td></td>
</tr>
</tbody>
</table>
similar age achieve a thickness of over 41M. whilst at Crossakeel (Fig.2) they are absent (Philcox 1984). At Granard (Fig.2) a patchy distribution similar to Keel is seen (Philcox opp.cit.).

This patchy distribution is held to be the result of the infilling of the uneven Silurian topography.

The lithology is not found on the Northern Inlier indicating that this was a site of erosion at the time. This is in contrast to the Keel Inlier where its presence must indicate that the inlier had little or no topographic expression during deposition of the Microconglomerate.

G. Provenance.

The provenance of the Microconglomerate is believed to be substantially local in character. The mudstone fragments result from the re-working of overbank deposits. Much of the sand sized undulate extinction quartz is re-worked from the Silurian sandstones (P.2 & 3) that are abundant in the area. However it is considered feasible that these clasts may have been derived directly from the Pre-Cambrian and/or Ordovician of North-West Ireland if the drainage basins of the Microconglomerate streams were extensive enough. The angular to well-rounded fine grained cleaved fragments have their origin in the Silurian "slates". The survival potential of well cleaved mudrocks would probably be rather low so a local origin is probable. The jasper and chert probably also originated from these Silurian sediments/metasediments, being moderately common in Silurian rocks exposed at the western end of the Longford-Down Massif (Morris 1983). The well-rounded quartzite clasts (P.5) probably result from reworking of a Silurian conglomerate. The clasts presumably owe their roundness and high sphericity to
FIGURE 14. MICROCONGLOMERATE, DEPOSITION IN SUBSIDIARY CHANNELS, CORE K37.

Depth down hole (m)

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clast supported conglomerate</td>
<td>Full activation of channel.</td>
</tr>
<tr>
<td>Green siltstone fining up into mudstone</td>
<td>In-fill of channel by over-bank flow from fully active channels. Thin sandstones result from very high discharge in these channels. Red colour and presence of calcretes imply exposure. Basal lag</td>
</tr>
<tr>
<td>Red fine grained sandstone</td>
<td></td>
</tr>
<tr>
<td>Red sandy siltstone</td>
<td></td>
</tr>
<tr>
<td>Red siltstone with calcrete</td>
<td></td>
</tr>
<tr>
<td>Conglomerate of re-worked Silurian</td>
<td></td>
</tr>
<tr>
<td>Silurian mudstone</td>
<td></td>
</tr>
</tbody>
</table>
polycyclic working being originally derived from the Dalradian of north-west Ireland (Fig. 15). Turbidity currents are known to have been active in the west of Ireland in Wenlock times carrying gravels southward into deep marine basins (Anderton et. al. 1979).

The similar ratios of straight extinction to undulate extinction quartz (1:3.4 in Silurian and 1:3 in the Microconglomerate) argues for a common origin for both lithologies. However the greater amounts of straight extinction quartz in the Microconglomerate and the presence of moderately rounded, white (vein) quartz and microcline indicates that some of the material had an extrabasinal, probably granitic, origin. The granite body may have been any of those indicated on Fig.16. Since later fluvial systems (Chap.16) are thought to have flowed to the Keel area from the north-west and the drainage pattern at this time was dominantly from the north (MacDermot and Sevastopolou 1972) the Crossdowney and Glenamaddy Granites are considered the most promising areas.

The presence of porphyritic andesite clasts in the Microconglomerate possibly favours the Crossdowney area since andesitic rocks are found in the Lough Acanon and Shercock areas (Stillman 1981) 22 and 40KMs. from Crossdowney respectively (Fig.16).

H. Late Devonian/Early Carboniferous Ireland.

Fig.17 attempts to illustrate the depositional environment of the Microconglomerate in a Late Devonian, Ireland wide context.

Three types of erosional/depositional regime can be identified:

1) Areas of net erosion.
   Stable, usually mountainous areas.

2) Areas of predominant erosion and transport.
Fig. 15. Source areas for the Lower Clastic Units.
FIG. 16. POSSIBLE GRANITIC SOURCES OF SEDIMENT IN THE KEEL AREA.
Relatively thin (0-600M.) deposition.

3) Areas of net deposition.

Areas in section 2) probably developed from section 1) types by a combination of erosion of the Caledonian mountain chains and depositional infilling. This reduced the gradient and competence of streams leading to limited deposition. Since this process would have taken a considerable time it may confirm a Late Devonian age for areas of thin deposition including those at Keel.

A modern analogy of this situation may be that of the present day Himalayas with the High Himalayas represented by mountains such as the precursor of the Longford-Down Massif, foothills by areas such as Keel and the Indus-Ganges plain by the Munster Basin.

It is perhaps of interest to note that the area of the Ganges-Bramhaputra alluvial plain covers some 75,000 sq.KMs. (Coleman 1969) whilst the area of the map shown in Fig.9 is approximately 80,000 sq.KMs..
FIGURE 17. A POSSIBLE PALAEOGEOGRAPHY OF DEVONIAN IRELAND.

Areas of net deposition (over 500 metres)

Areas of predominant erosion and transport

Areas of net erosion, stable and mountainous.

Keel
CHAPTER 4
CHAPTER 4.

Basal Clastics

Philcox (1984) has described the lithologies lying between his Pale Beds and the Silurian unconformity as the Basal Sandstone (Fig. 8).

At Keel these same lithologies are termed the Basal Clastics and may be subdivided by the use of the Quartz Pebble Conglomerate as a marker bed. Due to this ease of differentiation and the inferred difference in depositional environment and petrology of the Lower Mixed Beds this local practice is continued for the purposes of this work.

<table>
<thead>
<tr>
<th>Lower Mixed Beds.</th>
<th>Upper Quartz Sandstone.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basal Sandstone</td>
<td>Carboniferous</td>
</tr>
<tr>
<td>Unit (Philcox 1984)</td>
<td>Quartz Pebble Conglomerate.</td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td>Lower Quartz Sandstone.</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td>Microconglomerate</td>
<td>Devonian</td>
</tr>
</tbody>
</table>

The Microconglomerate has also been placed in the Basal Clastics of Philcox (1984) but as stated in Chapter 3 has been accorded a separate identity in this work.
CHAPTER 5.

Lower Quartz Sandstone.

A. Introduction.

The Lower Quartz Sandstone conformably overlies the Microconglomerate in 8 of the 10 cores examined. In one other core it overlies the Silurian whilst in the remaining cores the Microconglomerate occurs with no succeeding Lower Quartz Sandstone.

Occasionally the portion of the Lower Quartz Sandstone immediately overlying the Microconglomerate contains short (< 10 cms.) lengths of conglomerates resembling the Microconglomerate and thin green shales thus forming a transitional contact.

Short lengths (< 30 cms.) of quartz pebble conglomerate are found on approaching the contact with the overlying Quartz Pebble Conglomerate.

In common with the rest of the Basal Clastics Unit the Lower Quartz Sandstone is of a probable Lower Carboniferous inortis-Siphondella age (Phillips and Sevastopolou 1986) (Fig.9).

B. Lithology.

The average thickness of the Lower Quartz Sandstone is 5.7M.. The varying thickness is reflected in the high standard deviation of 4.6M.. The thickness ranges from 1.8-15.5M.

The unit may be divided into six lithofacies. The percentage of the succession represented by each is given in brackets.

Lithofacies a) Clean sandstones (72.5%)
Lithofacies b) Mudstones (7.0%)
Lithofacies c) Shales (1.0%)

58
Lithofacies d) Dirty sandstones (4.6%)
Lithofacies e) Conglomeritic sandstones (10.02)
Lithofacies f) Conglomerates (5.0Z)

Lithofacies a) Clean sandstones

The petrology is that of a well sorted, angular, mature sublitharenite. The sandstones are usually medium grained but may occasionally be coarse grained.

Point counting results are given below as an average. The average has been recalculated after the exclusion of carbonate cements, usually a ferroan dolomite. The cement may occasionally reach 7Z. Point counting conditions are as Chap. 1.7b.

<table>
<thead>
<tr>
<th>Clast.</th>
<th>Average %</th>
<th>Std.Dev.%</th>
</tr>
</thead>
<tbody>
<tr>
<td>Polycrystalline undulate extinction quartz</td>
<td>5.0</td>
<td>5.3</td>
</tr>
<tr>
<td>Monocrystalline undulate extinction quartz</td>
<td>68.4</td>
<td>16.0</td>
</tr>
<tr>
<td>Monocrystalline straight extinction quartz</td>
<td>12.6</td>
<td>10.0</td>
</tr>
<tr>
<td>Polycrystalline straight extinction quartz</td>
<td>0.3</td>
<td>0.3</td>
</tr>
<tr>
<td>Chert</td>
<td>0.9</td>
<td>1.1</td>
</tr>
<tr>
<td>Rock fragments</td>
<td>6.2</td>
<td>2.8</td>
</tr>
<tr>
<td>Microcline</td>
<td>3.1</td>
<td>1.7</td>
</tr>
<tr>
<td>Plagioclase feldspar</td>
<td>2.6</td>
<td>0.8</td>
</tr>
<tr>
<td>Biotite</td>
<td>0.2</td>
<td>0.6</td>
</tr>
<tr>
<td>Matrix</td>
<td>0.7</td>
<td>1.3</td>
</tr>
</tbody>
</table>
The rock is frequently green in colour. The colour becomes less pronounced with depth in an individual bed. The plagioclase feldspar is variably weathered whilst the microcline is quite fresh only rarely showing kaolinisation along the cleavages. Mica inclusions may be seen in some undulate extinction quartz clasts.

Well developed cross-bedding is intermittantly present throughout the lithofacies. The sandstones tend to fine upwards in any particular bed. An erosional basal contact is common where the sandstones overlie the mudstones of lithofacies b).

Lithofacies b) Mudstones.

The mudstones tend to be green and sandy. No sedimentary structures were observed.

Lithofacies c) Shales

The shales are black and well laminated (3mm. lams.). No red coloured fine grained rocks have been noted in the Lower Quartz Sandstone.

Lithofacies d) Dirty sandstones

The grain size of these sandstones varies from fine to coarse grained. The grains are angular and show moderate to good sorting. They are submature lithic arenites.

Averaged point counting results are given below. Conditions are as Chap.1.7b.

<table>
<thead>
<tr>
<th>Clast</th>
<th>Average %</th>
<th>%Std.Dev.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Polycrystalline undulate extinction quartz</td>
<td>5.3</td>
<td>4.7</td>
</tr>
</tbody>
</table>
Monocrystalline undulate extinction quartz
Monocrystalline straight extinction quartz
Polycrystalline undulate extinction quartz
  Biotite
  Chert
  Microcline
Plagioclase feldspar
  Rock fragments
  Matrix

n=4

Some of the matrix may be a pseudomatrix (Nockolds et al. 1978) developed from the rock fragments. Occasionally the matrix is a silt sized quartz. The majority of the rock fragments are composed of mica rich mudstones. Some fine grained supra-crustal rock fragments are also present.

Two types of sandstones are present in the lithofacies;

1) Coarse grained cross-bedded sandstones with abundant rock fragments (chiefly mudstones) and

2) Fine grained sandstones with appreciable matrix. These types tend to be highly micaceous on the laminations. They show a planar lamination, ripple lamination is occasionally present.

The first (coarser grained) types are found at the bases of beds whilst the second (finer grained) types are found at the tops.
Lithofacies e) Conglomeritic sandstones (Chap 1.8f)

The matrix of the lithofacies is similar to that of the clean sandstones of lithofacies a). The majority of the clasts are of well sorted, angular green mudstone up to 4cms. in length. More rarely present are well rounded quartz clasts up to 6cms. in diameter which tend to become more common as the contact to the overlying Quartz Pebble Conglomerate is approached.

The sandstones are cross-bedded with a dip of 25-30 degrees. Clast numbers and sizes decrease with height in any particular bed.

Lithofacies f) Conglomerates.

Clasts of re-worked Silurian grits are seen where the Lower Quartz Sandstone directly overlies the Silurian. Conglomerates resembling the Microconglomerate occur near the contact of the two lithologies. Conglomerate beds with both mudstone chips and rarer quartz clasts occur throughout the unit. Conglomerates with a majority of quartz clasts tend to become more common as the contact with the Quartz Pebble Conglomerate is approached. No conglomerate bed exceeds 15cms. in thickness.

An erosive basal contact is frequently present. Occasionally the clasts show an imbrication.

C. Palaeontology.

No body fossils were found in the cores examined.

D. Interpretation.

The Lower Quartz Sandstone shows transitional contacts with both the underlying Microconglomerate and the overlying Quartz Pebble Conglomerate. The fluvial origin of the Microconglomerate has been indicated above (Chap.3). It is proposed to demonstrate a fluvial origin
for the Quartz Pebble Conglomerate in Chapter 6. These relationships and
the distribution of the Lower Quartz Sandstone in relation to the
Microconglomerate indicate that the sandstone is of a similar fluvial
origin. Further evidence is furnished by the uni-directional cross-
bedding, lack of flaser bedding, evidence of fluctuating energy (Visher
1972) and the lack of fossils (Sykes 1973). Since quartz arenites are
considered to present tentative evidence of marine deposition (Long
1978) their absence in the Lower Quartz Sandstone may indicate a non
marine depositional environment. An abundance of fine grained
intraclasts (mudstone clasts in the conglomerates) is not consistent
with shallow marine sedimentation (Rust and Jones 1987).

Log-probability plots (Visher 1969) were made of the clast size
distribution as seen in thin section. The thin section sizes were
converted to a sieve size distribution using the method of Friedman
(1958). Figures 18 A and B show the plots. The size interval for the
saltation population and the population break both indicate a fluvial
origin. The data also indicates a fluvial origin when processed via CM
plots (Passega 1964). Unfortunately only 4 samples could be analysed
using the above methods due to compaction of the grains in the other
samples.

The lithology does not fulfill the criteria for deposition in a
A) or in meandering streams (Moody-Stewart 1966, Rust 1978, Reineck and
Singh 1980).

A braided stream environment is proposed for the Lower Quartz
Sandstone. The lack of coarse grained material indicates that the unit
should be placed in the sand dominant braided stream environment of Miall (1978).

The noted cyclicity of the deposits indicates that the South Saskatchewan profile type (Miall opp. cit.) may best describe the Lower Quartz Sandstone. In addition to the cyclicity the South Saskatchewan type may form a gradational proximal-distal relationship with gravelly braided streams of the Donjek type (Miall opp. cit.). This point is considered to be of great importance and is further examined below.

Miall (opp. cit.) places an upper limit of 10\% on gravel content in the South Saskatchewan River profile. However the status of conglomeritic sandstones, which make up 10\% of the total 15\% coarse grained sediment in the unit at Keel, is uncertain in this definition. In any case Miall (opp. cit.) emphasises that the models are not boxes into which a particular river must fit exactly. It may well be that the Lower Quartz Sandstone river at Keel occupied a more proximal position to its source area than does the modern South Saskatchewan River.

Due to the overall rarity of the Lower Quartz Sandstone Markov Chain Analysis (Miall 1973) was carried out using all the relevant core examined (Fig.19). Nowhere in the core examined is such a complete succession actually present. It is probable that certain sub-environments have been included in the analysis which may distort it somewhat. Where this is thought to have occurred reference is made below. Comparison of Figs.19, 20 and 21 shows the close similarities between the Lower Quartz Sandstone, South Saskatchewan River and Battery Point Successions. Whilst with the limited amount of material available it may not not really feasible to distinguish sand flats and channel

64
**Figure 19. Lower Quartz Sandstone, Sequence as generated by Markov Chain Analysis (8 cores used).**

**Lithology**

- Mudstone of Lithofacies B)
- Shales. Lithofacies C)
- Fine grained rippled sandstone of Lithofacies D)

**Interpretation**

- Vertical accretion

- In-channel

- Lithofacies A). Clean sublitharenite. Cross-bedding at base

- Lithofacies E). Conglomeratic sandstone

- Lithofacies F) conglomerate

Channel lag
FIG. 20. COMPOUND BAR SEQUENCE, SOUTH SASKATCHEWEN RIVER, (AFTER MİALL 1978)

FACIES

FL
SR
SP
ST
SE
**Fig. 20 A. Key to Figure 20.**

**SE**
Erosional scours with intraclasts. Crude cross-bedding present.

**ST**
Medium to very coarse grained sandstone. May be pebbly. Trough cross-beds. Originate from dunes.

**SP**
Medium to very coarse grained sandstone. May be pebbly. Planar cross-beds. Originate from bars and sand waves.

**SR**
Very fine to coarse sandstone. Rippled.

**FL**
Sand, silt or mudstone. Fine laminations and very small ripples.

**Grain Sizes.**

- c  Coarse grained sandstone
- s  Medium grained sandstone
- f  Fine grained sandstone
- m  Mudstone
FIG. 21. BATTERY POINT SEQUENCE (AFTER CANT AND WALKER 1976)

<table>
<thead>
<tr>
<th>Facies</th>
<th>Sub-environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>G</td>
<td>Vertical accretion</td>
</tr>
<tr>
<td>F</td>
<td>Bar top</td>
</tr>
<tr>
<td>D</td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>In channel</td>
</tr>
<tr>
<td>B</td>
<td></td>
</tr>
<tr>
<td>A</td>
<td>Channel floor</td>
</tr>
<tr>
<td>SS</td>
<td></td>
</tr>
</tbody>
</table>
Fig. 21 A. Key to Fig. 21 (Battery Point sequence)

SS  Scoured erosion surface overlain by up to 0.25 M of coarse massive sandstone with intermixed mudstone clasts.

A  Coarse sandstone with scattered pebbles. Large but poorly developed trough cross-beds are present.

B  Medium grained sandstone with well defined trough cross-beds which are however smaller than those in A.

C  Large solitary planar cross-beds intercalated in facies A and B.

D  Small planar cross-beds possibly developed from sand waves.

F  Fine grained rippled sandstone or interbedded rippled sandstone and mudstone.

G  Fine sandstone with low angle to flat lamination.

Grain size scale

c  Coarse grained sandstone.

s  Medium grained sandstone

f  Fine grained sandstone

m  Mudstone
P. 9. Finingup character of the Quartz Pebble Conglomerate as seen in the core. The sequence lies above an erosion surface.

P. 10. Quartz Pebble Conglomerate. Shows the high proportion of quartzite clasts (Q) as compared to vein quartz (V). The imbrication indicates transport from the right (south). Note the erosion surface (E-E). The hammer shaft is 30 cm. long.
P. II. Quartz Pebble Conglomerate. Typically sized exposure as seen on the Keel Inlier. Hammer shaft is 30 cms. long.

P.II. Quartz Pebble Conglomerate. Dolomitisation. Note the mudstone clast outlining the shape of a former quartz clast.
P. I3. Quartz Pebble Conglomerate. Dolomitisation. Note the remnants of a mudstone bed (M) confirming the replacement nature of the dolomitisation. The sample is No. 432.

P. I4. Quartz Pebble Conglomerate. Thin section of sample 432. Note the quartz grains "floating" in a sea of dolomite. Quartz (Q). Photomicrograph. Stained. Mag. x50. A grain of microcline is present. Both it and the quartz grains show signs of corrosion. The thin section has been prepared very thick to better show the quartz under crossed Polars.
P. 13. Quartz Pebble Conglomerate. Dolomitisation. Note the remnants of a mudstone bed (M) confirming the replacement nature of the dolomitisation. The sample is No. 432.

P. 14. Quartz Pebble Conglomerate. Thin section of sample 432. Note the quartz grains "floating" in a sea of dolomite. Quartz (Q). Photo-micrograph. Stained. Mag. x50. A grain of microcline is present. Both it and the quartz grains show signs of corrosion. The thin section has been prepared very thick to better show the quartz under crossed Polars.
sequences (Cant 1978) in the core at Keel due chiefly to the small overall percentage of fine grained material and the lack of lateral continuity of cores an attempt has been made below (Fig.22) to illustrate deposition on a vegetated island.

The South Saskatchewan River model of Miall (1978) can be considered equivalent to the Devonian Battery Point Formation of Cant (1978). Correlations of the various terms used in the literature referring to braided streams are shown in Table 1 below. Some of the correlations are occasionally rather approximate.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>1) Clean sandstones of Lithofacies a</td>
<td>Facies Sp, St</td>
<td>Facies C, B2</td>
<td>B and D.</td>
</tr>
<tr>
<td>2) Mudstones of Lithofacies b</td>
<td>Facies Fl</td>
<td>Facies F (part)</td>
<td>C</td>
</tr>
<tr>
<td>3) Shales of Lithofacies c</td>
<td>Facies Fl</td>
<td>Facies F (part)</td>
<td>C</td>
</tr>
<tr>
<td>4) Dirty sandstones of Lithofacies d</td>
<td>Facies Sr, Sh</td>
<td>Facies F</td>
<td>B1-B4</td>
</tr>
<tr>
<td>5) Conglomeritic sandstones of Lithofacies e</td>
<td>Facies St, Gm</td>
<td>Facies SS and A</td>
<td></td>
</tr>
<tr>
<td>6) Conglomerates of Lithofacies f</td>
<td>Facies Gm, Se</td>
<td>Facies SS</td>
<td>A</td>
</tr>
</tbody>
</table>
Observation of these various facies enables the depositional processes which controlled the deposition of the Lower Quartz Sandstone to be recognised (Figs. 23). The sequence shown, generated by Markov Chain Analysis, is thought to represent deposition in channels and on bar-tops followed by vertical accretion.

The conglomerates of lithofacies f) were deposited over an erosion surface. Conglomerates containing angular mudstone clasts are found overlying scour surfaces in the Hawkesbury Sandstone of the Sydney (Australia) Basin (Conaghan and Jones 1975). As more pebble sizes quartz and quartzite clasts became available these conglomerates became proportionally richer in these type clasts. Lag breccias at the base of sandy braided streams in the Rhondda Beds (facies A of Kelling, 1968) are commonly imbricated.

Occasionally conglomeritic sandstones of Lithofacies e) are found directly overlying the erosion surface. In this case they are considered to represent a poor development of Lithofacies f). In general however the conglomeritic sandstones overlie the conglomerates of Lithofacies f). As such they are thought to be equivalent to facies St of Miall (1978) and facies A of Cant (1978) (Table 1). These conglomeritic sandstones are thought to represent selectively preserved trough cross-beds deposited by sinuous crested dunes which were not eroded away on waning or low stage flow. The coarser grained varieties of Lithofacies d), with their abundant mudstone grains, are thought to represent a finer grained variety of Lithofacies e).

The conglomeritic sandstones of Lithofacies e) are overlain by the clean cross-bedded sandstones of Lithofacies a). Due to the extreme
difficulty of differentiating cross-bedding in cores Lithofacies a) cannot be subdivided. The sandstones were probably deposited by smaller dunes than those involved in Lithofacies e). Lesser amounts were probably deposited by linguoid or transverse bars and by sand waves (Fig. 23).

Although the sparse distribution of the Lower Quartz Sandstone does not permit the identification of bars, as is possible with the Quartz Pebble Conglomerate (Fig. 26), the bars were probably of a transverse type since finer grained and better sorted sediment favours their production (Smith 1970).

Vertical accretion deposits are represented by the finer rippled, dirty sandstones of Lithofacies d). Ripples are only occasionally developed, planar lamination is more usual. It is probable that the mica present has inhibited the production of ripples (Collinson and Thompson 1982). Similar flat laminated sandstones have been found at the tops of sequences in the Battery Point Sandstone (Cant 1978). However in these cases they overlie mudstones and have been interpreted as floodplain deposits (Cant opp. cit.). In the Lower Quartz Sandstone at Keel they underlie the mudstones and are therefore considered to be normal waning flow deposits.

All the above sandstones show a general fining upward character. Although Cant (opp. cit) states that there is little vertical trend in grain size in either the Battery Point Formation or in the Saskatchewan River sand bodies, sandstones fine up in the sandy braided Westwater Canyon Sandstone of New Mexico (Campbell 1976) and in the sandy braided Rhondda Beds (Kelling 1968).
Figure 22, Lower Quartz Sandstone: Deposition on a Vegetated Island, Core K 63.

Lithology

- Clean sublitharenite becoming conglomeritic towards the base.
- Permenant flooding and possible conversion to channel.
- Green sandy mudstone
- Black shale.
- Intermittant flooding of the island.
- Vertical accretion.
- Moderately clean sublitharenite.
- Medium grain micaceous sandstone
- Moderately clean Sla.
- Micaceous sandstone
- Sandy mudstone
- Clean sublitharenite.
- Conglomeritic in base.
- Channel fill. Increase in mica and decrease in clasts reflect decline in velocity as channel shallows and island starts to emerge.
- Channel lag.
FIG. 23. SUB-ENVIRONMENTS IN A BRAIDED STREAM.
(AFTER MOODY-STEWART 1966)

Deposition on (occasionally vegetated) braid bars and islands.

Deposition of coarse sediment in active channels

Deposition of fine grained material in abandoned channels
The increase in the green colour of the sandstones in an upward direction in any particular bed is thought to be the result of increased amounts of green mud matrix being present as energy levels fell.

Cant (1978) has remarked that where only rippled sandstones are present the sequence probably marks a sand flat deposit (Fig. 23). Where the rippled sandstone is overlaid by finer grained lithologies this is thought to represent an overbank deposition on islands above the level of the sand flat (Fig. 23). This is presumably a function of the greater preservation potential of the island deposits.

Figure 22 is an attempt to illustrate what may be an overbank deposit on a vegetated island. The sequence actually appears in the core examined. Unfortunately such is the broken up state of the shale in the core section that rootlet traces could not be discerned. The fact that these mudstones and shales only occur in one core may reflect the fact that there were few islands in the Lower Quartz Sandstone "river" at Keel.

The lower sandstones and conglomerates were deposited by the processes outlined above. The laminations in the shale indicate that when the island was flooded only small amounts of material were deposited at any one time. The shallow water depth over the island and the frictional effects of the vegetation would assist in trapping the sediment. The black colour is due to the high levels of organic matter derived from plants growing on the island. Flooding of islands takes place every 2.2 years, on average, on the South Saskatchewan River (Cant 1978).

The overlying mudstone was deposited during a very major flood. The
green colour of the mudstone indicates that the island was below the water table after this event. The lack of dessication features in the mudstone reinforces this view. Recolonisation by vegetation, except for those liking wet swampy ground, would be prevented. This major flood was probably a response to the increasing amounts of water becoming available as the marine transgression progressed. Aggregation of neighbouring channels would also be a factor.

The succeeding sandstones mark where the island became a channel again. Such re-activations of channels after periods as islands were probably common. Vegetated islands in the South Saskatchewan River stand only 1-2m. above the sand flats (Cant opp.cit.) and would therefore be vulnerable to major floods and subsequent re-activation as channels.

F. Provenance.

The provenance of the Lower Quartz Sandstone is dealt with in Chapter 8. However, it is thought that its relationship with the Microconglomerate is worthy of mention.

As will be stated (Chap.7) the main source area of the lowermost clastic units, bar the Microconglomerate, in the Keel area were the Dalradian metasediments and Caledonian granites of north-west Ireland as opposed to the substantially local provenance of the Microconglomerate. The increased precipitation resulting from the northward moving transgression (Clayton and Higgs 1979) enabled streams to transport sediment from the north-west (Fig.15). Localisation of the Lower Quartz Sandstone streams to channels used by the Microconglomerate meant that no Silurian material was available for erosion although some Microconglomerate may have been re-worked. The lower amounts of chert in
the Lower Quartz Sandstone may have been due to this lack of Silurian exposure.

The only Lower Quartz Sandstone core containing significant amounts of Silurian clasts is the one core where the Lower Quartz Sandstone directly overlies the Silurian.

The more consistent flow of the Lower Quartz Sandstone streams did not allow for as much mudstone deposition and thus less clasts were available from this source.

The increased amounts of microcline indicate increased levels of erosion in the granite source area.

Physically the rivers were of greater size than in Microconglomerate times and the lack of red mudstones indicates that the channels were below the water table for longer intervals thus not allowing long exposure and oxidation of their ferrous iron component.
CHAPTER 6
CHAPTER 6.

Quartz Pebble Conglomerate.

A. Introduction.

This lithology divides the Upper and Lower Quartz Sandstone units. As such it forms a marker bed in the Keel area. The unit generally conformably overlies the Lower Quartz Sandstone.

Its lower boundary is transitional and is taken at an arbitrary junction where lengths of conglomerate of over 30cms. dominate the succession.

Occasionally it may directly succeed the Microconglomerate with apparent conformity and more rarely it may immediately overlie the Silurian.

Quartz conglomerates showing close affinities with the Quartz Pebble Conglomerate, both in petrology and structural setting, are known from other areas of Ireland (Caldwell 1958, Dixon 1972) and from the Spanish Pyrenees (Nagtaagal 1969).

The unit is of a probable Lower Carboniferous inortus-Siphondella age (Phillips and Sevastopolou 1986).

To the west of Keel this unit is composed of 3 separate pebble beds becoming 4 in number at Strokestown (Fig.1) further west (Patterson 1970) unlike the more consistently conglomeritic nature of the succession around Keel.

The relative hardness of this highly quartzitic lithology has preserved the Northern and Keel Inliers and they topographically dominate the surrounding flat countryside.

B. Lithology.
The unit is essentially a pebble conglomerate (Wentworth 1922) with subsidiary conglomeritic sandstones, sandstones and mudstones.

In the cores examined it ranges from 4.5-65 M. in thickness with an average thickness of 21.3 M.

Although clast supported conglomerates (Lithofacies a below) make up the bulk of the unit a total of 6 lithofacies are clearly recognisable. The percentage of the unit represented by each is given in brackets.

Lithofacies a) Clast supported extraformational conglomerates (59\%)
Lithofacies b) Clast supported intraformational conglomerates (<1\%)
Lithofacies c) Conglomeritic sandstone (16\%)
Lithofacies d) Sandstone (21\%)
Lithofacies e) Mudstone and shales (3\%)
Lithofacies f) Calcareous beds (<<1\%)

Lithofacies a) Clast supported extraformational conglomerates.

This Lithofacies is a clast supported, moderately to well sorted polymict conglomerate with a medium to coarse grained sandstone matrix.

The clasts may be divided into 3 catagories;

1). Quartzitic clasts.

Mainly vein quartz and metaquartzite with subsidiary chert and jasper. The clasts are rounded to well rounded but of low sphericity. Most are disc shaped (Zingg 1935). The quartzite and vein quartz clasts range from less than 1 cm. to 20 cm. in diameter. Chert and jasper occur as small (generally less than 4 mm. diameter) granules within the matrix.

2). Angular mudstone clasts.

These may be up to 5 x 3 cm. in size in the cores but probably
occur in larger sizes (exceeding the diameter of the core). They may be green, red or black. The mudstone clasts are occasionally seen to be deformed by the quartzitic clasts. A cleavage is not present.

The numbers of mudstone clasts as a percentage of total clasts varies widely. Considerable lengths of the Quartz Pebble Conglomerate have no mudstone clasts whilst over 40% of the total number of clasts in some short lengths of core may be mudstone and thus approaching a composition similar to the intraformational conglomerates of Lithofacies b). Very rarely carbonate rich clasts are present. These are thought to have been derived from Lithofacies f). They must have a very local origin since the solubility of carbonate in fresh water is high and this fact would preclude a long transport history.

3). Angular Silurian derived fragments.

This provenance is implied from the presence of a cleavage in these fine grained fragments that is lacking in clasts from category 2. A local origin for the clasts is proposed since the survival potential of such fragments in this high energy environment would be doubtful over any length of time. In addition they have only been seen to occur in core from holes where the Quartz Pebble Conglomerate directly overlies the Silurian.

Analysis of the conglomerate has been confined to hand specimens and exposures since the large clast size would make microscopic examination of little value. Thin sections were however prepared from the "milky" vein quartz which confirm its igneous origin via the use of cathodeluminescence.

