1971

The Mogul Base Metal Deposits County Tipperary, Ireland

Robert Alexander Graham

Follow this and additional works at: https://ir.lib.uwo.ca/digitizedtheses

Recommended Citation
https://ir.lib.uwo.ca/digitizedtheses/457

This Dissertation is brought to you for free and open access by the Digitized Special Collections at Scholarship@Western. It has been accepted for inclusion in Digitized Theses by an authorized administrator of Scholarship@Western. For more information, please contact tadam@uwo.ca, wlswadmin@uwo.ca.
THE MOGUL BASE METAL DEPOSITS
COUNTY TIPPERARY, IRELAND

by
Robert Alexander Fergus Graham
Faculty of Science

Submitted in partial fulfillment
of the requirements for the degree of
Doctor of Philosophy

Faculty of Graduate Studies
The University of Western Ontario
London, Canada
June 1970

© Robert Alexander Fergus Graham 1970
ABSTRACT

The Mogul lead-zinc-silver deposits occur in the long-known mining district of Silvermines, Co. Tipperary, in the Irish Republic. They were discovered in 1963, brought into production in 1968 and, with reserves of about 14,000,000 short tons, have a minimum expected life of 14 years.

The deposits occur in relatively undeformed Lower Carboniferous carbonate rocks formed in a broad, shallow-water shelf environment between a rapidly subsiding trough to the south and a shoreline to the north. They are adjacent to a fault-bounded inlier of highly folded, argillaceous rocks of Silurian age, lying to the south.

The main deposit, the Upper orebody, is an oval-shaped, stratiform, massive sulphide body which has a high pyrite content and is relatively rich in zinc. It is underlain by a smaller, lead-rich deposit, the Lower orebody, consisting of galena-sphalerite veins and disseminations. Adjacent to the Upper orebody, on the same stratigraphic level but higher structurally, is a stratiform, massive barite lens (Magcobar).

The Upper orebody occurs at the base of a thick, brecciated and dolomitized succession. The occurrence within the breccia of stratification planes, thin interbedded shales, graded bedding, and double brecciation indicate its
sedimentary nature and formation by processes of resedi-
mentation. Similar features in the sulphide body show that
it has undergone the same processes and thus forms part of
the sedimentary sequence.

The mineralogy of the orebodies is relatively simple:
dominantly pyrite, sphalerite and galena. Textures in the
Upper orebody are fine-grained and intricate and indicate
relatively rapid precipitation. A wide range of colloform
types is developed and microscopic examination of these
reveals that they form by precipitation from solution rather
than by crystallization of gel globules. Textures in the
Lower orebody are simpler and coarser grained, particularly
at greater depths, and indicate relatively slow crystal
growth under more uniform conditions than in the Upper ore-
body. Such conditions are further indicated by a smaller
range in hardness (and thus composition) of galena, and a
more uniform distribution of iron in sphalerite.

Framboids are common in the Upper orebody and host-rocks.
They represent a distinct, early type of pyrite and have
well defined geometrical features which are partly related
to enclosing rock-type. They may have formed by pyrite
crystallization at points within minute globules of FeS
gel. Later colloform pyrite commonly nucleates on fram-
boisd.

Cobalt to nickel ratios in pyrite in the Upper orebody
are less than one, which is typical of sedimentary pyrite in
areas lacking volcanic associations, and suggest a sedi-
mentary origin for both elements. Much of the iron in pyrite
may also be derived from the sedimentary environment.

Sulphur isotope ratios in barite from the Magoobar deposit are close to those for Lower Carboniferous seawater whereas those from sulphides of the Upper orebody show extreme enrichment in the lighter isotope, indicative of prolific bacterial activity. Sulphides in the Lower orebody show less pronounced fractionation.

It is concluded that the orebodies formed from thermal, connate water which leached lead, zinc and barium from Lower Palaeozoic argillaceous rocks and was channeled upwards along the Silvermines Fault. Sphalerite and galena were deposited when the water encountered sulphur-bearing seawater in fractures near the surface. Submarine exhalations of the water into a restricted basin with anaerobic bottom conditions caused further rapid precipitation of sphalerite and galena in association with pyrite. Part of the precipitate took the form of a cryptocrystalline sulphide mud on the sea-bed and instability, probably related to fault movement, caused repeated sliding of the mud and brecciation of consolidated structures to form a layer of assorted sulphide fragments enclosed in a sulphide matrix. Barium migrated outwards from the basin and was precipitated as barite on encountering sulphate ions at a higher level.
ACKNOWLEDGEMENTS

The writer wishes to convey sincere thanks to Dr. G.G. Suffel for suggesting the topic, supervising the research and critically reviewing the manuscript. Thanks are also due to Dr. R.W. Hutchinson and Dr. W.R. Church for much discussion and advice.

The writer also wishes to thank the Management of International Mogul Mines Ltd., and in particular Dr. W.W. Weber and Mr. P.S. Cross, whose close cooperation enabled the project to be undertaken.

Sulphur isotope work was arranged by the kind cooperation of Dr. D.F. Sangster.

Financial assistance from the Ontario Government and the National Research Council is gratefully acknowledged.
# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Certificate of examination</td>
<td>11</td>
</tr>
<tr>
<td>ABSTRACT</td>
<td>iii</td>
</tr>
<tr>
<td>ACKNOWLEDGEMENTS</td>
<td>vi</td>
</tr>
<tr>
<td>TABLE OF CONTENTS</td>
<td>vii</td>
</tr>
<tr>
<td>LIST OF TABLES</td>
<td>xii</td>
</tr>
<tr>
<td>LIST OF ILLUSTRATIONS</td>
<td>xiii</td>
</tr>
<tr>
<td>CHAPTER 1 - INTRODUCTION</td>
<td>1</td>
</tr>
<tr>
<td>Recently discovered mineral deposits in the Irish Republic</td>
<td>1</td>
</tr>
<tr>
<td>Historical summary</td>
<td>3</td>
</tr>
<tr>
<td>Ore reserves</td>
<td>5</td>
</tr>
<tr>
<td>Previous work</td>
<td>7</td>
</tr>
<tr>
<td>Outline of present study</td>
<td>10</td>
</tr>
<tr>
<td>CHAPTER 2 - GEOLOGICAL AND PALAEOGEOGRAPHICAL SETTING</td>
<td>12</td>
</tr>
<tr>
<td>Regional geology</td>
<td></td>
</tr>
<tr>
<td>a) Pre-Carboniferous</td>
<td>12</td>
</tr>
<tr>
<td>b) Lower Carboniferous: stratigraphy and palaeogeography</td>
<td>13</td>
</tr>
<tr>
<td>c) Lower Carboniferous igneous activity</td>
<td>17</td>
</tr>
<tr>
<td>d) Post-Lower Carboniferous</td>
<td>19</td>
</tr>
<tr>
<td>Geological setting in the vicinity of the Mogul mine</td>
<td>20</td>
</tr>
<tr>
<td>a) Succession</td>
<td>20</td>
</tr>
</tbody>
</table>
b) Structure

   c) Positions of the Mogul and Magcobar orebodies

CHAPTER 3 - THE G ZONE OREBODIES: DIMENSIONS AND METAL DISTRIBUTIONS

Dimensions

Metal distributions

   a) Lower orebody

   b) Upper orebody

CHAPTER 4 - THE UPPER OREBODY AND ASSOCIATED MINERALIZATION

General aspects of the orebody and host rocks

   a) Horizons below orebody

   b) The orebody

   c) Dolomite Breccia

   d) Post-consolidational features

Mineralogy

   a) Sulphides and sulphosalts

   b) Other minerals

Textural features

   a) General remarks

   b) Colloform textures involving principally pyrite

      1) Type 1. Dendritic
      2) Type 2. Concentrically banded
      3) Type 3. Cylindrically shaped
      4) Type 4. Irregularly shaped with black centres
      5) Type 5. Framboidal clusters
      6) Type 6. Euhedral zoned structures

   c) Colloform structures involving sphalerite and galena
d) Types of colloform structures involving other mineral groups

1) Type 1. Pyrite and sphalerite
2) Type 2. Galena and pyrite
3) Type 3. Calcite and dolomite associated with sulphides
4) Type 4. Several minerals

<table>
<thead>
<tr>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>101</td>
</tr>
<tr>
<td>101</td>
</tr>
<tr>
<td>104</td>
</tr>
<tr>
<td>104</td>
</tr>
<tr>
<td>106</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>e) Other textural features</th>
</tr>
</thead>
<tbody>
<tr>
<td>1) Pyrite</td>
</tr>
<tr>
<td>2) Marcasite</td>
</tr>
<tr>
<td>3) Sphalerite and galena</td>
</tr>
<tr>
<td>4) Chalcopyrite, tennantite and boulangierite</td>
</tr>
<tr>
<td>5) Quartz, barite and carbonates</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>111</td>
</tr>
<tr>
<td>111</td>
</tr>
<tr>
<td>114</td>
</tr>
<tr>
<td>115</td>
</tr>
<tr>
<td>115</td>
</tr>
</tbody>
</table>

Mineralization in minor fault zone
Number 3 zone mineralization

CHAPTER 5 - THE LOWER OREBODY

<table>
<thead>
<tr>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>122</td>
</tr>
<tr>
<td>122</td>
</tr>
<tr>
<td>123</td>
</tr>
<tr>
<td>124</td>
</tr>
<tr>
<td>125</td>
</tr>
<tr>
<td>129</td>
</tr>
</tbody>
</table>

CHAPTER 6 - ASSOCIATED ORE DEPOSITS

<table>
<thead>
<tr>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>133</td>
</tr>
<tr>
<td>133</td>
</tr>
<tr>
<td>134</td>
</tr>
<tr>
<td>140</td>
</tr>
</tbody>
</table>

CHAPTER 7 - ANALYTICAL AND ASSOCIATED RESULTS

<table>
<thead>
<tr>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>143</td>
</tr>
<tr>
<td>143</td>
</tr>
<tr>
<td>145</td>
</tr>
<tr>
<td>148</td>
</tr>
<tr>
<td>149</td>
</tr>
</tbody>
</table>
CHAPTER 8 - FRAMBOIDS

Terminology 151
Distribution of frambooids 151
Textural features 152
  a) Shape 154
  b) Individual grains 157
  c) Size analyses 163
Mineralogy of frambooids 168
Grouping of frambooids 169
Overgrowths on frambooids 170
Discussion 170

CHAPTER 9 - DISCUSSION 183

Sedimentary features 183
  a) Features in the Mogul and Magcobar deposits 183
  b) Comparison with features in other areas 188
  c) Summary 191
Textures and mineralogy 192
  a) Colloform textures 192
    1) Previous work 192
    2) Discussion of evidence from the Upper orebody 198
    3) Paragenetic sequences in the Upper orebody 202
    4) Other colloform textures 203
    5) Erection of colloform textures 204
    6) Recrystallization of colloform textures 204
    7) Summary 205
  b) Other textural features 206
    1) Upper orebody 206
    2) No. 3 zone 206
    3) Lower orebody 207
Minor elements

a) Iron content of sphalerite 208

b) Silver and antimony content of galena 209

c) Nickel and cobalt content of pyrite 210

Metal distribution 212

Sulphur isotopes 215

Precipitation of sulphides and source of metals 218

a) Precipitation 218

b) Transport 221

c) Source 221

Summary of ore formation processes 223

CHAPTER 10 - CONCLUSIONS 226

REFERENCES 228

VITA
LIST OF TABLES

<table>
<thead>
<tr>
<th>TABLE</th>
<th>page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Lower Carboniferous succession in Ireland</td>
<td>14</td>
</tr>
<tr>
<td>2 Succession at the Mogul mine</td>
<td>23</td>
</tr>
<tr>
<td>3 Nickel and cobalt content of pyrite</td>
<td>144</td>
</tr>
<tr>
<td>4 Silver and antimony content of galena</td>
<td>146</td>
</tr>
<tr>
<td>5 Sulphur isotope results</td>
<td>150</td>
</tr>
</tbody>
</table>
# LIST OF ILLUSTRATIONS