In the limited number of exposures in the Keel area a distinct
bimodal distribution of clasts by composition was observed. Vein quartz clasts, recognisable by the milky colour imparted to them by abundant fluid inclusions (Folk 1968), were generally of a smaller average size (2-3cms.) as compared to the much larger grey, green and occasionally red metaquartzite clasts (average 14cms.) (Fig.28).

Examination of exposures indicated that approximately 25% of the exposure consisted of vein quartz with metaquartzite making up almost all the rest. The largest clasts were seen to have a diameter of 20cms.

In addition clast sizes were measured along the A (long) axis from photographs, after the method of Rust (1972), taken of 0.5M. square grids at various small exposures on the Keel Inlier. Computation of the total area of the sub 5cm. fraction reveals that this constitutes 33% of the total area. This percentage is made up of vein quartz (See Fig.28).

The main factor influencing clast size is the energy of the transporting medium, which sorts the clasts, not abrasion (Folk opp.cit.). As no difference has been noted in resistance to abrasion between vein (granitic) quartz and metaquartzite it would appear that the bimodal distribution may be inherited from the primary source material and not developed during transport.

The distribution is not thought to result from differential uplift of the source areas but more probably from the fact that the original quartz veins from which the vein quartz was derived were very much smaller, they are frequently a matter of millimeters wide, than the substantial bodies of metaquartzite which gave rise to the quartzite pebbles.

Lithofacies b) Clast supported intraformational conglomerate.
This lithofacies is essentially composed of mainly angular green occasionally red mudstone clasts in a sandstone matrix similar to that of Lithofacies a).

A similar conglomerate is found at the base of cycles in the Triassic of the Cheshire Basin (Thompson 1970).

A marked erosion surface is often found below this lithofacies exhibiting scour and fill structures. These surfaces are most frequently seen overlying the mudstones and shales of Lithofacies e). These surfaces are similar to the erosional contacts described by Allen (1983) but as to whether they are concordant or discordant cannot be established due to the lack of lateral control.

The clasts overlie the erosion surfaces in a layer usually only one clast thick but occasionally this may be two or three clasts thick.

Although volumetrically insignificant the lithofacies is considered to have considerable environmental significance.

Lithofacies c) Conglomeritic sandstone (Chap.1.8f).

The fraction of the sandstone clasts over 2mm. in size is primarily composed of metaquartzite and vein quartz with subsidiary jasper and chert. Mudstone clasts are locally present. The quartzite and vein quartz clasts tend to be of a much smaller average size than those in Lithofacies a). The matrix is a very clean sublitharenite, its petrology is similar to that of Lithofacies d) but is of a generally coarser grain size (coarse grained rather than medium to coarse grained). There is a clear tendency for clast numbers to increase rapidly with depth in any particular bed.

Erosional basal contacts are developed when the sandstones overlie
the mudstones and shales of Lithofacies e).

**Lithofacies d) Sandstone.**

Petrologically the sandstone is a well sorted, very clean, medium to coarse grained mature sublitharenite. Finer grained lithic wackes are occasionally present.

The grains are subangular although pressure solution may obscure some grain boundaries. Subangular grains are present even when protected from compaction effects by the presence of a carbonate cement. In these circumstances only point and occasionally concave-convex contacts are developed.

The results of point counting of the sandstones are given below as a recalculated average (with standard deviation) after the exclusion of carbonate cements. These cements, normally ferroan dolomite, occasionally comprise 35% of the lithofacies. Point counting conditions are as in Chap.1.7b.

<table>
<thead>
<tr>
<th>Clast</th>
<th>Average %</th>
<th>% Std.Dev.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Polycrystalline undulate</td>
<td>22.5</td>
<td>30.4</td>
</tr>
<tr>
<td>extinction quartz</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Monocrystalline undulate</td>
<td>50.7</td>
<td>26.2</td>
</tr>
<tr>
<td>extinction quartz</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Monocrystalline straight</td>
<td>5.6</td>
<td>2.7</td>
</tr>
<tr>
<td>extinction quartz</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Polycrystalline straight</td>
<td>2.7</td>
<td>5.9</td>
</tr>
<tr>
<td>extinction quartz</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Chert</td>
<td>11.9</td>
<td>17.1</td>
</tr>
<tr>
<td>Rock fragments</td>
<td>4.3</td>
<td>4.4</td>
</tr>
<tr>
<td>--------------------</td>
<td>-----</td>
<td>-----</td>
</tr>
<tr>
<td>Microcline</td>
<td>1.2</td>
<td>0.9</td>
</tr>
<tr>
<td>Plagioclase feldspar</td>
<td>0.9</td>
<td>0.7</td>
</tr>
<tr>
<td>Biotite</td>
<td>0.2</td>
<td>0.3</td>
</tr>
</tbody>
</table>

n=7

Chert in the above samples frequently shows pre-depositional veining testifying to its non-authigenic origin. Rock fragments include mudrock fragments of probable local derivation from the mudstones and shales of Lithofacies e) and other supra crustal rock fragments. Stretched quartz clasts of possible gneissic or schistose origin are present. The presence of mica inclusions in some quartz clasts indicate a probable metamorphic source. Whilst the clasts of plagioclase feldspar occasionally show extensive kaolinisation the microcline appears fresh.

When overlying the mudstones and shales of Lithofacies e) the sandstones frequently show erosional basal contacts. Similar erosional contacts may occur when they overlie another sandstone.

**Lithofacies e) Mudstones and shales.**

This lithofacies comprises the fine grained deposits of the Quartz Pebble Conglomerate. The bulk of the material is a green, chloritic sandy mudstone. Subsidiary green shales, black shales and black mudstones are also present. Rare indications of bioturbation are seen, rootlet traces are also present. Mica is prominent in both the mudstones and shales. Pyrite is frequently present, this is assumed to be authigenic (Folk 1968).

Quite commonly thin lenses of a medium to fine grained sandstone similar to those of Lithofacies d) occur within the mudstones. The
mudstones and shales mainly occur in 30cm. to 2M. thick beds but as they frequently show evidence for erosion they may have originally much thicker. In some cases beds may be over 5M. in thickness. Beds of red mudstone are extremely rare in the cores studied but clasts of this material are quite common in the conglomerates of Lithofacies a).

Lithofacies f) Calcareous beds.

Only 0.9M of this lithofacies was found. It all occurred in one bed in one core. It consists of very carbonate rich seams in a red mudstone giving the red mudstone a distinctly brecciated appearance. Due to the brecciated appearance, lack of possible limestone sources, high carbonate content and the indications of sub-arial exposure given by the red colouration of the mudstone the lithofacies is thought to represent a calcrete horizon.

C. Sedimentary structures

Introduction.

For reasons set out above (Chap.1.6d) the use of sedimentary structures, when combined with petrology, in the establishment of lithofacies (Miall 1977, Rust 1972) is only possible to a limited degree in core from angled holes. The author considers that it is somewhat more feasible in orientated core from vertical holes.

Sedimentary structures as found in each of the lithofacies (a-f) are set out below.

Lithofacies a) Clast supported extraformational conglomerates.

In the rare exposures of the Quartz Pebble Conglomerate examined a clear imbrication of the pebbles of Lithofacies a) is seen (P.8 and 10). An imbrication may also be discerned in the core but is very much less
clear owing to a lack of lateral continuity (P.9). Pebbles are seen to show horizontal bedding in core from vertical holes.

Measurements were made of the pebble imbrication at the exposure where P.8 was taken. Dips averaged 21 degrees, all dips were to the south. The dip of the principal bedding surface was 3 degrees to the south.

A well marked fining upward cycle is present at the above exposure (P.10). Small pebbles (2cms.) occur at the base fining upward into coarse sandstone. Overlying the sandstone is a scoured surface (EE) above which a new cycle commences with large (15cms.+ ) pebbles again fining upward into 6cms. pebbles. The remainder of the cycle is missing. Fining upward cycles are also detectable in the core.

As noted above (Chap.5.A) attempts to measure the orientation of the A axis of the conglomerate pebbles on the Keel Inlier (P.11) met with only limited success due to the low numbers of clasts whose A:B ratio exceeded the 3:1 ratio considered necessary to achieve meaningful results (Tucker 1982). However a rose diagram plotted out from the results indicates that the dominant flow direction of the transporting medium was East to West or visa versa (183-3 Degrees, Arithmetic Mean of the A axis) assuming that the A axis of the pebbles lay transverse to the flow (Collinson and Thompson 1982, Doeglas 1962).

Lithofacies b) Clast supported intraformational conglomerates.

Imbrication and flat bedding are observable in the core. Fining upward cycles similar to those in Lithofacies a) and having a maximum length of 0.75M, are seen in the core where the mudstone clasts are seen to fine upwards into coarse grained sandstone.
Lithofacies c) Conglomeratic sandstone.

Whilst the pebbles are usually scattered through the sandstone in a seemingly random manner they often show a tendency to become less numerous and smaller vertically upwards in any particular bed.

The lithofacies exhibits planar lamination which is frequently more pronounced towards the base of a bed especially when accompanied by an increase in pebble numbers. The planar lamination is frequently succeeded by poorly developed unidirectional cross-bedded sandstones similar to those seen in Lithofacies d).

Lithofacies d) Sandstones.

Planar bedding is developed in this lithofacies frequently being followed by unidirectional cross-bedding. Taking an average grain size of 0.3mm. this would imply a decline in the velocity of the depositing medium from 2.9KM./Hr. (Upper Plane Bed) to 2KM./Hr. (Sand Waves) (Middleton and Southard 1977). This does not take account of the upward decline in grain size also seen. Little ripple lamination has been found succeeding the cross-bedding. Two factors are thought to account for this:

1) In many areas of the postulated environment current velocities would have remained above the critical figure for the creation of ripples.

2) Ripples formed would be vulnerable to erosion by subsequent floods.

It is noticeable that the sandstone becomes finer grained and less clean as the contact with the overlying shales and mudstones of Lithofacies e) is approached.

90
Lithofacies e) Shales and mudstones.

The green mudstones which make up most of this lithofacies are massive. Although there is some disruption in both mudstones and shales, interpreted as bioturbation and/or rootlet traces, this is not extensive. Some of this disruption however could conceivably be caused by water expulsion. Where thin sandstone lenses are developed within the mudstones they show uneven basal contacts. These may have been produced by erosion or by loading effects.

Normally only where Lithofacies e) is present i.e. not eroded away by renewed flow is ripple lamination found in the underlying sandstones of Lithofacies d).

Lithofacies f. Calcareous Beds.

Apart from the erosive top contact caused by an overlying sandstone and the extensive brecciated/disrupted appearance already mentioned no sedimentary structures are present. Rootlet traces are seen and along these the red mudstone has become green probably as a result of porosity changes allowing the circulation of reducing ground water.

D. Carbonate cementation.

Very coarsely crystalline non-ferroan calcite is present as a vein filling cement.

E. Possible carbonate replacement of quartz.

In one core drilled to the north-west of the Keel Inlier (K 59) is seen a section of 6 metres (P.13) of the Quartz Pebble Conglomerate almost completely replaced by very coarsely crystalline slightly ferroan dolomite (P.12). Only eroded clasts of quartz and microcline remain "floating" in a sea of dolomite (P.14). It is noticable in the thin
sections prepared from this material that the bulk of the remaining clasts are composed of microcline. Having regard to the preponderance of quartz clasts, compared to quartz, found in the unaltered Quartz Pebble Conglomerate this would suggest that the microcline was less susceptible to replacement than the quartz.

It is beyond the scope of this report to speculate on the mechanism of the process but some points and inferences may be drawn;

1) Ghosts of former clasts and mudstone bands can be seen in the core (P.12 and 13). This suggests a replacement origin for the dolomite and would preclude an origin for the bed as some kind of calcrete horizon.

2) At a pH of 9-9.5 silica is dissolved and calcite precipitated (Blatt et.al. 1980).

3) A solution rich in calcium would tend to have a high pH i.e. to be alkaline.

It is suggested that hot fluids rich in calcium and magnesium ascended faults in the area. Upon encountering the quartz rich Pebble Conglomerate silica was dissolved and calcium and magnesium almost simultaneously precipitated as the fluid chemistry and/or temperature changed. The quartz was thereby replaced by dolomite. An initial replacement by calcite and a subsequent dolomitisation by a second fluid is also possible.

Fluids of a pH above 9 are rare in nature but Walker (1962) has speculated that such fluids could exist out of contact with the atmosphere and provides evidence of the replacement of chert by carbonate (Walker opp.cit) and the frosting of quartz by carbonate.
replacement (Walker 1957). R.L.Folk (pers. comm.) also envisages the participation of highly alkaline fluids but admits that such relatively large scale replacement is unique in his experience. Banks (1973) has reported the possibly diagenetic alteration of terrigenous mudstones to ferroan calcite.

Since the dolomitisation of the Quartz Pebble Conglomerate is only found in core from one hole it would appear that fluid of the correct chemistry was restricted to the sub-surface of area the to the North-West of the Keel Inlier only.

Fluids ascending faults and carrying magnesium ions derived from basinal shales have been proposed as the mechanism responsible for the creation of the Woo Dale Dolomite of Derbyshire (Schofield and Adams, 1986).

F. Palaeontology.

Although burrowing and rootlet traces are seen in the fine grained sediments of Lithofacies e) and f) no body fossils have been detected throughout the Quartz Pebble Conglomerate at Keel. It is felt that normal marine deposits would show some indication of the prolific fauna usually associated with them (Sykes 1973). No traces of even the smallest shell fragments were detected after staining for calcite. Since fossil fragments are abundantly preserved in overlying lithologies it is thought to be unlikely that the fossil debris was simply dissolved away e.g. to form early fringing cements (Nockolds et.al. 1978) since such early cements have not been observed in the unit and moldic porosity after fossils is not present. Any suggestion that the Quartz Pebble Conglomerate is the product of a stressed marine environment must be
refuted since such environments usually have a prolific if restricted fauna. In addition such environments e.g. lagoons, anoxic deep basins would not have the general high energy conditions under which the conglomerate was obviously formed.

6. Interpretation.

The generally coarse nature of the deposits coupled with structures indicative of velocities within the upper flow regime imply deposition from very powerful currents. The presence of a fine grained lithofacies shows that these currents were of a fluctuating nature. This is stated to be indicative of fluvial processes (Visher 1972). The lack of fossils indicates a non-marine origin (Sykes 1973). The lack of bi-polar cross-bedding, flaser bedding etc. indicates uni-directional flow. The scour surfaces referred to above tend to be associated with fluvial processes (Visher 1972) as do disc shaped pebbles (Folk and Ward 1957).

Thus it is apparent that the Quartz Pebble Conglomerate is of a water lain terrestrial origin.

Only three environments of deposition can be envisaged;

1) Alluvial fans.
2) Meandering streams.
3) Braided streams.

Collinson (1978 A) and Bull (1972) have proposed criteria for the recognition of alluvial fan deposits in the geological record. The Quartz Pebble Conglomerate at Keel does not fulfill these requirements. Therefore an origin in an alluvial fan environment is not considered feasible.

Numerous authorities (e.g. Moody-Stewart 1966, Rust 1978, Reineck
and Singh 1980) have postulated characteristics for meandering streams. Again the Quartz Pebble Conglomerate does not meet these criteria. Therefore an origin as the product of a meandering stream is rejected.

The requirements for a gravelly braided stream origin have been summarised by:

1) Doeglas (1962). Erosion surfaces tend to be more common in low-sinuosity (braided) streams.

2) Moody-Stewart (1966), Thompson (1970). Braided streams have over 50% gravel (2mm.+), whilst having restricted thicknesses of fines.

These points are reinforced by work on the (meandering) Mississippi (Fisk 1947) where true channel deposits are almost absent and on the (gravelly braided) Durance and Ardeche where the alluvium is almost entirely coarse grained (Doeglas 1962). Schumm (1968) states that bedload river channels tend to have less than 5% fine grained material.

3) Fine grained sediments in low sinuosity streams tend to be green/grey in colour in contrast to the red colour of high sinuosity (meandering) streams (Moody-Stewart 1966).

4) Marked vertical changes in grain size are common in braided streams (Long 1978).

With regard to all the above factors a gravelly braided stream environment is proposed for the Quartz Pebble Conglomerate Unit.

Whilst accepting the restraints imposed by the frequently unclear nature of the sedimentary structures present it is now possible to categorise each lithofacies according to the schemes propounded by Miall (1978) and Rust (1978) for braided streams.

It is not thought that a correlation table such as was constructed
in Chapter 5 is appropriate in this case since some of the lithofacies fit into as many as three of the facies of Miall (1978) or of Rust (1978). It is felt therefore that there should be some explanation in the text in these cases.

a) Correlation with the facies of Miall (1978) and of Rust (1978).

Lithofacies a) and b) Clast supported extraformational and intraformational conglomerates.

The imbrication and the flat bedding noted in both the lithofacies indicate that both should be assigned to the Gm facies of Miall (1978). This is stated to be a characteristic deposit on longitudinal bars and of lag and sieve deposits. Rust (1978) has proposed that the notation Se be used for deposits such as Lithofacies b) where intraclasts are present overlying an erosion surface. The cross-bedded gravels of facies Gt and Gp (Miall 1978) may be present but the identification of cross-bedded gravels in cores is extremely difficult (Chap 1.6d).

Lithofacies c) and d). Conglomeritic sandstones and sandstones.

The planar laminated portions of both lithofacies are placed in the Sh catagory of Miall (opp.cit.). The sandstone/conglomeritic sandstone was deposited under conditions of planar bed flow.

The cross-bedded portions may be accredited to any of the facies S1, Sp or St of Miall (opp.cit.). The type of cross-bedding is not capable of sufficient differentiation to more exactly subdivide them.

The rare rippled sandstones are placed in facies Sr of Miall (opp. cit.) deposited under lower flow regime conditions.

Lithofacies e) Mudstones and shales.

The massive mudstones are assigned to facies Fsc of
Miall (opp.cit.) and represent deposition in backswamps.

The finely laminated shales appear to conform to the facies F1 of Miall (opp.cit.), these deposits are the result of overbank or waning flow deposition.

**Lithofacies f) Calcareous beds.**

As mentioned above the deposit is thought to represent a calcrete horizon. These type deposits have been designated P by Miall (opp.cit.) and are interpreted as fossil soils.

Their rare preservation in the fossil record (presumably due to their high solubility) has been commented upon by Allen (1983)

b) **Discussion.**

The dominance of facies Gm, unclassified crossbedded sandstones (Sl, Sp or St) and the presence of Sh sandstones indicates that the Quartz Pebble Conglomerate at Keel may be interpreted as similar to the Donjek River Model of Miall (opp.cit.). This is thought to be typical of most cyclic braided stream deposits and more particularly of distal gravelly streams. The Keel succession may also be equated with the Gill model of Rust (1978) in which fining upward cycles are of great importance and which is also assigned an origin on distal braided streams and braid plains.

It might be thought when comparing the Keel and Donjek successions that the amount of conglomerate in the Keel succession is too low to warrant a Donjek River classification. However it has been proposed (Miall 1978) that if the percentage conglomerate in a complete section lies between 10 and 90%, a Donjek River classification is justified.

Developing the ideas of Miall (opp. cit.) of a graditional
proximal-distal relationship between his models it is thought probable that the proximal equivalent of the Quartz Pebble Conglomerate to the west and northwest i.e. towards its source area (Chap. 8) will prove to be of a Scott River type (Miall opp. cit.). This relationship is also of great importance in the understanding of the environments of deposition of both the Upper (Chap. 7) and Lower (Chap. 5) Quartz Sandstones.

Mialls model brings out the environment of the Quartz Pebble Conglomerate as a distal gravelly braided stream. However the origins of the sub-successions, as seen in the core at Keel, and their relationships to the sub-environments i.e. channels and bars in various areas of the braid tract may best be described in the light of work carried out on the Donjek itself (Williams and Rust 1969) and the Durance and Ardeche Rivers (Doeglas 1962).

To avoid needless repetition the channel types of Williams and Rust (1969) and of Doeglas (1962) are correlated (Table 2) below with the channel types of the author.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>First Order</td>
<td>Primary</td>
</tr>
<tr>
<td>2</td>
<td>Second Order</td>
<td>-</td>
</tr>
<tr>
<td>3</td>
<td>Third Order</td>
<td>-</td>
</tr>
<tr>
<td>4</td>
<td>-</td>
<td>Secondary</td>
</tr>
</tbody>
</table>

Deposition in Type 1 channels.

These deposits comprise extensive lengths of clast supported extraformational conglomerates of Lithofacies a) with minor amounts of the clast supported intraformational conglomerates of Lithofacies b). They were deposited in the main channel where water would run all year.
At low discharge fine grained sands would be deposited in these channels. Should the water velocity allow it the mudstones and shales of Lithofacies e) might even be deposited. However the rarity of such fine grained deposits in the succession may argue for a rather more proximal character than in the ideal Donjek River Model of Miall (1978). No movement of the coarse grained sediment would take place at this low discharge stage.

At high discharge the coarse grained sediments start to move and the finer grained sediments are progressively winnowed away.

At falling stage the coarser grained material will gradually become stationary from movement as bed load. Sand will settle and fill in the interstices between the pebbles. If sufficient sand is available it will form a graded, cross-bedded layer on top. Finally, the finest grained material will be deposited over this. If the supply of pebbles from upstream should exceed the rate of removal then new layers of pebbles will be formed on the bed (Doeglas 1962). The fine grained sediment will have a very low preservation potential (Long 1978) and it is probable that it and the sand on top of the pebbles plus most of the sand between the pebbles will be eroded away at the next flood stage when the pebbles start to move. Only pebbles deep in the pile, very large pebbles and sand infilling porosity very deep in the sediment pile will not be eroded or moved and will remain in situ.

Miall (1977) makes the point that since large bedforms are the least susceptible to erosion the deposits of large floods are the most likely to be preserved.

Inspection of Fig.29 reveals areas where the Quartz Pebble
Conglomerate is only thinly developed. In addition very little fine grained material is found in cores from these holes. These areas are thought to mark the sites of Type 1 channels. Whilst braided rivers commonly have two or more channels (Miall 1977) a single dominant channel is usually apparent (Rust 1972). These two categories of channels are indicated on Fig. 29.

Deposition in Type 2 channels.

Deposits in these type channels, which cross the major bars (Fig. 24), are not clearly recognizable in the core at Keel. However since these channels only show widespread fluvial activity at flood stages and are topographically higher than Type 1 channels it must be assumed that a smaller proportion of the coarse material i.e. pebbles would be deposited in them. They are probably represented by those core sections with rather more sandstone and fine grained material than the norm.

Deposition in Type 3 channels.

Similarly with Type 2 channels the identification of these type channels in core is uncertain. The sediments would be expected to show a greater percentage of fine grained material than Type 2 channels.

Deposition in Type 4 channels.

The deposits in these type channels are mainly composed of the mudstones and shales of Lithofacies e). A prominent feature is the presence of the thin sandstone bands mentioned above. The fine grained deposits show a sharp contact to the clast supported extraformational conglomerates of Lithofacies a) with no indication of fining upwards.

The deposits are thought to result from the blocking of a Type 1, 2 or 3 channel at its upstream end. In this case the presence of the
Fig. 24. Hierarchical organisation of channels and bars. (After Williams and Rust 1969)

1st, 2nd, 3rd (left) = Channel (Orders)

1st, 2nd, 3rd = Bar (Orders)

Fig. 24 Hierarchical organisation of channels and bars

After Williams & Rust (1969)
Fig. 26. Water levels and current directions in active and upstream cut-off secondary channels of braided rivers, modified after Doeglas (1962).
underlying conglomerate may favour the blocking of a Type 1 channel as was originally established by Doeglas (1962). Denny (1967) suggests an alternative mechanism whereby a braid channel may be captured by another channel exhibiting rapid headward erosion in the unconsolidated sediment. This would have a similar result in leaving the channel cut off from the main channel.

Except at the downstream end (Fig.26), where free communication is maintained with the Type 1 channel, the Type 4 channel will be dry during low water. At the downstream end fine grained sediment will be brought into the Type 4 channel from the Type 1 channel (Fig.26) to be deposited on the pebbly base of the former Type 1 channel.

At high water stages water may enter the Type 4 channel by overflowing the levee blocking the upstream end. Dependant on the velocity of the overflow sands of variable grain size will enter the channel over the levee. Upon reaching the standing body of water in the channel, the level of which will also have risen, the sands will be deposited over the fine grained material. Upon cessation of the flow the deposition of fine grained sediment via the downstream end from the Type 1 channel will resume.

This type of deposit is moderately common in the Quartz Pebble Conglomerate at Keel. It is chiefly distinguished by fine grained material overlying much cleaner and coarser sediment with no indication of a gradational relationship and its greater thickness when compared to those mudstone sections which resulted from deposition on bars.

At Keel the thickest sequence ascribed to the infilling of a Type 4 channel argues for a channel depth of 5.9M. however 2m. is a more
Figure 27. Quartz Pebble Conglomerate. Deposition in a Type 4 Channel.

<table>
<thead>
<tr>
<th>Depth down hole (m)</th>
<th>Lithology</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>19.3</td>
<td>Medium grain sublitharenite</td>
<td>Re-activation of channel.</td>
</tr>
<tr>
<td></td>
<td>Grey-green mudstone. Occ. sandy beds.</td>
<td>Mainly deposition from downstream end.</td>
</tr>
<tr>
<td></td>
<td>Red micaceous sandy mudstone. Red fine grain sublitharenite</td>
<td>Some input over levee.</td>
</tr>
<tr>
<td></td>
<td>Yellow-green mudstone.</td>
<td>Over-flow over the levee.</td>
</tr>
<tr>
<td></td>
<td>Sandy mudstone with occ. shale</td>
<td>Deposition from downstream end.</td>
</tr>
<tr>
<td></td>
<td>Lithic wacke.</td>
<td>Final input from upstream end.</td>
</tr>
<tr>
<td>22.4</td>
<td>Medium grained sublitharenite</td>
<td>Partial re-opening of the channel.</td>
</tr>
<tr>
<td></td>
<td>Coarse grained conglomeritic sandstone. Black mudstone</td>
<td>Deposition from downstream end.</td>
</tr>
<tr>
<td></td>
<td>Lithofacies a) conglomerate</td>
<td>Normal channel lag.</td>
</tr>
</tbody>
</table>
P. 15. Contact of the Upper Quartz Sandstone and the Lower Mixed Beds. Shows the abrupt change of sedimentary "style". Sudden change in grain size and colour. Inch rule as scale.

P. 16. The slightly erosional nature of the contact between the Upper Quartz Sandstone and the Lower Mixed Beds. Note the lack of transitional lithologies. The lines indicate the position of this sequence in the portion shown in P. 15.
P. 15. Contact of the Upper Quartz Sandstone and the Lower Mixed Beds. Shows the abrupt change of sedimentary "style". Sudden change in grain size and colour. Inch rule as scale.

P. 16. The slightly erosional nature of the contact between the Upper Quartz Sandstone and the Lower Mixed Beds. Note the lack of transitional lithologies. The lines indicate the position of this sequence in the portion shown in P. 15.

P. 18. Navan Micrite. Preferential dolomitisation of burrows in a lime mudstone. The dolomitising fluids may have gained entry to the sediment via the large burrow on the right. Stained. Plain light. Mag. x 50. Burrow denoted by B.

P. 18. Navan Micrite. Preferential dolomitisation of burrows in a lime mudstone. The dolomitising fluids may have gained entry to the sediment via the large burrow on the right. Stained. Plain light. Mag. x 50. Burrow denoted by B.
P. 19. Navan Micrite. Serpulid worm tubes showing an early isopachous cement and a later equant cement. The "matrix" has been dolomitised. Stained. Plain light. Mag. x 50.

P. 20. Navan Micrite. Serpulid worm tubes. Note the lack of shells etc. for the worms to encrust on. This may provide tentative evidence for the former existence of a hard ground. The lack of worm tubes in the core sections to the left and right may indicate that there were reduced sedimentation rates in the serpulid worm section giving rise to the hard ground.
P. 19. Navan Micrite. Serpulid worm tubes showing an early isopachous cement and a later equant cement. The "matrix" has been dolomitised. Stained. Plain light. Mag. x 50.

P. 20. Navan Micrite. Serpulid worm tubes. Note the lack of shells etc. for the worms to encrust on. This may provide tentative evidence for the former existence of a hard ground. The lack of worm tubes in the core sections to the left and right may indicate that there were reduced sedimentation rates in the serpulid worm section giving rise to the hard ground.
The Keel example chosen (Fig.27) illustrates the infilling well.

A) 60cms of Lithofacies a) conglomerate forms the base and represents the Type 1 channel before damming to form a Type 4 channel. The dam/levee starts to form after the deposition of the conglomerate.

B) 3mm. of Lithofacies e) black mudstone. This denotes the first deposit to be carried into the channel from the downstream (open) end.

C) 15cms. of Lithofacies c) coarse grained conglomeritic sandstone marks a partial re-opening of the channel. The 120cms. of Lithofacies d) type medium grained sublitharenite and 2cms of lithic wacke represent the waning flow deposits of the partial re-opening.

D) 90cms sandy mudstone of Lithofacies e). Due to its sandy nature this probably represents the final input via the upstream end under normal circumstances before the levee reaches its full height. After the deposition of this mudstone sediments would only enter the channel via the upstream end during occasions of very high discharge.

E) 90cms. yellow green mudstone. This represents input from the Type 1 channel via the downstream connection.

F) 10cms. red fine grained sandstone followed by 60cms. red micaceous sandy mudstone. These were the products of overflow over the levee. Taking consideration of the green mudstones above and below this horizon and the sharp contacts it would appear that the red colour is not an effect of sub-arial exposure but a pre-depositional feature which has been retained.

G) 170cms. grey green mudstone. Occasional sandier beds occur
FIG. 28. CLAST SIZE DISTRIBUTION IN THE QUARTZ PEBBLE CONGLOMERATE.
FIG. 29. POSITIONS OF CHANNELS AND BARS AS INFERRED FROM ISOPACHYTES OF THE QUARTZ PEBBLE CONGLOMERATE.
indicating minor high water stage flow over the barrier and deposition in the fine sediments derived from the downstream end of the channel. The absence of ripple lamination in the "over-barrier" sandstones may indicate deposition along the convex inner side of the channel (Doeglas 1962).

Modern examples of this type of channel are frequently vegetated and exhibit water escape structures (Williams and Rust 1969). The disruption of shale laminae and rootlet traces in the mudstones of Lithofaciese) may indicate that this was the case at Keel.

Deposition on braid bars.

Two major and three minor bars are seen at Keel (Fig.29). It is probable that these bars are dissected by Type 2 and 3 channels. Fig.29 indicates that the bars at Keel are of a longitudinal type. These are the most abundant bar types in the Donjek River (Williams and Rust 1969) and in gravelly streams in general (Miall 1977). Other bar types e.g. linguoid bars are rare in both cases.

From Fig.29 the largest of the Keel bars measures approximately 450 x 200M. Bars several hundred feet in length are seen on the Donjek River (Williams and Rust 1969) and bars up to 2.5KM. long and 45m. high have been observed in other braided streams (Rust and Koster 1984). These very large types may however represent several coalesced bars. On the Donjek a length/width ratio of 3.9 is found compared to an average figure of 2.7 for the Keel bars.

Initial formation of the bars was probably caused by clast accretion over an obstacle or a channel lag (Miall 1977). Finer material was trapped between the clasts and more bedload was deposited downstream.
Figure 30. Quartz Pebble Conglomerate, Deposition on Bar Tops, Core KA 165.

Lithology

- Highly conglomeritic sandstone.
- Green mudstone.
- Rippled fine-grain sandstone.
- Sublitharenite. Cross-beded.
- Conglomeritic sandstone

Interpretation

- Flooding of bar.
- Overbank flow onto bar.
- Channel deposition passing up into deposition on the bar top.
- Channel deposition showing decline in grain size and flow velocity.
in the lee of the bar. Growth of the bar is thus maintained in a downstream direction. Clast size tends to decrease in a downstream direction as well as with increasing bar height (Smith 1974).

A sub-succession thought to be typical of bar deposition is shown in Fig. 30. Although many core sections of what are construed to be bar deposits are incomplete, presumably due to erosion of the topmost sediments by the next flood, a fairly complete cycle can occasionally be seen.

Sediment movement only occurs on bars at high water stages. Upon waning flow deposition commences.

A) 30cm. of Lithofacies a) conglomerate is deposited over an erosion surface. The clasts show a crude imbrication. The scattered mudstone clasts present are derived from Lithofacies e) material deposited after a previous flood. The mud dried out to form dessication chips. The curled up edges of the chips made it easier for the water to get under the chips and move them.

B) 1m. of cross-bedded sublitharenite (Lithofacies d)). Horizontal bedding is present at the base. The grain size fines upward (coarse to medium).

C) 35cm. of sublitharenite (Lithofacies d)) and conglomeritic sandstone (Lithofacies c)). Although no obvious scour is present the sandstone is of a markedly finer grain than that of the underlying sandstone and is not as clean. In addition 10cm. of conglomeritic sandstone is present at the base. The succeeding 20cm. are cross-bedded and an upward decrease in grain size from medium to fairly fine is noted. The top 5cm. of fine sandstone is rippled. It shows a green
colouration which is presumably the result of an admixture of green mud on deposition. The colour fades with depth to grey/white.

D) 5cm. green mudstone of Lithofacies e). The top contact is an erosional one to the succeeding conglomeritic sandstone.

Positions in the cores where abundant amounts of Lithofacies b) intraformational conglomerates occur are regarded as important since they mark the sites of the commencement of "bar" cycles. They probably owe their origin to occasions when the water velocity was high enough to permit coarse sand to move over the bar but not high enough to move pebbles over it. Dessication chips from the mudstones of Lithofacies e) would be incorporated into a matrix of coarse sand with no admixture of pebbles.

H. Palaeoclimate.

Certain inferences may be drawn from the character of the depositional environment of the Quartz Pebble Conglomerate:

(1) The presence of calcrete horizons indicates that evaporation, at least on occasions, exceeded annual rainfall (Flugel 1982) and that the mean annual temperature was in the range 16-20 Degrees Centigrade and that the annual rainfall was a seasonally distributed but not excessively peaked 100-500 mm. (Allen 1974). Such conditions occur today in latitudes of 35 Degrees or less.

(2) The lack of dune sands hint that the area was probably not desert although the poor preservation potential of such unconsolidated sediments in conditions of renewed flow is appreciated. However desert sands have been recorded from within the sediments of other braided streams deposited in arid conditions (Thompson 1970).
The source area may have been arid and subject to flash floods to provide the material for the braided streams (Collinson 1978. A). However, this does not mean that the depositional area was necessarily so.

The above points tend to agree with the general thinking (Sevastopolou 1981) that Ireland lay near to the Equator in Lower Carboniferous times.

I. The relationship of the Quartz Pebble Conglomerate to the Upper and Lower Quartz Sandstones.

As stated above (Chap.5) and below (Chap.7) both the Lower and Upper Quartz Sandstones units are interpreted as the deposits of sandy braided streams. The fact that the Quartz Pebble Conglomerate, interpreted as the deposit of a distal gravelly braided stream, separates (Fig.8) these two moderately similar lithologies is explainable in terms of tectonic uplift and erosion of the source area.

A proximal to distal relationship between pebbly braided and sandy braided streams has been demonstrated (Miall 1977, 1978). Circumstances in which pebbly braided streams prograde over sandy braided tracts (Johnson and Vondra 1972) and vice versa are well known (Steel and Aasheim 1978, Stewart 1969). Cases of braided streams prograding over meandering streams are also recorded (Halstead and Nanda 1973).

These events have been linked to faulting (Johnson and Vondra 1972) and to mountain building (Halstead and Nanda 1973).

The deposition of the Quartz Pebble Conglomerate is interpreted in similar terms. Increased fluvial power through a marine regression is not considered likely. A regression on a large enough scale to account
for such a dramatic increase in clast size has not been detected in the Irish Dinantian (MacCarthy and Gardiner 1987).

As stated in Chap. 8 the source area for the Quartz Pebble Conglomerate and both Upper and Lower Quartz sandstones lay in the Dalradian of north west Ireland. It is thought that movement along one of the numerous large Caledonian trend faults in the area resulted in uplift which rejuvenated rivers draining the area and possibly increased precipitation. Evidence has been produced for movement (MacDermot et al 1987) on at least some of the faults e.g. the North Curlew and Ox Mountain Faults during this time. This increased power resulted in both the erosion of and the transportation away from the area of much coarser grained material. With regard to the sheetlike form (Rust 1978) of the deposits it is thought that the onset of Quartz Pebble Conglomerate sedimentation marked the establishment of a braid plain as opposed to the braided stream deposits of the Lower Quartz Sandstone. It has been noted (Rust and Koster 1984) that braidplain deposits form in response to major tectonic uplift. The presence of conglomeratic beds within the Lower Quartz Sandstone near its junction with the Quartz Pebble Conglomerate may mark the first manifestation of this uplift or be a response to higher discharges resulting from it since the influence of individual tectonic events is lost away from the immediate area (Rust and Koster opp.cit.).

Thus the coarse Quartz Pebble Conglomerate beds prograded over the deposits of the Lower Quartz Sandstone.

Additional lines of evidence for uplift are as follows;

1) The presence of reworked Cambrian, Ordovician and Silurian
acritarchs in the Courceyan Ballvergin Shale (Clayton et al. 1980) and the differing petrology of the shale (contains angular quartz, feldspars and mica) as compared to the Ferbane Mudstone and Ringmoylan Shale which sharply overlie and underlie it. This is thought to indicate that rapid uplift was taking place in the west of Ireland, probably in Galway or Mayo (Phillips and Sevastopolou 1986). Since the ages of the Ballyvergin Shale and the Quartz Pebble Conglomerate are comparable (Inortatus-Siphonodella conodant zone) it would appear possible that they represent responses to the same tectonic event.

It might be considered that the Ballyvergin Shale represents some kind of fine grained distal marine equivalent of the Quartz Pebble Conglomerate. However perusal of Fig.31 indicates that the two sediment pulses originated in different directions and that no Ballyvergin Shale or sandstone equivalent to the Quartz Pebble Conglomerate is seen in the Ferbane Mudstone at Cloghan or Windmill Hill which lie on a direct line between the depositional areas of the Ballyvergin Shale and Quartz Pebble Conglomerate.

2) Supplementary evidence may be found to the east of Keel where one or more thick sandstones occur in the mainly shaley Basal Transition Beds, thought likely to be equivalent to the Lower Clastic Units at Keel (Philcox 1984). Philcox (op. cit.) considers that these may be tongues of the Lower Clastic Units extending eastwards. One of these sandstones reaches a thickness of 12M. at Sion Hill (Fig.1). It may be of significance that whilst Sion Hill is directly in the path of the Keel "river" were its course to be extrapolated seaward (Fig.32) these same type sandstone interbeds thin to the north and east of Sion Hill e.g. at
FIG. 31. DISTRIBUTION OF THE BALLYVERGIN SHALE AND THE QUARTZ PEBBLE CONGLOMERATE.

B Ballyvergin
C Conglomerate
Q Quartz Pebble
S Templemore
T Silverynnes
W Loughrea
C Cloonan
WH Windmill Hill
BE Mount Belley
D Dunmore
S Strokestown
NC Newtown Cashel
SH Shona Hill
O Oldcastle
B Ballinalack
K Keel

Km
0 20

FIG. 31. DISTRIBUTION OF THE BALLYVERGIN SHALE AND THE QUARTZ PEBBLE CONGLOMERATE.
Figure 32, Course of the "Keel" River Extrapolated to Seaward.

- Sandstone thinning
- Probable palaeo-shoreline
- Course of River
- Oldcastle
- Silvermines
- Loughrea
- Sion Hill
- Keel
- Oldcastle
- Shoreline
- Probable palaeo-highland
- Course of River

KM
Oldcastle they only reach a thickness of 1M. This is held to be an effect of these areas being located away from the mouth of the Keel "river" and thus less effected by the increased amounts of sediments resulting from the postulated uplift.

Finally after deposition of the Quartz Pebble Conglomerate erosion of the uplifted source area combined with the effects of the marine transgression continuing to the south served to reduce the power of the rivers supplying the Keel area with sediment. The Quartz Pebble Conglomerate deposits retreated back up the sediment tract and sandy braid conditions were restored. The pebble beds in the immediately overlying Upper Quartz Sandstone probably owe their origin to those factors outlined above for the pebble beds in the Lower Quartz Sandstone.
1) MICROCONGLOMERATE

Ephemeral streams reworking Silurian sedimentaries. Deposition by braided streams and possibly alluvial fans in the Silurian peneplain. Only limited transport involved.

2) LOWER QUARTZ SANDSTONE

Increased precipitation giving increased and more sustained stream power. Erosion of metamorphics in North West Ireland commences. Cleaner and finer sediments result.
3) QUARTZ PEBBLE CONGLOMERATE

Movement on fault gives streams the power to erode, transport and round large quartz clasts.

4) UPPER QUARTZ SANDSTONE

Massif eroded down and effects of marine transgression: Reversion to Lower Quartz Sandstone conditions.
CHAPTER 7.

Upper Quartz Sandstone

A. Introduction.

This lithology conformably overlies the Quartz Pebble Conglomerate. The boundary to the Quartz Pebble Conglomerate is taken at an arbitrary junction where lengths of conglomerate over 30cms. long cease to appear.

In all cores examined the Upper Quartz Sandstone succeeds the Quartz Pebble Conglomerate.

In common with the Lower Quartz Sandstone and the Quartz Pebble Conglomerate the unit is of a probable Lower Carboniferous inortatus-Siphonodella age (Phillips and Sevastopolou 1986).

B. Lithology.

The unit is a mature, well sorted, usually medium grained, sub angular to angular (moderate sphericity) grey/white sublitharenite with subsidiary conglomerates, dirty sandstones, shales and mudstones.

The average thickness of the unit in the cores examined is 23M. (std.dev.9.8M.). The thickness varies from 2-41M.

The Upper Quartz Sandstone may be conveniently divided into 5 lithofacies. The percentage of the core examined which each makes up is given in brackets.

Lithofacies a) Clean sandstones (75\%)
Lithofacies b) Conglomerates and conglomeritic sandstones (0.8\%)
Lithofacies c) Dirty sandstones (12\%)
Lithofacies d) Mudstones (8.9\%)
Lithofacies e) Shales (3.3\%)
Lithofacies a) Clean sandstones.

Point counting results are given below as an average with a standard deviation. The average is a recalculated one after the exclusion of carbonate cements. Cements formed up to 21% of some thin sections. They were mainly composed of ferroan dolomite. The average cement figure was 2.1%.

Point counting conditions are as Chap.1.7b.

<table>
<thead>
<tr>
<th>Clast</th>
<th>Average %</th>
<th>±Std.Dev.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Polycrystalline undulate extinction quartz</td>
<td>3.7</td>
<td>2.2</td>
</tr>
<tr>
<td>Monocrystalline undulate extinction quartz</td>
<td>76.7</td>
<td>19.0</td>
</tr>
<tr>
<td>Polycrystalline straight extinction quartz</td>
<td>0.1</td>
<td>0.3</td>
</tr>
<tr>
<td>Monocrystalline straight extinction quartz</td>
<td>2.6</td>
<td>2.1</td>
</tr>
<tr>
<td>Plagioclase feldspar</td>
<td>2.1</td>
<td>2.1</td>
</tr>
<tr>
<td>Microcline</td>
<td>3.7</td>
<td>3.6</td>
</tr>
<tr>
<td>Chert</td>
<td>4.0</td>
<td>4.4</td>
</tr>
<tr>
<td>Rock fragments</td>
<td>6.0</td>
<td>6.4</td>
</tr>
<tr>
<td>Matrix</td>
<td>0.8</td>
<td>2.1</td>
</tr>
<tr>
<td>Biotite</td>
<td>0.2</td>
<td>0.6</td>
</tr>
<tr>
<td>Muscovite</td>
<td>0.1</td>
<td>0.5</td>
</tr>
</tbody>
</table>

Although no pyrite was observed in thin section it was noted frequently in the core. The common red weathering of the rock probably owes its origin to oxidation of the pyrite. The pyrite is thought to
have a diagenetic origin since the postulated depositional environment was an oxidising one. Folk (1968) states that practically all pyrite is diagenetic. The Plagioclase feldspar invariably showed some indication of kaolinisation. No rock fragments of possible Silurian origin were noted. The rock fragments were of sandy mudstone, usually red/brown in colour, occasionally green. The fragments were frequently mica rich. An abundant matrix was occasionally present but may have been to some degree a pseudomatrix (Nockolds et al. 1978). The quartz clasts occasionally show sutured boundaries but only where no carbonate cement has developed. Thus it is thought that an early cement prevented compaction of the grains. Rare "stretched" clasts are present. Infrequent overgrowths on the quartz clasts exist. Stylolites are moderately common in the core possibly resulting from dissolution of the carbonate cement.

Lithofacies b) Conglomeritic sandstones (Chap. 1.8f) and conglomerates.

The clasts in the lithology are of both an intraformational and extraformational origin. The intraformational types are green/black angular mudstone fragments up to 3.5 cm. long. The extraformational clasts are of rounded vein quartz and quartzite up to 3.5 cm. in diameter, small (0.5 cm.) jasper and chert clasts are present. The quartzitic clasts occur only near the junction of the Upper Quartz Sandstone with the Quartz Pebble Conglomerate. The mudstone types are found throughout the unit. The numbers and size of clasts show a tendency to decrease upward in any particular bed and often grade downwards into conglomerates. The conglomerates are clast supported, well sorted, polymict types. The matrix of the conglomerates is composed
of sublitharenite similar to the sandstones of Lithofacies a) but somewhat cleaner. Very rarely, well rounded clasts of calcareous rock are present.

**Lithofacies c) Dirty sandstones.**

The lithofacies may be divided into fine grained and coarse grained varieties. The fine grained variety is a lithic arenite occasionally grading into a lithic wacke. It is frequently highly micaceous and occurs towards the top of any particular cycle (Fig.34). A tendency exists for this variety to become more common with proximity to the overlying Lower Mixed Beds (Fig.8).

The coarse grained variety contains abundant mudstone grains. It inevitably occurs near the base of a cycle (Fig.34). As such it probably marks the transition between the conglomerates/conglomeritic sandstones of Lithofacies b) and the clean sandstones of Lithofacies a).

**Lithofacies d) Mudstones.**

The mudstones are normally black or green in colour. Red mudstones are occasionally found. The mudstones tend to be sandy, the amounts of this sand decreases towards the top of any particular mudstone bed. Occasionally small (<0.5cm) quartz clasts are found in the mudstones. Where the mudstones are found in association with the shales of Lithofacies e) they overlie the shales. Interbedded thin (2mm.) rather dirty fine grained sandstones similar to those of Lithofacies c) and showing slightly erosive bases are common. Pyrite is frequently present. Mudstones may form thin (<2mm.) partings within the sandstones. Very rarely irregular carbonate patches are found within the red mudstones giving the mudstones a brecciated appearance. These are interpreted as
Lithofacies e) Shales.

The shales are inevitably black but weather to shades of brown, yellow, green or white. Very occasionally thin interbeds of siltstone/very fine grained sandstone are developed. Rare small (<0.5cm.) quartz clasts are sometimes present.

C. Carbonate cementation.

Occasional micro-crystalline equant ferroan calcite filling pores is found. This was deposited by meteoric or connate water below the water table. The interparticle cements are referred to under lithology.

D. Sedimentary structures.

Lithofacies a) Clean sandstones.

Unidirectional cross-bedding is ubiquitous and is very tentatively identified as mainly of a trough type, some planar cross-bedding is also present. Overturned foresets are occasionally developed. Abundant erosion surfaces are present. The sandstones show a distinct fining upward character in any particular bed.

Lithofacies b) Conglomeritic sandstones and conglomerates.

Planar bedding is occasionally developed in the conglomeritic sandstones.

Lithofacies c) Dirty sandstones.

The finer varieties show rare water escape structures. Planar bedding and ripples are present.

Lithofacies d) Mudstones.

The mudstones are massive and individual mudstone sequences may be up to 57cm. thick. An erosive top contact is common.
Lithofacies e) Shale.

The shales are generally well laminated. Laminations are 1-8 mm thick. The beds may be up to 10 cm thick. An erosive top contact is common.

E. Palaeontology.

No body fossils have been detected in the Upper Quartz Sandstone. Occasional burrowing is seen in the mudstones of Lithofacies d). The burrows are shallow (5 cm.) and do not possess spreiten. However they are too few and of insufficient variety of form to classify them into an association (Frey and Pemberton 1984). Rootlet traces tend to be moderately abundant in the mudstones of Lithofacies d).

Infrequently the clean sandstones of Lithofacies a) show burrow mottling. The burrows are filled with green mudstone. The mottling becomes more common as the the contact with the Lower Mixed Beds is approached.

F. Interpretation.

With the exception of some aspects which are outlined below (Sect. G) the Upper Quartz Sandstone bears a remarkable similarity to the deposits of the Lower Quartz Sandstone. This resemblance is exemplified by a comparison of Figs. 19 and 34. This indicates that the two lithologies may share a common depositional environment i.e. a sandy braided stream interpreted as similar to the South Saskatchewan River model of Miall (1978). The precise mechanism which brought about the return to sandy braided streams from the gravelly braided streams of the Quartz Pebble Conglomerate has been examined in Chap. 6 H. In addition to the above similarity the following features are seen in the Upper...
FIGURE 34. MARKOV CHAIN ANALYSIS OF THE UPPER QUARTZ SANDSTONE UTILISING ALL THE AVAILABLE CORE.

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mudstone of Lithofacies D)</td>
<td>Vertical accretion</td>
</tr>
<tr>
<td>Shales of Lithofacies E)</td>
<td>Bar top deposition</td>
</tr>
<tr>
<td>Fine-grained sandstone of Lithofacies C)</td>
<td></td>
</tr>
<tr>
<td>Cross-bedded medium grained sublitharenite of Lithofacies A)</td>
<td>In channel deposition</td>
</tr>
<tr>
<td>Coarse grained sandstone of Lithofacies C)</td>
<td>Channel lag.</td>
</tr>
</tbody>
</table>
Quartz Sandstone core. The high percentage of cross-bedding is thought to be characteristic of marginal or non-marine environments (Long 1978). The deformation of foresets is thought to be a feature of fluvial environments (Coleman 1969, Thompson 1970). A red colouration of sedimentary rocks (Rust 1978) is characteristic of fluvial environments, particularly braided ones (Long 1978). Erosion surfaces are effects of sandy braided streams (Rust and Jones 1987).

With much more core available than was the case with the Lower Quartz Sandstone it has been possible to detect actual sequences in the core which are characteristic of specific areas of the braid tract. This was something that was very much more difficult in the Lower Quartz Sandstone. These sequences and the sub-environments to which they relate are outlined below. The sub-environments are indicated on Fig.23.

Deposition in channels (Fig.35).

1) 0.4M. highly conglomeritic very clean sandstone. The quartz pebbles decrease in size and number with decreasing bed depth. The bed was created by the in-channel deposition of large dunes.

2) 2.3M. coarse to medium grained very clean sublitharenite. Fines upward. Occasional low angle (trough ?) cross-bedding poorly developed. Occasional quartz pebbles. Caused by the in-channel deposition of moderately sized dunes.

3) 1.2M. medium grained sublitharenite. Low angle (trough ?) cross-beding. Very occasional quartz pebbles. Caused by the in-channel deposition of small dunes.

4) 2cm. quartz-wacke with ripple lamination. Fine-grained and green. This probably represents the final deposit of the channel
Figure 35, Upper Quartz Sandstone, Channel Fill Sequence, Core K59.

Lithology

<table>
<thead>
<tr>
<th>Depth down hole (m)</th>
<th>Very clean coarse grained sublitharenite.</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Grey-green mudstone with thin sandstone beds.</td>
</tr>
<tr>
<td></td>
<td>Red silty mudstone with rootlet traces.</td>
</tr>
<tr>
<td></td>
<td>Red fine grained sandstone</td>
</tr>
<tr>
<td></td>
<td>Yellow-green mudstones with shales. Mudstones sandy in bases.</td>
</tr>
<tr>
<td></td>
<td>Green fine grained rippled quartz wacke</td>
</tr>
<tr>
<td></td>
<td>Medium grained sublitharenite.</td>
</tr>
<tr>
<td></td>
<td>Coarse to medium grain sublitharenite -te. Occasional quartz clasts. Cross-bedded. Very clean conglomerite sandstone</td>
</tr>
</tbody>
</table>

Interpretation

<table>
<thead>
<tr>
<th></th>
<th>Channel re-activation.</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Products of over-bank flow from major active channels. Sandstones results of major discharges in these channels.</td>
</tr>
<tr>
<td></td>
<td>Final deposition of channel fill proper.</td>
</tr>
<tr>
<td></td>
<td>In channel deposition of small dunes.</td>
</tr>
<tr>
<td></td>
<td>In deposition of dunes.</td>
</tr>
<tr>
<td>Basal lag</td>
<td></td>
</tr>
</tbody>
</table>

I31
sequence proper.

5) 4.5M. yellow-green mudstone. This represents the bulk of the overbank deposits but was not the product of one flow. 15cm. lengths of well laminated green shales occur testifying to small floods between larger ones. At the base of the beds separated by the shales the mudstones are sandy and fine up into mudstones proper.

6) 0.2M. red fine grained sandstone.

7) 1.5M. red silty mudstone showing rootlet traces.

Beds 6) and 7) may represent the products, under waning flow, of one event. The Brahmaputra River, interpreted as a sandy braid, can deposit up to 16M. of sands and mud in one flood (Coleman 1969).

The red colour of the beds is probably syn-depositional as it would appear unlikely for discrete horizons with sharp top and basal contacts, to be subject to subariel exposure whilst the mudstones below and above appear to have been deposited under reducing conditions.

8) 1.5M. grey-green mudstone. Thin sandier beds are present occasionally showing erosional bases. These represent flow into the channel at high flood stages when slightly coarser sediments would be carried into the channel. The mudstones represent the low flow stages i.e. "background" stages.

9) 2M. red weathering very clean coarse grained sublitharenite. Occasional quartz pebbles. Probably represents the reactivation of the channel. The lack of a basal scour/erosion surface may indicate that the underlying mudstone was quite well consolidated when the sandstone was deposited since some scouring would be expected if the mudstones were still in a saturated condition.

I32
Letters 1-9 above correspond to those on the right hand side of Fig.35.

Deposition on sand flats (Fig.36).

1) No conglomerate or conglomeritic sandstone is developed at the base.

2) 5.2M. medium grained clean sublitharenite fining upwards. Low angle (trough ?) cross-bedding becoming obscure towards the base. Occassional 15cm. lengths of high angle cross-beds are present. These may be the result of cross channel bars giving planar cross beds. The trough (?) cross-beds result from dunes.

3) 90cm. impure fine grained lithic wacke. Highly micaceous. Planar laminated. The planar lamination may result from the suppression of ripple formation by the abundant mica (Collinson and Thompson 1982). Fining upward is apparent. The lithology becomes dirtier on approaching the contact with the overlying shale.

4) 30cm. black shale. The black colour is probably due to a high organic content from vegetation growing on the sand flat. Fine grained vertical accretion deposits may be up to 1M. thick in the South Saskatchewan River (Cant 1978).

5) 15cm. Impure fine grained lithic wacke. Highly micaceous. Well laminated. Represents a flooding of the island and thus possible conversion of it back to a channel.

G. Differences between the Upper and Lower Quartz Sandstones.

Although both sandstones show a similar depositional environment they do exhibit differences which may elucidate some important climatic and hydraulic contrasts between them.
FIGURE 36. UPPER QUARTZ SANDSTONE, SAND FLAT DEPOSITION
Core DDB 9.

Lithology

- Lithic wackes and lithic arenites.
- Black shale
- Planar laminated micaceous lithic arenite.
- Clean sublitharenite with occasional mudstone partings. Cross bedding appears mainly planar but some of trough type may be present.

Interpretation

- Permanent flooding of the sand flat.
- Occasional flooding of the flat at high discharge.
- Deposition on flat. Decrease in grain size reflects upward growth of the flat and less frequent flooding.
- Deposition on flat at high discharge. Trough cross-beds may represent cross channel bars. Mudstone partings indicate periods of low discharge.
1) The lower amounts of Silurian mudstone in the Upper Quartz Sandstone indicates that little local Silurian strata was exposed and available for erosion.

2) The greater amounts of dirty sandstone, shale and mudstone in the Upper Quartz Sandstone probably indicate lower stream velocities especially in the upper i.e. younger parts. The lower amount of conglomerates and conglomeratic sandstone may confirm this view. Additionally the much greater amounts of cross-bedding in the Upper Quartz Sandstone may indicate a generally lower energy regime. This is opposed to the Lower Quartz Sandstone much of which shows planar bedding. Since the grain size of the sandstones in the both units are comparable the Lower Quartz Sandstone is therefore interpreted as being deposited under higher energy conditions.

It is thought therefore that the Upper Quartz Sandstone river was of lower overall velocity than that which deposited the Lower Quartz Sandstone. This is not thought to have been due to a decreased water level but rather to an easing of the gradient. The river was entering a coastal plain where much lower energy conditions prevailed.

H. Palaeogeography.

Philcox (1984) has commented on the fact that the Basal Transition Beds (Chap.9) are absent at Keel. He speculates that these apparent marine beds thin and eventually die out due to a local thickening of the fluvial Upper Quartz Sandstone and Quartz Pebble Conglomerate in the Keel area. This is believed by the author to be the result of a local palaeohigh. Fig.37 shows the distribution of the Basal Transition Beds. Areas where the beds were not deposited i.e. where Upper Quartz
Fig. 37: Postulated palaeohigh in the Keel area in basal transition beds times.
Sandstone types continued to be deposited, are believed to mark the site of an extensive braid plain in the Newton Cashel-Keel-Ballinalack area. To either side of the braid plain marine conditions prevailed. Whilst these were normal in the Navan and Oldcastle areas (Fig.37), to the west of Keel evaporites were being deposited in the Strokestown area which may therefore have been a shallow gulf or protected by a barrier system.

The maintenance of fluvial conditions in the Keel area at this time indicates the existence of a palaeohigh. It is thought that this may have had an influence on sedimentation patterns right through the Courceyan and into the Chadian.
CHAPTER 8
CHAPTER 8.

Provenance of the Lower Quartz Sandstone, Quartz Pebble Conglomerate and Upper Quartz Sandstone.

Due to the factors set out below the Basal Clastic Units are inferred to have shared a similar provenance.

In the establishment of the location of this provenance five lines of reasoning are used:

1) Sedimentary structures, both on a small and large scale.

2) Pebble shape.

3) Petrology.

4) Heavy minerals and

5) Areal extent and thicknesses of the units.

a) Sedimentary structures.

The fact that the core was not orientated makes the calculation of palaeocurrents from small scale structures such as cross-bedding impossible. However certain larger scale structures may be used for the elucidation of these palaeocurrents.

1) Pebble orientation in the Quartz Pebble Conglomerate.

Pebble and cobble sized clasts are known to preferentially assume an orientation perpendicular to the current (Doeglas 1962, Rust 1972).

As stated above (Chap.7.C) the orientation of the long (a) pebble axis was measured. The arithmetic mean of these readings indicates a provenance area to the east or west of Keel. The wide "spread" of these figures as seen in the rose diagram (Fig.38) must be considered normal as they are in line with those produced by Doeglas (1962, Figs. A-D).

2) Longitudinal bars.
Figure 38. Pebble orientation in the Quartz Pebble Conglomerate.

10 Clast number
90 Direction of long axis
Scale 10 clasts = 1 centimeter,
P. 21. Calp. Calcite (C) resulting from the dedolomitisation of dolomite (D).
Stained. Plain light. Mag. x 50.

P. 27. Bioclastic Limestone Unit. Sponge spines replaced by ferroan calcite. Note the fragment of silicified bryozoan (B) in the top right. Unreplaced spines are present elsewhere in this thin section. Stained. Plain light. Mag. x 50.


P. 24. Shaly Pales. Storm lags. Packstone/siltstone couplets. Packstones have erosional bases and some degree of fining-up is present. Length of sections 70 cms.

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P. 27. Bioclastic Limestone Unit. Sponge spines replaced by ferroan calcite. Note the fragment of silicified bryozoan (B) in the top right. Unreplaced spines are present elsewhere in this thin section. Stained. Plain light. Mag. x 50.

P 21. Calp. Calcite (C) resulting from the dedolomitisatation of dolomite (D).
Stained. Plain light. Mag. x 50.

Longitudinal bars lie normal to the direction of flow (Rust 1972, Miall 1977). Observation of Fig. 29 and calculation of the arithmetic mean of the long axes of the bars indicates that flow was either from the north-west (320 degrees mag.) or the south-east (140 degrees mag.).

3) Stream channels.

Calculation of the arithmetic mean of the orientation of the stream channels shown in Fig. 29 gives results which closely resemble those for the longitudinal bars. They indicate flow from the north-west (315 degrees mag.) or south-east (145 degrees mag.). The range of stream channel orientation has been found to be small at 60 degrees (Williams and Rust 1969) and to be symmetrical about the trend of the composite stream channel i.e. the whole of the river tract including the active and inactive areas (Williams and Rust opp. cit.).

4) Pebble imbrication.

Numerous studies e.g. Doeglas (1962), Rust (1972), Reineck and Singh (1980) have revealed that flat pebbles in a braided stream environment tend to possess an imbrication dipping upstream. Bluck (1967) has demonstrated that this may be the structure showing the least variance of any structure. P.10 shows the only one of the small exposures of the Quartz Pebble Conglomerate examined to reveal an imbrication. Direction and angle of dip of the pebble imbrication was measured at this exposure. Taking account of the local dip the readings indicate a palaeocurrent direction from the south.

In view of factors examined above and below this is considered unlikely. It is known that in some cases imbrication dip is towards the centre of the channel due to the influence of the banks on large scale

I44
eddies (Teissayre 1975).

If this is the case at Keel a palaeocurrent direction from east or west is indicated.

b) Pebble shape.

The clasts of the Quartz Pebble Conglomerate are very well rounded. This feature is consistent with a long transport history (Thompson 1970).

c) Petrology.

The predominance of metaquartzite pebbles, strongly undulate quartz grains and occasional "stretched" quartz argues for a provenance area consisting mainly of high grade metamorphic rocks. Some undulate extinction quartz clasts must undoubtably be derived from the underlying Silurian greywackes over which the streams were flowing. Many of these Silurian rocks were themselves initially produced from the products of the erosion of the Dalradian Highlands to the north and north-west (Anderton et. al. 1979).

The approximate 20/80% "split" between straight and undulate extinction quartz probably reflects the proportions of igneous and metamorphic quartz respectively (Basu et. al. 1975). Cathodeluminescance studies of samples from the Quartz Pebble Conglomerate and Upper and Lower Quartz Sandstones tend to confirm this view.

The presence of vein quartz clasts, straight extinction quartz grains and the alkali-feldspar microcline points to a granitic component within the provenance area although the main mass of the intrusion may not have been exposed at the time, only the veins emanating from it.

The micas, as with the plagioclase feldspars, may have originated
in either the metamorphic or igneous source rocks.

Chert occurs in many Silurian and Ordovician successions throughout Eire e.g. in the Stroketown area (Morris 1983) and in the South Connemara Group of South Galway (Holland 1981). It has been noted in the Silurian at Keel. Thus its provenance probably lies in the Silurian and Ordovician rocks over which the "Keel River" and its tributaries were flowing. A proximal rather than distal source is preferred having regard to the softer nature of chert as compared to other quartz varieties (Folk 1968). The well rounded nature of much of the chert however mitigates against a very local source.