<table>
<thead>
<tr>
<th>FIGURE</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Geological sketch map of Ireland</td>
<td>2</td>
</tr>
<tr>
<td>2</td>
<td>Map of the Silvermines district</td>
<td>6</td>
</tr>
<tr>
<td>3</td>
<td>Distribution of the Waulsortian facies in Ireland</td>
<td>16</td>
</tr>
<tr>
<td>4</td>
<td>Isopach map of the Waulsortian facies in Ireland</td>
<td>18</td>
</tr>
<tr>
<td>5</td>
<td>Geological map of central Ireland</td>
<td>21</td>
</tr>
<tr>
<td>6</td>
<td>Geological map of the Mogul mine vicinity</td>
<td>24</td>
</tr>
<tr>
<td>7</td>
<td>Section through the Upper and Lower orebodies</td>
<td>25</td>
</tr>
<tr>
<td>8</td>
<td>Detrital quartz grains in shale (photo)</td>
<td>29</td>
</tr>
<tr>
<td>9</td>
<td>Isopach map of the Upper orebody</td>
<td>33</td>
</tr>
<tr>
<td>10</td>
<td>Relation between depth and Zn:Pb ratio, Lower orebody</td>
<td>35</td>
</tr>
<tr>
<td>11</td>
<td>Relation between depth and Zn:Pb ratio, Lower orebody</td>
<td>36</td>
</tr>
<tr>
<td>12</td>
<td>Variation diagram of Pb/Zn for the Lower orebody</td>
<td>37</td>
</tr>
<tr>
<td>13</td>
<td>Variation diagram of Pb/Ag for the Lower orebody</td>
<td>39</td>
</tr>
<tr>
<td>14</td>
<td>Variation diagram of Zn/Ag for the Lower orebody</td>
<td>39</td>
</tr>
<tr>
<td>15</td>
<td>Plan of lead distribution in Upper orebody</td>
<td>40</td>
</tr>
<tr>
<td>16</td>
<td>Plan of zinc distribution in Upper orebody</td>
<td>41</td>
</tr>
<tr>
<td>17</td>
<td>Plan of silver distribution in Upper orebody</td>
<td>42</td>
</tr>
<tr>
<td>18</td>
<td>Vertical distribution of Pb, Zn and silver in Upper orebody, west part</td>
<td>44</td>
</tr>
<tr>
<td>19</td>
<td>Vertical distribution of zinc in Upper orebody, central part</td>
<td>45</td>
</tr>
<tr>
<td>20</td>
<td>Vertical distribution of lead in Upper orebody, central part</td>
<td>46</td>
</tr>
<tr>
<td>21</td>
<td>Vertical distribution of silver in Upper orebody, central part</td>
<td>47</td>
</tr>
<tr>
<td>22</td>
<td>Vertical distribution of Pb, Zn and Ag in Upper orebody, eastern part</td>
<td>48</td>
</tr>
<tr>
<td>23</td>
<td>Variation diagram of Pb/Zn for Upper orebody</td>
<td>49</td>
</tr>
<tr>
<td>24</td>
<td>Variation diagram of Zn/Ag for Upper orebody</td>
<td>50</td>
</tr>
<tr>
<td>25</td>
<td>Variation diagram of Pb/Ag for Upper orebody</td>
<td>50</td>
</tr>
<tr>
<td>26</td>
<td>Interbedded sphalerite and shale (photo.)</td>
<td>54</td>
</tr>
<tr>
<td>27</td>
<td>Interbedded sphalerite and shale (photo.)</td>
<td>54</td>
</tr>
<tr>
<td>28</td>
<td>Soft-sediment folding in a galena band (photo.)</td>
<td>54</td>
</tr>
<tr>
<td>29</td>
<td>Sedimentary banding in a galena-shale layer (photo)</td>
<td>54</td>
</tr>
<tr>
<td>30</td>
<td>Folded, interbedded pyrite and shale (photo.)</td>
<td>56</td>
</tr>
<tr>
<td>31</td>
<td>Laminated sphalerite-shale deformed by pyrite clasts (photo.)</td>
<td>56</td>
</tr>
<tr>
<td>32</td>
<td>Sulphide fragments in shale (photo.)</td>
<td>56</td>
</tr>
<tr>
<td>FIGURE</td>
<td>page</td>
<td></td>
</tr>
<tr>
<td>--------</td>
<td>------</td>
<td></td>
</tr>
<tr>
<td>73</td>
<td>83</td>
<td>Dendritic colloform pyrite (photo.)</td>
</tr>
<tr>
<td>74</td>
<td>83</td>
<td>Dendritic colloform pyrite (photo.)</td>
</tr>
<tr>
<td>75</td>
<td>83</td>
<td>Dendritic colloform pyrite (photo.)</td>
</tr>
<tr>
<td>76-78</td>
<td>84</td>
<td>Sequence showing increasing dominance of concentric structure in colloform pyrite (photo.)</td>
</tr>
<tr>
<td>79</td>
<td>84</td>
<td>Elongated, concentrically banded colloform pyrite (photo.)</td>
</tr>
<tr>
<td>80</td>
<td>86</td>
<td>Sharply fractured, colloform pyrite (photo.)</td>
</tr>
<tr>
<td>81</td>
<td>86</td>
<td>Envelopment of early, concentric, pyrite structures by later bands (photo.)</td>
</tr>
<tr>
<td>82</td>
<td>86</td>
<td>Radial structure in colloform pyrite (photo.)</td>
</tr>
<tr>
<td>83</td>
<td>86</td>
<td>Coarse radial texture in colloform pyrite (photo.)</td>
</tr>
<tr>
<td>84</td>
<td>87</td>
<td>Interbanded radial and non-radial pyrite in colloform structure (photo.)</td>
</tr>
<tr>
<td>85</td>
<td>87</td>
<td>Radial texture in colloform pyrite (photo.)</td>
</tr>
<tr>
<td>86</td>
<td>87</td>
<td>Radial texture in colloform pyrite (photo.)</td>
</tr>
<tr>
<td>87</td>
<td>87</td>
<td>Colloform pyrite with variable grain size (photo.)</td>
</tr>
<tr>
<td>88</td>
<td>89</td>
<td>Concentric layers of cubic pyrite (photo.)</td>
</tr>
<tr>
<td>89</td>
<td>89</td>
<td>Distinctive layer occurring at same place in several colloform pyrite structures (photo.)</td>
</tr>
<tr>
<td>90</td>
<td>89</td>
<td>Pyrite growths at edge of colloform structure (photo.)</td>
</tr>
<tr>
<td>91</td>
<td>89</td>
<td>Irregularly shaped colloform pyrite structures (photo.)</td>
</tr>
<tr>
<td>92</td>
<td>90</td>
<td>Colloform pyrite with a granular internal texture (photo.)</td>
</tr>
<tr>
<td>93</td>
<td>90</td>
<td>Colloform pyrite surrounding galema (photo.)</td>
</tr>
<tr>
<td>94</td>
<td>90</td>
<td>Colloform pyrite surrounding euhedral pyrite (photo.)</td>
</tr>
<tr>
<td>95</td>
<td>90</td>
<td>Colloform pyrite surrounded by euhedral pyrite (photo.)</td>
</tr>
<tr>
<td>96</td>
<td>92</td>
<td>Porous, black, colloform pyrite (photo.)</td>
</tr>
<tr>
<td>97</td>
<td>92</td>
<td>Porous, black, colloform pyrite (photo.)</td>
</tr>
<tr>
<td>98</td>
<td>92</td>
<td>Rounded group of framboinds (photo.)</td>
</tr>
<tr>
<td>99</td>
<td>92</td>
<td>Zoned, radial pyrite (photo.)</td>
</tr>
<tr>
<td>100</td>
<td>95</td>
<td>Colloform galena-sphalerite (photo.)</td>
</tr>
<tr>
<td>101</td>
<td>95</td>
<td>Concentrically banded sphalerite (photo.)</td>
</tr>
<tr>
<td>102</td>
<td>95</td>
<td>Colloform galena-sphalerite (photo.)</td>
</tr>
<tr>
<td>103</td>
<td>95</td>
<td>Skeletal galena in sphalerite (photo.)</td>
</tr>
<tr>
<td>104</td>
<td>96</td>
<td>Skeletal galena in sphalerite (photo.)</td>
</tr>
<tr>
<td>105</td>
<td>96</td>
<td>Skeletal galena in sphalerite (photo.)</td>
</tr>
<tr>
<td>106</td>
<td>96</td>
<td>Relationship of fibrous sphalerite and galena crystals (photo.)</td>
</tr>
<tr>
<td>107</td>
<td>96</td>
<td>Relationship of fibrous sphalerite and galena crystals (photo.)</td>
</tr>
<tr>
<td>108</td>
<td>97</td>
<td>Dendritic, skeletal galena (photo.)</td>
</tr>
<tr>
<td>109</td>
<td>97</td>
<td>Galena globules in sphalerite (photo.)</td>
</tr>
<tr>
<td>110</td>
<td>97</td>
<td>Galena globules in sphalerite (photo.)</td>
</tr>
<tr>
<td>111</td>
<td>97</td>
<td>Galena globules in sphalerite (photo.)</td>
</tr>
<tr>
<td>112</td>
<td>98</td>
<td>Sphalerite globules in galena (photo.)</td>
</tr>
<tr>
<td>FIGURE</td>
<td>Description</td>
<td>Page</td>
</tr>
<tr>
<td>--------</td>
<td>-----------------------------------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>113</td>
<td>Galena-sphalerite intergrowths (photo.)</td>
<td>98</td>
</tr>
<tr>
<td>114</td>
<td>Concretionary, layered sphalerite (photo.)</td>
<td>98</td>
</tr>
<tr>
<td>115</td>
<td>Sphalerite globules (photo.)</td>
<td>98</td>
</tr>
<tr>
<td>116</td>
<td>Concentrically banded sphalerite (photo.)</td>
<td>100</td>
</tr>
<tr>
<td>117</td>
<td>Concentrically banded sphalerite (photo.)</td>
<td>100</td>
</tr>
<tr>
<td>118</td>
<td>Pyrite radiating from sphalerite nucleus (photo.)</td>
<td>100</td>
</tr>
<tr>
<td>119</td>
<td>Interbanded pyrite and sphalerite (photo.)</td>
<td>102</td>
</tr>
<tr>
<td>120</td>
<td>Figure 119 in oblique light (photo.)</td>
<td>102</td>
</tr>
<tr>
<td>121</td>
<td>Relationship of colloform pyrite to sphalerite (photo.)</td>
<td>102</td>
</tr>
<tr>
<td>122</td>
<td>Part of Figure 121 in oblique light (photo.)</td>
<td>102</td>
</tr>
<tr>
<td>123</td>
<td>Colloform sphalerite and granular pyrite (photo.)</td>
<td>103</td>
</tr>
<tr>
<td>124</td>
<td>Dendritic pyrite and sphalerite (photo.)</td>
<td>103</td>
</tr>
<tr>
<td>125</td>
<td>Fragments of colloform pyrite in sphalerite (photo.)</td>
<td>103</td>
</tr>
<tr>
<td>126</td>
<td>Galena-pyrite intergrowth (photo)</td>
<td>103</td>
</tr>
<tr>
<td>127</td>
<td>Pyrite encrusting galena crystals (photo.)</td>
<td>105</td>
</tr>
<tr>
<td>128</td>
<td>Angular pyrite grains in galena (photo.)</td>
<td>105</td>
</tr>
<tr>
<td>129</td>
<td>Pyrite globules in galena (photo.)</td>
<td>105</td>
</tr>
<tr>
<td>130</td>
<td>Colloform pyrite-calcite (photo.)</td>
<td>105</td>
</tr>
<tr>
<td>131</td>
<td>Colloform pyrite-calcite (photo.)</td>
<td>107</td>
</tr>
<tr>
<td>132</td>
<td>Colloform pyrite with calcite globules (photo.)</td>
<td>107</td>
</tr>
<tr>
<td>133</td>
<td>Dendritic calcite in sphalerite (photo.)</td>
<td>107</td>
</tr>
<tr>
<td>134</td>
<td>Colloform calcite-sphalerite (photo.)</td>
<td>107</td>
</tr>
<tr>
<td>135</td>
<td>Euhedral dolomite in sphalerite (photo.)</td>
<td>108</td>
</tr>
<tr>
<td>136</td>
<td>Various globules in colloform pyrite (photo.)</td>
<td>108</td>
</tr>
<tr>
<td>137</td>
<td>Sphalerite and composite globules in pyrite (photo.)</td>
<td>108</td>
</tr>
<tr>
<td>138</td>
<td>Finely zoned sphalerite-galena globules (photo.)</td>
<td>108</td>
</tr>
<tr>
<td>139</td>
<td>Globules cross-cutting bands in colloform pyrite (photo.)</td>
<td>109</td>
</tr>
<tr>
<td>140</td>
<td>Galena globules encrusted by sphalerite and pyrite (photo.)</td>
<td>109</td>
</tr>
<tr>
<td>141</td>
<td>Euhedral pyrite grains (photo.)</td>
<td>109</td>
</tr>
<tr>
<td>142</td>
<td>Interfering, euhedral pyrite grains (photo.)</td>
<td>109</td>
</tr>
<tr>
<td>143</td>
<td>Granular pyrite (photo.)</td>
<td>110</td>
</tr>
<tr>
<td>144</td>
<td>Granular pyrite formed by close packing of spheroids (photo.)</td>
<td>110</td>
</tr>
<tr>
<td>145</td>
<td>Galena intergrown in pyrite crystal (photo.)</td>
<td>110</td>
</tr>
<tr>
<td>146</td>
<td>Pyrite in shale (photo.)</td>
<td>110</td>
</tr>
<tr>
<td>147</td>
<td>Subhedral pyrite with parallel fracturing (photo.)</td>
<td>113</td>
</tr>
<tr>
<td>148</td>
<td>Fossil replaced by pyrite (photo.)</td>
<td>113</td>
</tr>
<tr>
<td>149</td>
<td>Pyrite enclosing dolomite grains (photo.)</td>
<td>113</td>
</tr>
<tr>
<td>150</td>
<td>Fractured and veined colloform pyrite (photo.)</td>
<td>113</td>
</tr>
<tr>
<td>151</td>
<td>Intensely fractured marcasite (photo.)</td>
<td>117</td>
</tr>
<tr>
<td>152</td>
<td>Granular sphalerite-dolomite (photo.)</td>
<td>117</td>
</tr>
<tr>
<td>153</td>
<td>Boulangierite in galena (photo.)</td>
<td>117</td>
</tr>
<tr>
<td>154</td>
<td>Quartz encrusting pyrite (photo.)</td>
<td>117</td>
</tr>
<tr>
<td>155</td>
<td>Pyrite fragments, no. 3 zone (photo.)</td>
<td>120</td>
</tr>
<tr>
<td>156</td>
<td>Bedded sulphide, top of Lower Dolomite (photo.)</td>
<td>120</td>
</tr>
<tr>
<td>Page</td>
<td>Description</td>
<td></td>
</tr>
<tr>
<td>------</td>
<td>-----------------------------------------------------------------------------</td>
<td></td>
</tr>
<tr>
<td>165</td>
<td>Veining in Lower Dolomite (photo.)</td>
<td></td>
</tr>
<tr>
<td>166</td>
<td>Pyrite grains veined by sphalerite (photo.)</td>
<td></td>
</tr>
<tr>
<td>167</td>
<td>Pyrite aggregate veined by sphalerite (photo.)</td>
<td></td>
</tr>
<tr>
<td>168</td>
<td>Pyrite partly replaced by galena (photo.)</td>
<td></td>
</tr>
<tr>
<td>169</td>
<td>Galena-sphalerite-dolomite relationships (photo.)</td>
<td></td>
</tr>
<tr>
<td>170</td>
<td>Laminated sulphide, B zone (photo.)</td>
<td></td>
</tr>
<tr>
<td>171</td>
<td>Laminated chert-hematite (photo.)</td>
<td></td>
</tr>
<tr>
<td>172</td>
<td>Chert concretions (photo.)</td>
<td></td>
</tr>
<tr>
<td>173</td>
<td>Chert concretion deforming laminations (photo.)</td>
<td></td>
</tr>
<tr>
<td>174</td>
<td>Colloform chert encrustation (photo.)</td>
<td></td>
</tr>
<tr>
<td>175</td>
<td>Radiating pink barite (photo. colour)</td>
<td></td>
</tr>
<tr>
<td>176</td>
<td>Barite enclosed by pyrite (photo.)</td>
<td></td>
</tr>
<tr>
<td>177</td>
<td>Colloform sulphides in a vein (photo.)</td>
<td></td>
</tr>
<tr>
<td>178</td>
<td>Vickers hardness of galena</td>
<td></td>
</tr>
<tr>
<td>179</td>
<td>Framboid and euhedral pyrite (photo.)</td>
<td></td>
</tr>
<tr>
<td>180</td>
<td>Non-spherical framboids (photo.)</td>
<td></td>
</tr>
<tr>
<td>181</td>
<td>Well defined framboids (photo.)</td>
<td></td>
</tr>
<tr>
<td>182</td>
<td>Poorly defined framboids (photo.)</td>
<td></td>
</tr>
<tr>
<td>183</td>
<td>Partly disintegrated framboid (photo.)</td>
<td></td>
</tr>
<tr>
<td>184</td>
<td>Disintegrated framboid (photo.)</td>
<td></td>
</tr>
<tr>
<td>185</td>
<td>Framboids surrounded by fine-grained pyrite (photo.)</td>
<td></td>
</tr>
<tr>
<td>186</td>
<td>Pod-shaped aggregate of pyrite grains in shale (photo.)</td>
<td></td>
</tr>
<tr>
<td>187</td>
<td>Number of individual grains in framboids: histogram (photo.)</td>
<td></td>
</tr>
<tr>
<td>188</td>
<td>Number of individual grains in framboids (photo.)</td>
<td></td>
</tr>
<tr>
<td>189</td>
<td>Large framboids with small grains (photo.)</td>
<td></td>
</tr>
<tr>
<td>190</td>
<td>Framboid group (photo.)</td>
<td></td>
</tr>
<tr>
<td>191</td>
<td>Framboids with some cubic grains (photo.)</td>
<td></td>
</tr>
<tr>
<td>192</td>
<td>Framboids with irregularly shaped grains (photo.)</td>
<td></td>
</tr>
<tr>
<td>193</td>
<td>Framboid with a granular texture (photo.)</td>
<td></td>
</tr>
<tr>
<td>194</td>
<td>Framboid with a granular texture (photo.)</td>
<td></td>
</tr>
<tr>
<td>195</td>
<td>Framboid with concentric pattern (photo.)</td>
<td></td>
</tr>
<tr>
<td>196</td>
<td>Small framboids in shale (photo.)</td>
<td></td>
</tr>
<tr>
<td>197</td>
<td>Framboid containing galena (photo.)</td>
<td></td>
</tr>
<tr>
<td>198</td>
<td>Framboids containing sphalerite (photo.)</td>
<td></td>
</tr>
<tr>
<td>199</td>
<td>Framboids with pyrite between grains (photo.)</td>
<td></td>
</tr>
<tr>
<td>200</td>
<td>Size distribution of framboids: histogram</td>
<td></td>
</tr>
<tr>
<td>201</td>
<td>Size distribution of framboids: histogram</td>
<td></td>
</tr>
<tr>
<td>202</td>
<td>Size distribution of framboids: histogram</td>
<td></td>
</tr>
<tr>
<td>FIGURE</td>
<td>Description</td>
<td>Page</td>
</tr>
<tr>
<td>--------</td>
<td>-----------------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>203</td>
<td>Size distribution of framboïds: histogram</td>
<td>165</td>
</tr>
<tr>
<td>204</td>
<td>Size distribution of framboïds: histogram</td>
<td>166</td>
</tr>
<tr>
<td>205</td>
<td>Framboïd group (photo.)</td>
<td>171</td>
</tr>
<tr>
<td>206</td>
<td>Framboïd group (photo.)</td>
<td>171</td>
</tr>
<tr>
<td>207</td>
<td>Framboïds encrusted with pyrite (photo.)</td>
<td>171</td>
</tr>
<tr>
<td>208</td>
<td>Framboïd encrusted with pyrite (photo.)</td>
<td>171</td>
</tr>
<tr>
<td>209</td>
<td>Framboïds encrusted with pyrite (photo.)</td>
<td>172</td>
</tr>
<tr>
<td>210</td>
<td>Framboïds enclosed by granular pyrite (photo.)</td>
<td>172</td>
</tr>
</tbody>
</table>
Recently discovered mineral deposits in the Irish Republic.

Until 1960 systematic mineral exploration had never been undertaken in Ireland but in 1961 a Canadian company found the sizeable base metal deposit of Tynagh, Co. Galway, and this led to a phase of intensive exploration in the Irish Republic with many Canadian companies participating. In the following few years several new deposits were discovered. Some of these proved to be of economic size and grade and as a result three new base metal mines are now in operation. Mogul's deposit, mined for its lead, zinc and silver is the largest of the three and was the latest to be brought into production. It consists of a large stratiform massive sulphide body in Lower Carboniferous limestone with underlying vein-type mineralization.

The Mogul mine is situated in the Silvermines district of Co. Tipperary (Fig. 1) in the central-southern part of Ireland. The other operational mines are the Tynagh mine, referred to above, which produces lead, zinc, copper and silver and the Gortdrum mine, 20 miles south of Silvermines, producing copper, silver and mercury. In addition a large barite mine (discovered in the late 1950's) is in operation adjacent to the Mogul mine. A feature
Fig. 1. Geological sketch map of Ireland showing the locations of recent mining development.
common to these deposits and several others of marginal or non-economic grade is that they all occur in calcareous sediments of Lower Carboniferous age.

**Historical Summary.**

A review of the mining history of the Silvermines district has been given by Rhoden (1958). In addition to the large Mogul deposit, the Silvermines district contains scattered Pb, Zn, Cu, Ag mineralization for over 2 miles along a belt close to the line of the Silvermines Fault which downthrows Lower Carboniferous limestone against Devonian sandstone. The sandstone on the south side of the Fault is moderately well exposed and mineralised veins occurring in it have been amenable to discovery by local inhabitants throughout history and traces of ancient workings have been reported. The earliest documented mining took place at the beginning of the 17th century. Galena bearing veins were the earliest to be exploited, originally for lead, but it was soon discovered that the galena also contained silver which was subsequently extracted. This phase of activity ended in the mid 17th century and no further mining operations took place until the early 19th century. From then, until late in the century, fairly extensive small scale mining was carried out at several centres along the belt and concentrates of lead, zinc, copper and silver were produced. The amounts mined were small, for example between 1853 and 1874, the mine at Shallee produced 1392 tons of lead and in a seven year period the mine at Gortnadyne
produced 420 tons of lead.