Similarly jasper is common in clastic sedimentary rocks and its derivation from Silurian rocks is thought probable. Jasper occurs in the Lower Palaeozoic rocks in the Stroketown area (Morris 1983) and the jasper may therefore have been derived from this locality.

The clasts of Silurian mudstone found in the Lower Quartz Sandstone and Quartz Pebble Conglomertae are certainly of a local provenance. The long term survival of such clasts in a high energy braided stream environment must be considered very doubtful. A significant fact may be their tendency to appear where the enclosing lithology e.g. Quartz Pebble Conglomerate directly overlies the Silurian. Due to the unweathered nature of the Silurian where this occurs it is considered that these localities were sites of active erosion prior to the deposition of the overlying lithology.

The presence of chloritic mudstones in the Quartz Pebble Conglomerate may be significant since chlorite schists are known to be present in the Dalradian of Western Ireland and were probably much more
common at higher levels of the tectonic pile (Graham 1983).

The abrupt rise in the amounts of Silurian rock fragments noted between the Lower Quartz Sandstone and the Quartz Pebble Conglomerate is believed to be an indicator of the fact that much more of the Keel area became subject to erosion as the Quartz Pebble Conglomerate rivers with their greater eroding power spread out over the plain.

It is thought that very little of the Silurian strata remained available for erosion in the Keel area by Upper Quartz Sandstone times.

It would appear probable that the fine-grained mudstones and shales in the units owe their origin to erosion and attrition by renewed flow of the overbank deposits of the respective units.

T-tests (Chatfield 1978) were performed on all the main components of the three units in a test for any significant difference between them. No significant difference (at the 5% level) was found. It is therefore inferred that the three units shared the same provenance area.

A significant difference, at the 2.5% level, for the Lower and Upper Quartz Sandstones was noted with regard to straight extinction quartz. Analysis of the correlation co-efficient of this parameter indicated that the percentage of straight extinction quartz falls with decreasing age of the sediment (co-efficient = -.658, probability of no correlation 7.03%). Analysis of each individual unit indicated a very strong correlation for straight extinction quartz with regard to the age of the sediment;

<table>
<thead>
<tr>
<th>Unit</th>
<th>Correlation Co-efficient</th>
<th>Probability of no correlation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lower Quartz</td>
<td>-.78</td>
<td>Negligable</td>
</tr>
</tbody>
</table>

I47
<table>
<thead>
<tr>
<th></th>
<th>Co-efficient</th>
<th>Probability</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sandstone</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Quartz Pebble</td>
<td>-.77</td>
<td>Negligable</td>
</tr>
<tr>
<td>Conglomerate</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper Quartz</td>
<td>-.32</td>
<td>33.3%</td>
</tr>
<tr>
<td>Sandstone</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Whilst a trend is noted for the percentage of undulate extinction quartz to rise with decreasing age of the sediments (co-efficient .398, probability of no correlation 13.1%) the trend is not nearly as strong as with straight extinction quartz. It is thought likely that the discrepancy may be accounted for by increased amounts of intraformational rock fragments and chert. The correlation co-efficient for rock fragments/chert and age with regard to the Upper and Lower Sandstone is -.714, the probability of there being no correlation is 5.1%.

To summarise, it appears that the percentage of straight extinction quartz falls with the decreasing age of the sediments. The percentage of undulate quartz, chert and intraformational rock fragments rises with decreasing age. Both these situations apply especially when comparing the Lower Quartz Sandstone / Quartz Pebble Conglomerate to the Upper Quartz Sandstone. It is felt that this increasing importance of undulate (metamorphic) quartz and chert and the corresponding decreasing importance of straight extinction (plutonic) quartz may be due to the postulated fault movement in Quartz Pebble Conglomerate times (Chap.6.h)
effecting the Dalradian metamorphic source area but not effecting the granitic area which must therefore have lain between the fault and Keel.

In the light of the above, it is clear that the three lithologies probably shared the same source area but changes took place in this provenance during the deposition of the units.

d). Heavy Minerals

A small programme of heavy mineral analysis was undertaken on the samples from the lower clastics units (Lower Quartz Sandstone-Quartz Pebble Conglomerate-Upper Quartz Sandstone) and Lower Mixed Beds. The following features were noted:

i) Rutile was present. Whilst this may occur in both igneous and metamorphic rocks, it tends to be unstable in low grade metamorphis. Thus, if the rutile is of metamorphic origin, a high grade terrain is indicated.

ii) Zircon shows a steady increase with a decreasing age of the sediments.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lower Mixed Beds</td>
<td>35%</td>
</tr>
<tr>
<td>Upper Quartz Sandstone</td>
<td>12%</td>
</tr>
<tr>
<td>Quartz Pebble Conglomerate</td>
<td>10%</td>
</tr>
<tr>
<td>Microconglomerate</td>
<td>3%</td>
</tr>
</tbody>
</table>

Zircon may be derived from both igneous and metamorphic rocks. The above figures do however indicate that one of the sources (igneous or metamorphic) was becoming increasingly important as a source of sediment. Having regard to the increase in undulate extinction quartz it is thought that the influence of the metamorphic source was becoming greater.
iii) Where zircon was abundant in the sample, approximately 85% tended to be of anhedral character, with the remainder subhedral. Magmatic zircon tends to have a marked crystal shape, whilst metamorphic zircon tends to be rounded up to very high grades of metamorphism (Blatt et al 1980). Therefore, it would appear that approximately 85% of the sediment is of a metamorphic provenance with the remainder emanating from an igneous terrain. Those figures show a close relationship with those for undulate (metamorphic) and straight (igneous) extinction quartz given above.

e). Areal Extent and Thickness of Lithologies

Fig. 39 shows that in the Keel area the units under discussion tend to thicken to the west (correlation co-efficient -.57, probability of no correlation 8.7%). Thinning is noted to the north (Fig. 39). The lack of the unit along the south of the Longford Down Massif probably indicates that the Massif was not the source of the quartz in the units.

The thinning to the east represents the more distal parts of the braid plain whilst the thinning to the north possibly represents its less active areas.

A similar, though less well developed picture is seen with regard to the Quartz Pebble Conglomerate. Figure 40 shows that Quartz Pebble Conglomerate lithologies tend to be restricted to an east-west belt running through Keel. By far the thickest developments of the lithology are seen at Keel and Ballinalack (Philcox 1984) possibly indicating that the most active areas of the braid tract were located here. This situation is repeated for the entire lower clastic units where the thickest successions are seen in the Keel and Ballinalack areas (Fig.
Fig. 39: Thicknesses of the Lower Clastic Units in the Irish Midlands showing thickening to the West and thinning to the North.
Fig 40. Existence of quartz pebble conglomerate lithologies, Landward.
It is considered possible that the succession at Tynagh, Silvermines and Templemore may be the products of a completely different river system to that at Keel.

The existence of the Quartz Pebble Conglomerate as bands (Chap.6) in the Strokestown area and to the east of Keel, may confirm that these localities were less active areas of the Keel braid plain.

A 46-92M. conglomerate containing pebbles of quartzite and vein quartz occurs in the Carrick Syncline (Caldwell 1958) 40 KM. to the north-west of Keel (Fig.15). It is associated with coarse sandstones and pebbly grits and comprises the base of the Boyle Sandstone Group overlying the Silurian unconformity. Although the Boyle Sandstone Group has been dated as Arundian (C2S1) in age (Coldwell opp.cit.) it may represent a later equivalent of the Quartz Pebble Conglomerate deposited back along the sediment path as the marine transgression proceeded and the source area was further worn down.

Approximately 21M. of associated fans and braided streams (Allanahy Member, Maam Formation) which appears to closely resemble the Scott River model of Miall (1978) outcrop in Clew Bay, County Mayo (Phillips and Clayton 1980).

Its precise age is uncertain but is known to predate the Late Tournasian (Phillips and Clayton opp. cit.). The clasts of the conglomerate include vein quartz, quartzite and chert and are stated to have a provenance in the Dalradian to the north of Clew Bay (Fig.41). The clasts are only slightly rounded (implying a shorter travel history) which may be a significant fact having regard to the well rounded nature of the clasts in the Quartz Pebble Conglomerate.
This conglomerate may represent the Scott River type proximal equivalent to the Donjek River type represented by the Quartz Pebble Conglomerate following the ideas of Miall (1978) of a proximal/distal relationship between the two models.

Provenance area.

Excluding intraformational clasts such as mudstone fragments and clasts of a partially local origin such as chert and jasper it is considered by the author that the major constituents of the three units have their origin in an area of mixed metamorphic and granitic rocks at a considerable distance from Keel and to the west/north-west or east/south-east of it.

Figs.16 & 15 indicate the granites in the Keel area and the metamorphic and granitic areas of Western Ireland which may have contributed quartzitic clasts to the Keel succession. These possible source areas are discussed below:

(1) The Galway Granite must be rejected as a source area since the "Keel river" would be required to;

a) Run east parallel to the seashore (Fig.41) , an unusual occurrence today.

b) Flow across the palaeslope which was to the south.

c) Streams are known to have run to the south during the Devonian.

Provenance areas for Old Red Sandstone deposits in Kerry (Capewell 1951) lay to the north, probably in the area of the Galway coast and may have been even further north.

(2) The Kentstown-Bellewstown Granite must be eliminated as it lies
Fig. 41. Possible courses of the Keel River.

Metamorphics and Granites of Donegal
Metamorphics and Granites of Mayo and Galway
Clew Bay
Galway
Carrick on Shannon
Keel
Longford-Down massif
Leinster massif

OM Ox Mountains
CM Curlew Mountains

Legend:
- : Highland
- : Course of River

Scale:
0 15 30 45 km
0 12 24 36 miles
to the east of Keel and no formations similar to the units under discussion are known at Navan (Philcox 1984) which lies between the granite and the Keel area.

(3) The buried granite pluton (Jacob 1985) north of Cloghan may have been the source of some of the clasts but not all. No metaquartzites are known from holes drilled in the area (Philcox 1984). The Lower Carboniferous overlies Silurian phyllites and greywackes. Thus there is no source for the quartite clasts.

Additionally in the course of marine transgression (Clayton and Higgs 1979) the area would have been cut off from Keel by the advancing sea and the supply of sediment cut off. If so, however this might explain the decline in the amount of straight extinction quartz in the Upper Quartz Sandstone (Chap.7). The Leinster Granite is rejected on similar grounds.

(4) The Crossdowney Granite is rejected as a primary source area since the Basal Clastic Unit thickens away from the Longford Down Massif and not towards it as would be expected if it were a main source area. In addition metaquartzites are not present in the area.

(5) The buried grano-diorite (Brown and Williams 1985) in the Glenamaddy area is a possible source for at least some of the straight extinction quartz at Keel. However, the lack of a source of quartzite rules it out as a major provenance area.

The major provenance area for the Basal Clastics Unit is believed to lie in the Dalradian metasediments of North West Ireland (Fig.15).

Major Dalradian quartzites e.g. the Ards Quartzite outcrop in Donegal and North Mayo. (Phillips 1981) and it would seem likely that in
pre-Carboniferous times the Dalradian also outcropped in the area between these two areas since to the south-east of Donegal Bay Carboniferous rocks postdating the lower clastic units lie directly on Pre-Cambrian strata (Dixon 1972). Since however, Caledonian granitoid plutons occur in close proximity to both these quartzite terrains closer identification of the provenance area is impracticable.

However, having regard to the rare "stretched" quartz clasts found (Chap.5.D) the presence of the Annagh Gneiss Complex in North Mayo may be significant.

The distance from the source area to Keel (130KM) might be considered a problem given the size of the Quartz Pebble Conglomerate clasts at Keel. However, cobble sized clasts have been identified in braided streams 300-500km from their source (Collison 1978 A).

To summarise; the primary provenance area for the quartzite clasts and grains in the Upper and Lower Quartz Sandstone and Quartz Pebble Conglomerate Units at Keel lies in the Dalradian rocks which probably formed highlands in what are now Galway and Mayo (Fig.15). The north Mayo area is slightly favoured. However, if the basal Boyle Sandstone is related in some way to the Quartz Pebble Conglomerate then Galway would be indicated.

The vein quartz clasts and grains were derived from the granites of the above areas probably with some input from more local intrusions e.g. the Glenamaddy Granite.

The probable courses of the "Keel River" are shown in Fig.41. The river would have had to flow around the Ox and Curlew Mountains which probably formed highlands at that time (Dixon 1972) and round the south
of the Longford-Down Massif, also highland in Lower Carboniferous times (Anderton et al. 1979). The river is indicated as flowing through one of the several "passes" which probably existed between the Ordovician volcanic hills in the Strokestown area.

**Possible dimensions of the "Keel River".**

The maximum channel depth in the Upper Quartz Sandstone was approximately 17M. This assumes a 60% compaction rate for the mudstones (Selley 1982) and a 15% loss for the sandstones (Blatt et al. 1980). 17M. is a very high figure for the Keel system and for braided streams in general. However it is emphasised that this is an absolute maximum figure for the cores examined. This is compared to an average bankfull depth of 15M. for the Brahmaputra River (Coleman 1969). This is not to suggest however that the Keel River was necessarily of similar dimensions to the Brahmaputra since the average bankfull depth of the River Tana is also 15M and this river is of modest dimensions at only 350KM. long (Collinson 1970) and 1.25KM. wide.

However the river must have been of quite respectable dimensions since in the Keel area its braid plain is a minimum of 2.8KM. wide (Fig.29). The Brahmaputra can be 13KM. wide (Coleman 1969). Using the formula of Long (1968) and a depth of 17M a maximum width of 1020M is estimated for the Keel River.

The main Keel River channel is thought to have been centered on Keel during the deposition of the Quartz Pebble Conglomerate and the Upper Quartz Sandstone due to the decline in thickness of both lithologies to the north-west and south-east of Keel (Fig.39). This trend is confirmed by correlation analysis. In the Keel area the
deposits of the Upper Quartz Sandstone cover an area of approximately 130 sq.KM. This is far from a large figure for a braided stream. The Hawkesbury Sandstone of Australia covers an area of 20,000 sq.KM. (Rust and Jones 1987) and the Westwater Canyon Sandstone of New Mexico may cover 16,000 sq.KM. (Campbell 1976). Both these formations are interpreted as the products of braided streams.
CHAPTER 9
CHAPTER 9.

Basal Transition Beds.

In many areas to the east and west of Keel a series of shales, siltstones, sandstones and minor limestones occur between basal sandstone types such as the Upper Quartz Sandstone and the base of the Navan Micrite (Chap. 11). These are the Basal Transition Beds. The unit thins out in the Newtown Cashel and Moyvore areas and does not appear in the successions at Keel and Granard (Philcox 1984) (Fig. 37). Correlation of the thickness of the lithology with distance from a north-south line drawn through Keel gives a co-efficient of .66 with a probability of no correlation of 4.2% thus confirming the thinning away from Keel. The probable reasons for this thinning are examined in Chap. 8.

It is considered that the Lower Mixed Beds at Keel (Chap. 10) may be time and partially environmentally equivalent to the Basal Transition Beds elsewhere since the Lower Mixed Beds at Keel show marked marginal marine characteristics.
CHAPTER 10.

**Lower Mixed Beds.**

A. Introduction.

Overlying the Upper Quartz Sandstone are the mudstones, variably clean sandstones and sandy limestones of the Lower Mixed Beds.

Philcox (1984) has placed this lithology in the Basal Sandstone Unit (Fig. 8). However for the purposes of this report the Keel practice of assigning the Lower Mixed Beds a separate identity (Fig. 8) has been retained. This is due to the marked difference, on petrological and depositional environment grounds, between the Lower Mixed Beds and the underlying Upper Quartz Sandstone.

The contact with the Upper Quartz Sandstone in the cores examined is very sharp (Ps. 15 and 16). The clean sandstones of the Upper Quartz Sandstone are succeeded by regularly interbedded mudstones and siltstones with less clean and finer grained sandstones.

In the cores examined the Lower Mixed Beds range from 1 to 10.5 metres in thickness with an average thickness of 6.6 metres (standard deviation 3.5 metres).

It is proposed that owing to the marked difference in petrology and sedimentary environment an extended period of non-deposition occurred between the deposition of the Lower Mixed Beds and of the Upper Quartz Sandstone.

Possible evidence for this proposition is examined in Section F.

The unit is interpreted as the result of one of the major marine transgressive cycles outlined by MacCarthy and Gardiner (1987). On the
basis of time and palaeogeography cycle 4 is suggested since in cycle 3 times the coastline lay approximately 120 Km. to the south (MacCarthy and Gardiner opp.cit.). Any marginal marine environment would have been of extremely wide dimensions for the sediments which later formed the Lower Mixed Beds to have formed part of it.

In cycle 5 times shelf limestones were being deposited in the Keel area (MacCarthy and Gardiner opp.cit.).

B. Lithology.

As would be expected, owing to its transitional position between terrigenous and true marine deposits, lithologies within the Lower Mixed Beds tend to be associated in a seemingly random manner. Five main lithofacies may be identified.

<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>% of Lower Mixed Beds.</th>
</tr>
</thead>
<tbody>
<tr>
<td>a) Fine grained sandstones</td>
<td>55</td>
</tr>
<tr>
<td>b) Siltstones</td>
<td>10</td>
</tr>
<tr>
<td>c) Mudstones and shales</td>
<td>33</td>
</tr>
<tr>
<td>d) Sandstone and lime mudstone</td>
<td>2</td>
</tr>
<tr>
<td>e) Lime mudstone</td>
<td>2</td>
</tr>
</tbody>
</table>

N.B. The sandstone/lime mudstone fining upward sequences of lithofacies d) tend to be closely interbedded fine grained sandstones (lithofacies a)) and siltstone (lithofacies b)) with rarer intercalations of lime mudstone (lithofacies e)).

To simplify descriptions and avoid repetition the sedimentary structures of each lithofacies are described under lithology. This procedure will avoid circumstances where, under the system employed in
previous chapters, a sandstone of comparable lithology would have to be described two or three times since its sedimentary structures are very different.

**Lithofacies a) Fine grained sandstone.**

These are fine to (rarely) medium grained sublitharenites. The clasts are subangular and moderately well sorted.

Point counting details are given below as an average after the elimination of cements. These are normally composed of ferroan dolomite. Point counting conditions are as in Section 1.7b.

<table>
<thead>
<tr>
<th>Clast</th>
<th>Average %</th>
<th>Standard deviation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chert</td>
<td>0.3</td>
<td>0.4</td>
</tr>
<tr>
<td>Plagioclase feldspar</td>
<td>2.4</td>
<td>2.0</td>
</tr>
<tr>
<td>Monocrystalline undulate</td>
<td>26.0</td>
<td>6.7</td>
</tr>
<tr>
<td>extinction quartz</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rock fragments</td>
<td>18.0</td>
<td>8.3</td>
</tr>
<tr>
<td>Monocrystalline straight</td>
<td>1.4</td>
<td>0.3</td>
</tr>
<tr>
<td>extinction quartz</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Polycrystalline undulate</td>
<td>1.1</td>
<td>1.5</td>
</tr>
<tr>
<td>extinction quartz</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The rock fragments are of dark mudstone. These are significantly better rounded than the quartz clasts. Occasionally the sandstones may be conglomeratic. In these cases the clasts may be up to 8 mm. long and of an angular shape. Where the conglomeratic sandstones occur in proximity to the lime mudstones the clasts may be of lime mudstone. Microcline having a fresh appearance is present.

Intermittently the lithology can be very clean and similar to the
sandstones of the Upper Quartz Sandstone. However the majority of the lithofacies a) sandstone contains significant amounts of rock fragments (10% +) and matrix.

Banded sequences of clean and matrix rich sandstones are common.

Cross bedding is frequently present, occasionally with reactivation surfaces. Thin "wisps" of mudstone are common on the foresets and are interpreted as flaser bedding. Very well marked 5-8 mm. laminations with heavy minerals in the base of the laminations are seen.

Occasionally very high angle wavy bedding is present with rare minor scour and fill structures. Locally the sandstone are red and green in colour.

Lithofacies b) Siltstones.

These are moderately well sorted with frequent bands and "wisps" of mudstone. The wisps are often composed of lime mudstone where the siltstones occur in close proximity to the lime mudstone. Dolomite (after the lime mudstone) is occasionally present. The grains in the siltstones are subangular. Point counting results are given below after recalculation for the removal of the ferroan calcite cement which may form up to 1% of the lithology.

<table>
<thead>
<tr>
<th>Clast</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Monocrystalline undulate</td>
<td>72.2</td>
</tr>
<tr>
<td>extinction quartz</td>
<td></td>
</tr>
<tr>
<td>Monocrystalline straight</td>
<td>5.2</td>
</tr>
<tr>
<td>extinction quartz</td>
<td></td>
</tr>
<tr>
<td>Matrix</td>
<td>15.7</td>
</tr>
<tr>
<td>Rock fragments</td>
<td>6.9</td>
</tr>
</tbody>
</table>
Muscovite mica is common, especially in the mudstone "wisps", and shows a well marked orientation.

Cross bedding, including herringbone types, is occasionally well developed.

The siltstones tend to be black in colour or more rarely green while red siltstones are uncommon.

**Lithofacies c) Mudstones and shales.**

Although commonly occurring in beds 15-30 cms. thick lengths up to 6.7 metres are occasionally seen. The mudstones are commonly sandy and show lenses of sandstone. The shales are extremely fissile and micaceous. A tendency is noted for beds to have massive sandstone at the base and fissile shale at the top.

Bioturbation is common. The lithology is almost inevitably black or more rarely green in colour. Red mudstones are only locally developed.

**Lithofacies d) Fining upward cycles.**

Occasionally fining upward cycles consisting of clean, fine grained sublitharenites grading up into lime mudstone through siltstones are seen.

**Lithofacies e) Lime mudstone.**

This lithofacies is normally very heavily bioturbated. Abundant moderately well sorted quartz clasts are present in the burrows along with subsidiary dolomite and microcline. The actual lime mudstone also shows dolomitisation effects but these are usually confined to areas adjacent to the burrows. Dolomite is also present in serpulid worm tubes. The mudstone itself contains variable amounts of silt sized quartz grains. These are subangular and show indistinct margins.
Frequently abundant pelloids are present and the lithology may then be classified as a pelloidal packstone.

Calcite filled cavities are present and are similar to the birds-eyes described by Shinn (1968).

This lithology bears a strong resemblance to that of the Navan Micrite (Chap. 11) seen stratigraphically above.

C. Carbonate cementation.

Very finely crystalline pore filling equant ferroan dolomite is noted.

Coarsely crystalline equant ferroan dolomite is seen to postdate baryte mineralisation.

Very coarsely crystalline non-ferroan calcite is present as vein fill.

The above cements are thought to be of a meteoric or connate water origin deposited both above and below the water table.

D. Palaeontology.

The first body fossils of the Keel succession are found in the Lower Mixed Beds. Shelly bands composed of bivalve debris occur in the sandstones of lithofacies a). The debris is well sorted and clasts of lime mudstone are incorporated in the bands. Erosional basal contacts are present. Scattered bivalves are moderately common in both the sandstones of lithofacies a) and siltstones of lithofacies b). These scattered bivalves are not abraded to any degree and show little indication of mechanical sorting e.g. a prefered orientation.

Fossil evidence tends to become more common near the top of the Lower Mixed Beds i.e. close to their boundary with the Navan Micrite.
The lime mudstones/pelloidal packstones of lithofacies e) lack body fossils but micritised serpulid worm tubes are found in the quartz grain infill of the burrows.

The bioturbation is most prominently developed near the top of the Lower Mixed Beds i.e. near the contact with the overlying Navan Micrite and in the lime mudstones. Situations where a heavily bioturbated section is succeeded by core showing little, if any, bioturbation are common.

E. Interpretation.

Features noted in the clastics of the Lower Mixed Beds give valuable indicators as to their environment of deposition.

The presence of bivalve debris implies an aqueous environment, sedimentary structures such as cross bedding and scour and fill may also suggest this.

The presence of fining upward cycles indicate waning flow conditions in subaqueous environments.

Bi-directional flow i.e tidal influences is inferred from the presence of mud "wisps" (interpreted as flaser bedding) in the sandstones and the lenses of sandstone in the mudstones interpreted as lenticular bedding. Both of these features tend to occur in intertidal and subtidal zones (Reineck and Singh 1982) though such features are known to occur in other non-tidal environments (Raaf and Boersma 1971). The fact that occasional flasers of lime mudstone are present indicates proximity to marine environments since no terrestrial carbonates are found in the vicinity of the Lower Mixed Beds. Herringbone cross-bedding
most readily occurs in tidal environments (Collinson and Thompson 1982) and the banded sequences of matrix rich and matrix poor sandstone in lithofacies a) may indicate bi-directional flow.

The presence of mudstone clasts probably indicates some degree of sub-aerial exposure and the burrow mottling of the sandstones seen in lithofacies a) is restricted to shallow marine and estuarine/lagoonal environments (Heckel 1972).

A marine environment is inferred from the presence of heavy minerals in the planar laminations (swash lamination) of the sandstones (lithofacies a)), a feature associated with beaches exposed to wave action (Clifton 1969).

Reactivation surfaces suggest fluctuating energy and are commonly found in tidal flat environments although they are not diagnostic of such (Reineck and Singh 1980).

The presence of the above features suggest a tidal flat environment for the Lower Mixed Beds.

This intertidal environment is demonstrated with examples from the actual core.

1. Channel sequences.

These are well developed in the Lower Mixed Beds. Fig.42 shows a graphic log of such a channel sequence.

At the base of the section is a channel lag (represented by (a) in the log). In this particular example it is made up of rounded mudstone clasts and more angular lime mudstone clasts similar to the polymict intertidal channel lags reported by Roehl (1967). The sorting of the clasts is generally poor.
FIGURE 42. LOWER MIXED BEDS, CHANNEL SEQUENCE, CORE K 64,

**Depth down hole (m)**

<table>
<thead>
<tr>
<th>Depth</th>
<th>Lithology</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>125-5</td>
<td>Greenish-black lenticular bedded shale, Black pyritic mudstone. Shale</td>
<td>Overbank deposits.</td>
</tr>
<tr>
<td></td>
<td>Fine grained planar bedded micaceous lithic arenite.</td>
<td>Possibly top of point bar.</td>
</tr>
<tr>
<td></td>
<td>Fine grained sandstone lens. Medium grained sublitharenite fining up into lithic arenites. Decline in grain size, presence of mica and change from cross-bedding to planar bedding may reflect declining energy as channel migrated.</td>
<td>Short term fluctuations.</td>
</tr>
<tr>
<td></td>
<td>&quot;Banded&quot; clasts of mudstone</td>
<td>MAIN CHANNEL FILL.</td>
</tr>
<tr>
<td></td>
<td>Rounded mudstone and angular lime mudstone clasts.</td>
<td>Result of storms or stacked channels.</td>
</tr>
<tr>
<td>1309</td>
<td></td>
<td>Channel lag.</td>
</tr>
</tbody>
</table>

I69
The core section in question does not lie in close proximity to any bed of lime mudstone in the core so the lime mudstone clasts are thought to have been derived from lagoonal sediments lying to seaward (Fig.43) possibly as a result of storms. The large size (0.5-5cms.) of the clasts suggests such a high energy mechanism even though velocities of 1.5 metres/sec. have been recorded in tidal flat channels (Reineck and Singh 1980) and would have thus been able to move them according to the data of Sundborg (1956). The angular shape of the clasts suggests that in fact water velocities within the channel were not sufficient to give sustained movement and thus round the clasts to any significant degree.

The sandstones fine upwards with clasts becoming less abundant. A tendency is noted in many of the channels for the clasts to appear in repeated bands. This may furnish additional evidence that the primary depositional mechanism for these clasts was "events" such as storms. However this may also indicate "stacked" channels where a channel meandered over a pre-existing one and eroded the upper sediments leaving only the lag as a relic.

Although not the case in the example of Fig.42 the mudstone clasts are often accompanied by shelly debris, this is predominantly non-rounded, poorly sorted bivalve debris. Such shelly debris may form the entire content of the lag in some cases.

The main body of the channel (represented by (b) in Fig.42) is composed of medium grained sandstone of lithofacies a). This tends to become finer grained and micaceous and to develop planar bedding towards the top. This is held to reflect decreasing velocity due to lateral movement of the channel. Thin, fine-grained sandstone lenses are found
Fig. 43. A plan view of the geometry of a modern tidal flat complex. (After James 1984).

Keel interpretations

Open shelf or basin

Reefs

Tidal flat

Tidal channel

Tidal flat

Lagoon (100's of metres to 10's of kilometres wide)

Tidal flat

Land

Upper mixed beds

Navan micrite

Lower mixed beds

Lagoon

Island

Lime sand shoals

Micrite
throughout indicating shorter term declines in velocity probably due to tidal effects. The cross-bedding sporadically developed throughout possibly results from the megaripples which develop in tidal channels (Reineck and Singh 1980). The cross-bedding is more abundant towards the base of the bed. Thus the planar bedding at the top of the bed may be due to a decline in stream power to plane bed conditions from megaripples (Reineck and Singh opp.cit.). Although not seen in this particular section of core reactivation surfaces occur in similar sequences in other Lower Mixed Beds cores at Keel.

A bed of finely laminated (1-2mm.) shale overlies the sandstone. This is signified by (c) in Fig.42. A 2cm. thick merging contact of dirty fine-grained sandstone is present. The shale laminations are straight, parallel and continuous. The undisturbed nature of the lamination indicates a lack of bioturbation possibly as a result of rapid deposition.

Overlying (c) is a bed of black pyritic mudstone (signified by (d) in Fig.42). This may indicate the initiation of overbank deposition from the channel. A major flood, as opposed to (e) below, is considered the most likely depositional mechanism.

At the top of the section is a greenish-black shale ((e)in Fig.42). Within this shale are thin beds of siltstone and fine-grained sandstone. Such deposits are common on tidal flats (Reineck and Singh opp.cit.) and presumably result from overbank flow which occasionally became powerful enough to transport sand sized sediment onto the flat in a manner similar to that of crevasse splays on rivers such as the Mississippi.

Fig.44 shows a second channel sequence. This is probably a smaller
FIGURE 44, LOWER MIXED BEDS, CHANNEL SEQUENCE, CORE K67.

LITHOLOGY

Depth down hole (m)

62-9

Sublitharenite with clasts at base.

Sandstone showing slumping.

Green sandy mudstone
Fine grained micaceous sandstone.
Planar bedded.
Bioturbated.
Flaser bedding present.

Medium grained sublitharenite
Cross-bedded.

Angular clasts green mudstone

Black mudstone

INTERPRETATION

Lag of overlyi ng channel.

Slumping of channel wall.

Temporary decline in stream power.

Channel-fill.
Reduction in grain-size and increase in mic -a and planar bedding indicat e declining en -ergy.

Basal Lag

I73
channel than that described above. The finer grain size and less mineralogically mature nature of the sediment imply this.

At the base (a) are angular shards of green mudstone forming a basal lag. The lack of lime mudstone clasts may suggest a greater distance from the lagoon in comparison to the first type of channel above (Fig. 42).

Overlying the lag are micaceous fine grained sandstones (shown as (b) in Fig. 44). Cross bedding is common and planar bedding is developed towards the top of the bed. Mudstone flasers are moderately common indicating a tidal influence. The sandstone is often highly bioturbated while interbedded sections of non-bioturbated sandstone are developed which are interpreted as representing periods of rapid deposition. No escape structures were noted however in the top of the bioturbated portions.

Quite commonly developed in both channel types are banded sandstones. The bands are composed of green, mud rich sandstones and white mud free sandstones. The mud rich bands probably represent conditions where the stream power dropped so abruptly that both sediment moved as bed load (sand) and that in suspension (mud) were deposited together.

A thin bed of green sandy mudstone ((c) in Fig. 44) is located within the main sandstone body. This may mark a reduction in stream power possibly due to;

1) A temporary abandonment of the channel or
2) A reduction of flow caused by a very high tide.

Immediately overlying the mudstone is a 15cm. thick portion of
P. 29. Bioclastic Limestone Unit. Brachiopod valve deforming sediment implying that the sediment was very soft at the time of deposition. Sequence youngs to the right.

P. 30. Bioclastic Limestone Unit. Storm lag exhibiting an erosional base (E-E) to the underlying shale and a marked fining-up. Sequence youngs to the left. Scale is a new penny.
P. 31. Waulsortian Mudmound Complex at Carrickbuoy Quarry. Mound facies (M) with encrinitic Flank beds (F).

P. 32. Contact of the Mound and Flank facies as seen in the core. The two pieces were originally joined. This shows the very abrupt change. Note the pronounced stylolites in the Mound portion. Yellow scale is 12.5 cms. long.
P. 33. Waulsortian Mudmound Complex at Carrickbuoy Quarry. Parts of two mounds separated by the Flank beds (F) are seen. Mound facies (M).

P. 34. Stromatoids in the Mound facies of the Waulsortian Mound Complex. Note the flat bases and the high depositional dip of many of the stromatoids. Mud filled stylolite to the left. Scale is five pence piece.
sandstone showing wavy, very high angle dips and a possible erosional base. This is interpreted as slump bedding. The slump probably occurred as the water level was falling in response to the ebb tide and water was being withdrawn from the sediments. Similar type slumps have been known to occur in river channel banks (Coleman 1969).

Overlying this channel sequence is the lag of another channel ((d) in Fig.44). The water body which deposited this lag probably eroded away any fine grained vertical accretion deposits which had developed in the underlying channel sequence.

Channel sandstone deposits make up the greater part of the Lower Mixed Beds. The 55% of the lithology composed of sandstone compares well with the findings that the majority of tidal flat sediments will be reworked by laterally migrating channels in a comparatively short time (58% in 68 years, Reineck and Singh 1980).

2. Beaches.

Rarely sandstones of lithofacies a) exhibit swash lamination (Clifton 1969). This is stated to be characteristic of the upper-swash zones of beaches (Clifton opp.cit). The existence of these sandstones within an environment which does not appear subject to significant wave action, as is indicated by the fine grained nature of much of the Lower Mixed Beds sediments excluding the channel sandstones, is thought worthy of comment. The swash lamination is thought to be due to a local long term effect rather than to swash activity after storms. Such storms would be likely to leave indications in more than the isolated cores in which swash lamination is found. In addition it is hard to visualise a storm having such long term effects as to produce swash lamination in
FIGURE 45. LOWER MIXED BEDS, BEACH SEQUENCE, CORE DDB 7.

<table>
<thead>
<tr>
<th>Depth down hole (m)</th>
<th>Lithology</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Navan Micrite</td>
<td></td>
</tr>
<tr>
<td>Rippled and burr owed micaceous lithic arenite.</td>
<td>Lower shoreface.</td>
<td></td>
</tr>
<tr>
<td>Fine grained silt an sublitharenite -te. Shows swash (beach) laminat-ion. Occasional clasts present.</td>
<td>Beach.</td>
<td></td>
</tr>
<tr>
<td>Rippled micaceous laminated lithic arenite with shale s.</td>
<td>Foreshore backshore.</td>
<td></td>
</tr>
</tbody>
</table>
such a depth of sediment. Rather it is considered that the swash lamination was produced where the shoreline was exposed to wave action possibly produced by wind where the lagoon was wide enough to give the wind sufficient "fetch" or where the shoreline was directly opposite a barrier island inlet (Chap.12). In such a position the "damping" effects of a lagoon upon waves would not be nearly as great. That such localised high energy conditions can occur within barrier protected shorelines is indicated by the existence of ooids in the Laguna Madre of Texas (Rusnak 1960).

Fig.45 illustrates a length of core in which swash lamination is found.

At the base are rather impure, micaceous fine grained laminated sandstones with ripples. These are interpreted as representing the foreshore/backshore. Overlying this are fine grained white clean sublitharenites showing swash lamination. The length of this area of the core (corrected for the drilling angle) is 0.6 metres. This probably represents the sediment exposed at low tide where the wave energy lasts for the longest time (Reineck and Singh 1980). Succeeding these clean sandstones are calcareous, shaley, micaceous sandstone with ripples and burrow mottling. These are interpreted as representing the lower shoreface. At the top of the sequence is the overlying Navan Micrite.

3) Tidal flats.

Substantial portions of the Lower Mixed Beds are formed of mudstones, siltstones and fine grained sandstones that are thought to have been deposited on tidal flats.

The rippled sandstones interbedded with siltstones testify to
Fig. 46. Mechanism for a transgressive limestone in the Lower Mixed Beds.

NM2 Navan Micrite (2nd transgression)

LM2 Lower Mixed Beds (Prograding)

NM1 Navan Micrite (1st transgression)

LMB1 Lower Mixed Beds
**Lithology**

- Navan Micrite with mudstone bands.
- Fairly clean sublitharenite. Conglomeritic in base with angular black mudstone fragments.

**Interpretation**

- Infilling of channel.
- Decline in sand sized material indicates declining influence of sediment from main channels. Sandstone beds indicate major floods from major channels.
moderately high energy levels on the flood and ebb tides with periods of little to no water movement when deposition of fine grained material from suspension would occur.

Siltstones with mudstone bands were probably deposited further to landward than were the flaser bedded sandstones mentioned above since energy levels tend to be lower in the vicinity of the high water mark (Reineck and Singh 1980).

Mudstones with siltstone lenses probably represent the lowest energy regime of all marking suspension very near the high water mark.

Flasers of lime mudstone are often developed within both the fine sandstones and siltstones of the tidal flat deposits. These indicate deposition on the flats to landward of the lime mudstone bed within the Lower Mixed Beds themselves (Fig.46) or of the actual Navan Micrite. They were thus precursors of these later transgressive beds. Similar thin limestone lenses have been noted in transgressive intertidal and shallow subtidal deposits in the south of Ireland (Sleeman 1977).

Occasionally thick deposits of mudstones are present (Fig.47). Thick mudstones (up to 100 metres thick) have been noted in marginal marine facies in the south of Ireland (Kuipjers 1975) and have been interpreted as lacustrine deposits. Due to their intertidal position it is thought that the particular beds at Keel represent similar pond-like features possibly formed by the infilling of cut-off intertidal channel meanders. The upward increase in the amount of fine grained material reveals that the channel was abandoned slowly indicating deepening and eroding of chutes and swales rather than breaching of the meander neck. The sandstone lenses indicate periods of high stage flow when the
channel would receive coarser grained material as overflow from the main channel. The location of bivalves in discrete bands could also indicate intermittent deposition since there would be a tendency for only the top layer of any particular deposit to be colonised before deposition of the next layer. The black colour of the mudstone would suggest that the sediment was anoxic and that possibly vegetation was growing in the shallow water and that the dead plants were not oxidised after falling into the water.

The fining upward cycles of lithofacies d) are thought to have been deposited on the tidal flats and reflect waning energy conditions as the water and sediment overflowed from the channels and flooded the flats. The water eminating from the lagoon to seaward of the flats would carry lime mud in suspension. Fining upward cycles have been noted in tidal flat deposits in the south of Ireland (Graham 1975).

F. Lime mudstone interbed.

In the majority of the cores examined a thin lime mudstone bed 0.3-1.0 metres thick is present located approximately halfway through the Lower Mixed Beds succession. The bed is thought to represent a continuation of the transgressive event which produced the intertidal Lower Mixed Beds lying below. After deposition of the lime mudstone the transgressive event ceased. The intertidal deposits then prograded seawards over the lime mudstones (Fig.46).

Due to the heavy bioturbation and the presence of pelloids and birds-eyes the lime mudstone is assigned to Standard Microfacies type 19 of Wilson (1975) indicating deposition in restricted ponds on tidal flats. The planar character of the birds-eyes indicates that they result
from dessication of the exposed sediment. Thus at least some of the limestone is not of a sub-tidal origin. However the presence of serpulid worm tubes, indicative of salinities below 50 parts per thousand (Flugel 1982), suggest that exposure was not long term since long intervals of evaporation and drying out would not permit colonisation by the worms. The restriction of the worm tubes to the silt filled burrows indicates that the sediment was probably soft since the tubes are only found in what would have been fairly solid sediment i.e. the silt fill of the burrows.

G. Relationship of the Lower Mixed Beds to the Upper Quartz Sandstone.

Transition beds between the Lower Mixed Beds and the Upper Quartz Sandstone have been reported (Patterson 1970). However these have not been detected in the cores examined by the author.

Due to the marked petrological and environmental differences between the two units it is thought that a substantial amount of non-deposition is represented at the contact. The sharpness of this contact (Ps. 15 and 16) and the lack of transitional beds is thought to be the result of two factors:

1) The braid plain at Keel was the result of braid switching over a considerable length of time. Braid switching over long distances in short periods of time is well documented (Holmes 1965). The Brahmaputra River was 98 KM. away from its present course only 200 years ago (Coleman 1969). At any one point in time such a river would be flowing in one braid system and thus localised to one part of the plain. A marine transgression would have the effect of depositing sediment of an apparently incongruous nature upon the braided stream deposits of the
Upper Quartz Sandstone since the transitional environments i.e.
meandering streams resulting from the transgression would not be present
at that point.

2) There would presumably have been a substantial amount of re-
working of the Upper Quartz Sandstone by the marine transgression.
CHAPTER 11
CHAPTER II.

Navan Micrite

A. Introduction

Phicox (1984) has termed the lithologies lying between his Basal Clastics and the base of the Shaley Pales as the Pale Beds (Fig.8). However due to the ease of recognition of an extensive limestone unit lying at the base of the Pale Beds it is proposed to follow the Keel practice and differentiate this limestone unit as the Navan Micrite. Where the unit has been substantially dolomitised it is known as the Bioclastic Dolomite at Keel (Patterson 1970).

The remainder of the Pale Beds of Philcox (1984) are dealt with in Chap.12 as the Upper Mixed Beds (Patterson 1970).

The contact of the Navan Micrite with the underlying Lower Mixed Beds is sharp but is considered to be conformable both on the grounds of stratigraphy and the presence of the micrite bed within the Lower Mixed Beds (already considered). This earlier micrite may indicate the presence of a lime mudstone creating environment to seaward of the Lower Mixed Beds.

The presence of the Navan Micrite marks the renewed onset of the marine transgression (probably cycle 5 of MacCarthy and Gardiner 1987) by which means the limestones of the Navan Micrite advanced over the Lower Mixed Beds. Terrigenous sediment was now trapped in the Lower Mixed Bed type environments to the north of Keel and extensive carbonate production commenced.

The Navan Micrite is the first lithology in the Keel area to contain fossils of an undoubted marine affinity.
In the cores examined the Navan Micrite ranges from 2-31m. in thickness with an average of 14.6m. (Std.Dev. 8.7M.).

Observation of Figs. 48 & 49 reveals a pronounced thickening of the lithology from approximately 2M. in the Keel area to 31M. to the East and 29M. to the west. This occurs together with a complimentary increase and decrease in the thickness of the Lower Mixed Beds.

The thinning at Keel is thought to be caused by one or both of the following:

(1) The postulated palaeohigh still existed in the Keel area. Lower Mixed Beds type sedimentation would continue to dominate the area and lagoonal type sediments would not be deposited until after they had been deposited to the east and west of Keel. This would lead to thicker Lower Mixed Beds sediments since they would be deposited for a longer period of time at Keel.

(2) The area was still the site of a major fluvial system in Lower Mixed Beds times. Such a river would bring more terrigenous sediment into the Keel area and thus possibly adversely effect the growth of lime secreting organisms by increasing the turbidity of the water. This would influence photosynthesis and additionally irritate filter feeding animals.

B. Lithology

The lithology is primarily calcareous and may be subdivided into six lithofacies:

(a) Peloidal wackestones/grainstones and algal mudstones.

(b) Oolitic grainstones.
(c) Rudstones.
(d) Bioclastic packstones.
(e) Bioclastic grainstones.
(f) Dolomites.

Lithofacies (a).

These peloidal wackestones/grainstones form the bulk of the rocks in the Navan Micrite. On a local basis a major component can be of an algal origin.

The lithofacies is of a predominantly pale tan colour.

The lithofacies usually shows a tremendous amount of burrowing. These burrows tend to be of a low angle and non-lined and are frequently filled with coarse silt. This may be due to one or both of the following mechanisms:

(1) As an organism burrowed the peloids produced represent a substantial volume reduction of the original sediment since the fine grained mud would be compacted to form the peloids. This would lead to a net volume concentration of the incompressible silt particles already present within the mud and which would not be ingested by the organism.

Intense reworking of fine muds into grain supported pellets has been reported from Shark Bay, Western Australia (Hagan and Logan 1974).

(2) Movement of animals over the sediment surface might cause the silt particles to be set in motion. These might then roll around and fall into the burrows.

The burrows are often preferentially dolomitised probably due to the enhanced porosity associated with them (P.18). Roehl (1967) recorded
a segregation of dolomite crystal sizes with the smaller crystals being
confined to inter-burrow areas.

Laminar fenestrae filled with calcite and dolomite but lacking internal sediment are variably common. These are interpreted as birds-eyes (Shinn 1968).

Silt is abundant in much of the Navan Micrite, occasionally forming up to 35% of the lithology and grains "float" in the micritic matrix. Frequently the grains are euhedral indicating an authigenic origin around pre-existing much smaller "seed" crystals.

Thin bands of mudstone and shale up to 25 cm. thick occur in this lithofacies of the Navan Micrite. These are rare in the upper part of the unit.

Very rarely thin (8 cm.) beds of fine-grained sandstone and siltstone, showing an erosional base are present.

Rare clasts (over 2 mm. in diameter) with a crudely rounded shape and concentric, irregular laminae around a nucleus are present, which are interpreted as oncoliths.

Occasionally the sediment is unburrowed. As regards this variable intensity of burrowing 0-100% bioturbation has been noted in tidal flat deposits and is thought to indicate variable deposition rates (Gietelink 1973). It is hoped that a tidal flat origin for the Navan Micrite can be demonstrated.

The peloids are thought to be of a faecal origin due to their moderate to good sorting and high length to width ratio (up to 3:1). They range up to 0.5 mm. long but in the main are much smaller.

Bivalve and brachiopod debris is found in this lithology but tends
to occur high in the sequence i.e. near the contact with the overlying Upper Mixed Beds (Fig.8).

The pale colour of the lithology is due to the intense bioturbation. Although abundant organic matter may have been present this was consumed by the numerous deposit feeders. In addition the constant bioturbation would keep the sediment in contact with the oxidising lagoonal waters for a considerable time (Wilson 1975).

Few sedimentary structures are seen, many that were originally present were probably destroyed by the intense bioturbation. Occasionally fine, frequently wavy, lamination is seen mainly towards the top of the unit. Rarely a high angle slurried appearance of the lithology is noted. In very rare instances infilled fissure like features are present. These are interpreted as mudcracks. These frequently cannot be detected in outcrops of known tidal flat deposits (Sellwood 1972), and so their apparent rarity in the Keel core is not surprising.

Lithofacies (b)

The oolitic grainstones are confined to rare 1.5m. lengths and tend to be developed near to the top and to the bottom of the Navan Micrite.

Although micritised the ooids retain sufficient of their structures e.g. several laminae, radial structure to indicate that they are normal marine ooids (Flugel 1982). They are well sorted and generally the ooids make up 40-50% of the lithofacies. The largest ooid observed is 0.75mm. in diameter but the norm is approximately 0.5mm. The ooid nuclei are generally 0.1-0.2mm. in diameter. Large quartz nuclei (4mm.) are noted but the euhedral character of the quartz probably indicates that an
authigenic process and growth of the nuclei within the ooids took place after their formation. Ooids containing these large nuclei tend to be of a similar size to those not having an authigenic quartz nuclei. The nuclei material corresponds to that available in the area i.e. quartz, feldspar, carbonate grains and shell fragments. Compound ooids occur showing no indication of pressure welding, thus they are true compound ooids and not a post-depositional feature. Only one generation of ooids is recognised. A dolomite cement is present. The calcite precursor of this cement must, in some circumstances, have been deposited early in the diagenetic cycle since in some cases overcompaction has occurred whilst in others the early cement has protected the ooids from compaction.

**Lithofacies (c). Rudstones**

The rudstones of this lithofacies appear in localised beds up to 0.4M. thick throughout the Navan Micrite. The lithofacies is composed of micritised bioclastic debris (cortoids) and lithoclasts in a coarse equant ferroan calcite cement (P.22). In many cases this cement is seen filling micrite envelopes. Infrequently dolomite fills the envelope. Occasionally where actual algal borings are seen these are seen to perpendicular to the shell surface. The bioclastic debris includes fragments of brachiopods (punctate and impunctate), crinoids, bryozoans, foraminifera and bivalves. The lithoclasts include lumps of oolitic grainstone (P.17) and sandy peloidal wackestones typical of lithofacies (b) and (a). Both bioclasts and lithoclasts are usually well rounded and sorted and may be up to 6mm. long. In some cases lithoclasts can make up over 90% of the clasts present. Occasionally in these cases
the clasts are angular.

Lithofacies (d) Bioclastic packstone

Very rarely (only one example was found) a thin (0.2M.) bioclastic packstone is present. The bioclasts are made up entirely of bivalves.

Microcrystalline dolomite forms a matrix and appears to have replaced micrite since micrite is found as an undolomitised discontinuous rind surrounding the bivalves. A poorly developed orientation of the grains is present indicating some water movement at the time of deposition.

Lithofacies (e) Bioclastic grainstones

This lithofacies shows only a slight development in the Navan Micrite.

The lithofacies contains abundant poorly sorted and angular bioclasts and lithoclasts. The bioclasts include abundant foraminifera (Endothyranopsis sp.), brachiopod debris, bivalve shells and calcispheres. All these show micritisation bar some brachiopod debris. Unmicritised crinoidal debris is also present.

The poorly sorted and rounded lithoclasts are composed of peloidal wackstones similar to those of lithofacies(a).

Lithofacies (f) Dolomites

Apart from the dolomitisation in burrows in lithofacies (a) complete dolomitisation of the Navan Micrite (Bioclastic Dolomite of Patterson 1970) is common. In these cases the lithology takes on a black colour. The dolomite is usually ankeritic and has a microcrystalline fabric. Any pre-existing fabric is lost.

C. Carbonate cementation
Very coarsely crystalline dolomite fills in birds-eye porosity and forms a cement to the grainstones. Preferential dolomitisation is present in burrows (P.18). Occasionally tectonic veins show a very finely crystalline non-ferroan dolomite followed by coarsely crystalline non-ferroan calcite which is in turn followed by barytes mineralisation. Very coarsely crystalline non-ferroan calcites infill breccias.

Serpulid worm tubes show two generations of cement, an isopachous acicular marine calcite cement followed by an equant, moderately crystalline, calcite infill (P.19). Where the tubes have been fractured the second generation cement has been dolomitised but not the first. Possibly this selective dolomitisation has been facilitated by the larger grain size of the equant cement.

D. Interpretation

The environment is interpreted as one of quiet water deposition. This is indicated by the presence of micrite in the wackestones and the moderate sorting of some peloids. The apparent good sorting of some of the peloids is probably due to pellet production by the same size and species of animal rather than by sorting effects. Peloids tend to occur preferentially in low energy subtidal and shallow intertidal zones (Flugel 1982). Again the existence of low angle, non-lined burrows in lithofacies (a) may indicate low energy conditions (Frey and Howard 1969) as may the absence of water formed sedimentary structures even where bioturbation has not taken place.

In this generally low energy environment local areas of moderate to high energy did occur as indicated by the presence of oolitic grainstones and well washed rudstones. The rare oncoliths in the
wackestones and grainstones of lithofacies (a) resemble the Type R oncoliths of Logan et al. (1964) and again indicate localised high energy.

The existence of an extensive sand body (Fig. 8) overlying the Navan Micrite indicates, in a transgressive regime, a barrier to normal marine influences lying to seaward of the Navan Micrite depositional environment.

The presence of ooids limits the depth of deposition to 100M. (Flugel 1982), but the high percentage of ooids indicates a maximum depth of 15M. (Flugel opp. cit.) since production of ooids is presumably maximised with decreasing depth i.e. higher energy and increased light thus increasing ooid growth. However it should be noted that Loreau and Purser (1973) claim that that currents retaining ooids in favourable sites for their growth are also very important in giving high ooid densities.

The existence of ooids indicates that water temperatures exceeded 20 Degrees Centigrade (Flugal 1982).

The laminar fenestrae have been interpreted as birds-eyes. These tend to be restricted to the supratidal and occasionally the intertidal zones, they never occur in subtidal environments (Shinn 1968). Similarly mud-cracks indicate dessication of the sediment caused by sub-ariel exposure.

The presence of algae indicates water depths not exceeding 200M. Calcispheres, as found in the Navan Micrite, are thought to be the fruiting bodies of Dasycladaceae which are not found below 100m. and are more usually restricted to the shallow subtidal zone (Flugal 1982).
addition associations of algae and organisms such as foraminifera, bryozoans and serpulids are criteria for this zone (Flugal opp.cit.).

Cortoids, as are found in the grainstones of lithofacies (b) and (e) and in the rudstones of lithofacies (c), are held to indicate deposition in water depths of 15-20M. (Swinchatt 1965). Perpendicular borings occur in the intertidal and subtidal zones down to 20M. (Budd and Perkins 1980).

Hardgrounds may have been developed as suggested by the existence of serpulid worm tubes with no trace of shells etc. for them to encrust on (P.20). The existence of marine cements associated with the worm tubes may be of significance here.

Stenohaline organisms such as crinoids and brachiopods tend to be localised to discrete portions of the Navan Micrite succession on a seemingly random basis. This may indicate that salinity varied locally. This is characteristic of barred lagoons (Reineck and Singh 1980). Faunal changes due to increasing salinity have been noted in lagoons on the Trucial Coast, South-East Persian Gulf (Purser and Evans 1973). Variable salinities have been noted in Shark Bay, Western Australia where summer salinities may reach 55 parts per thousand in the lagoon (Davies 1970 A).

The lithoclasts described are probably the result of intense burrowing of the sediment and subsequent rounding by current action. They are common in intertidal environments (Wilson 1975).

Owing to the above factors i.e. intertidal and subtidal environments, variable salinities and generally quiet marine conditions the Navan Micrite is considered to be the product of deposition behind a
bar sand (the precise nature of this bar sand is examined in Chap.12). Philcox (1984) has commented on the close relationship between the Navan Micrite and the bar sand. Both are restricted to the northern portion of the Irish Midlands (Fig.54) and it would appear that where the one lithology is present the other will be and vice versa. A link on depositional environment grounds is therefore likely.

Having established the general nature of the depositional environment i.e. a lagoon it is now possible to more exactly link each of the lithofacies to specific areas of the lagoon.

Lithofacies (a) Peloidal wackestones and grainstones.

These constitute over 90% of the sediments.

The peloidal grainstones are assigned to Standard Microfacies 16 of Wilson (1975). As such they indicate deposition in very warm, shallow water with only moderate circulation, a shelf lagoon is indicated.

The peloidal wackestones, which appear to be associated with the birds-eyes, are assigned to Standard Microfacies 19 of Wilson (opp.cit.). Deposition in very restricted bays is indicated where sediments would be exposed to dessication at low tide stands and the water would probably be of abnormal salinity. The upward increase in these peloidal wackestones towards the boundary with the overlying Upper Mixed Beds indicates that the lagoon was being filled up with sediment and that circulation was being increasingly restricted. Sub-ariel exposure became common giving rise to laminar birds-eyes through sediment dessication (Shinn 1968). The thin laminations, though generally uncommon, are seen in increasing numbers towards the top of the Navan Micrite. These laminae are thought to be of an algal origin.
The algae which appears in the mudstones of lithofacies (a) may be the remnants of a formerly much more extensive cover. The greater part of this may have broken down to furnish the lime mud subsequently re-worked into peloids by burrowing organisms. Very significant rates of production of lime mud by algae and serpulid worms (mean figure 118 grms./metre squared/annum.) have been reported from Florida Bay (Nelsen and Ginsberg 1986).

The thin mudstones and sandstones normally associated with the wackestones and grainstones of lithofacies (a) may reflect sporadic fluvial inputs into the lagoon. Such processes are postulated to have occurred in the Carboniferous of South-West England (Walker 1969). Alternatively they may have been formed by sand being swept into channels from shallow sand flats (represented at Keel by the Lower Mixed Beds) and deposited in fans at the channel ends as is known to happen in Shark Bay (Hagan and Logan 1974).

The proposition that they represent the distal ends of washovers is not considered likely since the usual location of the beds near the contact of the Navan Micrite with the Lower Mixed Beds seems to indicate a landward source for the material rather than derivation from a sand body to seaward of the lagoon.

The occasional high angle slumps are thought to occur from the sides of intertidal channels. They were caused by the collapse of saturated sediments during falling water stages as the tide retreated. Similar phenomenon have been observed in other water channels.

The presence of quartz in the Navan Micrite is due to wind blown silt from the Lower Mixed Beds to landward or the bar sand to seaward.
Much of this silt served as seeds for the subsequent growth of authigenic quartz. The floating nature of the grains appears to rule out any depositional mechanism involving water since the grains would tend to form lenses. Wind blown sand is common in carbonate sediments in Shark Bay (Davies 1970 A).

Lithofacies (b) Oolitic grainstones

This lithology appears in two sub-environments:

1. Where located directly over the Lower Mixed Beds they represent the basal high energy transgressive deposits. The shallow waters would permit the formation of ooids where tidal/current or wind conditions lead to sustained water movement.

The rarity of these grainstones is thought to indicate that suitable conditions were only locally present e.g. opposite tidal inlets where tidal influences would be strongest or where the lagoon was very wide. Lagoons in the Persian Gulf may be up to 40KM. wide (Purser and Evans 1973) and under these conditions the wind would develop enough fetch to agitate the waters on the landward side of the lagoon. Oolitic sands are developed in the Khor Odaid Lagoon, Persian Gulf in wind exposed localities with only moderate water agitation (Loreau and Purser 1973). Oolitic sands and microcoquinas are also locally present on the landward side of the barrier protected Laguna Madre, Texas (Rusnak 1960). They occur in areas of slow deposition where wind influence is strong. Ooids can also form where associated with river deltas on the landward side of lagoons e.g. Wooramel Delta, Shark Bay Davies 1970 B.

2. Oolitic grainstones are found in discrete thin beds throughout the Navan Micrite. These indicate that local high energy shoals existed
in the lagoon possibly as a result of coarse sediment being amalgamated by currents into shoals which then developed ooids. The high density (occasionally over 90%) of ooids in the lithofacies implies deposition in depths occasionally less than 7m. (Flugel 1982). Oolites containing approximately 20% ooids are present. It is suggested that these ooids were developed in deeper water or were transported off a shoal and deposited in surrounding lower energy environments.

**Lithofacies (c) Bioclastic rudstones**

These lithologies are thought to mark the sites of the channels which would have dissected the tidal flats and probably continued through the subtidal portion of the lagoon to link up with the channels breaching the bar sand to seaward. As such they are assigned to Standard Microfacies 24 of Wilson (1975).

The presence of cortoids testifies to the shallow nature of these channels. The size of the smallest clast observed (1mm.) implies current velocities of 3KM./Hr. (Sundborg 1956), well within the velocity of 5.4 KM./Hr. recorded for some tidal gullies (Reineck and Singh 1980). Dillon (1970) has recorded the fastest velocities in lagoons in the vicinity of tidal inlets.

The lack of quartz grains in the rudstones is probably a result of them being winnowed away by the high water velocities.

The fact of large individual lithoclasts of differing composition being present implies that at least one type was transported some distance. The possibility that the lithoclasts emanated from channel wall beds of differing lithologies in vertical proximity is acknowledged however.
The water in many of these channels was probably of normal salinity since the debris of stenohaline organisms e.g. crinoids and brachiopods is found. Whilst realising that the crinoids may have been introduced after death due to buoyancy effects this is not thought to be the case with the brachiopods. These probably remained substantially in situ after death. These stenohaline channels were probably located near to tidal inlets where the lagoonal waters remained at normal salinity throughout the tidal cycle. Further into the lagoon variable salinities would be found (Hagan and Logan 1974, Fischer 1961). Variable salinities on a seasonal basis have been noted in Laguna Madre by Shephard and Moore (1955) who also found stenohaline types (echinoids) localised near barrier island inlets.

It is however possible that the lime muds outside the channels was too soft for the crinoids to root as intense burrowing can produce a thixotropic sediment with 50-60% water (Johnson 1978). In addition the channels may have been the only locations where there was enough water movement to provide nutrients for the filter feeding crinoids.

The non micritised state of the crinoids might hint that they were transported. However this is considered to be probably an effect of selective micritisation in which echinoderms are little, if at all, effected by micritisation (Flugal 1982). Similar mixtures of micritised and unmicritised shell debris have been reported from the Jurassic Gigas Beds (Schmidt 1965).

Where the clasts comprise a high proportion of lithoclasts it is probable that these were produced in areas of high bank failure by the process outlined above concerning lithofacies (a).
Thin (2cm.) beds of this type occasionally overlie the Lower Mixed Beds. In this case they represent the basal beds of the transgression in a similar fashion to some oolitic grainstones of lithofacies (b). Similar basal conglomerates have been found in Laguna Madre (Rusnak 1960). However energy levels were probably not nearly as high as in the case of the oolitic grainstones and it is tempting to propose that in these instances a bar sand was already in place and that only parts of the lagoon were experiencing high energy conditions.

Lithofacies (d) Bioclastic packstones

The lithofacies is assigned to the Standard microfacies 8 of Wilson (1975). Its packstone nature rather than the wackestone and abundant restricted fauna possibly indicates a mass mortality or slow sedimentation rates. A shelf lagoon environment with quiet water conditions below normal wave base is proposed.

Lithofacies (e) Bioclastic grainstones

This lithology is assigned to the Standard microfacies 11 of Wilson (opp.cit.). Deposition, probably on shoals, in areas of constant water movement at or above wave base is probable. The absence of ooids may indicate that current energy was only sufficient to winnow away the fine sediment and not high enough to create ooids. Winnowing of sediment by wind in shallow areas of Shark Bay has been reported (Hagan and Logan 1974). Since the beds do not show grading it is thought unlikely that the beds were storm generated.

The presence of stenohaline debris e.g. crinoids and bryozoa may indicate that the water was of normal salinity but the ease of transport of such debris must be considered. The presence of brachiopod debris is
significant and may indicate that salinities were indeed normal. The significance of the abundant foraminifera is unclear. Salinity was probably not a factor. It may be that the combination of well aerated water and moderate movement formed a favourable environment.

**Lithofacies (f) Dolomites**

The dolomites may be developed in any of the other lithofacies. No preferential dolomite according to depositional environment as outlined by Flugal (1982) has been detected.

The dolomitisation probably resulted from the precipitation of gypsum in the supratidal zone of the flat. The resulting fluids would then have the high magnesium to calcium ratio required to dolomitise calcium carbonate. This process is known to be taking place on sabkhas in the Persian Gulf at present in a type of evaporative pumping mechanism (Bathurst 1975).

No indications of supratidal deposits such as gypsum, collapse breccias etc. have been found in the Navan Micrite at Keel possibly due to factors examined below (Sect.E). However some indirect evidence of the one time presence of gypsum is available.

1. Evaporites are present in the equivalent Micrite Unit (Philcox 1984) at Strokestown (Fig.1) 20 KM. north-west of Keel. Thus suitable climatic conditions probably existed in the Keel area at the time.

2. Dedolomitisation is noted in dolomite from the Keel area (P.22). The dissolution of gypsum can give rise to the high calcium/magnesium ratio needed for dedolomitisation (Blatt et.al.1980).