Around 1870, cheaper metals from abroad became available and resulted in the cessation of mining in the Silvermines district. Further investigations of the area were carried out during World War I and again in 1929 and 1930 but it was not until 1949 that mining restarted and was enabled to continue because of increased metal prices during the Korean War. The renewed mining was by the Irish Exploration Co. which later became known as Silvermines Lead and Zinc Co. Ltd. Their efforts were concentrated on the galena veins in the sandstone at Shallee near the western end of the belt. They also attempted to produce zinc oxide concentrate from a calamine deposit at the eastern end but mining and treatment difficulties forced the abandonment of this project. At Shallee mining continued until 1958, with a break between 1953 and 1955, and altogether 355,000 long tons of ore were produced. The grade of the ore was low, averaging about 1½% Pb and about 0.6 oz./ton Ag, and mining became uneconomic after the Korean War.

Geophysical surveys and diamond drilling between 1952 and 1957 led to the discovery of a large massive barite body adjacent to the Mogul orebody, at that time still undiscovered. The barite body was leased to Magnet Cove Barite Corporation and is presently being mined (Magcobar).

In 1962, after the discovery of the base metal orebody at Tynagh, Co. Galway, Consolidated Mogul Mines Ltd. (now International Mogul Mines Ltd.) made an agreement with Silvermines Lead and Zinc Company to incorporate under the name
Mogul of Ireland, owned 75% by Mogul and 25% by Silvermines. Altogether Mogul of Ireland now holds prospecting and mineral rights over an area of 31½ square miles covering a rectangular strip of land astride the mineralized zone of the Silvermines Fault.

Late in 1962, intensive systematic exploration was started in this area. Geological mapping, geochemical soil-sampling surveys, induced polarization, gravimetric and electromagnetic surveys were all carried out during the next year. Finally anomalous areas were drilled and by 1964 three ore zones (K, B & G) had been outlined (Fig. 2). Early in 1965, exploratory drilling was suspended and efforts were directed towards development of the G zone, the largest of the three. Preparations were made for underground mining and a concentrator was built on the site. In the summer of 1968, the mine went into production. Further exploration drilling was started late in 1969 to outline more precisely the B zone and this work is still in progress.

**Ore reserves.**

The G zone contains two orebodies referred to as the Upper and Lower orebodies. Drill-indicated reserves for these are as follows:

<table>
<thead>
<tr>
<th></th>
<th>SHORT TONS</th>
<th>%Pb</th>
<th>%Zn</th>
<th>oz./ton Ag</th>
</tr>
</thead>
<tbody>
<tr>
<td>UPPER</td>
<td>9,332,100</td>
<td>2.4</td>
<td>9.2</td>
<td>0.74</td>
</tr>
<tr>
<td>LOWER</td>
<td>2,068,800</td>
<td>4.5</td>
<td>3.4</td>
<td>1.13</td>
</tr>
<tr>
<td>TOTAL</td>
<td>11,400,900</td>
<td>2.8</td>
<td>8.2</td>
<td>0.8</td>
</tr>
</tbody>
</table>
Fig. 2. Sketch map of the Silvermines district showing the location of deposits and their relation to the Silvermines Fault.
Figures for the B and K zones are as follows:

<table>
<thead>
<tr>
<th></th>
<th>SHORT TONS</th>
<th>Pb</th>
<th>Zn</th>
<th>oz./ton Ag</th>
</tr>
</thead>
<tbody>
<tr>
<td>B zone</td>
<td>2,000,000</td>
<td>4.0</td>
<td>6.0</td>
<td>1.0</td>
</tr>
<tr>
<td>K zone</td>
<td>500,000</td>
<td>1.0</td>
<td>6.5</td>
<td>0.5</td>
</tr>
</tbody>
</table>

The mill is designed to operate at 3000 tons per day and this rate is being achieved.

**Previous work.**

References to the Silvermines deposits date back as far as the early 17th century and, with the exception of a few which have appeared since 1962, all refer to the long known, vein-type mineralization. Perhaps the most useful of the older work is that contained in the Memoirs of the Geological Survey of Ireland. Wynne (1861) examined the deposits for the Survey and wrote up a fairly detailed field report of all the deposits known at that time.

In 1907, Russel examined a suite of specimens collected from the old mines and provided a list of the minerals occurring in each deposit. He did not however discuss the genesis of the deposits.

The only major study of the Silvermines deposits was carried out by Rhoden in the late 1950's, (Rhoden, 1958; 1959). In the first paper, he presented a detailed account of the Silvermines Fault system and the stratigraphy of the area, and also described the various deposits. His work showed that the Fault system was a complex made up of numerous faults with three dominant trends, E.N.E., E.S.E. and N.-S., and that the last two of these trends represent tension fractures.
which opened up in response to a regional E.N.E. couple prevailing in southern Ireland after the Hercynian folding. He considered the whole fault system, and in particular the tension faults, as the channelling system for uprising hydrothermal solutions from an undisclosed source. He stated that all mineralization was later than the fracturing and postulated for it a Triassic to early Jurassic age. He also pointed out that no mineral zoning is evident in the deposits as a whole. In his 1959 paper Rhoden described the mineralogy of the deposits, which he considered to be leptothermal, and presented a paragenetic sequence based on textural relationships.

The first paper to mention the recently discovered Mogul deposits was written by Weber in 1964. His main purpose was to describe the techniques used in the discovery of the deposits but he also briefly described the Upper and Lower orebodies of the G zone. He suggested that the Upper orebody was syngenetically precipitated in a chemically favourable basin and then subjected to recrystallization and other processes connected with consolidation and possible metasomatic waters. He considered the Lower orebody to be a replacement deposit with associated silicification and dolomitization controlled by the Silvermines Fault zone.

In a presentation in 1968 Weber described the geology of the Silvermines area and included much new information revealed by extensive exploration drilling. He also amplified his earlier descriptions and theories of origin of the orebodies indicating that the sulphides of the Upper orebody formed as gels in a restricted basin, hinged on the fault plane.
and that recurrent movement on the fault during deposition caused sliding of the gels, giving rise to the abundant slump structures and breccia trains now found in the ores. He considered that biogenic activity may also have been important in the formation of the ores. The metals may have been originally deposited in the Lower Palaeozoic sediments and reached the Carboniferous through periods of recycling.

Pereira (1967), made reference to the Mogul deposits in a paper comparing Lower Carboniferous deposits in Ireland with Cretaceous deposits in Iran and stressed the importance to ore deposition of coastal sedimentation and marine transgression, active tectonic movement during deposition, fumarolic and volcanic associations, and carbonate sedimentation. In earlier papers, Pereira (1963; 1963a) suggested that the Irish deposits formed in coastal lagoons with fumaroles the main source of materials, but deposition possibly aided by bacterial action. He thought that deep euxinic basins with restricted circulation were not necessary for the development of sulphides provided there was sufficiently great organic activity to produce locally reducing conditions on the sea bed.

Snelgrove (1966), in a two-part paper briefly described Irish Carboniferous deposits and discussed Pereira's ideas on their origin. He pointed out that their volcanic associations and relatively high silver content set them apart from the Mississippi Valley type deposits. He also considered that the Hercynian folding may have had considerable influence in remobilizing syngenetic deposits.
Russel (1958), suggested that the Carboniferous base metal deposits in Ireland were located at the intersections of major north-south upper mantle fissures with east or north-easterly trending faults of Caledonian age. He proposed that magma intruded at the intersections initiated a circulation of pore waters which leached metals from the Lower Palaeozoic sediments and deposited them in Carboniferous rocks or on the Carboniferous sea floor. He thought the mantle fissures were related to continental drifting in Devonian and Carboniferous times.

In summary it can be said that ideas on the origin of the Silvermines deposits have changed radically since the discovery of the large stratiform orebody. Until that time, the theory of epigenetic emplacement after the Hercynian folding was unquestioned. Since then all authors have supported a syngenetic origin for at least the Upper orebody. There is general agreement that it formed in a localised basin near an active fault and possibly near a shoreline. Biogenic activity is thought to have played a part in its formation and as a source of metals volcanic activity, underlying argillaceous sediments and recycling of previously deposited metals have been called on. The vein-type mineralization is thought to have been deposited either by solutions rising to the Lower Carboniferous sea floor or by remobilization of the syngenetic deposits at a later date.

Outline of present study.

The old mines in the Silvermines district have been
described in detail previously and although further work could be done on them, it was decided to restrict the present study mainly to Mogul's G zone orebodies. The main object was to determine the origin of this zone. It was considered important because it has features similar to both Mississippi Valley type deposits and massive sulphide deposits of the copper-zinc type.

Because no previous detailed research had been done, it was decided to carry out a general study. With Weber's work (1968) as a background some further field work was done in the Mogul mine and in the Magcobar open pit barite mine to study ore-sediment relationships, and sedimentary structures and to collect specimens. Extensive collections were also made from G zone exploration drill core and a few specimens were taken from cores of the K and B zones.

In the laboratory textural relationships of the various sulphides and associated sediments were studied in detail to establish the paragenesis of the deposits. Metal distributions were plotted to illustrate zoning, some minor-element analyses on sulphides were made and sulphur isotope ratios were determined.
CHAPTER 2
GEOLOGICAL AND PALÄOGEOGRAPHICAL SETTING

Regional.
(a) Pre-Carboniferous.

From Cambrian to Silurian most of Ireland formed part of the Caledonian geosyncline and during that time a thick sequence of eugeosynclinal sediments with local volcanics was laid down. At the end of the Silurian these rocks were strongly folded along an E.N.E. trend in the southern part of Ireland but were scarcely metamorphosed. Some associated post-orogenic granitic intrusive activity did occur but the closest of these intrusions to the Silvermines district are in Co. Galway, 60 miles to the north-west, and in Co. Wicklow, 60 miles to the east.

In Devonian times palæogeographic relations over the Irish region changed considerably. Almost the whole area became land of high, but subsiding relief, on which terrigenous deposits of red sandstones and conglomerates accumulated. These lie with strong unconformity on the underly ing Lower Palæozoic rocks. The extreme southern part of Ireland lay on the northern flank of a rapidly subsiding trough which extended east-south-eastwards and underwent rapid clastic sedimentation. Towards the end of the Devonian relief had become almost completely subdued and the sea began
to transgress northwards so that the uppermost Devonian beds are marine. Sporadic andesitic igneous activity did occur during the Devonian but occurrences are far to the north or south of the Silvermines district.

(b) Lower Carboniferous: stratigraphy and palaeogeography.

A comprehensive account of the stratigraphy and palaeogeography of this period is given by George (1958). During the Lower Carboniferous or Dinantian the marine transgression already initiated continued so that in central and southern Ireland there is a conformable passage from Devonian to Carboniferous.

A generalised stratigraphic succession for the Dinantian is given in Table 1.

During the Dinantian the trough in the south remained in existence and continued to accumulate clastic sediments derived mainly from the south. Much of the rest of the Ireland was covered by a shallow shelf sea but parts of the north-west remained land. The extreme east of the country just south of Dublin also remained above water and formed part of a landmass extending eastwards across the Irish Sea (Leinster Massif).

The lowest Dinantian group, the Lower Limestone Shale, is extensively developed over central-southern Ireland and is fairly uniform in character. It forms a continuation of the uppermost Devonian and consists of interbedded sandstones, siltstones, shales and a few thin limestones. The group grades upwards into the Lower Limestone which also is extensive
### TABLE 1

Generalised Lower Carboniferous or Dinantian succession in Ireland.

<table>
<thead>
<tr>
<th>Group*</th>
<th>Epoch</th>
<th>Thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper Limestone</td>
<td></td>
<td>1500 ft.</td>
</tr>
<tr>
<td>Middle Limestone (Calp)</td>
<td>Visean</td>
<td>up to ca. 800 ft.</td>
</tr>
<tr>
<td>Waulsortian Limestone</td>
<td>- - -</td>
<td>0-3000 ft.</td>
</tr>
<tr>
<td>Lower Limestone</td>
<td>Tournaissian</td>
<td>600-800 ft.</td>
</tr>
<tr>
<td>Lower Limestone Shale</td>
<td></td>
<td>ca. 150 ft.</td>
</tr>
</tbody>
</table>

*"group" is a term used locally but not in the strict sense."
**TABLE 1**

Generalised Lower Carboniferous or Dinantian succession in Ireland.

<table>
<thead>
<tr>
<th>Group*</th>
<th>Epoch</th>
<th>Thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper Limestone</td>
<td></td>
<td>1500 ft.</td>
</tr>
<tr>
<td>Middle Limestone (Calp)</td>
<td>Visean</td>
<td>up to ca. 800 ft.</td>
</tr>
<tr>
<td>Waulsortian Limestone</td>
<td></td>
<td>0-3000 ft.</td>
</tr>
<tr>
<td>Lower Limestone</td>
<td>Tournaissian</td>
<td>600-800 ft.</td>
</tr>
<tr>
<td>Lower Limestone Shale</td>
<td></td>
<td>ca. 150 ft.</td>
</tr>
</tbody>
</table>

* "group" is a term used locally but not in the strict sense.
and fairly uniform. It is composed mainly of well bedded, dark bioclastic limestones with some interbedded black shales. Cherty beds are common in its uppermost part, and form the base of the stratiform sulphide mineralization of the Mogul Upper orebody.

The overlying Waulsortian Limestone is of importance because most of the Mogul Upper orebody occurs in its lower part and it also contains the greater part of the Tynagh orebody. The group has been studied in detail by Lees (1961; 1964). He regards it as a carbonate mudbank complex and discourages continuation of the long-used term "reef". It is analogous to a reef only in that geographically it occupies a broad belt separating the deep trough on the south from a more lagoonal type of basin to the north (Fig. 3). Unlike a true reef it did not form a wave-resistant barrier and did not have a rigid structure.

The main part of the group is formed by a complex of coalescing calcilutite mounds with diameters in the order of hundreds of yards and heights probably not exceeding 50 ft. above the surrounding sea floor. The mounds contain abundant fenestellid bryozoans and local pockets with a rich shelly fauna and are thought to have been supported by a framework of plants which are not preserved. Secondary porosity was commonly developed in the mounds at an early stage and the cavities formed were then filled by geopetal calcite muds or later sparry calcite ("reef tufa").

The base of the Waulsortian Limestone is diachronous and somewhat younger in the south-west. Its thickness is
Fig. 3. A generalised map showing the distribution in Ireland of the Waulsortian limestone and other lithofacies at the close of Waulsortian times. (After Lees, 1961).
variable ranging from over 3000 ft. west of Limerick to less than 500 ft. in the Silvermines district (Fig. 4). The top of the group is marked by a return to normally stratified limestones and in the transitional zone thin interbedded cherts are widely developed.

The overlying Middle Limestone is widespread and consists of dark to black, massively bedded limestones containing a large amount of argillaceous and carbonaceous impurity. The Upper Limestone is a uniform, compact, well bedded limestone with a widely varying fossil content.

(c) Lower Carboniferous Igneous activity.

Several minor igneous centres were active in Ireland during the Dinantian and, in the southern half of the country at least, these were confined to the Visean (Charlesworth, 1964). Extrusives, including lavas and tuffs, predominated over intrusives.

The most important of these centres occurs in Co. Limerick 20 miles south-west of the Silvermines district (Ashby, 1939). There were two phases of explosive extrusive activity, one at the top of the Waulsortian Limestone and the other at the top of the Middle Limestone. The volcanics have an alkaline affinity and include ultra-mafic picrite-basalts, olivine-basalts, analcite-basalts, trachybasalts, trachyandesites and trachytes. Tuffs and agglomerates were abundantly developed during both phases. Associated intrusives are of minor importance.

Igneous activity is associated with both the Gortdrum and Tynagh orebodies. At Gortdrum, thin altered feldspar-porphyry
Fig. 4. An isopach map of the Waulsortian limestone. The thicknesses are in feet and are approximate. The dashed lines are less accurate than the solid lines. The dotted lines represent approximate positions of coast lines at the end of Waulsortian times. (After Lees, 1961).
dykes of uncertain age occur in the zone of epigenetic mineralization (Thompson, 1967) and at Tynagh, tuff and iron-formation are laterally equivalent to the ore-bearing Waulsortian Limestone (Derry et al., 1965; Schultz, 1966). It has also been reported that a shale horizon at the top of the Lower Limestone at Silvermine is tuffaceous (J. Gordon-Smith in a personal communication to Snelgrove, 1966).

(d) Post-Lower Carboniferous.

At the end of the Dinantian the sea withdrew from most of Ireland and the Upper Carboniferous is represented only by localised brackish water, clastic sediments and coal measures. Sediments of Post-Carboniferous age (Pleistocene excepted) do not occur in the southern and central parts of Ireland.

Of some significance to the present study are the effects of the Hercynian orogeny which occurred at the end of the Carboniferous. The zone of intense folding affected only the extreme south part of Ireland where strong east-west folds were developed in Carboniferous and Devonian sediments. North from this belt the effects of folding decreased fairly rapidly and its trend changed to E.N.E. to coincide with the Caledonian trend in the basement. In the central part of Ireland the main effect of the Hercynian folding combined with the effects of normal faulting was to produce in the Carboniferous cover a number of periclinal structures with cores of Lower Palaeozoic and Devonian rocks which now form highland areas. There is evidence (see later) to show that these periclines with associated normal faulting originated in the Lower
Carboniferous but there is no published evidence to prove that they rose enough to form islands in the Lower Carboniferous sea. Derry et al. (1965) believed they were in existence as topographic highs during the Devonian and that Devonian and later sediments were "draped over" them. However Schultz (1966a) later rejected this because he thought the highs were formed entirely by post-Carboniferous folding. Schultz's idea was not accepted by Derry (Derry, 1966).

The Hercynian folding was unaccompanied by igneous activity in Ireland (Charlesworth, 1964, p. 322) but in S.-W. England the Cornwall Granite and associated mineralization is of this age. In spite of the lack of igneous activity many minor Irish copper-lead-zinc occurrences are attributed to Hercynian "magmatic solutions from deep-seated sources" (Charlesworth 1964, p. 323).

**Geological setting in the vicinity of the Mogul mine.**

(a) Succession.

The Silvermines district lies on the fault-bounded northern edge of one of the largest of the periclinal inliers in southern Ireland. (Fig. 5). The inlier is irregular but elongated in an E.-W. direction and forms a highland ridge. It has a core of highly folded Silurian rocks surrounded, and at its highest point (over 2000 ft. above the surrounding plain), capped unconformably by Upper Devonian sandstones. In contrast the Carboniferous sediments, conformable on the Devonian, never occur at great elevations on the pericline. Whether this is due to erosion or non-deposition is uncertain.
Fig. 5. Geological map of the central part of Ireland.
Because of the lack of outcrop, information regarding the succession in the mine area has been gained largely from exploration drill core and to a smaller extent from underground workings and the Magcobar open pit mine. The succession has been previously described by Weber (1964; 1968) and thin sections of the various formations have been examined and described (private company report).

In the vicinity of the mine, rocks range from Silurian to Lower Carboniferous in age and a generalized succession is given in Table 2.

The distributions of the formations on surface and in section are shown in Figures 6 and 7.

The Silurian is represented by a series of highly folded mudstones, slates, siltstones with some intercalated coarser grained sediments. They are usually grey in colour but may be greenish or purplish.

The overlying Devonian sequence consists mainly of sandstones or locally quartzites, but also contains some conglomerates and greenish or greyish shales. They are thought to have been deposited in a coastal marine environment (Charlesworth, 1964, pp. 182-184; Weber, 1968; Rhoden, 1958).

The Basal Fragmentals is equivalent to the "Mixed Beds" of Rhoden or more generally the Lower Limestone Shale. At the Mogul mine it consists of alternating marine carbonaceous shales, calcareous siltstones, sandstones and thin limestones. Bioclastic debris is abundant and consists of fragments of crinoids, bryozoans, brachiopods, molluscs, ostracods, sponge spicules and corals.
**TABLE 2**

Succession in the vicinity of the Mogul mine.

<table>
<thead>
<tr>
<th>Formation*</th>
<th>Thickness.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Calp Limestone</td>
<td>?</td>
</tr>
<tr>
<td>Chert Horizon</td>
<td>100-200 ft.</td>
</tr>
<tr>
<td>Dolomite Breccia (Upper orebody)</td>
<td>150-350 ft.</td>
</tr>
<tr>
<td>Muddy Reef.</td>
<td>0-100 ft.</td>
</tr>
<tr>
<td>Muddy Limestone</td>
<td>0-650 ft.</td>
</tr>
<tr>
<td>Lower Dolomite (Lower orebody)</td>
<td></td>
</tr>
<tr>
<td>Basal Fragmentals</td>
<td>0-130 ft.</td>
</tr>
<tr>
<td>Devonian sandstones etc.</td>
<td>0-300 ft.</td>
</tr>
<tr>
<td>Silurian slates etc.</td>
<td>?</td>
</tr>
</tbody>
</table>

* The formation names are those that have been used by the mine geologists and which have been used in previous publications.
Fig. 6. Geological map of the vicinity of the Mogul mine showing the sub-surface positions of the various orebodies.
Fig. 7. Generalised north-south section through the Upper and Lower orebodies.
The two overlying formations, the Lower Dolomite and Muddy Limestone together form the equivalent of the Lower Limestone of other localities. This is a well bedded, dark, partly argillaceous, bioclastic limestone. The numerous fossil fragments include mainly crinoids but also bryozoans, foraminifera, ostracods and gastropods. Complete dolomitization has occurred in the lower half of the group (Lower Dolomite) with an accompanying increase in grain size. The dolomitization occurs only in the Silvermines district; otherwise the group characteristics are fairly consistent over a large area.

The Muddy Reef is a poorly bedded, nodular, fine-grained limestone with some crinoid remains. Chert becomes increasingly common upwards and occurs in nodules or as a replacement of fossils or matrix. Near the top there is a fairly widespread stratiform horizon composed entirely of chert up to about 20 ft. thick and above it a sporadically developed black or greenish shale. The shale is commonly interbedded with sulphides, mainly pyrite. Locally, bedded sulphides composing the base of the Upper orebody extend well down into the Muddy Reef.

The overlying Dolomite Breccia, containing the Upper orebody in its lowest part, is the lateral equivalent of the Waulsortian Limestone of other areas. It occupies an elliptical area, greater than 1 mile in length (Weber, 1968), centered over the Upper orebody and extending east over the B zone. The Dolomite Breccia is a soft-sediment slump breccia composed mainly of variably sized, angular and rounded
fragments of Waulsortian limestone with crinoid fragments in a fine carbonate matrix. The lateral extent of brecciation is greatest in the lower part and this, together with complexity of brecciation, tend to decrease upwards so that the lower peripheral parts of it are overlain by Waulsortian limestone. The Dolomite Breccia presents a massive or poorly stratified appearance in outcrop though closer examination reveals good evidence of bedding. A more detailed description of its sedimentary features is given in a later section.

The upper part of the Dolomite Breccia is interbedded with chert and finally passes up into a massive chert breccia, the Chert Horizon. The extent of the main brecciated chert is less than that of the Dolomite Breccia and it is also thinner. It does not extend as far east as the B zone but in many more distant areas thin stratiform cherts occur at the same horizon.

The Chert Horizon is succeeded by a black carbonaceous and argillaceous limestone, the Calp Limestone, which is not exposed in the mine workings or on surface. It is probably the equivalent of the Middle Limestone of other areas.

(b) Structure.

The Carboniferous strata lie on the southern limb of a broad syncline plunging gently towards the E.N.E. Close to the Silvermines Fault the northerly dip of the strata has been accentuated and reaches 60°. This dip rapidly decreases northwards from the fault to about 10°.

As shown in figure 7, all Carboniferous formations below the Dolomite Breccia pinch out southwards close to
the fault plane. This could be explained either by post-depositional sliding along bedding planes during uplift of the footwall fault block or by non-deposition south of the fault due to uplift contemporaneous with sedimentation.

Neither mechanism can be proved with certainty but evidence tends to favour the latter. Assuming the former to have occurred one would expect abundant evidence of slippage along bedding planes and folding and thickening where dips flatten to account for the thinning nearer the fault plane. Movement along bedding planes is clearly evident but only very close to the fault, and where the dip flattens the beds are undeformed and not thickened.

If the second alternative is assumed and the southern fault block was uplifted to form an island one would expect a facies change on approach to the fault line. This does not occur, with the possible exception of the green shale at the top of the Muddy Reef, which does contain abundant fragmental quartz and indicates possible near-shore conditions (Fig. 8). Derry et al. (1965) favour the idea that the inliers in central Ireland formed topographic highs during the Lower Carboniferous and it is possible that they were in the process of uplift but rarely got above sea level. The soft sediment slumpage in the Dolomite Breccia certainly points to contemporaneous fault activity.

The Silvermines Fault forming the northern boundary of the pericline is one of a series of normal faults with an E.N.E. trend. It appears from beneath the Shannon Estuary and extends from there to just east of the Silvermines area,
Quartz grains (light grey) in shale bed at the top of the Upper orebody. Polished section.
a distance of 25 miles. Its maximum downthrow (to the north) around Silvermines is 1100 ft. (Rhoden, 1958) and at the Mogul Mine its downthrow is 650 ft. (Weber 1968). Evidence from exploration drilling indicates that it is mainly a slump plane at the base of the Carboniferous and that it dies out at depth but the fact that its trend coincides with the Caledonoid trend suggests that it is associated with an underlying zone of weakness (Weber, 1968).

As already pointed out, the work of Rhoden (1958) shows that the Silvermines Fault zone is made up of a complex of shears, the two most important of which are E.S.E. and E.N.E. normal faults with northerly downthrows. Recent exploration drilling at the Mogul mine confirms this view. In the vicinity of the mine the E.S.E. trend is dominant and the main fault zone trends in that direction. However E.N.E. trending faults are still present and the combination of the two leads to a "saw-tooth" fault-trace, the details of which are not mapped. Both east and west of the mine the overall fault trend reverts to the normal E.N.E. direction.

(c) Positions of the Mogul and Magcobar orebodies.

The stratigraphic positions of the Upper and Lower orebodies are shown in Figure 7. The Lower orebody, mainly the vein type deposit, is best developed in the Lower Dolomite close to the fault. As the formation swings away from the fault the mineralization dies out gradually. The vein-type mineralization becomes more intense upwards and merges with the massive sulphide of the Upper orebody. The upper orebody itself is stratiform and most of it is contained within
the base of the Dolomite Breccia. A small ore zone of minor
significance the No. 3 zone, occurs stratabound in the Dolo-
mite Breccia about 50 ft. above the Upper orebody.

The B zone, also a stratiform body of massive sulphide,
occurs at the same stratigraphic level as the Upper orebody
at the base of the Dolomite Breccia, but at a much greater
distance from the fault. The K zone, a second vein-type
lead-zinc orebody, occurs in the Lower Dolomite. The Mag-
cobar barite deposit is a stratiform body, also at the base
of the Dolomite Breccia, but at a higher topographic level
than the two sulphide bodies.
CHAPTER 3
THE G ZONE OREBODIES: DIMENSIONS AND METAL DISTRIBUTIONS

Dimensions.

The limits of the Upper orebody follow fairly closely the limits of the massive sulphide body. The orebody has a strike length of about 2500 ft. and a maximum down-dip length of about 2000 ft. The south-west and south-east margins are straight and fairly well defined but the northern boundary is more poorly defined and has many tongues of ore protruding northwards (Fig. 9). The isopachs also reveal north-trending tongues within the orebody. The orebody is thickest in the south with a maximum thickness of 84 ft. and thins out gradually northwards.

The outlines of the Lower orebody depend on the intensity of sphalerite-galena veining in the Lower Dolomite and it is thus irregular in shape. Its greatest development is however, under the thickest part of the Upper body and it thins out more rapidly to the east than to the west.

Metal distributions.

Using assay results from the exploration drill core the distributions of lead, zinc and silver were plotted in plan and vertical section. Different metal ratios were plotted graphically for both orebodies. The exploration holes were
Fig. 9. Isopach map of the Upper orebody. The thicknesses are in feet.
drilled at intervals of 150 ft. or less along north-south lines spaced at 100 ft. intervals. The assays are averages for five foot core lengths in and peripheral to the ore zones. Figures for iron are not available.

In the Lower orebody as a whole, lead, 4.5%, predominates over zinc, 3.4%, and silver averages 1.13 oz. per ton. It contrasts with the Upper orebody (2.4% Pb, 9.2% Zn, 0.74 oz./ton Ag) in having a much higher Pb:Zn ratio. Its total lead value is 75% greater and silver 55% greater than in the Upper orebody.

(a) Lower Orebody.

In general the greatest lead values in the Lower orebody occur at depth and they decrease upwards. Zinc values tend towards the reverse trend. Figure 10 shows plots of the Zn:Pb ratio against depth for different north-south sections. The ratios were calculated using average figures for each ore intersection in each section. In most sections the Zn:Pb ratio increases upwards. When the values are plotted on a much more detailed scale, for example for average ratios over five foot intervals, the curves are extremely irregular showing that zinc enrichment can occur over short distances at depth. The increase in the Zn:Pb towards surface is probably brought out best when the average ratios for ore intersections in all sections are plotted against depth on one graph (Fig. 11). When lead values are plotted against those for zinc, a distinct antipathetic relationship is brought out (Fig. 12).

A plot of lead versus silver values shows a good direct
Fig. 10. Graphs showing variation of the Zn : Pb ratio with depth in the Lower orebody. Each graph is constructed using assay values from ore intersections in a north-south line of drill holes. The graphs are numbered according to the distance in feet of each such line from the west end of the orebody. Depths are shown in terms of mine plan elevations and on the same scale surface level varies between 5300 and 5500 feet.
Fig. 11. Variation of the Zn : Pb ratio with depth in the Lower orebody. Constructed using assay values of drilled ore intersections from all parts of the orebody. On the same vertical scale surface elevation varies between 5300 and 5500 feet.
Fig. 12. Variation diagram showing relationship between zinc and lead in the Lower orebody. Constructed using 53 points.
proportional relationship (Fig. 13) and as would therefore be expected a plot of zinc versus silver shows an antipathetic relationship (Fig. 14).

Very small copper values have also been found in the Lower orebody and most of these occur near its western end and in its deepest parts.

(b) Upper orebody.

The greatest base metal concentration in the Upper orebody occurs in the southern part coinciding with the area of greatest thickness. Plans showing the distribution of lead, zinc and silver are shown in Figures 15, 16 and 17 respectively.

Lead is most strongly concentrated in the southern part with a maximum of over 9% and it drops off fairly rapidly northwards though a few other local high spot do occur. Zinc is high in the area corresponding to high lead but increases northwards and reaches its greatest concentration, up to 22%, in a north-east trending belt north of the lead maximum. North of this, zinc values fall off but not as rapidly as do the lead values. The main silver concentration corresponds fairly well with that of galena but values fall off even more rapidly northwards than they do for galena. Some localised spot highs for silver in the northern part of the orebody do not correspond to lead spot highs.

The contoured distributions of lead and silver (Figs. 15 and 17) show some evidence of north trending highs and lows. These however are not so apparent in the zinc diagram (Fig. 16) and in the centre of the orebody they are
Fig. 13. Variation diagram showing relationship between lead and silver in the Lower orebody. Constructed using 53 points.

Fig. 14. Variation diagram showing relationship between zinc and silver in the Lower orebody. Constructed using 53 points.
Fig. 15. Plan showing the distribution of lead in the Upper orebody. Constructed using average values for complete ore intersections. Contoured at 1, 2, 3 and 5% Pb.
Fig. 16. Plan showing the distribution of zinc in the Upper orebody. Constructed using average values for complete ore intersections. Contoured at 5, 8, 12 and 16% Zn.
Fig. 17. Plan showing the distribution of silver in the Upper orebody. Constructed using average values for complete ore intersections. Contoured at 0.3, 0.6, 1.0 and 2.0 oz. per ton Ag.
overprinted by a fairly strong north-easterly trend.

The distributions of lead, zinc and silver were plotted in several north-south vertical sections through the Upper ore-body. This revealed some general trends, none of which is strongly developed.

In a section 400 ft. east of the west end of the body lead is more concentrated in the lower part and zinc in the upper part (Fig. 18). Silver clearly follows the lead in its distribution. In a section 1400 ft. from the west end, i.e. near the centre of the orebody, the relationships are more complex but zinc is still in greater concentration in the upper part (Fig. 19). Lead is high in the lower part but also becomes fairly high again near the top (Fig. 20). High silver values accompany the lower lead but lower silver occurs at the top (Fig. 21). At the east end of the orebody (2100 ft. from the west end), there is no clear zoning of the metals and the distribution of silver follows lead only to a small extent (Fig. 22).

Graphical plots of Pb/Zn, Zn/Ag and Pb/Ag were also made for the Upper orebody using drill core assay results averaged over total ore intersections (Figs 23, 24 and 25). The Pb/Zn and Zn/Ag plots do not show any clearly defined relationship in contrast with the fairly distinct antipathetic relationships shown by the same plots for the Lower orebody (Figs. 12 and 14). The Pb/Ag plot shows a direct proportional relationship but it is not so well defined as in the same plot for the Lower orebody (Fig. 13). One feature of interest, evident on comparing the two Pb/Ag graphs, is that
Fig. 18. North-south section across the Upper orebody showing the distribution of lead, zinc and silver. Section line located 400 feet east of the west end of the orebody. Length of section 600 feet. Vertical scale greatly exaggerated.
Fig. 19. North-south section across the southern part of the Upper orebody showing the distribution of zinc. Section line located 1400 feet east of the west end of the orebody. Length of section 600 feet. Vertical scale greatly exaggerated.
Fig. 20. North-south section across the southern part of the upper orebody showing the distribution of lead. Section line located 1400 feet east of the west end of the orebody. Length of section 600 feet. Vertical scale greatly exaggerated.
Fig. 21. North-south section across the southern part of the Upper orebody showing the distribution of silver. Section line located 1400 feet east of the west end of the orebody. Length of section 600 feet. Vertical scale greatly exaggerated.
Fig. 22. North-south section across the Upper orebody showing the distribution of lead, zinc and silver. Section line located 2100 feet east of the west end of the orebody. Length of section 800 feet. Vertical scale greatly exaggerated.
Fig. 23. Variation diagram showing relationship between zinc and lead in the Upper orebody. Constructed using 149 points.
Fig. 24. Variation diagram showing relationship between zinc and silver in the Upper orebody. Constructed using 149 points.

Fig. 25. Variation diagram showing relationship between lead and silver in the Upper orebody. Constructed using 149 points.
the Ag: Pb ratio is higher in the Upper than in the Lower orebody, although, because the Lower body is so much richer in galena it has a higher total silver content.

Copper which occurs locally in measurable amounts in the Lower orebody is not measurable in the Upper body but copper-bearing minerals have been detected microscopically in its lower part near the southern end.

In summary lead and zinc in the Lower orebody show an antipathetic relationship and zinc increases upwards. Silver shows a distinct direct proportional relationship to lead. In the Upper orebody lead and silver are concentrated near the southern end, i.e. near the contact with the Lower orebody, and they also have a tendency to concentrate near the base. Zinc is concentrated farther north and preferentially in the upper part. The antipathetic Pb/Zn and Zn/Ag relationships do not occur. The sympathetic Pb/Ag relationship is not as strong as in the Lower orebody and there appears to be a tendency for it to break down northwards and eastwards from the south-west margin.
CHAPTER 4
THE UPPER OREBODY AND ASSOCIATED MINERALIZATION

General aspects of the orebody and host rocks.