The possibility that the dolomitisation was produced by fluids ascending faults as in the case of the Quartz Pebble Conglomerate is
also considered possible however.

E. Overall sequence of the lithofacies of the Navan Micrite

The sequence (Fig. 50) corresponds well to that proposed by James (1984) for his shallowing upward low energy tidal unit developed on a low energy sub tidal unit.

The basal oolites and lithoclastic rudstones mark the initial part of the transgression. The lithoclastic rudstone clasts were probably reworked from the underlying Lower Mixed Beds. Peloidal grainstones of lithofacies (a) frequently form the basal bed of the Navan Micrite at Keel. A similar situation exists at Shark Bay where pellet and skeletal packstones form the basal sheet (Hagan and Logan 1974). It is possible that thin supratidal sheets were deposited in advance of these beds (Fig. 51). However no trace of these beds nor relics such as collapse breccias etc. has been found. It can only be assumed that the transgression was too rapid for the formation of such lithologies before the strand line became establishment to the north of Keel. Subtidal deposits were now formed in the lagoon which was protected by a bar sand (Fig. 52). The subtidal area probably acted as the sediment source from which lime mud was transported onto the intertidal flats to landward (James 1984).

The source of the calcium carbonate, apart from the skeletons of the macrofauna, is unclear. Algal production has been mentioned above. Probably of even greater importance were the photosynthetic activities of the algae. This would decrease the amounts of carbon dioxide in the water and thus lead to the precipitation of calcium carbonate.

Sedimentation rates now began to exceed the rate of sea level rise
Fig. 50. Generalised overall sequence of the Navan Micrite.

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Subenvironment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Supra-tidal (possible)</td>
<td></td>
</tr>
<tr>
<td>Pelloidal wackestones with Birds-eyes.</td>
<td>Intertidal</td>
</tr>
<tr>
<td>Pelloidal grainstones and algal boundstones.</td>
<td>Subtidal</td>
</tr>
<tr>
<td>Occasional mudstones (m).</td>
<td></td>
</tr>
<tr>
<td>Oolitic grainstones and lithoclastic rudstones of basal beds.</td>
<td></td>
</tr>
</tbody>
</table>
possibly due to a halt or slowdown in the rate of sea level rise or an increase in calcium carbonate production rates.

The intertidal flats then began to prograde over the subtidal deposits (Fig. 51). Such a process is known to be taking place in the lagoons of the Persian Gulf (Evans et al. 1973). Supratidal deposits may or may not have been present as the prograding process may not have gone on long enough for them to have become established in the Keel area before the area was flooded. Alternatively the bar sand may have removed any supratidal deposits when it transgressed across the lagoon (Chap. 12). A lack of supratidal deposits has also been noted in some Upper Devonian/Lower Carboniferous successions in southern Ireland (MacCarthy 1974).

Eventually the subtidal (source area) became so restricted (Fig. 51) as to be incapable of providing further sediment and progradation ceased.

Finally the bar sand transgressed over the lagoonal sediments eroding much of them as it went in a manner similar to that described by Wilkinson (1982). Transgression rates must have been quite high since substantial amounts of the lagoonal sediments were left intact.

The above model of Wilkinson (op. cit.) is preferred to that of Ginsberg (1971) which proposes constant sea level rise. If this were the case at Keel the bar sand would tend to transgress over the lagoonal subtidal sediments to landward before the intertidal beds had prograded over them. Consequently bar sand deposits would be found directly overlying subtidal deposits. This is not the case at Keel.
Fig. 51. Accretion of carbonate from subtidal source areas onto tidal flats. (Partly after James 1984)
CHAPTER 12
CHAPTER 12.

**Upper Mixed Beds**

A. Introduction

Sharply overlying the Navan Micrite at Keel are the Upper Mixed Beds (Patterson 1970). These are equated to the Pale Beds (Fig. 8) of Philcox (1984) with the addition of the Upper Sandstone (Philcox opp.cit.). The local term Upper Mixed Beds has been retained since it is felt that it best conveys the sense of the complex of depositional environments that gave rise to the various lithologies.

The contact with the underlying Navan Micrite/Bioclastic Dolomite is placed where sandstone beds begin to enter the succession. Lime mudstones, oolites, bioclastic rudstones and grainstones are however present within the lower part of the Upper Mixed Beds.

Occasionally the Upper Mixed Beds are absent at Keel. A similar situation is present at Corlea (Crowe 1986) 10 KM. to the west of Keel. The significance of such omissions is discussed below.

As may be seen from Fig. 53 the Upper Mixed Beds at Keel may be divided into two broad portions;

1. A thick sandstone body including the Upper Sandstone of Philcox (1984) forms the top part of the Upper Mixed Beds.

2. A mixed succession of sandstones, shales, siltstones and limestones forms the lower part.

The average thickness of the Upper Mixed Beds in the core examined at Keel is 40M. (Std.Dev. 16.5M.). The thickness ranges from 14-73M..

To the south the Upper Mixed Beds may be correlated with the Ballystein Limestone (Philcox opp.cit.). As such they form much of the
Fig. 52. Main morphological components of the barrier model. (After Blatt, Middleton and Murray 1980)
**Fig. 53. General subdivision of the Upper Mixed Beds.**

<table>
<thead>
<tr>
<th>UNIT</th>
<th>SUB-UNIT</th>
<th>LITHOLOGY</th>
<th>ENVIRONMENT</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td></td>
<td>Upper Sandstone. Clean sandstone, oolitic at base.</td>
<td>Barrier</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td></td>
<td>Argillaceous lime mudstone</td>
<td>Lagoon</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Moderately clean sandstone with mudstone bands (m)</td>
<td>Distal washovers</td>
</tr>
<tr>
<td>I</td>
<td></td>
<td>Clean sandstones</td>
<td>Proximal washovers</td>
</tr>
</tbody>
</table>

Navan Micrite
carina conodont zone (Phillips and Sevastopolou 1986).

The Upper Sandstone of Philcox (1984) which forms a part of the first sub-division referred to above (Fig. 53) comprises a sheet stretching across the north Midlands of Ireland and covers an area possibly up to 4,700 sq. KM. in extent. Fig. 54 indicates a thickening of the Upper Sandstone to the south (correlation co-efficient 0.43, probability of no correlation 18.5%). A thinning from west to east is also seen substantiated by a correlation co-efficient of 0.61 (probability of no correlation 10.5%). Although the Upper Sandstone forms only a part of the Upper Mixed Beds and the completeness of the succession is unknown in other areas it is felt that these thickness relationships may be indicative of the provenance of the unit.

A more local thinning (Fig. 54), on a line passing through Lough Sheelin-Granard-Keel and Newton Cashel, is also present possibly indicating the continued existence of a palaeohigh in the area.

B. Lithology

Although the Upper Mixed Beds are of a very variable character six main lithofacies may be recognised. Where appropriate, sedimentary structures are included in the description

Lithofacies (a) Clean sandstones (mainly sublitharenites)
Lithofacies (b) Unclean sandstones (lithic arenites and wackes)
Lithofacies (c) Siltstones, shales and mudstones
Lithofacies (d) Bioclastic rudstones and grainstones
Lithofacies (e) Oolitic grainstones
Lithofacies (f) Lime mudstones and other carbonate sediments.

Lithofacies (a) Clean sandstones
The sandstones are fine to very fine grained, clean, well sorted, angular, mature sublitharenites with occasional quartz arenites.

Point counting results are given below as an average. This average has been recalculated after the exclusion of carbonate cements which occasionally make up 45% of the lithology.

Point counting conditions are as Chap. 1.7 b.

<table>
<thead>
<tr>
<th>Clast</th>
<th>Average %</th>
<th>% Std. Dev.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Monocrystalline straight</td>
<td>13.8</td>
<td>4.5</td>
</tr>
<tr>
<td>extinction quartz</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Monocrystalline undulate</td>
<td>75.5</td>
<td>8.0</td>
</tr>
<tr>
<td>extinction quartz</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Polycrystalline undulate</td>
<td>0.9</td>
<td>0.9</td>
</tr>
<tr>
<td>extinction quartz</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Microcline</td>
<td>0.8</td>
<td>0.6</td>
</tr>
<tr>
<td>Plagioclase feldspar</td>
<td>3.1</td>
<td>4.5</td>
</tr>
<tr>
<td>Rock fragments</td>
<td>5.7</td>
<td>3.8</td>
</tr>
</tbody>
</table>

Frequently mudstone clasts have been distorted to form a pseudomatrix (Nockolds et al. 1978). Grain contacts are concavo-convex but point contacts are developed where abundant cement is present. Corrosion of the quartz clasts by this cements is common. Similar features have been noted in many parts of the world (Johnson and Vondra 1972, Selley 1964). Occasionally up to 50% deformed and micritised ooids may be present. It is shown below that these oolitic sandstones appear to have a relationship with the oolites of lithofacies (e). Rarely the ooids and bioclastic debris form discrete 1-3mm. thick bands. Associated with the well formed, well sorted normal ooids are fragments of crinoids.
and brachiopods.

Bioturbation is variable. The rock fragments are of brown mudstone and other supra crustal rock material. The brown mudstone is probably of a local provenance since its survival potential in the postulated environment would be low. The feldspars are variably weathered, plagioclase is frequently heavily kaolinised whilst the microcline shows only minor weathering.

The petrology of the sandstones appears essentially similar to that of the lower clastic units (Lower Quartz Sandstone - Quartz Pebble Conglomerate-Upper Quartz Sandstone) with regard to the preponderance of undulate quartz as opposed to straight extinction quartz and the presence of microcline. This similarity was borne out when the point counting results were subjected to a "T" test (Chatfield 1978). No difference was found at the 95% significance level.

The lack of chert in the sandstones of the Upper Mixed Beds is probably related to the less stable and durable nature of chert (Folk 1968). The lack of polycrystalline quartz grains reflects the more sustained high energy in the postulated environment. Polycrystalline grains break down to single grains under prolonged abrasion (Folk opp.cit.).

It would appear from this data that the lower clastic units (see above) and the Upper Mixed Beds shared the same general provenance.

Planar lamination is developed towards the contact with the overlying Shaley Pales and underlying Navan Micrite. Very occasionally angular shale or lime mudstone clasts up to 1 x 1.5cm. in size form conglomeritic beds up to 10cm. thick. Thin (2cm.) mudstone partings are
frequently present. Ripple lamination and cross bedding are moderately common in the lithology. Occasionally beds of cross-bedded sandstone up to 75cm. thick are found in sections of the siltstones of lithofacies (c), these sandstones frequently have erosional bases, planar lamination, cross-bedding and bioturbated tops. Herring-bone cross-bedding is occasionally found near the top of the Upper Mixed Beds near its contact with the Shaley Pales.

The clean sandstones of the Upper Mixed Beds tend to be localised into a very thick bed at the top of the lithology. A thinner bed, never more than 6m. thick occurs at the base (Fig. 53). This lower bed may not always be present.

**Lithofacies (b) Unclean sandstones**

These sandstones are lithic arenites and wackes often showing parallel lamination. They are confined to the extreme top of the Upper Mixes Beds and to their lower half (Fig. 53). Bioturbation is abundant throughout. Frequent 1-2cm. beds of black mudstone are present in addition to flasers of shale and mudstone.

**Lithofacies (c) Shales, siltstones and mudstones**

Though developed throughout the Upper Mixed Beds by far the greatest development of this lithofacies occurs in the lower parts of the Upper Mixed Beds. Beds of these materials tend to be thin. They rarely exceed 1M. in thickness and generally are very much thinner.

The siltstones frequently develop thin mud flasers and show occasional slump features.

The lithology is often interbedded with the carbonate sediments of lithofacies (f).
Lithofacies (d) Bioclastic rudstones and grainstones

These rudstones and grainstones are confined to the top 20% of the Upper Mixed Beds (Fig.53).

They are composed of very well rounded and sorted bioclastic debris often showing a well developed imbricate structure. Quartz clasts are absent. The bioclasts have been strongly micritised. Often a micrite rind is left infilled with coarsely crystalline equant ferroan calcite. The bioclasts consist of crinoidal, brachiopod, bivalve and bryozoan debris.

Lithofacies (e) Oolitic grainstones

The oolitic grainstones of lithofacies (e) are found only at the base of the thick sandstone forming the top of the Upper Mixed Beds (Fig.53). Philcox (1984) has divided this thick sandstone into two parts. The top two-thirds (approx.) has been termed the Upper Sandstone (Philcox opp.cit.) whilst he has placed the remainder (oolitic portion) in his Pale Beds.

Only the presence of ooids appears to differentiate the one sandstone body from the other. It is proposed that the entire sandstone sequence at the top of the Upper Mixed Beds (Fig.53) be called the Upper Sandstone for the purposes of environmental interpretation. However it must be emphasised that the term is not used in the sense of Philcox (opp.cit.).

The ooids in the lithology are well rounded and sorted with quartz and, rarely, feldspathic nuclei. The euhedral nature of some of the nuclei and the calcite inclusions within them argue that some of the quartz is of an authigenic origin.
Although micritisation is widespread, enough structure remains to show that the ooids are of normal marine origin. The average size is about 0.7 mm. Occasionally compound ooids are present. Some of the ooids show compaction features whilst others in the same slide do not. This may indicate an early lithification of the sediment. Marine cements are not however seen. Ooids can form over 90% of the lithology and as mentioned above a gradational relationship to the oolitic sandstones of lithofacies (a) is present. Some minor dolomitisation of the ooids is also seen.

Heavily micritised and rounded fossil debris of brachiopod origin is present together with intraclasts up to 8 mm. in diameter. These comprise moderately rounded fragments of algal lime mudstone, birds-eye micrite and lime mudstone containing brachiopod debris.

Occasional intraclasts contain ooids. These are not thought to have been derived from the immediate area since they contain much more silt sized quartz than is present in the remainder of the sample. It would appear that these are possibly another indication of some degree of early lithification of the sediment. These intraclasts were probably created due to the intense burrowing of the sediment and then rounded by current action (mud balls of Wilson 1975). Some may owe their origin to "ripped-up" in storms.

Lithofacies (f). Lime mudstone and other carbonate sediments

The lithofacies contains lime mudstones, bioclastic wackestones and algal bindstones.

A major proportion of the limestones in the succession is almost completely dolomitised and the identity of the limestone precursors is
thus obscured. The dolomite is mainly microcrystalline with some ranging up to finely crystalline. However enough of the original lithology remains to show that they fall into one of the basic types outlined above.

The lime mudstones tend to contain silt sized quartz and occasionally shows weak bioturbation and peloids. Planar, calcite filled cavities are variably present. Ostrocods, bivalve debris and calcispheres make up the biota. These show no preferential orientation.

The bioclastic wackestones occasionally grade into packstones. They also contain silt sized quartz. Peloids are present and the lithology is well burrowed. The biota consists of sparse ostrocods and abundant bivalve debris of different sizes. Brachiopod and crinoidal debris showing evidence of abrasion is present along with micritised ooids. The debris is non orientated and poorly sorted.

The algal mudstones show very little quartz when compared to the other limestone types. This may be a consequence of increased sedimentation rates in the algal mudstones compared to the other lithologies. This may have been due to high production rates of calcium carbonate by the algae. The lithology is well burrowed and peloids are present. Small serpulid worm tubes are seen, being locally abundant. Bivalves and calcispheres plus worn brachiopod debris make up the biota. The debris is rarely micritised.

Much of the limestone is rather argillaceous when compared to the Navan Micrite. The amount of argillaceous material shows a tendency to increase towards the contact with the overlying thick sandstone (Upper Sandstone).
C. Carbonate cementation.

Moldic porosity is occasionally filled with a series of cements. An authigenic quartz cement is followed by coarsely crystalline dolomite. These are followed by "cockscomb" barytes.

Coarsely crystalline ferroan dolomite occurs as a cement in the sandstones and as vein fill.

Occasionally a cement of compositionally zoned calcite is seen to follow dolomite. The "zoned" cement was probably developed as a result of changes in the chemistry of the circulating groundwaters possibly due to a fluctuating water table.

D. Palaeontology

Lithofacies (a) Clean sandstones

In general very little fossil evidence is available in lithofacies (a) apart from scattered and abraded remnants of crinoids and ooids.

Rare instances of highly fossiliferous clean sandstones are found. These types are confined to sequences interpreted as washovers (see below). The fossils form a hash of abraded crinoidal, bryozoan, brachiopod and bivalve debris. All show micritisation effects bar the crinoidal debris, possibly a result of preferential micritisation (Flugal 1982). A pronounced orientation and imbrication is apparent. Micritised ooids are present in addition to angular 4mm. long chips of lime mudstone.

Since the fossil content of these types bears little resemblance to that of the surrounding limestones and siltstones it is thought that the faunal remains are not indigenous. These sandstone types are categorised as the carbonate intrabasinal sandstones of Zuffa (1980).
In addition fossiliferous sandstones are found in the trangressive sandstones (see E.4).

Lithofacies (b). Unclean sandstones

Rare bivalves and ostracods are found.

Lithofacies (c). Shales, sandstones and mudstones

No body fossils detected.

Lithofacies (d). and (e).

Owing to the high energy conditions postulated for these environments (Sect.E) it is considered doubtful that any of the faunal content is of an in situ origin. It is probable that the bioclasts were brought into the depositional site and then reworked to their present well rounded and sorted form.

E Interpretation

The position of an extensive body of clean, well sorted sublitharenite and associated lithologies showing marine depositional features and biota lying between the lagoonal sediments of the Navan Micrite and the open marine shelf sediments of the Shaley Pales and Bioclastic Limestone suggests that the Upper Mixed Beds were derived from an offshore bar sand.

The bar sand and its associated sediments have been separated in the introduction into units 1 and 2 (Fig.53) respectively. Contrary to normal practice these units will be interpreted from the top (unit 1) downwards since it is felt that this will make the succession more easily understood.

Unit 1

The bar sand is interpreted as the remains of a barrier island. It
is interpreted as such since the bar sand was separated from the shoreline by the lagoonal deposits of the Navan Micrite and the derived quartz clasts in the contiguous lagoonal deposits indicate that the bar sand was emergant. It is thought that the quartz clasts were transported by wind from the barrier island.

In common with many barrier island sequences in the geological past e.g. the Silurian of south-west Wales (Bridges 1976), the Cretaceous Pee Dee and Black Creek Formations of the Carolina Coast Plain (Swift 1968) to the present day barrier islands of the American Atlantic seaboard (Fischer 1961) little or nothing of the actual barrier has escaped destruction by the transgressing sea during the process of shoreface erosion (Swift 1968). Transgressive barrier islands are however occasionally preserved (Curry et.al. 1969).

In the particular case of the Upper Mixed Beds at Keel only a thin sandstone (up to 4m. thick but normally much thinner) resulting from shoreface erosion and the occasional remains of tidal inlets and shoreface deposits is preserved of the actual barrier island sequence. Although substantial portions of what are interpreted as back barrier deposits remain, it is quite conceivable that some of these also may have been re-worked. Certainly it appears likely that the lagoonal sediments of the Navan Micrite were considerably eroded by the transgressing barrier leaving a "ravinement zone" (Stamp 1922) made up of the thin sandstone referred to above.

Cores in which definite traces of barrier island sedimentation occur e.g. channel lags, ooids etc. mark positions of stillstand and upward or progradational growth of the barrier in intervals of slowdown or
halt in the rate of marine transgression. Other cores showing only the variably thick ravinement sandstone mark areas transgressed by the barrier between one still-stand position and the next when transgression re-commenced. Distances between positions of still-stand can be tens of Kilometres (Swift 1975).

Since the Upper Mixed Beds at Keel, and similar equivalent beds in other areas of the Irish Midlands (Philcox 1984), overlie lagoonal sediments, where sand would not have been available from those sediments to replace that lost by shoreface erosion and that deposited to seaward during still-stand, it is thought that the replenishment sand originated externally.

The similar petrology of the clean sandstones of lithofacies (a) to those of the lower clastic units has been remarked upon above. Coupled with the increasing thickness of the Upper Sandstone unit of Philcox (1984) east to west (Fig.54) this is thought to indicate a sediment source, possibly via longshore drift, from a fluvial system draining the Dalradian Highlands to the north-west and entering the sea to the west of Keel. The correlation co-efficient for this thickening is -.613 (Probability of no correlation 10.5 %). A thinning is also noted to the north of Keel with a correlation co-efficient of .43 (probability of no correlation 18.5 %). A similar thinning is noted in the Gulf of Mexico where a westward drift of material supplies the Gulf Coast barriers (Rusnak 1960).

The extensive amount of lagoonal sediment may indicate that the tidal range was moderate (Fischer 1961). Alternatively the same preservation may occur in cases of rapid transgression (Swift 1968).
FIG. 54. THICKNESS VARIATIONS IN THE UPPER SANDSTONE OF PHILCOX (1984) SHOWING THICKENING OF THE LITHOLOGY TO THE WEST AND THINNING TO THE NORTH.

APPROXIMATE SOUTHERN LIMIT OF THE UPPER SANDSTONE.

THICKNESS OF UPPER SANDSTONE (IN METERS).
Lengths of the (unit 1) Upper Mixed Beds core indicate specific areas of the barrier island model that have escaped destruction. These sequences are described below.

1 Tidal channels (Fig.55)

The tidal channel sequences may be split into six component parts; these are described below:

1. A massive sandstone of lithofacies (a) with mudstone bands, less than 1 cm. thick, of lithofacies (c). This represents the underlying back-barrier sediments. The association of clean sandstones with thin shales in this setting might suggest a washover origin for this particular unit.

2. Variably bioturbated rudstones of lithofacies (d) interbedded with thin (1 cm.) clean sandstones. The fossil debris is worn and well sorted. These rudstones are thought to represent fossil lags.

3. Mid grey fine grained sublitharenites of lithofacies (a). The base of the bed shows marked bi-directional (herring-bone) cross-bedding. This is thought to occur as a result of water flowing through the inlet on the flood and ebb tides.

4. The mudstone is possibly a consequence of deposition in an eddy or backwater.

5. Oolitic grainstones of lithofacies (e) and oolitic sandstones of lithofacies (a). The percentage of ooids shows a definite increase in an upward direction from approximately 10% to over 90%. Variable ooid percentages have been linked to depths of formation (Flugel 1982). In the Persian Gulf it has been noted that in the areas of levees bordering barrier island tidal channels ooids increase in size and abundance with
FIGURE 55. UPPER MIXED BEDS, TIDAL CHANNEL SEQUENCE, CORE K 64.

Lithology

Cross-bedded fine grained sublitharenites. A decrease in cleanliness and an increase in mudstones is seen in an upward direction.

Oolitic grainstones and sublitharenites with a substantial content of ooids. Ooid numbers increase upward.

Mudstone
Fine grained sublitharenites. Bi-directional cross-bedded in base.

Bioturbated mudstones with clean sandstones.

Massive sublitharenite with thin mudstones.

Interpretation

Accreting barrier sands.

Tidal inlet levee.

Quiet backwater

Tidal inlet deposits with lags.

Backbarrier.
increasing proximity to the levee (Loreau and Purser 1975). Thus it appears that this particular part of the sequence marks the transition of a levee as the inlet migrated along the barrier island. Cases are known where a levee only exists on one side of a channel on its seaward end (Loreau and Purser opp.cit.). The intraclasts possibly result from "rip-ups" as a channel migrated within the lagoon. These clasts were then transported out of the lagoon on the ebb tide and became resident in the vicinity of the levee away from the main channel where water velocity was not great enough to carry them out of the inlet but could move and round them.

(6) These are cross-bedded fine-grained sublitharenites of lithofacies (a). A decrease in cleanliness and the increase of mudstone bands is noted as the top of the bed is approached. Cross-bedding is present in the bed but the majority of the sandstone is massive. This may be due to bioturbation. The sandstone is thought to mark accreting sediments as the tidal inlet migrated. This would result in lower energy states in an upward direction giving the reduced cleanliness and the mudstones. As the sands migrated they were deposited on top of the tidal inlet deposits. P.23 shows sandstone actually deposited on top of ooids probably reflecting lateral migration of the inlet bank.

The presence of tidal inlets indicates that tidal conditions were probably not microtidal i.e. below 2m.(Elliot 1978).

Tidal channel sequences are very rare in the Keel core. The abundance of such inlets is a function of;

a) The number of inlets per unit distance.

b) The rate and constancy of migration.
c) The nature of regression and the rate of barrier regression relative to inlet migration.

(Dickinson et al 1972)

Padre Island, Gulf of Mexico, does not have an inlet in 180KM. of barrier island (Rusnak 1960). Bridges (1976) has calculated that at slow inlet migration rates an inlet might occur at a given spot once every 64,000 years in a distance of 200KM. He also makes the point that some inlets move back and forth over the same tract and do not migrate over the full length of the barrier. Thus the detection of only one inlet in the Keel core examined is not remarkable.

2 Shoreface

Shoreface sequences are moderately common in this unit of the Upper Mixed Beds.

The sediments deposited on the back-shore of a beach resemble those of the shoreface and might be confused with them since both show signs of fluctuating energy when compared to the consistently higher energy conditions of the intertidal zone.

However in a transgressive regime, such as at Keel, a preserved backshore sequence would have lithologies representing the swash zone and shoreface overlying it. This is not the case at Keel. In addition backshore sediments show a very poor potential for preservation during marine transgressions.

The following features are recognisable in shoreface sequences in the Keel core (Fig.56);

1) Fine grained, very well sorted, supermature sublitharenites of lithofacies (a). Occasionally small mudstone clasts are found. Small
Figure 56. Upper Mixed Beds. Shore-Face Sequence. Core DDB 6.

Depth down hole (m)

16.8

Lithology

3 Bioclastic rudstones with sandstones.

2 Shales interbedded with sandstones.

1 Clean Sublitharenite

Interpretation

Shore-face. Bioclastic lenses represent storm lags.

Deposition above fair weather wave base.

Foreshore.
Pebbles have been reported from near the low water mark in foreshore sediments on high energy coasts (Clifton et al. 1971). Thus it is possible that lower foreshore sediments are present. The occasional presence of swash zone lamination in a similar position, very low in the sequence, may confirm this.

Bioturbation is sparse in this lowermost section of the core. Lamination tends to become increasingly well marked with distance up the section. Cleanliness declines in a similar manner. This increase in lamination is thought to represent the increasing depth of water and the consequent decline in energy levels. An increase in the amount of bioturbation is noted with distance up the section. Increases in bioturbation with distance from the shoreline are seen at Sapelo Island, Georgia (Elliot 1978).

2) Shales of lithofacies (c) with 0.3cm. laminations probably marking prolonged intervals of calm weather enabling deposition to take place at or just above fair weather wave base.

3) Bioclastic rudstones of lithofacies (d). The bioclasts are only moderately sorted but well rounded and micritised. The bioclasts comprise debris from crinoids, brachiopods and bryozoa. They represent the typical fauna of the Bioclastic Limestone Unit (Chap. 14) that, it is believed, was developing to seaward. As such they are thought to have been derived from seaward by storms. Such debris has been reported to have been transported from depths of 25m. onto barrier islands (Pierce 1970). Owing to the rounding of the clasts it is thought that the transport of the debris was not carried out in one episode but rather as a series of steps through progressively shallow water where the debris...
would be rounded during and after storms. There would be a period of calm before the debris was remobilised by the next storm. The erosional bases commonly developed under these beds may confirm this storm origin. Transported and mixed assemblages are common in nearshore zones (Kreisa 1981). It is thought that these Upper Mixed Beds types are equivalent to the storm lags of Brenner and Davies (1973).

3 Washovers

Where barrier sequences are preserved a major portion of the lower half of unit 1 is thought to consist of washover material.

The following core sequence (Fig.57) represents a series of such washovers.

1) Beds of shell debris up to 20cms. thick appear low in the section. The debris is well sorted and rounded. Clasts of lime mudstone are incorporated with the shell debris. Erosive bases are seen on the beds. These beds occur in a fine grained lithic arenite of lithofacies (b).

2) This section consists of a mid grey, fine-grained lithic arenite of lithofacies (c). 2cm. bands of black mudstone are common. These are thought to represent the normal back barrier sediments. 0.5cm. low sphericity black clasts, composed of what appears to be the same material, in the overlying sandstone implies erosion of the mudstone by the sandstones. Cross-bedding is occasionally visible but the very extensive bioturbation seen throughout has probably destroyed most of the sedimentary structures which were previously present.

Section 1 probably represents proximal washovers with the basal shell beds fining up into the fine-grained sandstone as the velocity of
Figure 57. Upper Mixed Beds, Washover Sequence, Core K 61.

UNIT LITHOLOGY INTERPRETATION

Fine grained lithic arenites with thin black mudstones. Abundant bioturbation.
The sandstones represent washovers of a distal type.

Mudstones represent the normal back-barrier sediments.

Fine grained Proximal type lithic arenites with beds of shell debris. The sandstones
-es fine-up.
the flow decreased. Section 2 is probably of a more distal character
only intermittently deposited with normal back barrier sediments being
deposited between events.

Precisely why, in an overall transgressive regime, distal washovers
should overlie proximal types is unclear. The reverse situation would be
considered more normal. Possibly the washovers were created during a
temporary seaward progradation of the barrier.

The existence of washovers and tidal inlets is not mutually
exclusive. Both are known to occur on Galverston Island, Texas (Reineck
and Singh 1980).

Both inlet and shoreface sequences commonly have an overlying
sandstone body closely resembling the transgressive sandstone sheet
described below.

4 Transgressive sandstones

Where traces of the barrier island are not present the equivalent
lithology is a 0.8-4m. thick sequence of sandstone which rests on unit 2
of the Upper Mixed Beds (Fig.53) or on the Navan Micrite. An erosive
basal contact is developed. In some instances the sandstone may be
absent and the mudstones and argillaceous limestones of the Bioclastic
Limestone unit will immediately overlie unit 2 or the Navan Micrite.

The sandstones are of Lithofacies (a) type but show a distinct
trend to become dirtier as the basal contact is approached and the
sandstones eventually grade into less clean lithofacies (b) type
sandstones. The sandstones are locally very fossiliferous with abundant
reworked crinoidal debris. Local concentrations of gastropods and
bivalve shells are also seen. The lithology is frequently rippled and
lamination is developed where the commonly found bioturbation has not destroyed it.

As noted above the sandstone is thought to mark the passage of the re-worked barrier island sediments over the lagoon and back barrier. The sandstone thus marks a ravinement. The fossiliferous nature of some of the sandstones probably reflects re-working of these underlying lithologies. The fact that gastropods and bivalves are more common in these lithologies as opposed to the Upper Mixed Beds and the Bioclastic Limestone Unit suggests this. Lenses of fossils from underlying lithologies have been described from the Peedee/Black Creek Formations contact, thought to mark a ravinement (Swift 1968).

No sequence of sedimentary structures or lithologies characteristic of barrier islands has been found in the transgressive sandstones.

Unit 2

This unit comprises the mixed succession lying below the barrier sand of unit 1 (Fig.53). It is equivalent to the lower part of the Pale Beds of Philcox (1984) (Fig.8).

The succession appears to represent a back barrier facies in which sediments derived from the sand body to seawards are closely associated with sediments of a Navan Micrite type.

The succession is uncommon in the Keel core, appearing only rarely when the thin transgressive sandstone referred to above is present. This may be due to erosion during the transgression but is more likely due to the fact that such back barrier deposits would not be developed except at stand-still stages.

Although the succession is extremely variable both on lithological
and bed thickness grounds a gross pattern is discernable.

Two sub-units are recognisable (Fig.53):

Sub-unit 1). These are fine grained, moderately clean sandstones of lithofacies (a). Planar and cross bedding are both developed locally. Interbedded black (well laminated) shales and mudstones are present plus rare lime mudstone beds. The basal contacts of the sandstones with these mudstones are erosive. Flaser bedding is occasionally developed as well as rare slumps. The sandstones tend to be bioturbated and to become cleaner as the boundary to the Navan Micrite is approached. Fossils are locally abundant being chiefly crinoidal, brachiopod, bivalve and bryozoan debris, this debris is well sorted and rounded. Rounded clasts of dark mudstone are also present plus compacted ooids. The maximum development of the component is 9m.