The host rocks of the Upper orebody contain many features, and in particular sedimentary structures, which are important in determining the origin of the body. These have been examined and a comparison with structures in the orebody itself shows that the orebody is an integral part of the sedimentary succession.

(a) Horizons below orebody.

The upper part of the Muddy Reef just below the topmost shale and chert beds contains some concretionary sulphide, almost exclusively pyrite. The concretions have a flattened spheroidal shape and vary in size up to about two inches in diameter. They are either simple and composed entirely of pyrite or compound with a limestone centre and an outer pyrite shell. In the same part of the succession chert occurs commonly either in nodules, as a replacement of fossils or in the form of a matrix also possibly a replacement. Rocks in this part of the succession show very little deformation either of a sedimentary or tectonic nature.

The overlying chert horizon also shows little sign of deformation but, in the shale above, brecciation occurs and is probably penecontemporaneous with sedimentation and due to
slumping. This marks the base of a succession of several hundred feet, including the Upper orebody, which has been strongly brecciated mainly during sedimentation, and in which brecciation is the most characteristic feature.

As already mentioned the shale below the massive sulphide shows a discontinuous distribution, possibly due to erosion, and it contains several features of significance in determining the formation of the orebody. The thin greenish bed contained within the shale is composed of rounded to angular quartz grains 0.02 to 0.05 m.m. in diameter in a sericitic matrix. Its material may have been derived from erosion of Devonian sandstone uplifted on the southern side of the fault, and it is the only bed in the brecciated succession containing detritus derived from significantly older rocks, i.e., the detritus in the rest of the succession is derived from penecontemporaneous sediments.

Parts of the shale contain sulphide and in less deformed parts this occurs as laminae interbedded with the shale. Pyrite, sphalerite and galena all occur in significant quantities and in fact form ore in the upper part. Figures 26 and 27 show sphalerite interbedded with fine dolomitic shale. The sphalerite laminae have a thickness comparable with laminae in the shale and they are deformed to the same extent as the shale by minor slump folds and pre-consolidational faults.

Figure 28 shows a narrow band rich in fine-grained galena which is contorted on a small scale although difficult to show photographically because of lack of contrast. Figure 29
Fig. 26. Finely interbedded sphalerite (light grey) and shaly dolomite showing pre-consolidational faulting.

Fig. 27. As fig. 26 but showing soft-sediment folding in addition to faulting.

Fig. 28. Fine-grained galena showing soft-sediment contortions.

Fig. 29. Photo-micrograph of part of fig. 28 showing the platy habit of small galena grains (light grey) and the curving of galena-clay laminae around larger pyrite clasts.
illustrates part of a polished section of the same specimen and shows laminae of shale and galena curving around a pyrite fragment.

Finely interbedded pyrite is common and a specimen containing folds is shown in Figure 30.

Shales close to the base of the orebody show abundant evidence of deformation penecontemporaneous with deposition. They contain fragmental in addition to interbedded sulphide. The occurrence of both together is illustrated in Figure 31 where finely interbedded sphalerite and shale laminae are highly contorted by soft sediment slumping and also bent around what appear to be pyrite clasts.

Unlaminated layers of shale with unsorted sulphide and other fragments also occur just below the orebody (Figs. 32, 33 and 34). The fragments are composed dominantly of pyrite or dolostone and less commonly of fine grained sphalerite. Galena fragments do not occur but small amounts of galena are contained in some of the other sulphide fragments. Fragment outlines vary considerably and are usually very irregular. Some are angular but others are rounded in places with diffuse boundaries indicating various stages of consolidation at the time of fragmentation. Rarely do the fragments exceed 1 cm. in diameter but an exception in the form of a large fine-grained sphalerite fragment is shown in Figure 34.

Figure 33 shows an unbroken pyrite spheroid with a radiating texture and also a dolostone fragment with a pyrite encrustation. These features suggest diagenetic growth of pyrite after brecciation. They occur only rarely below the orebody but are much more common in and above it.
Fig. 30. Folded, interbedded pyrite and shale just below the base of the Upper orebody.

Fig. 31. Finely interlaminated sphalerite (lightest grey) and shale distorted by larger clasts of pyrite. Large dolostone fragment in upper left-hand corner.

Fig. 32. Sphalerite (sp) and other fragments in a dark clay matrix. Base of Upper orebody.

Fig. 33. Unsorted pyrite and dolostone fragments in a dark clay matrix. Base of the Upper orebody.
**Fig. 34.** Shale enclosing large fragments composed mainly of sphalerite. Base of Upper orebody.

**Fig. 35.** Solitary coral enclosed in a pyrite bed in interbedded pyrite and shale at the base of the Upper orebody.

**Fig. 36.** Section through an elliptical sulphide "ball" showing fairly straight internal laminations traversing it. Base of Upper orebody.

**Fig. 37.** Gradational contact from shale below to massive sulphide of the Upper orebody. Width of field of view about 4 feet.
Fig. 34. Shale enclosing large fragments composed mainly of sphalerite. Base of Upper orebody.

Fig. 35. Solitary coral enclosed in a pyrite bed in interbedded pyrite and shale at the base of the Upper orebody.

Fig. 36. Section through an elliptical sulphide "bail" showing fairly straight internal lamination: traversing it. Base of Upper orebody.

Fig. 37. Gradational contact from shale below to massive sulphide of the Upper orebody, width of picture in view about 3 feet.
In many instances the pyrite-rich beds contain fossil fragments, particularly crinoids and solitary corals (Fig. 35). The fossils are often converted to dolomite and partially infilled by pyrite but pyrite does not replace shell material.

Close to the main fault where the shale is steeply dipping it displays a strong fissility parallel to bedding planes probably related to post-consolidational movement on the fault. Another feature of possible tectonic origin in the same region is the occurrence of sulphide "balls". These are well rounded, sub-spherical pieces of massive sulphide varying in size from a few inches to three feet in diameter. They have a smooth and sometimes polished outer surface and are enclosed by shale. Their internal structure (Fig. 36) resembles that in the overlying massive sulphide and is not related to their external shape showing that they are not concretions but probably blocks of sulphide broken from the overlying body and rolled and rounded off in the shale during slippage associated with faulting.

(b) The Orebody.

The Upper orebody can be termed a "massive sulphide" and a large part of it is 80-100% sulphide. The concentration of sulphide is greatest in the south-central part and tends to drop off slightly with diminishing thickness towards the edges. From marginal areas outwards beyond ore limits there is a fairly rapid decrease in the content of sulphide.

The lower contact is variable in character. In some places the massive sulphide rests in sharp contact with
underlying rocks almost devoid of sulphide and in others the amount of sulphide in the shales beneath increases upwards to merge into massive sulphide (Fig. 37).

The stratigraphic level of the base also varies. In the southern part the base is coincident with the base of the Dolomite Breccia but northwards it transgresses upwards through about 20 ft. In the central-western part thin massive sulphide bands occur well down in the Muddy Reef and indicate early development of a small basin undergoing intermittent sulphide deposition.

Brecciation is a dominant feature in the orebody. Although most of the breccia is unstratified traces of stratification occur in it at the base (Fig. 38). Breccia fragments vary from rounded with diffuse outlines (Fig. 39) to angular (Fig. 40) and vary in size from microscopic to, in one instance, 2 ft. in diameter. They are composed of pyrite or sphalerite or aggregates of several minerals but, as in the shale beneath, they are never composed of galena alone. There is also evidence of multiple brecciation, i.e. fragments composed of cemented smaller fragments (Fig. 41). Over the greater part of the orebody the fragments are embedded in fine-grained sulphides sometimes with admixed dolomite, clay or chert (Fig. 42).

Evidence that some of the sulphide was deposited originally as fine laminae can be seen in some of the fragments and in a few localities throughout the orebody small pockets of finely laminated sulphide, including pyrite, sphalerite and galena, occur in an only slightly deformed state (Fig. 43).
Fig. 38. Stratified sulphide breccia at the base of the Upper orebody. Length of pencil (lower centre) 3 inches.

Fig. 39. Rounded pyrite fragments in breccia illustrated in fig. 38.
Fig. 38. Stratified sulphide breccia at the base of the Upper orebody. Length of pencil (lower centre) 3 inches.

Fig. 39. Rounded pyrite fragments in breccia illustrated in fig. 38.
Fig. 40. Angular pyrite fragments in associated massive sulphides. 3.5 cm. of core.

Fig. 41. Larger angular pyrite fragments in general of smaller angular pyrite fragments. 3.5 cm. of core.

Fig. 42. Fragmental pyrite in a matrix of sphalerite and dolomite.
Fig. 40. Angular pyrite fragments in brecciated massive sulphide. Upper orebody.

Fig. 41. Large pyrite fragment composed of smaller angular pyrite fragments.

Fig. 42. Fragmental pyrite in a matrix of sphalerite and dolomite.
Fig. 43. Finely interlaminated sulphides, mainly pyrite, with some interbedded clay rich laminae (black). Upper orebody.

Fig. 44. Finely laminated pyrite, sphalerite and dolomite distorted by larger, colloform pyrite clasts. Where laminations are most constricted between clasts (lower left) only pyrite remains. Upper orebody.
Fig. 43. Finely interlaminated sulphides, mainly pyrite, with some interbedded clay rich laminae (black). Upper orebody.

Fig. 44. Finely laminated pyrite, sphalerite and dolomite distorted by larger, colloform pyrite clasts, where laminae are most constricted between clasts (lower left only pyrite remains). Upper orebody.
Deformation of finely inter-banded pyrite and sphalerite between fragments of colloform pyrite is shown in Figure 44. Where these bands are most constricted between fragments, sphalerite bands pinch out. Localised thin beds of black shale or very fine grained clastic dolostone also occur in the orebody.

Apart from brecciation and fineness of grain, which characterise most of the orebody, there are other varied characteristics recognisable in hand specimen, but which show no consistent areal or vertical distribution. Parts of the ore are very compact, structureless and extremely fine grained. They are composed of pyrite and smaller amounts of galena and sphalerite. Relative proportions of the minerals affect slightly the overall colour of the hand specimen which otherwise resembles a pyrite aggregate.

Other parts are composed of slightly brecciated colloform sulphides with pyrite predominating. This type of ore contains many small cavities and $H_2S$ is released from these when they are cut open. Colloform sphalerite-galena-rich ore also occurs. It is however much more compact and non-porous.

Another common ore type consists of fragmental or colloform pyrite and sphalerite embedded in a black consolidated mud, rich in very fine grained sphalerite. Around the lateral margins of the orebody similar fragments and admixed dolostone fragments occur in a fine-grained dolomite matrix.

The upper contact of the orebody is usually much more sharply defined than the lower contact. It is overlain either directly by brecciated dolostone or by a thin shale band
and then brecciated dolostone (Fig. 45). A small amount of sulphide, in part bedded, occurs in the shale or dolostone for several feet above the contact but it is usually pyrite and rarely makes ore.

Parts of the contact are planar and fragments in the upper few inches of ore display lamination and flattening parallel to the contact. In such instances the ore separates readily from the overlying rock. In other parts however the contact is undulating and locally tongues of sulphide project upwards into the overlying rocks (Fig. 46). Contacts in these areas are not always sharp and sulphide, mainly pyrite, occurs in the lower few feet of the Dolomite Breccia. Where it does so it is not fragmental in nature but forms a matrix to the dolostone fragments (Figs. 47, 48). This contrasts with the sulphide below the orebody which usually occurs as fragments.

(c) Dolomite Breccia.

The greater part of the Dolomite Breccia formation consists of an unsorted accumulation of dolostone fragments with a generally massive appearance. However closer examination reveals in some places the occurrence of sedimentary structures which are of particular importance in determining the mechanism of formation of the breccia and also of the orebody.

The massive appearance in outcrop is in places broken by poorly developed bedding planes (Fig. 49) which have formed between layers differing slightly in shape and size of fragments or in matrix content. The fragments vary from angular to rounded and some have stylolitic margins. For the most
Fig. 45. Sharp contact (marked by dashed line) between massive sulphide of the Upper orebody and overlying shale. Width of field of view about 10 feet.

Fig. 46. Tongue of sulphide projecting upwards from the Upper orebody into the overlying Dolomite Breccia. Knife is on Dolomite Breccia.
Fig. 45. Sharp contact (marked by dashed line) between massive sulphide of the Upper orebody and overlying shale. Width of field of view about 10 feet.

Fig. 46. Tongue of sulphide projecting upwards from the Upper orebody into the overlying Dolomite breccia. Knife is on Dolomite breccia.
Figs. 47 and 48. Dolostone fragments (light grey) enclosed by pyrite (darker grey). Near top contact of Upper orebody.

Fig. 49. Exposure in the Magcobar open pit mine of the Dolomite Breccia showing poorly developed bedding planes dipping toward left of the picture.
part they are unsorted as regards size (Figs. 50, 51). They are nearly all composed of a fine-grained dolomitized pure limestone and although their identity is somewhat obscured by brecciation and recrystallization many of them resemble Waulsortian limestone. Neighbouring fragments may be variable or similar in appearance and locally several matching pieces of a single rock type occur close together (Figs. 53, 52). The matrix of the rock is a hardened dolomitic mud.

Thin beds of fine dolostone fragments or clay particles occur as intercalations in the Dolomite Breccia. Such bands are usually strongly deformed (Fig. 54), but in some parts they have escaped deformation and have retained depositional structures, particularly graded bedding (Fig. 55). Graded bedding in the coarser fraction is extremely rare but one example was found in which a coarse graded layer was deposited on top of, and deformed, a fine muddy layer (Fig. 56).

As in the sulphide body there is evidence of multiple brecciation in the Dolomite Breccia. Figure 57 shows part of a layer of doubly brecciated fragments and figure 58 shows a laminated breccia layer which has been fractured and disorientated.

Fossils are uncommon in the Dolomite Breccia and are usually restricted to layers devoid of large dolostone fragments. Their remains are extremely fragmentary and appear to consist mainly of crinoid ossicles. They are always converted to dolomite and due to recrystallization their boundaries and other skeletal structures are commonly very indistinct. The fragments usually lie with their long axes parallel to
Fig. 50. Angular dolostone fragments in the Dolomite Breccia.

Fig. 51. Varibly shaped fragments, mostly dolostone, in the Dolomite Breccia. Pyrite fragment marked py.

Fig. 52. Matching dolostone fragments in the Dolomite Breccia.

Fig. 53. Fractured but unseparated fragment of dolostone in the Dolomite Breccia.
Fig. 54. Contorted, fine-grained dolomite beds (bottom left of photograph) in the Dolomite Breccia.

Fig. 55. Graded beds of dolomite grains in the Dolomite Breccia.
Fig. 50. Finely laminated shaly dolomite layer, in the Dolomite breccia, resting on coarse breccia (b). Upper part of the same layer contorted by deposition of overlying coarse graded layer (a). Specimen in a fits immediately below that in b.
Fig. 57. Double brecciation in Dolomite Breccia. Dolostone fragments composed of smaller fragments.

Fig. 58. Disrupted layer of laminated breccia (outlined) in the Dolomite Breccia.

Fig. 59. Conformable and cross-cutting pyrite bands (light grey) in the Dolomite Breccia.

Fig. 60. Conformable and cross-cutting pyrite bands (black) in a bedded part of the Dolomite Breccia. Dolostone fragments in pyrite band at bottom left are slightly displaced.
bedding in the enclosing rocks. Sometimes fossil fragments are associated with pyrite and, as in units below the orebody, pyrite may fill cavities in the fossils but does not replace them.

In the vicinity of the Magcobar mine, sulphide, almost entirely pyrite, occurs in the lower part of the Dolomite Breccia and shows varying relationships to the breccia. In finely laminated parts it occurs as fine interlaminated bands but also as cross-cutting seams (Fig. 59).

When viewed in more detail (Fig. 60) these pyrite bands are seen to cross-cut each other and transgress from one bedding plane to another along minor faults. A thicker conformable pyrite band contains slightly displaced fragments of bedded dolostone.

Pyrite also occurs in other very erratically shaped pods or lenses. That for example in Figure 61 occurs in a hollow in a fairly coarse breccia layer and is overlain by a much finer grained layer. The upper part of the pyrite has radiating and concentric internal structures and a botryoidal surface. This surface is enclosed by very fine grained dolomite matrix.

Similar structures but on a much larger scale occur in the same vicinity where pyritic sulphide fills elongated troughs on the upper surface of the massive barite body. The troughs trend in a north-westerly direction and the maximum depth of sulphide found in them is about two feet.

Pyrite also occurs as small isolated spheres with a strongly radiating texture (Fig. 62) and as thin colloform
Pyrite (upper centre) filling a hollow in coarse breccia. Upper surface of the pyrite, in contact with overlying fine-grained sediment, is smoothly curved. Dolomite breccia at Magcobar mine.
1. Small pyrite spheres with barely discernible radial structure in the Dolomite Breccia.

2. Dolostone fragment (medium-grey, just left of center), with a thin encrustation of colloform pyrite (light grey). Dolomite Breccia.
Fig. 65. Pyrite grains (black) in a cavity filling in unaltered Mamlacottian limestone.

Fig. 66. Strong jointing commonly developed in massive sulphide. Upper orebody.

Fig. 67. Fault (paralleling dashed line) at top of the Upper orebody. Sulphide (light) deformed by fracturing and shale above (dark) by folding. Displacement of about 2 feet not discernible from photograph.
Fig. 5. Pyrite grains (black) in a cavity filling in unaltered waulsortian limestone.

Fig. 6. Strong jointing commonly developed in massive sulphide, Upper orebody.

Fig. 7. Fault (parallel slickensided plane at top of the Upper orebody). Sulphide alteration deformed by fracture. Sulphide in fault zone alternate by folding. Displacement at small scale not discernible from photograph.
coatings, again with a radiating texture, around dolostone fragments (Fig. 63). In one part angular pyrite fragments were found along with dolostone fragments but this is unusual in the Dolomite Breccia.

The process of dolomitization has been carried to completion in all parts of the Dolomite Breccia above the orebody with the exception of the uppermost part where fragments may have calcitic cores. A few isolated patches of unaltered Waulsortian-type limestone have also been found in the top part of the formation. These display the complex secondary cavity filling structures or "reef tufa" described by Lees (1964), (Fig. 64). Some of the cavities contain grains of pyrite (Fig. 65). Farther east, above the B zone, undolomitised breccia and unaltered Waulsortian-type limestone are more common.

(d) Post-consolidational features.

The orebody and host rocks have been affected only to a very minor extent by events, tectonic or otherwise, occurring after their final consolidation. Fissility in shale near the main fault and possibly the formation of sulphide "balls" below the orebody have already been ascribed to a tectonic origin. The only other notable feature of tectonic origin is the development of joints and faults. Most of the massive sulphide body is strongly and irregularly jointed (Fig. 66). Some joint planes are coated with a very thin layer of coarsely crystalline sphalerite and galena but otherwise jointing has had little effect on the ore. A few faults were found in the orebody but they are small and of little
significance. One of them (Fig. 67) is a normal fault with a downthrow of about 2 ft. affecting the upper contact of the orebody. The massive pyritic sulphide has clearly been fractured and displaced whereas the overlying shale has accommodated by folding. The fault plane in the sulphide is marked by a zone about an inch thick composed of finely granulated pyrite. Another fault, possibly with greater displacement, was intersected by a drift in the Dolomite Breccia well above the orebody. Along its plane there is a massive sulphide band, up to a foot wide, containing pyrite, sphalerite and galena. These sulphides are unbrecciated but contain a central plane with horizontal slickensides.

Fracturing grading into brecciation (Fig. 68) post-dating consolidation occurs in the Dolomite Breccia immediately overlying the thickest part of the massive sulphide. The breccia fragments, commonly not far removed from their original position, are angular and greatly variable in size. Their cementing medium is a coarse-grained white dolomite containing a few crystals of dark brown sphalerite. The sulphide body below the brecciated zone contains high-angle, vertically slickensided surfaces (Fig. 69) which are curved to varying degrees with respect to a vertical axis. The brecciation and slickensiding may be related in that late subsidence in the sulphide caused collapse and fracturing of the overlying Dolomite Breccia.

Mineralogy.

(a) Sulphides and Sulphosalts.
In the Upper orebody pyrite, sphalerite and galena are by far the most abundant minerals and apart from non-metals they are almost without exception the only minerals visible to the naked eye. Other sulphides or sulphosalts, which were identified microscopically or by x-ray methods are, chalcopyrite, tennantite \( \text{As}_2 \), \( \text{Cu}_2\text{S}.2\text{(Cu. Fe)}\text{S}.2\text{Sb}_2\text{S}_3 \), boulangarite \( \text{5PbS}.2\text{Sb}_2\text{S}_3 \) marcasite, bournonite \( \text{Cu}_2\text{S}.2\text{PbS}.\text{Sb}_2\text{S}_3 \) guitermanite \( \text{3PbS}.\text{As}_2\text{S}_3 \) and pyrrhotite.

Chalcopyrite and tennantite are mainly confined to the basal part of the orebody near the south end and occur as minute fracture filling in pyrite although chalcopyrite also occurs in sphalerite. Boulangarite occurs mainly as aligned growths in galena, also in the southern part of the orebody. Bournonite was found only in one specimen collected from the ore dump. It occurs as a small aggregate of crystals about 1 cm. across on the edge of a coarse-grained dolomite-filled vug in massive pyrite. Guitermanite was found only in one locality and was also associated with coarse white dolomite at the upper contact of the southern part of the massive sulphide. Marcasite occurs erratically in small amounts throughout the orebody. Pyrrhotite was found as lamellae and blebs in sphalerite along the fault zone in the Dolomite Breccia.

(b) Other minerals.

The remaining minerals which make up a small proportion of the orebody are in order of abundance, dolomite, chert, clay minerals, calcite, barite and quartz. All, except possibly quartz, are present locally in amounts sufficiently large to be identified in hand specimen. Barite has a tendency to be
Fig. 68. Late, coarse brecciation in the Dolomite Breccia just above the thickest part of the massive sulphide. Fragments cemented by coarse-grained, white dolomite.

Fig. 69. Steeply dipping slickensided surfaces in massive sulphide below area of late brecciation in the Dolomite Breccia.

Fig. 70. Fragmentary, colloform pyrite (white) in a sphalerite matrix. Upper orebody.

Fig. 71. Large, dendritic pyrite structures in a fine-grained sphalerite-dolomite matrix. Some concentric structures also present. Upper orebody.
Fig. 68. Late, coarse brecciation in the Dolomite Breccia just above the thickest part of the massive sulphide. Fragments cemented by coarse-grained, white dolomite.

Fig. 69. Steeply dipping slickensided surfaces in massive sulphide below area of late brecciation in the Dolomite Breccia.

Fig. 70. Fragmentary, colloform pyrite (white) in a sphalerite matrix. Upper orebody.

Fig. 71. Large, dendritic pyrite structures in a fine-grained sphalerite-dolomite matrix. Some concentric structure also present. Upper orebody.
purely descriptive definition of the term will be retained here but the quotation marks will be dropped.

The colloform structures in the Upper orebody may involve one or more of the following minerals: pyrite, sphalerite, galena, calcite and dolomite. Pyrite commonly occurs alone in colloform aggregates and shows a wide variety of forms which can be treated separately. Galena or sphalerite rarely occurs alone but both are intimately associated and can be considered together. Various other combinations of two or more minerals occur in colloform aggregates and these can be treated as a third group.

A variety of other textural features occurs and these are described together in one section.

Framboidal pyrite is common in the Upper orebody and according to Bastin's (1950) definition can be considered with the group of colloform structures. However because so many studies have been devoted solely to framboids and because they do not show any gradational relationships with any other colloform structures they are considered in a separate chapter.

(b) Colloform structures involving principally pyrite.

This group can be divided into six types which are in themselves fairly distinctive although there are some gradational relationships. Figure 70 shows a variety of these forms, typically fragmentary and in a sphalerite matrix.

(1) Type 1. Dendritic.

Colloform structures of the dendritic type are relatively uncommon. Their outward shape is ovoid and they
have long axes of up to about 1½ cm. in length (Fig. 71) but more commonly in the order of ½ cm. Cross sections of the structures are commonly frond-shaped (Fig. 72) and display a radiating or branching array of fibrous pyrite (Figs. 72, 73). In some instances several of the structures radiate from a central pyrite core (Fig. 74) or form in a group connected by coarser granular pyrite (Fig. 75). Usually a preferential growth direction is evident, for example, in Figure 74 those projecting upwards (with respect to the page) are most strongly developed, those projecting sideways are poorly developed and none project downwards. Because of effects of brecciation original orientation of the structures could not be determined with certainty.

(2) Type 2. Concentrically banded.

Members of this group are very common and display a wide variation in terms of external shape, internal structure, grain size and overall size. Many are single, curved spherical or hemispherical forms but others are composed of compound, coalescing, subspherical forms and have kidney-shaped outer surfaces. In size they may be up to 5 cm. across but more commonly ½ to 1 cm.

There is clearly a gradational relationship between these forms and the dendritic type as is demonstrated by the sequence of Figures 72, 76, 77 and 78 through which concentric structure becomes increasingly dominant and dendritic structure disappears. Locally, and also in common with the dendritic forms, there is a tendency towards elongation (Fig. 79) of the structures. The mechanism causing elongation
Figs. 76-78. Colloform pyrite structures, Upper orebody. Sequence illustrating decreasing dendritic and increasing concentric structure.

Fig. 79. Accentuated elongation in a concentrically banded pyrite structure, Upper orebody.
is more clearly evident in structures where it is not accen-
tuated (Fig. 80) and in which the individual bands are thic-
kest over the crest of the structure and thinnest on its flanks.
A great number of such forms observed with the highly curved
surfaces projecting upwards and in one instance an undeform-
med thin band was found containing many with flat lower sur-
faces and convex upper surfaces.

In general the most irregularly curved bands are those
near the centres of the structures and outwards the banding
becomes more smoothly curved. The centres may also be com-
posed of several separate concentric structures or even
crystals of a different mineral which are enveloped by the
outer bands (Fig. 81).

Radially crystalline pyrite is very commonly developed
in concentrically banded structures (Fig. 82) and its grain
size varies from coarse (Fig. 83) to fine. Different bands
within the one colloform structure may show radial structures
of differing grain size and some bands may have no radial
structure. In a few instances radial pyrite crystals are
separated by thinner, shorter radial crystals of sphalerite,
calcite or galena.

Closer examination of the relationship between bands
with and without radial structure (Fig. 84) shows coarse ra-
dial pyrite crystals of an inner band sharply terminated
by a band of extremely fine grained pyrite. Radial struc-
ture in pyrite of the succeeding outer band is initially fine
and radiates from point sources on the fine-grained band.

In another instance (Fig. 85) a fine-grained pyrite band
Fig. 80. Concentrically banded pyrite structure showing tendency toward elongation (upward in photograph). Structure has sharply fractured margin (left). Upper orebody.

Fig. 81. Envelopment of concentrically banded pyrite spheres by succeeding pyrite bands. Note coarse granular band which merges with the matrix pyrite. U. orebody.

Fig. 82. Radial structure in concentrically banded pyrite. Upper orebody.

Fig. 83. Coarse radial structure in a colloform pyrite structure. From the Dolomite Breccia, Magcobar mine.
Fig. 84. Enlargement of part of specimen in fig.82 showing the fine-grained band truncating inner, coarse, radial pyrite crystals. New radiating crystals nucleate on outer part of the fine-grained band. Upper orebody.

Fig. 85. Part of a concentrically banded pyrite structure with radiating crystals partly penetrating a fine-grained pyrite band (black). Upper orebody.

Fig. 86. Radiating pyrite crystals extending right through fine-grained pyrite bands (black) in a concentrically banded structure. Upper orebody.

Fig. 87. Fine-grained, concentrically banded pyrite structure with diffuse variation in grain-size indicated by colour variations. Upper orebody.
Fig. 84. Enlargement of part of specimen in fig. 82 showing the fine-grained band truncating inner, coarse, radial pyrite crystals. New radiating crystals nucleate on outer part of the fine-grained band. Upper orebody.

Fig. 85. Part of a concentrically banded pyrite structure with radiating crystals partly penetrating a fine-grained pyrite band (black). Upper orebody.

Fig. 86. Radiating pyrite crystals extending right through fine-grained pyrite bands (black) in a concentrically banded structure. Upper orebody.

Fig. 87. Fine-grained, concentrically banded pyrite structure with diffuse variation in grain-size indicated by colour variations. Upper orebody.
(etched and black) is irregular in shape and penetrated by terminating radial crystals of an inner band. The radial pyrite in the succeeding band is, as in the previous example, fine and tends to radiate from point sources on the fine grained band. Locally the radial pyrite crystals extend right through a fine-grained band (Fig. 86).

A variety of other features is also displayed by concentrically banded colloform pyrite and some of these are illustrated in Figures 87-93. Variation in grain size may not coincide with concentric layering (Fig. 87). The colloform structures can be composed of bands of cubic pyrite crystals (Fig. 88). Distinctive bands of fine grained pyrite can occur at the same relative position in several colloform structures (Fig. 89). Unusual skeletal forms of pyrite locally develop around the outer margins of the structures (Fig. 90). In fine grained parts of the dolomite breccia concentrically banded spheroids are irregular in shape and have diffuse outer boundaries (Fig. 91) or may be composed almost entirely of a polygonal pyrite aggregate (Fig. 92). Colloform pyrite may form a crustiform layer around another crystal or group of crystals (Fig. 93).

Another variations of concentrically banded structures involves both euhehedral grains and fine-grained colloform pyrite. Fine-grained, concentrically banded pyrite may encrust a euhehedral pyrite grain (Fig. 94) or may be surrounded by a euhehedral pyrite growth (Fig. 95).

(3) Type 3. Cylindrically shaped.

This rather uncommon form consists of long, cylin-
Fig. 88. Concentrically banded pyrite structure with bands containing many cubic crystals. Upper orebody.

Fig. 89. Concentrically banded pyrite structures with a similar fine-grained band (black) in each structure. Upper orebody.

Fig. 90. Concentrically banded pyrite structure with calcite crystals (black) coated by pyrite protruding from the outer surface. Upper orebody.

Fig. 91. Irregularly shaped, concentrically banded pyrite structures in the Dolomite Breccia.
Fig. 92. Concentrically banded pyrite structure with polygonal grains increasing in size outwards. Upper orebody. Etched with 1:1 HNO₃.

Fig. 93. Concentrically banded pyrite enclosing large, euhedral galena crystals. Upper orebody.

Fig. 94. Euhedral pyrite crystals surrounded by concentrically banded pyrite. Upper orebody. Etched, 1:1 HNO₃.

Fig. 95. Concentrically banded pyrite surrounded by a euhedrally zoned pyrite growth (right half of photograph). Upper orebody. Etched with 1:1 HNO₃.
Fig. 92. Concentrically banded pyrite structure with polygonal grains increasing in size outwards. Upper orebody. Etched with 1:1 HNO₃.

Fig. 93. Concentrically banded pyrite enclosing large, euhedral galena crystals. Upper orebody.

Fig. 94. Euhedral pyrite crystals surrounded by concentrically banded pyrite. Upper orebody. Etched, 1:1 HNO₃.

Fig. 95. Concentrically banded pyrite surrounded by a euhedral zoned pyrite growth (right half of photograph). Upper orebody. Etched with 1:1 HNO₃.
Fig. 96. Irregularly shaped pyrite structure with black, porous, pyrite centre. Upper orebody.

Fig. 97. Sub-spherically shaped pyrite structure with black, porous, pyrite centre. Grey areas are calcite. Upper orebody.

Fig. 98. Colloform pyrite structure composed of framboinds. Upper orebody.

Fig. 99. Part of a colloform structure composed of euhedral, radiating, pyrite crystals. Upper orebody. Etched with 1:1 HNO₃.
larger than for types previously described and they differ from banded structures with finer radial pyrite in that the radiating crystals are zoned, as revealed by etching. 

(c) Colloform structures involving sphalerite and galena. 

Colloform or other types of sphalerite when viewed using plane polarised vertical light show no internal structural features and in order to see these it was found best to use oblique reflected light. One disadvantage of this method is that it cannot be used with high power objectives which focus at a very small distance from the specimen surface. All photographs of sphalerite which show internal structure were taken in oblique light. 

Colloform textures involving sphalerite and galena are varied but cannot be classified into groups as conveniently as was done for pyrite forms. Except locally, colloform sphalerite and galena do not form discrete structures as does pyrite but instead the textures tend to merge with those of surrounding minerals (Fig. 100). 

Concentrically banded sphalerite with or without interbanded galena (Figs. 100, 101) is common. The bands however usually form as crustiform coatings on other surfaces rather than as concentric spheres, and a subordinate radial structure, probably originally fine radiating sphalerite crystals, is always developed. Several examples of radiating sphalerite were etched and in all the texture thus revealed, using vertical illumination, was a fine granular aggregate with a grain size in the order of 10 microns. 

Co-existing colloform sphalerite and galena usually
show an intricate relationship with both radiating and concentric features (Fig. 102). Another relationship which occurs is that which appears in section as one or more "Y"-shaped aggregates of galena surrounded by concentrically banded sphalerite (Figs. 103-105). Slight variations of this relationship are common but in general the texture appears as small regularly shaped galena crystals each surrounded by concentrically banded sphalerite. Examination of several sections has shown that enclosing sphalerite bears two types of relationship with the galena. In one (Fig. 106) the sphalerite fibres are aligned parallel to the faces of the galena and in the other (Fig. 107) sphalerite fibres radiate from points on the galena surfaces and especially from galena crystal terminations.

In some instances galena occurs in an elongated dendritic form and is enclosed in a layer of sphalerite (Fig. 108).

Another colloform texture developed on a much smaller scale consists of small globules of galena about 50 microns in diameter enclosed by sphalerite. The globules may be separate, coalescing or aligned in chains and many contain intricately intergrown sphalerite in their central parts (Figs. 109, 110, 111). The reverse of this relationship also occurs where globules of sphalerite, usually of smaller diameter, are enclosed by galena (Fig. 112). Although the galena shows intricate small-scale relationships with sphalerite it is now coarse-grained as is shown by cleavage traces (Fig. 113).
Fig. 100. Concentrically banded sphalerite-galena structure (left), barely discernible and merging into the surrounding mineral aggregate. Upper orebody.

Fig. 101. Concentrically banded sphalerite with a galena sphere at the centre of one of the structures. Upper orebody. Oblique light.

Fig. 102. Colloform sphalerite (grey) and galena (white) with concentric and radial texture. Upper orebody.

Fig. 103. "Y"-shaped skeletal galena crystals (black) in colloform sphalerite (white). Upper orebody. Oblique light.
Fig. 104. Enlarged view of skeletal galena crystals (black) shown in fig. 103. Upper orebody. Oblique light.

Fig. 105. Skeletal galena crystals (black) surrounded by concentrically banded sphalerite (white and grey). Upper orebody. Oblique light.