Sub-unit 2). This is composed of the various dolomite and limestone lithologies of lithofacies(f). As such it bears a striking resemblance to the Navan Micrite. Shale and mudstone beds up to 15cm. thick are developed.

Interpretation of Unit 2

The largely sandstone sub-unit 1) is thought to represent the initial manifestation of the barrier sand as, at the end of a transgressive event, it came to rest in the Keel area. The Navan Micrite was already in place having been protected to some degree by the barrier when it was in its more southerly position. The presence of planar bedding implies high fluid velocities of 60-120cms./sec. (Blatt et.al. 1980) whilst the fossil debris, little of which is common in the lagoonal environment in which it was deposited, argues for derivation
from outside the lagoon. The rounded nature of the clasts implies substantial re-working outside the generally quiet lagoonal environment. Since this fossil debris tends to be localised above erosion surfaces and having consideration of the above factors an origin for the sediment as the result of washovers is proposed. Transport of fossil debris from depths of 25m. onto barrier islands and in some cases through them via washovers has been noted (Pierce 1970).

The planar bedded sandstones probably resulted from washovers of a more proximal nature than those showing cross-bedding which were deposited on the distal margins of the washover fan. The siltstones and shales represent the "background" low energy deposits on the back barrier. The presence of flasers indicates a tidal influence. Slumps along the tidal flat/lagoonal boundary have been reported from Laguna Madre, Texas (Dickinson et al. 1972).

The rare lime mudstone bands within the sandstone may have been precursors of the prograding limestones described below. Interfingering of washover and lagoonal deposits has been noted (Dillon 1970).

After the barrier became stationary, due to a slow-down or halt in the rate of sea level rise, the barrier began to grow upward and may even have begun to prograde seaward.

Within the lagoon production of low energy lime mud started to reach a maximum again. Possibly the turbid nature of the water due to repeated washovers had depressed the production, thus few limestone beds are found in the sandstones. Deposition of lime sediments now began to exceed the rate of relative sea level rise and carbonates started to be deposited on top of the washover fans. The mudstone beds in the
limestones may represent the extreme distal ends of washovers from the barrier.

The absence of sand washovers in component 2) is worthy of mention since washover formation might be expected to carry on even if the barrier was growing upward. Two factors might explain this;

a) The barrier may indeed have started to prograde as barriers are known to do when enough sediment is available. This progradation would tend to take the "zone" of washovers with it so that washovers would no longer be deposited so far into the back barrier.

b) After the transgression had slowed or stopped conditions for the creation of washovers may have been less favourable. These less powerful washovers may not have reached into the lagoon, the sands etc. would have been deposited on the back barrier.

At the start of the next phase of the transgression the rate of sea level rise would have been relatively slow. the barrier island (Unit 1 above) transgressed slowly over Unit 2 depositing normal back barrier and barrier beach sediments. The limestone and mudstone clasts in the sandstones of lithofacies (a) may have their origin in the erosion of Unit 2 sediments by the transgressing barrier. As the rate of sea level rise increased sediment supply to the barrier was cut off as rivers started to deposit sediment in their estuaries instead of supplying it to the near shore marine environment (Bridges 1976). Shoreface erosion started to dominate and the thin transgressive sands of Unit 1 started to be laid down as the barrier migrated.

However shoreface erosion was not deep enough to erode away the
sediments of Unit 2 which were preserved along with parts of Unit 1.
CHAPTER 13
Shale Pales

A. Introduction

The Shaley Pales have been included in the Upper Mixed Beds in the local nomenclature (Slowey 1986). However owing to the distinctive nature of this lithology in the Keel area they are given a separate identity in this work in common with Philcox (1984) who correlates them with the Middle Ballysteen Limestone of southern Ireland.

In the core examined the Shaley Pales may be up to 62m. thick but are usually substantially thinner. Phillips and Sevastopolou (1986) have reported that this lithology becomes much sandier to the north and west in the Irish Midlands. This is thought to reflect sediment supply from the west by longshore drift in an analogous manner to that outlined for the barrier sands of the Upper Mixed Beds. It is considered unlikely that any significant amount of sediment was derived from the immediate Keel area since the barrier would have effectively starved the area in front of it by trapping the sediment behind it.

The first clearly definable bed of the Shaley Pales is a fissile and silty black shale. However between this shale and the clean sandstones of the Upper Mixed Beds are interbedded shales and sandstones.

B. Lithology

The succession is a series of sandstones, bioclastic grainstones, bioclastic packstones, shales and siltstones.

Philcox (1986) has subdivided the Shaley Pales at Keel into three parts. The author has added a fourth, the Transition Beds;
4) Upper Shaley Pales.
3) Middle Shaley Pales.
2) Lower Shaley Pales.
1) Transition Beds.

Whilst the main trend of Phicoxs subdivision is agreed with i.e. a general change from fine-grained sandstone to shales and bioclastic packstones it is felt that such rigid subdivisions are not really appropriate to such a variable succession, at least in the case of Keel.

1) Transition Beds.

This sequence may be up to 8m. thick. The thickness of individual sandstone beds declines upwards with increasing proximity to the Lower Shaley Pales. Individual beds of sandstone may reach 1m. in thickness. Amalgamated beds of a similar postulated origin are known to reach 5m. in thickness (Goldring and Bridges 1973). Cross-bedding and normal grading are occasionally developed in the sandstones. The shales are well laminated. The sandstones are unbioturbated and range from lithic wackes to sublitharenites. Clasts of lime mudstone are common in the sandstones especially near the bases.

2) Lower Shaley Pales.

These shales, which constitute the Lower Shaley Pales in the Keel area, average 0.7m. in thickness (std.dev. 0.5M.). The shale is fissile and silty and usually non-calcareous though thin wisps of lime mudstone are sometimes present. In addition thin laminae of silt and very fine-grained, rippled and flaser bedded sandstone are developed and show erosional bases with the shale. Normal grading of the silts and fine
sandstones is seen.

3) Middle Shaley Pales.

The Middle Shaley Pales may be up to 34M. thick but in most cases are considerably thinner.

This lithology is basically composed of bioturbated mudstones and siltstones and bioclastic lime packstones. The packstones are particularly prominent in the lower half of the sub-unit.

The packstones, which may show a rudstone texture at the base of a bed, consist of crinoidal and bioclastic debris with bivalve and brachiopod shells. The packstones are usually in erosional contact with the siltstones (P.24). Rarely, well sorted and rounded bioclastic grainstones are present composed of micritised crinoidal, brachiopod, bivalve and bryozoan debris along with rare ooids.

4) Upper Shaley Pales.

The Upper Shaley Pales may be up to 24M. thick at Keel. They are primarily bioclastic wackstones similar in faunal content to the packstones of the Middle Shaley Pales. Shales are present in variably thick (up to 3M.) beds. The shales tend to be unfossiliferous and have siltstone partings. Philcox (1984) has placed the top and the basal boundaries of the Upper Shaley Pales where shales over 0.7M. thick enter the sequence. This practice has been followed in this study.

C. Carbonate cementation.

A finely crystalline non-ferroan calcite is present as vein fill. A ferroan calcite cement infills intergranular and intraparticle porosity.

An isopachous acicular (marine) calcite cement is occasionally present with a later equant ferroan calcite cement infilling
intraparticle porosity.

**D. Interpretation.**

The bulk of the Shaley Pales is thought to represent deposition on a shelf starved of any great deal of coarse terrestrial material by the presence of the Upper Mixed Beds barrier. As remarked upon above most of the sediment was probably derived by longshore drift from the west. The fine silt and mud was separated from the sand of the barrier by storm and wave action and emplaced in deeper water to seaward.

1) Transition Beds

The beds are thought to be similar to the Transition Zone of Reineck and Singh (1980) and mark the link between the coastal sand deposits (Upper Mixed Beds) and the fine grained shelf deposits (remainder of the Shaley Pales).

The decline in the thickness of the beds and their (normally) graded nature indicate that the sandstones may represent storm deposits. Cross and graded bedding are frequently found in these type beds (Walker 1984). The interbedded siltstones and shales represent the background sedimentation. The well laminated shales indicate that deposition was mainly below the fair weather wave base of 5-15M. (Walker opp.cit.). The limestone clasts in the sequence may indicate deposition near a tidal inlet. Clasts might be transported into the inlet by tidal action from the lagoon to landward. The clasts may have been produced in the lagoon as a result of burrowing (mud balls of Wilson 1975). From here they might be moved seaward by strong currents, following storms, along with the sand. It has been predicted that hurricanes may produce velocities of 1M/sec. in the seaward returning storm surge ebb current (Forristall
The preservation potential of storm beds is high since the energy levels during fine weather are much lower than during the storms which created the beds.

The decrease in thickness of the beds is a function of distance from source and of water depth i.e. the proximal beds will generally be thicker than the distal. Thus beds will thin in an upward direction in a transgressive regime.

2) Lower Shaley Pales.

The flaser bedding within the siltstones indicates that the sediments were deposited in comparatively shallow water subject to tidal influences. The presence of ripples indicates the influence of traction currents possibly induced after intermittent storms, a similar mechanism is thought to be responsible for the siltstone lenses. The storms agitated the sediment and winnowed away the fines which were then deposited in deeper water leaving the silt. Alternatively they may represent the extreme distal end of storm surge deposits (above). Storm deposits have been found 40KM. away from where they originated (Reineck and Singh 1980).

It is considered feasible that the Lower Shaley Pales simply represent an extended period of time during which major storm activity was absent, only minor storms effected the sediment. The lack of bioturbation may, on the other hand, reflect rapid deposition of the shales during which the sediment was not burrowed. Sediments in these near shore zones tend to be heavily bioturbated (Reineck and Singh opp.cit.).
3) Middle Shaley Pales.

The Middle Shaley Pales are interpreted as silts upon which grew abundant crinoids and bryozoa. Within the sediment were brachiopods and bivalves. Occasional corals are also present. This comparative rarity of corals is commented upon in Chapter 14. The presence of the crinoids and corals indicates clear water conditions. Certainly mud sized sediment, such as occurs in the Lower Shaley Pales, was not present on a large enough scale to irritate the delicate filter feeders. This lack of very fine grained material and the presence of crinoids indicates that there was some water movement probably caused by current activity. Water depths of less than 100M. would be required for coral growth given that the water was not turbid.

The background sediment i.e. the siltstones with some fossil debris can be seen in P.24 lying between the bioclastic bands.

The bioclastic packstones are thought to the deposits created after winnowing of the silts after storms and are interpreted as similar to the storm lags of Brenner and Davies (1973). The fining upward character of the debris, the erosional bottom contacts and pronounced alignment of the fossil fragments parallel to the base of the bed tends to confirm this.

The rare grainstones interbedded with the siltstones are also interpreted as storm deposits. The interbedding of these coarse and relatively fine grained sediments testifies to sudden short term changes in energy levels such as would be created by storms. However the micritised, well sorted and rounded nature of the clasts suggests that the debris was subjected to considerably more work probably in shallower
water. It is proposed that the debris originated on the shelf, where crinoids etc. were common, and was transported onto the barrier island by waves and/or currents. An origin within lagoonal channels, where these organisms were also common, and subsequent transport into the tidal inlet is not ruled out however. In either case rounding and sorting of the debris was carried out in a high energy environment. The surf zone of the barrier beach or tidal inlets suggest themselves as such environments. Owing to the greater rarity of tidal inlets and the rarity of the grainstones tidal inlets are favoured. An origin on beaches would result in many more of these type sequences being present. The presence of ooids, already detected in tidal inlet sequences in the Upper Mixed Beds, may reinforce this reasoning. After rounding the debris was transported out onto the near shore shelf by storm surges in a similar manner to the sandstones of the Transition Beds. The fining upward association of the grainstones with siltstones and mudstones/shales is similar to storm generated sequences found in the Ordovician of south-west Virginia (Kreisa 1981) and the Devonian of New York State (Craft and Bridges 1987).

4) Upper Shaley Pales.

Although the packstones were formed in a similar manner to those in the Middle Shaley Pales i.e. storm generated, it is considered that the sediments of the Upper Shaley Pales reflect deposition at greater depth. This is suggested by the greater amount of fine-grained sediment. The diminished influence of storms i.e. the wave base was lowered less frequently due to the greater depth of water, is reflected by the lower number, thinner bedded and siltier nature of the packstones.
This follows a trend discernible in the upper parts of the Middle Shaley Pales of fewer interbedded packstones being present.

Fig. 58 shows insoluble residue and carbonate analysis of the Shaley Pales. They reflect the lower carbonate and higher insoluble residue contents of the Upper Shaley Pales and then the rise in carbonate and fall in insoluble residue of the Middle Shaley Pales. The rise in terrigenous silt in the Upper Shaley Pales is reflected in decreased carbonate and increased insoluble residue.
Figure 58. Bioclastic Limestone Unit and Shaley Pales. Insoluble residue versus depth down hole.

\[ Y = -71.54 + 0.105 \times X \]

Manuolitan Mound direction
FIG. 58, BIOCLASTIC LIMESTONE UNIT AND SHALY PALES, CARBONATE VERSUS DEPTH DOWN HOLE.

% CARB.  

\[ y = 223.6 + 0.169x \]

\[ r = -0.73 \]

Bioclastic Limestone Unit  
Shaly Pales

Waxsolution Mound direction

Depth down hole (meters)
CHAPTER 14
CHAPTER 14.

**Bioclastic Limestone Unit.**

A. Introduction.

The Bioclastic Limestone Unit conformably overlies the Shaley Pales and has been correlated with the Upper Ballysteen Limestone of south Ireland by Philcox (1984). The unit is placed in the late *carina* to early *anchoralis* zones (Phillips and Sevastopolou 1986). In the Irish Midlands the unit is frequently termed the Argillaceous Bioclastic Limestone but the local Keel name has retained in this study. Both terms are in fact considered misnomers as the bulk of the unit consists of variably fossiliferous shales and mudstones.

In the Keel core examined the Bioclastic Limestone Unit is up to 130M. thick but elsewhere in Ireland thicknesses of over 200M. are known (Philcox 1984, Slowey 1986).

Together with the Shaley Pales the Bioclastic Limestone Unit is thought to represent the major stage of a marine transgression (Philcox 1984). On palaeogeographic and time considerations cycle 5 of MacCarthy and Gardiner (1987) is suggested.

B. Lithology.

The lithology basically comprises fossiliferous calcareous mudstones and shales, wackestones and packstones. Occasionally black unfossiliferous shales are developed. A gross pattern is discernible as the succession is traced upwards towards the overlying Waulsortian Mudmound unit. This gross pattern is in line with the observation of Wilson (1975) that lithologies deposited in his Standard Facies Belt 2 tend to be uniform over great distances. At the base are fossiliferous
silty calcareous mudstones and wackestones. The fossil debris within these includes crinoidal and brachiopod material. These are succeeded by wackestones and packstones. These show a marked increase in fossil debris, particularly crinoids. Other fauna represented includes impunctuate brachiopods, bivalves, ostracods, bryozoa, rare corals and rare ooids. The bioclastic debris frequently shows a planar orientation and may comprise 30% of the lithology. Occasionally this figure may rise to 80%. In these cases erosional basal contacts are present and the debris is graded. This type of bed tends to be more common near the base of the unit. Apart from these cases the debris is generally poorly sorted and angular. The crinoidal fragments may reach dimensions of 8 x 6mm. In common with all the lithology these packstones and wackestones contain abundant silt and mud. This forms a matrix along with the occasionally abundant micrite. Quartz as replacement and matrix may comprise up to 40% of the lithology. Where silt sized the quartz is subrounded to subangular and is of low sphericity. The wackestones and packstones are interbedded with black calcareous mudstones showing a marked shaley lamination where the lithology is not bioturbated.

The wackestones and packstones are succeeded by more argillaceous, less fossiliferous wackestones interbedded with calcareous mudstone and shale beds which tend to be of a greater thickness than the foregoing lithogies.

Overlying these are wackestones with calcareous shale interbeds. Chert nodules are moderately common. Near the base of the overlying Waulsortian Mudmounds are found beds of very pure lime mudstone similar in appearance to the Mudmound lithology.
Fig. 58 shows an analysis of a series of Bioclastic Limestone Unit and Shaley Pale samples from one core. The purer nature of the crinoidal packstones and wackestones is reflected in the higher carbonate and the lower insoluble residue percentages after the low and high values respectively of the basal mudstones and wackestones. A fall in carbonate and rise in insoluble residue marks the presence of the muddier wackestones. A steady rise in carbonate and fall in insoluble residue reflects the onset of the Waulsortian Mudmound unit which may therefore have been influencing sedimentation at considerable distances from its own environment of deposition possibly by means of some micrite "escaping" from the mounds after production. The correlation coefficients for these figures are -.73 (carbonate) and .67 (insoluble residue) with a probability of no correlation of 4.3% and 6.8% respectively.

Bedding in the unit tends to be of an irregular character (P.25) and often shows ball and flow (Wilson 1975) structures. Occasionally the bedding is at a high angle and is thought to represent slumping.

Bioturbation (P.26) is extensively developed throughout the unit.

C. Carbonate cementation.

Medium and coarsely crystalline ferroan calcite fills veins, interparticle and occasional moldic porosity. This cement is considered to have been deposited from meteoric or connate water below the water table.

An interesting feature seen in thin section (P.27, 28) is the replacement of bryozoan fragments by silica and the deposition of non-ferroan calcite within the fragments. In addition spines, believed to be
siliceous sponge spicules, are replaced by ferroan calcite.

It is possible that the bryozoan fragments were replaced by silica since details of the structure are still visible. The solution would have been of pH below 9 (Blatt et al. 1980) and saturated for silica. Calcite would have been dissolved and silica precipitated as the solution cooled or its chemistry altered.

Replacement of the silica of the spines was carried out by a subsequent solution of high pH (above 9) which was saturated for calcite. This was precipitated as the solution temperature fell or the chemistry changed. A similar mechanism is probable in the replacement of some of the Quartz Pebble Conglomerate (Chap. 6). The selectivity of the spine replacement may be due to the greater porosity of the spine concentrations (P. 27). Replacement of brachiopod and ostracod fragments by quartz is also seen whilst replacement of the abundant crinoid fragments has not been observed. This may indicate a greater susceptibility of low magnesium calcite to replacement by quartz. Authigenic quartz overgrowths on silt are also seen.

D. Interpretation.

The angular nature of the bioclasts infers that transportation was negligible. The poor sorting of the clasts and the lamination of the shales indicates deposition below fair weather wave base. The lack of micritisation of the bioclasts might indicate that deposition was in fact well below the maximum fair weather wave base of 5-15M. (Walker 1984). However the inhibition of micritisation by rapid sedimentation is acknowledged. The presence of storm lags, as seen below, implies deposition above storm wave base. This may be as deep as 200M. (Ewing
1973). Thus, though water movement was not vigorous it must have been sufficient to supply food and oxygen to the prolific filter feeders present.

The profusion of crinoidal and bryozoan debris coupled with a rarity of corals is considered worthy of comment.

The presence of corals in the underlying Shaley Pales suggests that water depth was not a factor since the depth of deposition of that unit was well within the range of corals.

The requirement of both crinoids and corals for firm substrates (Heckel 1972) would suggest that this was not a factor considering the large numbers of crinoids present although the sea bottom was not particularly hard. The shell fragment in P.29 has deformed the sediment as it settled to the sea floor. The disarticulated condition and concave-up attitude of the shell suggests that there was some current activity as in life position it should be the other way up.

It is thought that the absence of corals may be due to the water being somewhat turbid near the bottom. The corals with their feeding apparatus near the bottom could not tolerate the irritation caused by the mud and silt or the detrimental effects on photosynthesis even though there may not have been a great amount of the material. The crinoids, being stalked, would be above this turbid lower layer. Although bryozoans are also intolerant of turbidity (Heckel opp.cit.) it is likely that they grew on the crinoids (Wilson 1975) and were thus not effected by the turbidity.

The rare ooids, though not showing signs of abrasion, may have been derived from the ooid rich portions of the Upper Mixed Beds (Chap.13) by
currents initiated by storms. The velocities of these currents may reach 60 cm./sec. in depths of 10-20 M. (Walker 1984) so would have been perfectly adequate to transport ooids. Ooids have been found in 120 M. water depths off the Yucatan Peninsula. There is a sharp rise in ooid numbers as the water shallows landwards. This is thought to be due to seaward dispersal of ooids from their shallow sites of generation (Logan, 1969). Alternatively a source on the Waulsortian Mudmounds lying basinwards of the Bioclastic Limestone environment is possible. This point is further examined below.

The lamination and black colour of some of the shales imply that anoxic conditions existed within the sediments on occasions. Possibly the water had progressed over the ramp (Chap.16) to such a distance that tidal activity was repressed and only storm activity was available to overturn the water column and re-oxygenate the sediment. Should there be an extended period between such storms stagnant water might accumulate and preclude any fauna. The unfossiliferous nature of some of the shales may confirm this.

The presence of packstones and the orientation of the clasts however indicate that there were periods of higher energy. The erosive bases and grading in the packstones (P.30) argue for a storm origin similar to those proposed for the storm lags of Brenner and Davies (1973). The character of the bedding (P.28) is thought to be the result of sedimentary boudinage. This is envisaged to be caused by the disruption of the patchily distributed shale and carbonate by stretching and flowage. The effects tends to be accentuated by the shales compacting regularly and the early cementing of the carbonate which then
resists compaction (sedimentary boudinage/ball and flow of Wilson 1975). The existence of the Ball and Flow is probably a reflection of the high amounts of terrigenous content in the sediment. Ball and flow structures are common in facies deposited on shelves (Wilson opp.cit.).

As noted above the Shaley Pales and the Bioclastic Limestone Unit are thought to represent the main phase of a marine transgression. This is thought to be of a eustatic origin (MacCarthy and Gardiner 1987). Rises in sea level due to melting ice are not considered likely since there were no polar ice caps at this time (Sevastopolou 1987). Possibly crustal doming prior to the early Atlantic rifting proposed by Haszeldine (1984) may have been the mechanism.

The Shaley Pales were deposited outside the barrier in a shallow marine environment. During the transgression the marine waters overtopped the barrier and cut down the amount of sediment available. This cut off in sediment supply is reflected in the high carbonate and low insoluble residue levels of the lower portions of the Bioclastic Limestone Unit. Two mechanisms probably operated:

a) Less terrigenous sediment was actually available due to source drowning and easing of river gradients giving a decline in stream power.

b) The less turbid water encouraged the growth of filter feeders such as crinoids. The carbonate produced by the organisms effectively diluted what sediment was available.

Conditions then became less favourable to the fauna as the water deepened. This may have been linked to a decline in current activity adversely effecting the filter feeders. Although progressively less sediment was available and was of a finer grain size it proceeded to
become more dominant at the expense of the carbonate.

Eventually the Waulsortian Mudmounds started to influence sedimentation and increased levels of carbonate helped to produce the purer lithologies which underlie the Mudmound Unit. The rise in carbonate levels and fall in insoluble residue in the Bioclastic Unit after the deposition of the muddier wackestones is illustrated by correlation co-efficients of .93 and -.96 respectively with a negligible probability of no correlation.

The chert rich Bioclastic Limestone lying just below the Waulsortian Mudmounds is frequently rich in siliceous spines. Thus it may be that the chert is derived from sponge spicules whose silica has been re-mobilised and coalesced to form the chert nodules.

E. Thickness variations.

Observation of Fig.59 suggests that the Bioclastic Limestone Unit thickens to the south away from the Keel Inlier. This is confirmed by a correlation coefficient of .72 (probability of no correlation 8.6%).

It is considered highly unlikely that factors such as distance from shorelines, minor regressions etc. could have such an effect on sedimentation patterns in such a relatively small area.

The thickening is thought to indicate that the inlier was having an effect on sedimentation during the deposition of the Bioclastic Limestone Unit. Movement along the main Keel Fault (Fig.9) is known to have postdated deposition of the lower clastic units (Lower Quartz Sandstone - Quartz Pebble Conglomerate-Upper Quartz Sandstone). However this movement is postulated to have occurred during deposition of the Waulsortian Mudmounds (Slowey 1986). It may now appear from isopachyte
FIG. 59. ISOPACHETES OF THE BIOCLASTIC LIMESTONE UNIT SHOWING THICKENING TO THE SOUTH-EAST OF KEELE.
evidence (Fig. 59) that movement may have occurred somewhat earlier.

Possibly movement along the fault created a palaeohigh by re-activating the Caledonian trend faults of the Keel Inlier. This same feature may have affected sedimentation patterns to some degree up to Bioclastic Limestone times as has been postulated in previous chapters. More marked movement in Bioclastic Limestone times gave rise to increasingly profound effects.

Alternatively it may be that the area south of the inlier subsided due to fault activity. The main Keel Fault is known to be a normal fault and to therefore have resulted from extensional forces, so the idea of subsidence is possibly more attractive than uplift of the inlier. As subsidence proceeded carbonate production increased accordingly thus giving greater thicknesses of the Bioclastic Limestone.

Two further lines of evidence favour the latter mechanism;

a) In core from holes to the north of the proposed fault little difference in lithology is seen. A difference would be expected if the rock was laid down as a more condensed sequence over a palaeohigh.

b) In general greater thicknesses of the Bioclastic Limestone are present where the proposed Calp basin occurs (Chap.16). Thinner sequences occur in areas where the Oakport Limestone (Chap.16) is present.

Thus it may be that this activity in Bioclastic Limestone times was the precursor of the Calp Basin.

A small sub-basin, within the larger basin outlined above, occurs to the south of Keel. This sub-basin affected sedimentation right up through Waulsortian Mudmound times.
CHAPTER 15
CHAPTER 15.

A. Introduction.

Conformably overlying the Bioclastic Limestone Unit is the Waulsortian Mudmound Unit. A degree of interfingering of lithologies is seen however in the uppermost beds of the Bioclastic Limestone Unit.

The term "reef" has been used in the literature (e.g. Nevill 1958) for this lithology. However the term is not correct since the lithology lacks the numbers of organic frame builders required by the strict definition of reef (see Wilson 1975, Selley 1982).

The term "mound complex" is adapted from Sevastopolou (1982) since the lithology encompasses the actual mound and associated off mound facies in varying proportions.

In the core examined at Keel the mound lithologies reach over 290M. in thickness. This does not allow for loss by erosion since the holes producing the thickest mound sequences were frequently spudded in the Waulsortian Mound. Thicknesses reach 600M. in the Cork City area and may be as much as 800M. around the Shannon Estuary (Sevastopolou opp.cit.). In these more southern areas the actual mound lithology dominates and off mound facies are of much less prominence.

The lithological unit is probably of a late anchoralis age in the Keel area but is of an early carina age in the extreme south of its range on the Kenmare-Cork line. Establishment of the mounds is known to have occurred at a much earlier date in the south of Ireland than in the Irish Midlands (Sevastopolou opp.cit.). It thus appears to be highly diachronous with the base becoming progressively younger to the north.
It would appear that the Waulsortian Mounds advanced behind the transgressing sea but continued to be deposited in sites further to the south which were now covered by much deeper water. This tendency of the Waulsortian Mounds to be deposited in varying depths of water is examined below.

B. Lithology.

Three basic lithofacies make up the lithology of the Waulsortian Mudmound Unit:

Lithofacies a) Thin bedded fossiliferous calcareous mudstone.

Lithofacies b) Bioclastic packstones.

Lithofacies c) Bioclastic wackestones.

Lithofacies a) Thin bedded fossiliferous calcareous mudstone.

The lithofacies shows an abundant fauna. Crinoidal debris tends to dominate, occasionally comprising 80% of the total bioclasts. Also present are fenestral bryozoa, ostracods, brachiopod debris, bivalves, foraminifera and rare corals. Calcispheres are also seen. The bioclasts are poorly sorted with angular clasts. Occasionally they show abrasion effects, being broken with angular edges. The largest bioclasts may be 25 sq.mm. in area (5 x 5mm.). Rarely 20mm. long crinoidal ossicles may be observed with up to 12 still articulated segments.

A tendency is noted for the bioclastic debris to appear in discrete thin laminations interbedded with relatively unfossiliferous mudstone. These laminations are 1-2mm. thick. Thicker unfossiliferous beds also occur up to 5cm. thick interbedded with the mudstones.

Very occasionally some micrite is present and a wackestone texture is developed.
Insoluble residue analysis of the mudstone beds gives figures up to 93% insoluble residue (average = 57.8%, s.d. = 18.3%, n = 14). Samples for analysis were selected from parts of the core not noticeably rich in fossil debris.

Occasionally thin (up to 15cm.) lenses of lithofacies c) wackestone are present within lithofacies a).

The fossiliferous mudstones of lithofacies a) merge gradually into the packstones of lithofacies b) showing increasing amounts of bioclastic debris and decreasing amounts of mud. This occurs as part of an overall transition to the wackestones of lithofacies c).

Lithofacies b) Bioclastic packstones

These packstones contain abundant bioclasts. Whilst sorting and rounding are poor they both appear better than in lithofacies a). Although the bioclastic debris contains ostracods, bryozoans, gastropods and bivalves crinoidal debris makes up by far the largest fraction. This may form over 90% of the total lithology and so constitute encrinites (Wilson 1975). The crinoidal debris occasionally shows sutured contacts and syntaxial overgrowths on the crinoidal debris are occasionally very well developed. Calcispheres are present.

Silt and mud may form up to 15% of the lithology but is usually very much less. Insoluble residue analysis of the lithofacies gives percentages as low as 1.6%.

Beds of similar bioclastic packstone in the exposure at Carrickbuoy (P.31) are approximately 10cm. thick.

The contact of the packstones of lithofacies b) to the wackestones of lithofacies c) is usually very sharp (P.32). Occasionally there is a
degree of interfingering of the two lithologies before the wackestones of lithofacies c) commence to dominate the succession.

Lithofacies c) Bioclastic wackestones.

These wackestones contain variable amounts of bioclastic debris within abundant micrite. The debris is both poorly sorted and rounded and may occasionally reach 3x2mm in size. The debris sometimes occurs in pockets and then has a packstone texture. The debris consists of bryozoa, ostracods, bivalves, foraminifera, sponge fragments, rare corals and rare algal fragments. Calcispheres are also present. The debris is occasionally micritised. In much of the lithology bryozoan debris forms a major part of the bioclasts which frequently constitute 20% of the lithology. The micrite matrix is often very rich in peloids which due to their "nested" distribution, similar general size and similar colour to the matrix are considered to be of faecal origin.

Carbonate and insoluble residue analysis reveals the consistently pure nature of the lithology. The average insoluble residue of 28 samples is 18% (s.d. = 11.2%) and the average carbonate is 74.4% (s.d. = 17.7%). Thin section analysis reveals that the majority of the insoluble residue is mud with small amounts of silt sized quartz.

Laminar porosity is common in the lithofacies and can (rarely) form up to 50% of the rock. It may be up to 5cm. long and 2cm. high. Geopetal micrite is present with a later radial fibrous calcite cement. These vugs are interpreted as stromotactids (P.35).

The lithofacies is often extensively dolomitised. The ferroan dolomite crystals are very coarsely crystalline and show an undulate extinction.
C. Interpretation.

Lithofacies a). Thin bedded fossiliferous calcareous mudstone.

By utilising the rare bioclastic wackestones in the lithofacies it is possible to place lithofacies a) in one of the standard microfacies of Wilson (1975). The diversity of the organisms and their homogenity indicates that the lithofacies be placed in standard microfacies 9. An origin in shallow neritic water at or just below wave base is implied. The occasionally abraded nature of the bioclasts infers that the wave base may have been periodically lowered by storms. Storm wave base may be as deep as 200M. (Ewing 1973). If we infer that fair weather wave base was 5-15M. (Walker 1984) a depth range of 5-200M. is probable. This is a similar depth to that inferred for the Bioclastic Unit. In fact the fossiliferous calcareous mudstones of lithofacies a) are considered to be the lateral equivalents of the Bioclastic Limestone Unit differing only in that lithofacies a) was deposited under conditions which were also suitable for the formation of the wackestones of lithofacies c).