Fig. 106. Radiating, acicular sphalerite crystals, (normal to concentric colour banding) growing parallel to crystal faces of galena (black). Upper orebody. Oblique light.

Fig. 107. Acicular sphalerite crystals radiating from projections on galena crystals (black). Upper orebody. Oblique light.
Fig. 108. Dendritic, skeletal galena (white) enclosed by sphalerite. Upper orebody.

Fig. 109. Galena globules (white) in sphalerite. Some development of chain globules. Upper orebody.

Fig. 110. Compound globules of galena (light grey) and sphalerite (dark grey), enclosed by sphalerite. Upper orebody.

Fig. 111. Partly coalescent galena globules (light grey) in sphalerite. Upper orebody.
Fig. 112. Sphalerite and sphalerite-galena globules in galena. Darkest grey is dolomite. Upper orebody.

Fig. 113. Coarse cleavage, indicated by dark triangular pits, in an area of finely intergrown galena and sphalerite. Upper orebody.

Fig. 114. Layered and concretionary sphalerite (medium grey) with some small galena crystals (g) at centres of concretions. Upper orebody.

Fig. 115. Group of sphalerite globules (white) within otherwise structureless sphalerite (dark grey). Upper orebody. Oblique light.
Fig. 112. Sphalerite and sphalerite-galena globules in galena. Darkest grey is dolomite. Upper orebody.

Fig. 113. Coarse cleavage, indicated by dark triangular pits, in an area of finely intergrown galena and sphalerite. Upper orebody.

Fig. 114. Layered and concretionary sphalerite (medium grey) with some small galena crystals (light grey) at centre. Upper orebody.
Some areas of sphalerite when viewed in vertical light show a concretionary or lensoid rather than colloform texture (Fig. 114). The centres of many spherical concretions contain some galena with a granular texture and some contain angular fragments of pyrite. This type of sphalerite when viewed in oblique light appears for the most part to be structureless and also almost colourless. In a few instances however oblique light reveals small aggregates of tiny sphalerite globules each with a diameter of about 40 microns (Fig. 115), and in one section some of the lenses of structureless sphalerite are encrusted by thin layers of radiating sphalerite.

Concentrically banded colloform sphalerite usually shows a darkening in colour outwards although there are exceptions to this (Fig. 117). The boundaries between bands differing in colour are usually not sharp (Fig. 116). Commonly there is not much change in the grain size of radiating sphalerite fibres from one band to another (Fig. 116) but in a few instances the radiating habit is interrupted by thin, well defined interbands of granular sphalerite (Fig. 117). Interstices between adjacent areas of colloform sphalerite are locally filled by a coarse-grained pale coloured sphalerite (Fig. 116).

Samples of all varieties of colloform sphalerite were analysed using an X-ray powder camera and in none was there any detectable wurtzite, i.e., all showed strong sphalerite lines.
Fig. 116. Colloform sphalerite with diffuse colour banding. Coarse granular sphalerite interstitial to colloform structures. Upper orebody. Oblique light.

Fig. 117. Large scale colloform sphalerite with alternating colour bands. Upper orebody. Oblique light.

Fig. 118. Pyrite with a radiating texture surrounding sphalerite (medium grey) containing slender laths of barite (dark grey). Upper orebody.
(d) Types of Colloform textures involving other mineral groups.

(1) Type 1. Pyrite and sphalerite.

Sphalerite in the form of a small colloform structure may form the nucleus for encrusting radial or concentrically banded pyrite (Fig. 118), or it may occur interbanded with colloform pyrite (Fig. 119). Examination of the specimen in Figure 119 under oblique light shows that the sphalerite band is composed of several radiating and concentric sphalerite aggregates nucleating on the lower pyrite surface (Fig. 120).

Colloform pyrite structures when surrounded by sphalerite always display convex margins against it but in some places smaller aggregates of non-colloform granular pyrite show scalloped or concave edges against sphalerite (Fig. 121). When the two relationships (Fig. 121) are examined at higher magnification and in oblique light (Fig. 122) it can be seen that the sphalerite is colloform and radiates from points on the colloform pyrite, whereas the non-colloform pyrite occurs in the interstices between the spheres of sphalerite thus giving it a scalloped appearance. Granular pyrite showing slightly irregular concave boundaries against sphalerite with less well developed colloform texture is quite common in the orebody and leads to a typical texture as shown in Figure 123.

Delicate, small-scale intergrowths of a dendritic type occur between pyrite and sphalerite (Fig. 124). Some of the dendrites (Fig. 124) are terminated by a very thin concentric pyrite band.

In a few instances partially fragmented colloform pyrite
Figs. 119, 120. Interbanded, colloform pyrite (py) and sphalerite (sp). Identical field of view in each photograph but fig. 119 is taken in vertical light and fig. 120 in oblique light. Oblique illumination shows that the sphalerite band is composed of concentrically banded structures nucleating on the outer surface of the inner (lower) pyrite band. Upper orebody.

Fig. 121. Colloform pyrite with a dendritic texture enclosed by sphalerite with small areas of pyrite with embayed margins (lower left).

Fig. 122. Enlarged view of outlined part of fig. 121 in oblique light showing embayed pyrite (py) filling spaces between spherical sphalerite structures (sp). Upper orebody.
Fig. 123. Granular pyrite with concave margins against rounded colloform sphalerite (dark grey). Upper orebody.

Fig. 124. Dendritic, colloform pyrite (light grey) and sphalerite with very delicate outer rim of pyrite. Upper orebody.

Fig. 125. Fragmental, and probably partly replaced, colloform pyrite (light grey) in sphalerite. Upper orebody.

Fig. 126. Subhedral intergrowth of pyrite (light grey) and galena. Lightly etched with dilute HCl. Upper orebody.
occurs surrounded by light coloured, fine-grained sphalerite, which reveals no structure even in oblique light (Fig. 125).

(2) Type 2. Galena and pyrite.

Galena occurs with pyrite in concentrically banded colloform structures and shows an intricate relationship with it somewhat similar to that between galena and sphalerite (Fig. 102). A variety of other textural relationships also occurs some of which are not truly colloform. Figure 126 shows a subhedral intergrowth in which most of the galena displays straight cube faces against pyrite and in Figure 127 large crystal faces of galena in a coarse-grained aggregate are coated by a thin layer of intricately textured pyrite.

Two further distinctive but uncommon textures comprise small angular grains of pyrite enclosed by galena (Fig. 128), and small spherical aggregates of pyrite in galena (Fig. 129). In the latter the pyrite spheres are usually in the order of 25 microns in diameter but locally (Fig. 129) they have attached to them a much larger spherical structure. The small spheres always contain a small irregular crystal of galena protruding inwards from the outer edge (Fig. 129).

(3) Type 3. Calcite and dolomite associated with sulphides.

Calcite occurs quite commonly in colloform aggregates with other sulphides but rarely forms a complete colloform structure on its own. The only instances of the latter are a few tiny globules about 50 microns in diameter enclosed by sphalerite.

Colloform textures involving other minerals include
Fig. 127. Subhedral galena grains thinly encrusted by pyrite (light grey). Lightly etched with dilute HCl. Upper orebody.

Fig. 128. Angular to sub-rounded, fragmental pyrite (light grey) in galena. Lightly etched with dilute HCl. Upper orebody.

Fig. 129. Pyrite globules in galena. Note the skeletal galena crystal protruding into each globule. Lightly etched with dilute HCl. Upper orebody.

Fig. 130. Sub-spherical structure composed of calcite (dark grey) and pyrite in an irregular intergrowth. Carbonate outside the structure is dolomite. Upper orebody.
pyrite and calcite irregularly intergrown in a spherical aggregate (Fig. 130) or regularly interbanded in a concentrically banded structure (Fig. 131), and small, ovoid, calcite globules occurring in a particular band in a colloform pyrite structure (Fig. 132). Calcite also occurs in a dendritic texture with sphalerite (Fig. 133) or in a dominantly concentrically banded structure with it (Fig. 134).

Dolomite rarely occurs with a colloform texture and, where in association with sphalerite, usually takes the form of euhedral crystals with numerous inclusions (Fig. 135).

(4) Type 4. Several minerals.

Colloform textures of this type involve small compound globules consisting of two or three minerals enclosed usually by a large colloform structure of pyrite but locally by fine-grained sphalerite. Figure 136 shows colloform pyrite enclosing globules composed of galena - sphalerite and galena-calcite-sphalerite. A few of the globules are composed of sphalerite alone (Fig. 137) and a few show concentric banding on an extremely fine scale (Fig. 138). They commonly occur astride surfaces separating bands of differing grain size in the pyrite (Fig. 139) and in some instances astride the boundary between the colloform pyrite and surrounding sphalerite without any deformation of either globules or colloform pyrite. They also occur grouped in partly coalescing chains (Figs. 136, 137). The globule diameters nearly all fall in the 20-40 micron range.

An instance was found (Fig. 140) in which inwardly directed, hemispherical galena globules are encrusted by a
Fig. 131. Concentrically banded, colloform structure consisting of calcite (black) and pyrite (white). Grey matrix is dolomite. Etched with dilute HCl.

Fig. 132. Part of a concentrically banded pyrite structure containing a band of calcite (black) and sphalerite globules. Upper orebody.

Fig. 133. Dendritic calcite growth (dark grey) in sphalerite (medium grey). Upper orebody.

Fig. 134. Concentrically interbanded calcite (dark grey) and sphalerite (medium grey). White is pyrite. Upper orebody.
Fig. 135. Euhedral dolomite crystals (dark grey) enclosed by sphalerite and containing many sphalerite inclusions. Upper orebody.

Fig. 136. Concentrically banded pyrite structure containing small globules composed of sphalerite (dark grey), galena (light grey) and calcite (black). Upper orebody.

Fig. 137. Sphalerite (dark grey) and sphalerite-galena globules in pyrite. Chain structure developed by some globules. Upper orebody.

Fig. 138. Very finely, concentrically zoned sphalerite-galena globules in galena. Upper orebody.
Fig. 139. Part of a concentrically banded pyrite structure with sphalerite-galena globules cross-cutting the banding. Sphalerite dark grey. Upper orebody.

Fig. 140. Coalesced galena globules encrusted successively by sphalerite (dark grey) and pyrite. Upper orebody.

Fig. 141. Zoned, pyritohedral pyrite crystals at edge of a massive pyrite band. Upper orebody.

Fig. 142. Marginal part of above band showing coalescence of pyrite crystals which retain zoning.
Fig. 143. Granular texture commonly displayed by pyrite in the Upper orebody.

Fig. 144. Granular texture formed by coalescence of spherical pyrite structures. Upper orebody.

Fig. 145. Subhedral pyrite crystal in the Dolomite Breccia with a spherical zone of intergrown galena (dark grey). Lightly etched with dilute HCl.

Fig. 146. Grains and lenses of pyrite in shale at the base of the Upper orebody.
layer of sphalerite and the remaining space filled by pyrite. Instances have also been found in which the interspace between touching colloform pyrite structures is encrusted by calcite and then filled by sphalerite.

(e) Other textural features.

(1) Pyrite.

Some areas of massive pyrite possess a granular texture. The texture is rather abnormal in that grain-size even in one polished section is extremely variable, grain boundaries are sinuous, angles at triple junctions are highly variable, quadruple junctions are common and in some instances a small grain is completely enclosed by a larger one. A similar granular texture is developed in some concentrically banded colloform pyrite structures and superimposed on the banding.

Where small amounts of pyrite occur surrounded by dolomite in the marginal parts of the orebody it is usually in the form of small (diameter in the order of 100 microns) zoned, pyritohedral crystals. The number of zones in each crystal is almost invariably two or three. Figures 141-143, taken within one small polished section, show a gradation from isolated crystals in dolomite to massive pyrite. At the edge of the massive pyrite (Fig. 142) zoning of crystals is still clearly discernible but within it (Fig. 143) the zoning disappears and a granular texture, somewhat more regular than that described above, develops.

A granular pyrite texture also develops by the close packing and cementation of small, concentrically banded, spherical
structures (Fig. 144).

In a few instances small, euhedral or subhedral pyrite crystals in the Dolomite Breccia enclose a central spherical zone with intricately intergrown galena (Fig. 145).

In shaly horizons at the base of the orebody pyrite occurs as scattered irregular grains and in narrow lenses (Fig. 146) or more commonly as small cubic or elongated prism-shaped grains. Pyritohedral grains are uncommon in shale.

Pyrite in concretions already referred to in the Muddy Reef occurs as a granular mosaic of subhedral grains, (Fig. 147), distinct from granular textures described above, or as more isolated subhedral, cubic grains separated by calcite. In the latter case fossil fragments can be found within the concretion and they may be partly replaced by pyrite (Fig. 148). Pyrite in the concretion (Fig. 147) shows a distinct set of parallel fractures, usually in one direction only, which resembles a cleavage. This feature occurs not only in concretions but also in several parts of the Upper orebody. Commonly the fractures reaching grain boundaries are filled by sphalerite.

Pyrite also occurs in small irregular patches with a "lace-like" texture enclosing many minute particles of dolomite (Fig. 149). It occurs as minute euhedral grains forming a partial matrix around dolostone fragments in the Dolomite Breccia and is common in stylolites as frambooids and larger grains. In the massive, compact, ore previously referred to it occurs as a close-packed aggregate of framboi with a small amount of sphalerite acting as a cement.
Fig. 147. Granular mosaic of subhedral pyrite grains in a concretion in the Muddy Reef. Parallel fracturing or cleavage in most grains. Etched with 1:1 HNO₃.

Fig. 148. Curved fossil fragment in the Muddy Reef partly replaced by small pyrite grains (white).

Fig. 149. Pyrite aggregate (white) enclosing many small dolomite grains and forming a lace-like pattern. Upper orebody.

Fig. 150. Part of a concentrically banded pyrite structure fractured and veined by sphalerite (medium grey). Upper orebody.
in the porous colloform type of ore are commonly filled or lined by strongly zoned subhedral pyrite.

An apparently late fracturing affects some forms of pyrite. It produces cracks and in places slight displacement of parts of a colloform structure. The fractures are usually filled by sphalerite, (Fig. 150) galena or dolomite. Instances occur of colloform pyrite, with much interbanded calcite, fractured and veined by dolomite. On the other hand there are a few examples of colloform sphalerite fractured and veined by pyrite.

A small thickness of unaltered Waulsortian limestone containing some sulphide occurs just below the Upper orebody near its northern margin. The sulphide, mainly pyrite with some coarse-grained sphalerite, is associated with coarse-grained calcite and occurs in a rounded patch a few centimeters across cross-cutting delicate structures of the fine-grained limestone. Some of the pyrite is in the form of small prism-shaped or even acicular crystals randomly arranged and some in larger prism-shaped crystals possesses a parallel fracturing resembling cleavage. The rest of the pyrite has fine-grained granular texture.

(2) Marcasite.

Marcasite is rather uncommon in the Upper orebody. Its most frequent occurrence is in association with fracturing and galena veining in pyrite, where it occurs patchily in the pyrite as fairly large grains. In many such instances the marcasite is intensely fractured (Fig. 151). In one instance a small granular pyrite aggregate was found
with a well defined rim of granular marcasite. One occurrence of a group of small globules of marcasite was found.

(3) Sphalerite and Galena.

At the top of the southern part of the orebody granular sphalerite and dolomite of about the same grain-size occur together (Fig. 152). The sphalerite is dark reddish-brown and the dolomite in contact with it consistently possesses euhedral grain boundaries against the sphalerite. Galena occurs less commonly in the Dolomite Breccia but where it does, it shows the same relationship to the dolomite as does sphalerite, i.e. dolomite is consistently euhedral against the galena.

Galena and less commonly sphalerite occur as fine fracture fillings in other minerals, especially pyrite, and also as coarse subhedral crystals along some joint planes in the massive sulphide.

(4) Chalcopyrite, tennantite, boulangerite.

Chalcopyrite and tennantite occur usually with galena in fine fractures in pyrite. Chalcopyrite also occurs as minute blebs or lamellae in sphalerite, and tennantite less commonly is found in colloform galena as small blebs around the edge of the galena aggregate in contact with surrounding minerals. Boulangerite always occurs in galena and usually in the type with a colloform texture. In contrast with tennantite it occurs throughout the galena and is usually in elongated blebs showing a preferred alignment (Fig. 153).

(5) Quartz, barite and carbonates.

Quartz occurs in a few localities where pyrite is
interbedded with cherty dolomite. There, some pyrite crystals or lenses are rimmed by a layer of fine acicular quartz radiating from faces or sides parallel to bedding (Fig. 154). No instances were found of quartz radiating from sides perpendicular to bedding. The acicular quartz crystals are usually straight but those in Figure 154 shows a complementary change in growth direction above and below the pyrite.

Barite occurs as large straight or curved prismatic crystals in a mainly dolomitic matrix, as fine acicular crystals in a galena or sphalerite matrix (Fig. 118), and as small rounded grains where enclosed by pyrite.

Buhedral dolomite in sphalerite has already been referred to and in a few instances this is zoned with some zones containing minute blebs of sphalerite. In a few localities the dolomite crystals contain cores of calcite which may be irregular in shape or may coincide with the form of the surrounding dolomite.

**Mineralization in minor fault zone.**

Reference has already been made to a minor fault zone containing sulphides and intersected in the Dolomite Breccia about 100 ft. above the Upper orebody. This mineralization displays interesting textures and minerals with both similarities to and contrasts with those in the underlying orebody, to which it may be closely related.

Pyrite and sphalerite are the dominant minerals commonly occur as interbanded colloform structures as granular aggregates. The colloform pyrite b...
Fig. 151. Intensely fractured marcasite. Upper orebody.