The factors which may have influenced this were:

1) The organic skin, the origin of which is discussed below and which is postulated to have been very important in the formation of the wackestones of lithofacies c), may not have been able to stand up to the relatively higher energy conditions in the shallower water in which the Bioclastic Limestone was deposited. Although below fair weather wave base anything growing on the bottom would be more susceptible to storm damage since storm wave base would extend to the bottom more frequently.

2) There may have been predation or competition from other organisms upon the organisms which produced the micrite for later
incorporation into the wackestones of lithofacies c). Lees and Miller (1985) have postulated an upper limit of 120M. depth for the formation of the mounds. This is near the theoretical maximum depth for the photosynthesis of blue-green algae in clear water (Brock 1976). Possibly competition from blue-green algae eliminated the micrite producing organism from the Bioclastic Limestone environment.

3) The more turbid water of the shallower Bioclastic Limestone environment adversely effected the growth of the micrite producing organism.

Lithofacies b) Bioclastic packstones.

These packstones were formed under conditions of water movement sufficient to impart a grain supported fabric to the sediment and are thus largely free of the mud found so abundantly in the mudstones of lithofacies a). Lithofacies b) is assigned to a special case of standard microfacies 12 of Wilson (1975). In this special case less winnowing of the sediment is required than in the case of the high energy coquinas which are normally found in standard microfacies 12.

Lithofacies c) Bioclastic wackestones.

Observations at Carrickbuoy Quarry (P.33) reveal that as in other parts of Ireland (Lees 1964) the wackestones of lithofacies c) form the actual mound which is surrounded by the packstones of lithofacies b) and the calcareous mudstones of lithofacies a).

Many authors (e.g. Wilson 1975) have pondered upon the origins of the Waulsortian Mudmounds. It may be as well to reiterate some points of importance pertaining to their depositional environment.

1) The lack of terrigenous mud in the mounds as compared to the
abundant amounts in lithofacies a) indicates that the mounds had a depositional relief on the sea floor i.e. they were above the mud which was accumulating in the area close to the sea bed and where the equivalent lithofacies a) mudstones were forming.

2) The clean packstones of lithofacies b) have obviously been subjected to current activity. However they lie immediately adjacent to the highly micritic wackestones of lithofacies c). It is highly unlikely that fine grained material such as micrite could accumulate whilst being subject to current activity unless some baffling/encrusting mechanism were present to prevent removal of the micrite.

3) The micrite was produced in the very near vicinity of the mounds. Little micrite is normally present in the low energy calcareous mudstones of lithofacies a) where significant quantities would be found if micrite production was general and not confined to the specific mound sites.

4) The attitude of the stromotactids indicate that mound slopes of 40 degrees were common (P.35). However slumping of the mound micrite is not common. This would appear to indicate that some type of binding/encrusting mechanism preventing failure of the slope was present.

The following environment is proposed for the deposition of the mound complexes at Keel. The environment is summarised in Fig.60.

The mounds established themselves on a basal bioclastic pile. This appears to be a ubiquitous feature of mudmounds world wide and throughout geological time (Wilson 1975). This basal pile would be the upper part of the Bioclastic Limestone Unit. Where lenses of mound
Figure 60. Postulated mechanism for the deposition of the Waulsortian mud mound complexes.

1) Colonisation stage.

Filamentous bacteria colonise a wackestone pile away from the influence of silt and currents. Micrite is precipitated by the bacteria from the water and is trapped by the bacteria.

2) Establishment stage.

The bacterial coating protects the mound from erosion. Eventually the mound reaches such a height as to be totally unaffected by the silt which only moves near the bottom. Currents flowing near the mound are speeded up by the Bernoulli Effect.

3) Terminal stage.

Changed conditions. The old mound is overwhelmed and a new mound is initiated.
material appear in the Bioclastic Limestone these possibly mark abortive attempts of the mounds to establish themselves. These may represent the sheet forms of Lees (1964) extending outwards from already established knoll forms (Lees opp.cit.) that lay to basinward. The lithofacies c) wackestone lenses within the mudstones of lithofacies a) may represent similar attempts at colonisation. The mounds tended to be restricted to particular areas in the Irish Midlands as at Keel. Read (1975) considers that isolated mounds develop where the rate of sea level rise is rapid. However in the case of Keel the author considers it more likely that fine grained terrigenous input from nearby land masses was responsible. The nearest land mass at this time was approximately 60KM. away from Keel (Sevastopolou 1982). It is thought that growth of the mounds was initiated on local palaeohighs, consisting of bioclastic debris, probably only a metre or so above the general sea floor. This height was however enough to keep the juvenile mound out of the mud and occasional silt which was being transported along the bottom. This fine grained sediment may have prevented the growth of the micrite producing organism. This may possibly have been some type of microbe as postulated by Lees and Miller (1985). The microbe is thought, by the author, to have possibly been some type of bacteria which produced filaments. Bacteria that produce filaments 40-50 microns long are relatively common today. However no bacteria posseses a calcareous component in its physiology so the production of micrite directly from the bacteria after it died is ruled out. Rather it is thought that the micrite was produced by precipitation from the carbonate saturated water when the bacteria respired removing carbon dioxide from the water. Bacteria utilising
carbon dioxide for respiration (anaerobic respirers) are known today e.g. *Clostridium aceticum*.

Reaction.

\[ 4 \text{H}_2 + 2 \text{CO}_2 = \text{CH}_3\text{COOH} + 2 \text{H}_2\text{O} \]

The micrite produced was then trapped by the bacteria, which probably formed a relatively tough coating on the mound, either by adhering to the sticky mucus coat commonly found on bacteria and/or by trapping in the bacterial filaments. Since the bacteria themselves were composed of proteins, lipids etc. with no parts capable of preservation in the fossil record no trace would be found of their former presence except for the micrite. The decay of the bacteria would however lead to the precipitation of more micrite since the products of the decay e.g. ammonia would raise the pH of the water even more.

The theory that the micrite was produced and/or trapped by calcareous algae is considered unlikely since Lees and Miller (1985) have advanced well reasoned arguments that deposition of the mounds took place initially in water as deep as 300M. This is well below the photic zone to which photosynthetic algae would be confined. In addition few traces of algae have been found in Waulsortian Mounds except those deposited in their very shallowest depths i.e. Phase D at above 120M. (Lees and Miller opp.cit.).

The ideas of King (1986) that the crinoids and bryozoa of the mounds trapped micrite from carbonate turbidites is considered very unlikely at Keel since:

1) There do not appear to be enough crinoids and bryozoa present to
trap the micrite in such large quantities.

2) On the postulated carbonate ramp setting turbidites would be very infrequent at best, and

3) Micrite would also be trapped by the abundant crinoids in the mudstones of lithofacies a) and the packstones of lithofacies b). As mentioned above in general very little micrite is present.

Water currents would tend to be deflected around the growing mound. This deflection of currents would tend to speed up the current in a variation of the Bernoulli Effect. The nearer the mound the faster would flow the water (Fig.60). These currents would not have flowed at all times since mud is occasionally found in thin lenses and shelter porosity of lithofacies b). Rather it is thought that the currents were probably a response to storms and actually brought terrigenous sediment with them. The result of these speeded up currents would be to winnow away mud from the immediate vicinity of the mound. The process would be most marked near the mound and the effect would decrease away from it. This would account for the progressive increase in mud away from the mound culminating in the calcareous mudstones of lithofacies a). The mound was protected from erosion by the binding activities of the filamentous bacteria forming a coating over the mound.

Mud being deposited away from the mound would prevent the establishment of the bacteria possibly by covering the bacteria and thus barring it from its sources of nutrients and carbon dioxide i.e. the sea water. With regards to nutrients the increased currents near the mound may have increased the productivity of the crinoids by imparting greater movement to the crinoid arms through the water and by bringing greater
amounts of nutrients to the site. Thus it is probable that thicker stands of better developed crinoids occurred near the mound giving the well washed packstones of lithofacies b). These packstones have been termed flank beds by Miller (1986).

The relative scarcity of crinoids in the mound (lithofacies c) packstones is possibly the result of the crinoids not being able to root in the relatively tough bacterial coating. Those cases where fossil debris is concentrated into lenses may mark occasions when storms lowered the wave base and rupture of the bacterial coating took place. Crinoids, bryozoa etc. might quickly colonise the rupture before renewed bacterial growth could seal the breach. As the crinoids died the resulting debris would form lenses. Eventually the bacteria would succeed in closing the breach.

Being comparatively near to the coastline and therefore to sources of terrigenous sediment, which could halt mound growth, cases of mounds being "killed" are seen at Keel. Similar restricted mound growth is seen in other areas adjacent to emergent land masses in Ireland at this time e.g. Leinster Massif and the Galway area (Wilson 1975). However further to the south such sources did not exist and the actual mounds (lithofacies c wackestones) form practically the entire sequence of the Waulsortian due to separate mounds aggregating.

The micrite producing bacteria might be killed off by being overwhelmed by mud. This might be brought about by:

1) Increased current strengths or increased erosion rates on land might make more sediment available. This might cause the "off-mound beds" i.e. lithofacies a) to be built up so as to physically submerge
the mound as the rate of accumulation of the off mound beds exceeded that of the mound.

2) A change in current direction might cause a cessation in the supply of nutrients and so kill the bacteria.

3) Competition factors as outlined in Chap. 15.3

Of course new mounds could be established on the site of an old one or else-where should factors prove favourable.

Lees and Miller (1985) suggest that Waulsortian Mounds formed in water depths ranging from over 300M. to less than 120M.. These depth variations were not detectable in the restricted numbers of Keel samples examined in thin section by the author. However it was noticeable that micritisation of the bioclastic debris was restricted to the upper (i.e. shallowest water) parts of the core. The ooids found in the Bioclastic Limestone Unit may have been derived from Phase D of Lees and Miller (opp.cit.) which at less than 120M. is the shallowest depth at which the mounds formed.

The process whereby the Waulsortian Mounds became less well developed in a northerly direction in Ireland reaches its logical conclusion in the area of the Northern Inlier. Here the mounds are replaced by a mid grey to black vuggy calcitic mudstone. Solitary rugose corals and crinoids are common. These are the Mound (Reef) Equivalent Beds of Slowey (1986). It must be assumed that there was so much sediment in the water or the water so shallow that there was no opportunity for the mounds to develop.

D.Local thickness variations.

Inspection of Fig. 61 shows a definite thickening of the Waulsortian
FIG. 61. ISOPACHYTES OF THE WAULSORTIAN MUDMOUND COMPLEX SHOWING THICKENING TO THE SOUTH-EAST OF KEEL, THICKNESSES IN METRES.

K - Keel

THE SOUTH-EAST OF KEEL.

Fig. 61. Isopachytes of the Waulsortian Mudmound Complex showing thickening to
Mound in a southerly direction away from the Keel Inlier. A correlation co-efficient of 0.67 (6.9% probability of no correlation) confirms this. This is thought to reflect the subsidence of the basin postulated to have been initiated during deposition of the Bioclastic Limestone Unit. Owing to the localised character of the thicker sequences a rise in sea level is not considered to have been responsible. The small sub-basin to the south-east still continues to influence sedimentation. Thicker sequences were laid down as limestone production increased in response to the subsidence. Similar rapid thickness changes in the Waulsortian at Tatestown, 3.5k north-west of Navan (Fig. 1) have been attributed to fault movement during deposition (Andrew and Poustie 1986). In addition much larger scale thickness variations are seen in the Limerick area (Lees 1961).

The more local thickening to the south-east of Keel indicated on Fig. 61 may mark the site of a major mound as opposed to the more general thickening on an east-west line along the inferred fault which possibly marks the sites of minor mounds and attempts of the major mound to colonise the surrounding area. The fact of there being a high proportion of Lithofacies c) calcareous mudstones in the area indicated to the west of the mound may confirm the "off-bank" nature of these sediments. This major mound would have had dimensions of 1500 x 1000 feet. This compares with diameters of over 1000 feet reported by Lees (1964). It is of course perfectly conceivable that this local thickening is the result of the amalgamation of two or more mounds growing in a favoured site. In Fig. 61 the inter-reef (mound) beds are the calcareous mudstones and bioclastic packstones of lithofacies a) and b).
E. Porosity in the wackestones of lithofacies c).

The porosity has been interpreted as being of a stromotactid variety. It was probably formed by the bacterial filaments creating shelter porosity which the micrite and later radial fibrous calcite cement preserved. Similar cements are known to be of a marine origin (Bathurst 1982). This early cementation also assisted to preserve the mound from erosion. The irregular shape of the "roofs" of the cavities (Fig.34) probably arose due to the collapse of semi-lithified sediment into the cavity as suggested by Wilson (1975).
CHAPTER 16
CHAPTER 16.

Post Waulsortian Mound beds.

A. Introduction.

In the Keel area the Waulsortian Mounds are conformably overlain by one of two lithologies, the Calp Limestone or the Oakport Limestone.

Both lithologies are of a Chadian age (George et al. 1976) as compared to the underlying Mound complex which is the last lithology of a Courceyan age in the area.

The Calp and Oakport Limestones have very different petrologies and so will be treated separately. However since their depositional environments are thought to be intimately linked their interpretation is dealt with on a joint basis.

B. Lithology.

1) Calp.

The Calp classically consists of a highly argillaceous lime mudstone. However on both chemical analysis and thin section grounds much of the Calp might better be classified as a calcareous mudstone. Insoluble residue analyses average 62.8% whilst carbonate analyses average 19.2%. Carbonate may be as low as 4% in some cases. The lithology is dark grey to black in colour with rare bioturbation. Silt sized quartz is occasionally moderately common. Fossils are fairly common in the mudstones and include brachiopods, bivalves, sponge spicules, ostracods and crinoidal debris. Calcispheres are also present. The bioclastic debris is very occasionally micritised. Much of the debris and fossils appears worn and of a generally smaller size when compared to that from the Mound and Bioclastic Limestone Units. Sorting
is moderate to poor. Occasionally the debris is silicified and chert nodules are common. High angle wavy bedding, interpreted as slump bedding, is occasionally abundant.

Interbedded with the calcareous mudstones, and locally of major prominence, are beds of bioclastic grainstones, bioclastic packstone and highly bioclastic calcareous mudstones. The bioclasts are predominantly of a crinoidal origin and are abraded and worn. The beds show a tendency for erosional basal contacts and gradational top contacts with the "normal" background mudstones. At the base they show a grainstone texture grading up into packstones and highly bioclastic mudstones. The beds are commonly normally graded and are of a variable thickness ranging from 1 cm. to, exceptionally, 1.5 m. per cycle. In the thinner beds a pronounced lamination is present. The laminae are from 2-4 mm. thick and appear to be composed of alternations of dark mud sized material and lighter silt sized material. Occasionally vertical thin (>0.1 mm.) non-lined burrows are visible in the fossiliferous mudstones on their lower contact to the "normal" mudstones. These burrows tend to end at the contact or to penetrate only a short distance into the underlying "normal" mudstone.

The above beds show a tendency to increase in both thickness and frequency with proximity to the top of the Calp facies. In addition these bioclastic beds appear very well developed around the Northern Inlier where dolomitisation is also prevalent. The (ankeritic) dolomite crystals may range from microcrystalline to very coarsely crystalline in size.

Much of the mud and silt sized terrigenous material in the Calp
mudstones was probably derived from land masses to the north. Such material has been reported from much deeper waters in the present day worlds oceans (Scholle 1971).

Very rarely small (4 cm.) pebbles of red sandstone are found in the "normal" mudstones, but not associated with the highly bioclastic beds. The nearest terrigenous source for these relatively large and dense objects lay approximately 100KM. to the north-east (Sevastopolou 1981). Consequently it is thought that they may have been rafted in on the roots of trees (although drop-stone textures were not observed) and are thus not considered to be of any depositional significance.

2) Oakport.

The Oakport Limestone is only present in the extreme west of the Keel area. It is a mid grey, medium grained peloidal grainstone. The peloids may be micritised ooids or be of a faecal origin. The lithology contains interbeds of darker coloured dolomite which may reflect better porosity along these bands. Very occasionally shale bands are present together with bands of fossil debris which show wavy erosional bases and normal grading. Sorting of the debris is good and apart from the crinoidal debris, which constitutes most of it, is micritised. Syntaxial overgrowths are present on some of the crinoidal debris. In contrast to the Calp lithology chemical analysis reveals that the Oakport is a very much purer lithology with an average insoluble residue of 21.5% and average carbonate of 82.6%.

C. Interpretation.

It is considered that at the end of Waulsortian times the mounds had grown up into the photic zone as the rate of carbonate accumulation
exceeded the rate of sea level rise. The sediments of the Bioclastic Limestone and Mound Units had by now succeeded in constructing an extensive carbonate ramp. At this stage competition, possibly by blue-green algae as outlined in Chap. 15.3, prevented growth of the micrite producing bacteria and thus growth of the mounds. Cessation in the rate of sea level rise made further upward growth of the mounds impossible since they could not grow into waters of less than 120M. A return to conditions of Bioclastic Limestone deposition now took place and bioclastic wackestones and calcareous mudstones and shales began to be deposited on top of the mounds. It is thought probable that this was a combination of the calcareous shales of lithofacies c) of the mound complex covering the mounds from the sides and Bioclastic Limestone types moving basinward over the mounds. Shallow ramp facies prograding over deeper ramp lithologies are known from the Ordovician of Virginia (Read 1980).

It was at this point that sedimentation of the Calp and Oakport commenced. Their mutually exclusive nature, differing depositional environments and sudden appearance in the sequence, particularly in the case of the Calp, indicate that substantial and rapid subsidence commenced over part of the area.

The Calp mudstones are considered to be of a fairly deep water origin. The black colour of the mudstones may imply that the bottom was somewhat anoxic and that carbon was building up in the sediment. The rarity of bioturbation may confirm this. The presence of fossils composed of carbonate infers deposition above the Carbonate Compensation Depth of 4-7000M. (Blatt et.al. 1980). Since the bivalve fragments
appear essentially unpitted deposition above the Aragonite Compensation Depth of 500M. (Blatt et al. opp.cit.) is probable. Although it is possible that the fossil debris in the thicker beds was protected from dissolution after rapid deposition as turbidites (see below) dissolution effects are not seen in the thin turbidites or on the sparse bioclasts in the "normal" mudstones. Therefore any idea that rapid deposition protected the bioclasts from dissolution is considered unlikely.

The interbedded grainstones, packstones and highly fossiliferous mudstones are interpreted as turbidites derived from the areas still remaining as shelves to the north and west of Keel. The worn nature of the bioclasts, the graded bedding (Scholle 1971) and the erosive basal contacts suggest this. Indistinct top contacts have been reported for carbonate turbidites (Scholle opp.cit.). Both the thicker and thinner beds are thought to be the distal type of turbidite described by Walker (1967). They show regular bedding, few scours or channels and the mudstone layers are well developed between the turbidites with little mixing. The beds are also well graded with sharp bases while the laminated tops grade into the overlying mudstone and. The thickness of some of the beds might indicate a proximal origin. However beds as thick as 27M. have been interpreted as distal turbidites (Scholle 1971).

The packstones may be assigned to the Standard Microfacies 4 of Wilson (1975) indicating deposition on slopes. The deposition of these lithologies (Allodapic Limestones of Meischner 1964) is characteristic of basin edges (Yarewicz 1977). Since talus etc. is not found, the slope may be comparable to the 2-3 degrees slope postulated for the deposition of the Mississippian Rancheria Formation (Yarewicz opp.cit.) where
similar type turbidites occur. Certainly the slopes would not have been as steep as those (up to 40 degrees) proposed by Davies (1977) for the deep water carbonates of the Canadian Arctic where abundant debris sheets were deposited.

The burrows found in the thinner turbidites appear similar to those found in the carbonate turbidites of Italy by Scholle (1971). They may represent the response of some animal to freshly deposited turbidites as it burrowed in looking for food. Seilacher (1962) considers such burrows to be probable evidence for turbidity current origin. The cessation of the burrows at or just below the grainstone/"normal" mudstone interface may indicate the boundary of aerobic and anaerobic conditions.

The very high proportion of crinoidal debris in the turbidite beds is probably due to their high internal skeletal porosity and consequent low bulk density (Maiklen 1968). This would render them more likely to be entrained by the turbidity current than denser bioclasts e.g. brachiopods.

The Oakport, due to its micritised clasts and lack of lime mud, is thought to have accumulated in areas of constant wave action at or above wave base at the platform edge. It is assigned to the Standard microfacies 11 of Wilson (1975) with deposition in shallow water of less than 10M. depth. Any fine grained sediment would have been winnowed away by waves or longshore drift. The fossil debris would have been largely derived from outside the area since the constantly shifting lime sand and the high energy conditions would not have constituted a very suitable environment. The shale beds within the lithology indicate that there were occasional periods of low energy. It may therefore be that
the environment was subject to extended periods of storms during which the debris was rounded and sorted. The erosional bases and normal grading of the fossil bands may confirm that periodic storm activity took place. During intervals of calm weather deposition of the shales took place. What little indigenous fauna is present probably established itself during these periods of calm weather.

Deposition of the Oakport is thought to be a consequence of the progressive shallowing of the carbonate ramp alluded to above (Sect.3) part of which had collapsed to form the Calp basin. The Oakport Limestone is thus interpreted as forming on the shelf which surrounded the basin. Similar relationships are seen in the Upper Cambrian and Lower Ordovician of Nevada (Cook and Taylor 1977) and the Cambrian of Utah (Brady and Rowell 1976).

The first manifestations in the Keel area of this "Calp" Basin may be found in Bioclastic Limestone times (Chap.14). Nolan (1987) has dated the differentiation of the Dublin Basin into blocks and basins to the Late Courceyian/Early Chadian. Rees (1987) has recognised subsidence in the Boyne Valley area also during the Early Chadian.

Due to the localised character of the subsidence producing blocks and basins it seems extremely unlikely that the cause of the relative rise in sea level was eustatic, the primary mechanism must have been tectonic.

This localised subsidence was probably caused by the recrudescence of ancient fault lines. This was attributable to either:

a) Simple subsidence or

b) Tension caused by incipient back arc spreading processes similar
to those experienced by the Devonian of South-West England.

The north-westerly trend of the basin (Fig. 62) suggests that subsidence probably took place in a graben-like fashion between pre-existing Caledonian trend faults. Movement is known to have been taking place on the North Solway Fault (which continues on into Ireland to form a southern margin of the Longford-Down Block) throughout the Dinantian (Anderton et al. 1979). Brown and Williams (1985) have postulated large north-easterly trending antiforms with volcanic cores and zones of weakness on their margins. One of these lines of weakness appears to occur in the Keel area where the Main Keel Fault may be involved and another line of weakness just to the north of Harberton Bridge (Figs. 9 & 63). It is these two lines of weakness which are thought to be involved in the creation of the "Calp" Basin.

Movement along the fault in the Keel area was probably the "trigger" for the deposition of the turbidites into the basin.

Fig. 62 illustrates that the dimensions of the basin may be defined by plotting out the occurrences of the Oakport (shelf) Limestone and the Calp (basinal) mudstones. The basin margin may be plotted with confidence in the south of County Dublin and in the Crossakeel, Oldcastle and Navan areas since boulder beds occur here presumably associated with movement of the margin. At Keel the margin runs between the Calp beds lying around Keel itself and the western end of the Keel Inlier where Oakport Limestone is found. The occurrence of numerous Calp turbidites around the Northern Inlier probably indicates that the margin lay just to the north of that inlier. The basin margins to the south and south-west are not so readily definable but it is felt that those
indicated may be accepted with some degree of confidence.
FIG. 62. DIMENSIONS OF THE CULP BASIN.

Boundary of the Culp Basin.

Boulder Conglomerate
Oakport Limestone
Culp Basin Shales

South Dublin
Raberton Bridge
Ballycastle
Emoore
Winmill Hill
Sieveorris
Lough
Mount Bellew
Stranorlan
Common

Newtown Cappel
Ballyfinnack
Barando
Doolin
Rossakeel
Lough Sheelin
Navan

Fig. 62. Dimensions of the Culp Basin.
Fig. 63. Residual gravity profile constructed after processing simple Bouger anomaly data. (After Brown and Williams 1985).
CHAPTER 17
CHAPTER 17.

Conclusions.

The Silurian rocks of the area were deposited as deep marine muds in the fore-arc basin area of an active subduction zone on the Northern side of the closing Iapetus Ocean. Coarse sandstones and silts were emplaced as turbidites into the basin. These were probably triggered by shock waves created by the subducting oceanic plate.

The Iapetus Ocean finally closed in Wenlock to early Devonian times and extensive highlands were created over the north of Ireland.

After a period of net erosion, deposition in the Keel area commenced with the Upper Devonian Microconglomerate. Evidence would suggest that this unit was deposited by small braided streams mainly draining the near vicinity and some localities immediately to the north and/or west (Fig.64).

FIG. 64. INFERRED DEPOSITIONAL ENVIRONMENT OF THE MICROCONGLOMERATE

Local mountainous areas

Possible alluvial fans

SILURIAN
The succeeding Lower Quartz Sandstone was deposited by sandy braided streams with more water available in the north of Ireland due to the northward moving transgression. Consequently sediment could be brought from the Dalradian Highlands of North-West Ireland and from granite terrains to the west of Keel (Fig. 65).

**Fig. 65, Inferred Depositional Environment of the Lower Quartz Sandstone.**

Overlying the Lower Quartz Sandstone is the Quartz Pebble Conglomerate. This unit is thought to be a response to movement along faults in the Dalradian Highland source areas. This movement increased gradients and made coarser sediment available for transport by the braided streams draining the area. These streams assumed the character of pebbly braided streams with this change in sediment load (Fig. 66).
Eventually the uplifted hinterland was eroded down. Due to the reduced stream power pebbles could no longer be carried. Only sand and smaller sized sediment was carried and ultimately deposited in the Keel area by streams of a sandy braided type. These were the deposits which became the Upper Quartz Sandstone (Fig. 67).
The marine transgression had by this time advanced into the vicinity of Keel. The Lower Mixed Beds were laid down as a series of marginal marine deposits in the Keel area (Fig. 68). The northward advance of the sea had effectively lowered the gradient between the mountain source to the north-west and Keel. This, together with the erosion of these mountains, only permitted the rivers draining them to remove fine sediments. Some cleaner sandstones suggest times of greater discharge. Thus the sediments of the Lower Mixed Beds are generally finer grained than those of the underlying post-Silurian lithologies.

As to whether a barrier island lay to seaward at this time is unclear. However from the evidence of lime mudstone interbeds within the Lower Mixed Beds it would appear likely that lagoonal limestones of a Navan Micrite type were being laid down to seaward so the presence of a barrier may be inferred.
With further marine transgression a lagoon was established behind a barrier island. Within the lagoon were deposited the lime muds and channel deposits which became the Navan Micrite (Fig. 69). The back barrier, washover, inlets and barrier itself gave rise to the sediments of the Upper Mixed Beds.

Fig. 69. Inferred depositional environment of the Navan Micrite.

The main stage of the marine transgression then advanced over the
barrier island. All traces of the barrier were removed apart from the shoreface deposits and back barrier. As the barrier was planed off its sediments were redistributed to seaward and landward. This left a thin sandstone sheet over the areas between the sites of still stand of the barrier (Fig.70).

**Fig. 70. Upper Mixed Beds, Transgressive Sandstone and Traces of Former Barrier Islands.**

Sediments which had been deposited to seaward of the barrier prograded over the barrier and advanced north under the influence of the transgressing sea.

The coastline and with it the sources of terrigenous sediment were now far to the north of Keel. Very little terrigenous sediment was deposited on the carbonate ramp which was now forming. Abundant crinoids

289
were growing on this ramp and it was the remains of these crinoids and other organisms, along with what terrigenous sediment was available, which produced the bioclastic wackestones, packstones and calcareous mudstones of the Bioclastic Limestone Unit. Under the influence of the transgression these lithologies were deposited upon the Shaley Pales as the facies belts moved northwards (Fig. 71).

**Fig. 71. Depositional environment of the Shaly Pales and of the Bioclastic Limestone Unit.**

Further to seaward on the ramp Waulsortian Mudmound complexes were being deposited in water depths down to 300M. (Fig. 72). The Mound complexes eventually transgressed north over the Bioclastic Limestone lithologies. Immediately north of the Keel area deposition of the Mound complexes ceased, possibly due to shallowing water and/or terrigenous sediment influence. Keel thus probably marks the northern limit of the Waulsortian Mound lithology in Ireland.
By the close of Waulsortian times the Mudmound complexes and the Bioclastic Limestone Unit had succeeded in constructing a carbonate ramp (Ahr 1973) extending from just north of Keel to the Cork-Kenmare line (Fig. 73). The Waulsortian mound lithology ends abruptly at this line which is thought to be fault controlled (Phillips and Sevastopolou 1986).

The ramp is not thought to have been of a distally steepened type (Read 1975) as toe of the slope talus material is absent in the mudstones of the now sediment starved South Munster Basin (Phillips and Sevastopolou 1986). Although carbonate turbidites have been described from there (Sevastopolou 1982) the high energy belt found to landward of the distally steepened portions of such ramps (Read 1975) is apparently absent. However Wilson (1975) makes the point that such high energy belts may be only a few miles wide as compared to the broad facies belts.
on the shelf. Such a narrow belt may not have been detected in the area of the Cork-Kenmare line.

Similar lines of reasoning rule out the presence of a carbonate platform.

Thus the ramp is interpreted as of a homoclinal type and is thought to be similar to the ramp with barrier bank complexes model of Read (1975).

Facies belts on these type ramps with their Keel representatives are:

<table>
<thead>
<tr>
<th>Facies belt of Read (1975)</th>
<th>Keel representative</th>
</tr>
</thead>
<tbody>
<tr>
<td>1) Tidal/supratidal complex</td>
<td>Lower Mixed Beds</td>
</tr>
<tr>
<td>2) Lagoonal carbonates</td>
<td>Navan Micrite</td>
</tr>
<tr>
<td>3) Barrier bank complex</td>
<td>Upper Mixed Beds</td>
</tr>
<tr>
<td>4) Deep ramp carbonates</td>
<td>Bioclastic Limestone Unit</td>
</tr>
<tr>
<td>5) Deep ramp buildups</td>
<td>Waulsortian Mudmound complexes</td>
</tr>
</tbody>
</table>

Examples of this model in the geological past include the Holocene of Shark Bay (Hagan and Logan 1974) and the Devonian Helderberg (Laporte 1969).

After the deposition of the Waulsortian Mudmound complexes parts of
Figure 73. Extent of the Waulsortian Mudmounds in Ireland.

- Waulsortian Mudmound Lithology
- Limits of Mudmound Lithology
- K: Keel
- CKL: Cork-Kenmare Line
- SMB: South Munster Basin
the ramp collapsed in response to tectonic forces. Basinal marine shales (Calp) were deposited together with calcareous turbidites from the surrounding shelf areas. On these shelf areas shallow water carbonates (Oakport Limestone) continued to be deposited (Fig 74).

**FIG. 74. INFERRED DEPOSITIONAL ENVIRONMENT OF THE CALP AND THE OAKPORT LIMESTONE.**

So in conclusion it would appear that the Courceyan/Chadian succession around the Keel Inlier was deposited on:

a) A fluvial/intertidal margin to the north of the Upper Devonian Sea and

b) later, a storm influenced homoclinal ramp which grew under the influence of a northward transgression and became structurally differentiated during the Early Chadian.

Figure 75 summerises the facies mosaic of the Carboniferous rocks of the Keel area.
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