Fig. 152. Granular dolomite (dark grey) and sphalerite in the Dolomite Breccia just above the Upper orebody.

Fig. 153. Regularly arranged boulangerite growths aligned in galena. Upper orebody. Etched with dilute HCl.

Fig. 154. Pyrite crystal (black, centre) with acicular quartz crystals radiating from its margins. Base of Upper orebody.
granular or a coarse, zoned radial texture and the inter-
banded sphalerite displays a strong radial structure in obli-
que light. The radiating sphalerite crystals are locally qui-
te wide, much more so than in any similar structures in the
main orebody. However the colloform sphalerite when etched
reveals a coarse granular aggregate of twinned crystals un-
related to the radiating texture. Most of it is dark red-
dish-brown and that with a colloform texture shows little tra-
ce of colour banding.

Blebs and lamellae of chalcopyrite are relatively common
in the sphalerite. In the colloform sphalerite particular
bands contain more chalcopyrite than others. Lamellae show
some tendency to form parallel to the radial and concentric
sphalerite structure but more commonly they form along grain
boundaries and twin planes related to the granular texture.
A small amount of pyrrhotite in the form of tiny blebs also
occurs in colloform sphalerite. No chalcopyrite occurs in
the areas with pyrrhotite and vice versa. Areas of sphale-
rite containing chalcopyrite and pyrrhotite are consistently
slightly darker in colour than other parts.

A small amount of galena occurs with the colloform spa-
lerite and pyrite and locally takes the form of radiating crys-
tals between radiating sphalerite crystals giving rise to "car-
rot-shaped" crystals similar to those described by Roedder
(1968, Fig. 23, p. 464). Galena also occurs interstitial-
ly in granular pyrite. It contains inclusions of arsenopy-
rite, tetrahedrite, boulangerite and bournonite. The arse-
nopyrite occurs as scattered rhomb-shaped crystals, the
Fig. 151. Intensely fractured marcasite. Upper orebody.

Fig. 152. Granular dolomite (dark grey) and sphalerite in the Dolomite Breccia just above the Upper orebody.

Fig. 153. Regularly arranged boulangerite growths aligned in galena. Upper orebody. Etched with dilute HCl.

Fig. 154. Pyrite crystal (black, centre) with acicular quartz crystals radiating from its margins. Base of Upper orebody.
boulangerite as thin growths parallel to cleavage traces and both the tetrahedrite and bournonite as scattered rounded grains.

A small amount of calcite occurs as dendritic colloform structures in sphalerite.

**Number 3 zone mineralization.**

The disseminated sulphides of the No. 3 zone consist of fragmental and also veinlet and interstitial filling types.

Pyrite, especially, occurs as fragmented colloform bands and irregular fragments mixed with fragments of dolostone in the Dolomite Breccia (Fig. 155). It also occurs as framboids, as euhedral cubic grains surrounding grains of dolomite, and intergrown with and surrounding colloform sphalerite and galena.

The colloform sphalerite is light coloured and intricately intergrown with pyrite. In transmitted or oblique light it appears fine grained but etching reveals, as in sphalerite in the main orebody, a coarse granular aggregate of twinned grains. A second type of sphalerite also occurs. It is medium brown, coarse grained and interstitial to dolomite grains which are euhedral against it. It usually contains a few blebs of chalcopyrite. The same type of sphalerite also occurs in small veinlets and its distribution is erratic.

Galena occurs as small delicate, dendritic structures which are rimmed by a fine layer of pyrite. It also occurs as interstitial or vein fillings.

Some marcasite occurs in the No. 3 Zone and has two
Fig. 155. Pyrite fragments (black) in brecciated dolomite, No. 3 zone.

Fig. 156. Interbedded sulphide (lighter grey) and shale at the top of the Lower Dolomite.

Fig. 157. Complex sulphide and carbonate veining in the Lower Dolomite.

Fig. 158. Strongly fractured and veined part of the Lower Dolomite. Barite veins (dark grey) cut by dolomite veins (white).
distinct forms. In one it is interbanded with thin flat layers of pyrite and each band is composed of thin columnar crystals aligned perpendicular to the banding. The other occurrence is in pyrite in association with fracturing and galena veinlets, similar to one of its forms in the main orebody.

Finer-grained portions of the Dolomite Breccia in the No. 3 zone show fairly well developed graded bedding involving mainly dolomite particles. Larger dolostone fragment in coarser fractions contain a small amount of calcite at their centres.
CHAPTER 5
THE LOWER OREBODY

General features.

The Lower orebody is a disseminated orebody in the broad sense and as such its limits, shown in Figure 7, are determined by base metal grade rather than by limits of mineralization. The total mineralized area is therefore a little larger than illustrated but conforms in general to the same configuration. Mineralization of ore grade is confined mainly to the Lower Dolomite. It is most intense near the surface and decreases gradually with depth but where the Lower Dolomite diverges from the main fault plane the intensity falls off rapidly.

The top of the Lower Dolomite is marked by a thin black shale horizon which contains abundant bedded pyrite. In the vicinity of the main fault the shale also contains sphalerite, galena and chalcopyrite which occur in particular beds especially those containing pyrite (Fig. 156). Locally, sphalerite, galena and chalcopyrite occur in a stratabound manner in more dolomitic beds within or just below the shale horizon. A few other thin beds of pyritic shale occur lower down in the Lower Dolomite and near the fault they also contain some sphalerite and galena. In hand specimen textures in these bands resemble those in the Upper orebody.
Bedded pyrite also occurs to some extent in shales of the Basal Fragmentals and disseminated pyrite, sphalerite, galena and chalcopyrite are fairly common in Devonian sandstone in the fault zone.

The principal zone of mineralization in the Lower Dolomite consists of fracture and interstitial fillings. The fracture fillings form a complex array of veins (Fig. 157) of many attitudes. Where bedding in the dolostone is distinct there is evidence of slight displacement along the sharp angular fractures (Fig. 159). The thickness of the fracture fillings may be up to several inches. In the vicinity where the Upper and Lower orebodies come in contact the intensity of mineralization increases to produce locally massive sulphide. Interstitial fillings occur between dolomite grains.

Mineralogy.

The minerals occurring in the Lower orebody are for the most part the same as those in the Upper orebody but proportions are different. They are galena, sphalerite, pyrite, chalcopyrite, arsenopyrite, loellingite, tennantite-tetraedrite, marcasite, dolomite, quartz and barite.

Galena and sphalerite are by far the commonest and their distribution has already been outlined in terms of lead and zinc distribution. Pyrite is common as thin bands interbedded with shale but it is much less common in the vein or interstitial fillings. It also occurs disseminated in the wallrocks marginal to veins and to a small extent in the more massive mineralization at the top of the orebody. Arsenopyrite
and loellingite are associated in many places with the pyrite marginal to veins, Arsenopyrite also occurs within the veins and is more widely distributed than loellingite which is restricted more to the central zone of the orebody. Members of the tennantite-tetrahedrite series have a widespread distribution but make up only a very small proportion of the vein fillings. Of the two, tennantite, confirmed by X-ray, is by far the commoner. Tetrahedrite is very restricted in occurrence and was identified by optical means only. Chalcopyrite occurs right across the orebody in small amount but tends to be more common in the west and in the lower parts. It usually occurs in veins in association with quartz. Marcasite is rare and has been found in only one instance associated with colloform pyrite near the top of the orebody.

Of the other three minerals dolomite is a common vein filling throughout the orebody. Quartz, as chalcopyrite, is more common in the west and in the lower part. It occurs mainly in veins but locally patchy silicification occurs in the wall-rocks marginal to them. Barite occurs commonly as coarse vein fillings at the extreme east end of the orebody but is rare in other parts.

Textural features.

The Lower orebody does not show such a wide range of textures as does the Upper orebody and they cannot be so conveniently categorised. In the Lower orebody there is a gradual change from fine grained, commonly colloform textures at the top to coarse granular aggregates in the lower parts.
However there are many exceptions to the overall trend.

(a) Upper part of the orebody.

In the more massive mineralization just below the Upper orebody colloform textures are common. There are a few examples of concentrically banded pyrite similar to those in the Upper orebody and there is also evidence of brecciation of these. A few colloform sphalerite aggregates with concentric and radial structure occur. More commonly, though, colloform textures are developed in encrusting layers rather than in the discrete spherical and hemispherical forms common in the Upper orebody and the layers are composed of different minerals (Fig. 160). Some extremely fine and intricate colloform textures are developed in parts of this upper zone. Contrasting with this there are also coarse colloform textures with subhedral zoned crystals of pyrite (Fig. 161).

Sphalerite with a radiating structure commonly occurs as an outer layer on colloform pyrite. It also occurs as "uncomformable" layers within colloform pyrite layers as shown in Figure 161 where sphalerite radiates outwards from points on the inner apparently broken pyrite surface and truncates pyrite bands on the right and possibly above as well. In the same section a similar pyrite layer is truncated against sphalerite and a thin layer of pyrite and some larger pyrite crystals occur along the contact with sphalerite (Fig. 163). These textures are distinctly different from sphalerite-pyrite textures in the Upper orebody where sphalerite is conformably interbanded with pyrite and rarely shows cross-cutting relationships.
Fig. 159. Angular fracturing and minor displacement of bedding in a specimen from the Lower Dolomite.

Fig. 160. Finely interbanded pyrite, galena, (both white) and sphalerite (grey) surrounding structureless sphalerite (lower right). Upper part of the Lower orebody.

Fig. 161. Colloform and zoned, subhedral pyrite transected by an irregular sphalerite veinlet (medium grey). Upper part of the Lower orebody.

Fig. 162. Fine-grained pyrite and dolomite containing a pyrite veinlet with quartz at its centre. Top of the Lower Dolomite.
One instance was found of a band of tennantite occurring in a colloform, concentrically banded aggregate of sphalerite and galena. The whole colloform structure is small and contained in a cavity in coarse-grained pyrite. Usually, however, in the upper part of the orebody tennantite (or tetrahedrite) occurs as thin stingers in pyrite and is usually in association with galena which occurs more widely in a similar form. Chalcopyrite and sphalerite also occur as fracture fillings in pyrite.

The bedded pyrite in shale at the top of the Lower Dolomite occurs in a variety of forms; anhedral grains, euhedral pyritohedral and cubic grains, and frambooids. It is fine grained and in a clay-mineral or cherty matrix. A few instances of sulphide veinlets cutting this pyrite zone were found. The margins of the veinlets are poorly defined and Figure 162 shows one consisting of an outer zone of pyrite and a quartz centre.

Where several sulphide minerals occur stratabound in more dolomitic shale beds, pyrite form is similar to that in the shale but galena and chalcopyrite in places partly pseudomorph euhedral dolomite crystals (Fig. 164) and occur along grain boundaries of granular dolomite. Pseudomorphing of dolomite is initiated around the edges of the grain and in no instance was the whole grain found to be replaced.

In the region where the vein-type mineralization in the Lower Dolomite merges upwards into more massive sulphide, pyrite grains and granular aggregates are irregularly veined by sphalerite and galena (Figs. 165, 166). The amount of the veining mineral varies and may occupy more than half of
Fig. 163. Zoned, subhedral pyrite fractured and rimmed by later pyrite and sphalerite (dark grey). Upper part of the Lower orebody.

Fig. 164. Galena and chalcopyrite (both white and indistinguishable from each other) pseudomorphing dolomite crystals. Upper part of the Lower orebody.

Fig. 165. Pyrite grains (white) veined and partly replaced by sphalerite. Upper part of the Lower orebody.

Fig. 166. Granular pyrite aggregate (white) veined and partly replaced by sphalerite. Upper part of the Lower orebody.
the original pyrite volume. In places the pyrite is prism-shaped with a well developed fracturing or cleavage and galena and sphalerite, if present in the section, invariably occur along the cleavages. As before, they may make up a considerable portion of what was formerly pyrite and locally the stage is reached where only a few narrow lath-shaped pyrite grains are dispersed in sphalerite or galena. Chalcopyrite bears a similar relation to pyrite but is much less common. It also occurs as an intergranular filling in areas of granular pyrite and in one instance was found pseudomorphing part of a poorly formed colloform aggregate.

(b) Lower part of the orebody.

The lower part of the orebody includes mainly the vein-type and disseminated mineralization in the Lower Dolomite. Vein structures and textures are not consistent. Their margins are either sharply or poorly defined; they may contain one or more minerals, and may be zoned or unzoned. Thickness and extent of an individual vein are extremely variable.

Where zoning is developed usually only two zones can be recognised and these are an outer dolomite and an inner sulphide zone. Locally however a more complex zoning is found to give, for example, from the outside, successive zones of dolomite, pyrite, quartz, dolomite and galena; or dolomite, galena and sphalerite; or quartz, galena, sphalerite and pyrite.

Pyrite is uncommon in veins but where it does occur it forms granular aggregates or occurs as isolated grains commonly prism-shaped with parallel fracturing and veined by
sphalerite and galena (Fig. 167). Where pyrite occurs as a distinct zone in a vein it is subhedral with well developed growth zoning directed inwards. Disseminated pyrite is common in the wall-rock close to veins where it occurs as small, isolated crystals, anhedral to euhedral in form.

Pyrite also occurs to a small extent as frambooids, both in the wall-rocks and surrounded by other sulphides in the veins. In one locality, where a pyritic shale bed is cut by a vein, pyrite grains indigenous to the shale are covered by a layer of secondary pyrite near the vein.

The mode of occurrence of arsenopyrite and loellingite is similar to that of pyrite. Both tend to occur marginal to veins as euhedral crystals. Arsenopyrite crystals are diamond shaped in section and can be distinguished from loellingite which occurs as greatly elongated prismatic crystals or, where present in large amount, as rosettes of columnar crystals.

In constrast with the above, sphalerite and galena never display euhedral form. In veins they occur in irregular granular aggregates and have sinuous margins against each other. Sphalerite is usually medium-brown but variations from almost colourless to dark reddish-brown do occur. It is coarse grained and does not display radiating colloform textures. Most sphalerite, except the palest variety, contains a little chalcopyrite in the form of thin lamellae aligned along twin planes or as films around grain boundaries. In one instance it was found occurring along a growth zone in a sphalerite crystal. Dolomite is commonly included by sphalerite and
Fig. 167. Pyrite with parallel fracturing or cleavage partly replaced by galena (medium grey). Upper part of Lower orebody. Lightly etched with dilute HCl.

Fig. 168. Marginal part of a vein in the Lower Dolomite. Dolomite (dark grey) mostly in euhedral crystals. Galena (white) surrounds dolomite and contains rounded grains of sphalerite (grey).

.. Fig. Sulphide from the B zone showing fine banding, probably of sedimentary origin.
galena in veins and almost invariably displays euhedral outlines against both.

Sphalerite-galena veins commonly have curving and poorly defined boundaries. The marginal zones consist of sphalerite and galena interstitial to subhedral to euhedral dolomite grains (Fig. 168). A fairly consistent feature of some of these zones is that galena, which is more abundant, occurs in contact with dolomite whereas sphalerite occurs as rounded grains completely enclosed by the galena (Fig. 168). Small amounts of tennantite (or tetrahedrite) occur in the vein-type and interstitial galena. It occurs as small rounded grains. Locally a few arsenopyrite grains are also enclosed by the galena.

Chalcopyrite occurs as fairly small rounded grains, associated with galena, in fractures in pyrite and as irregularly shaped patches within quartz veins. It is coarse grained. In two instances it was found to be strongly anisotropic with straight lamellar twins, but generally it is very weakly anisotropic and untwinned.

Barite occurring in veins is coarse-grained prismatic and that disseminated in the Lower Dolomite is also prismatic and euhedral against dolomite grains.
CHAPTER 6
ASSOCIATED ORE DEPOSITS

The other recently discovered deposits in the Silvermines area have been briefly examined. Some of their features are an aid in the determination of the process of formation of the major Mogul orebodies and will be summarised here. The distribution of the deposits, the B and K sulphide zones and the Magcobar barite deposits, is shown in Figure 6.

The B zone.

The B zone occurs at the same stratigraphic horizon as the Upper orebody and both the succession and textural features of the two deposits show strong similarities. It is underlain by thin shale and chert horizons, and overlain by the Dolomite Breccia which is almost identical to that already described but of lesser vertical extent. In its upper parts it becomes less brecciated, less dolomitic and finally passes up into unaltered Waulsortian limestone.

The ore zone itself is thinner and not so massive as the Upper orebody. It tends to be divided into bands of sulphide separated by shale or brecciated dolostone. The pyrite content is less than in the Upper orebody and the ratio of zinc to lead (3:2) is lower. The silver content
(1 oz. per ton) is slightly higher but when expressed as a ratio with lead it is slightly lower than in the Upper orebody.

Structures in the ore zone itself are similar to those in the Upper orebody but show less brecciation. Layers of undeformed shale and finely laminated sulphides (Fig. 169) are more common and particularly noticeable are fine interlaminations of galena and shale.

From the few sections examined microscopically the B zone mineralogy appears simpler than in the Upper orebody. The only sulphide minerals found were pyrite, galena, sphalerite and marcasite.

Textures are similar to those in the Upper orebody although very fine grained textures are more common and large colloform structures less common. Pyrite occurs as subrounded colloform structures similar to those shown in Figure 91 but containing some large marcasite grains in their inner parts. It also occurs in much smaller globular structures in association with sphalerite and galena and as subhedral to euhedral (pyritoehedral) grains in dolomite and surrounding fragments of dolostone. Pyrite in the form of framboids is not common.

The Magcobar barite deposit.

The Magcobar barite deposit is a stratiform massive barite lens, 20 to 30 ft. thick, containing 2,000,000 tons of ore which averages 85% BaSO₄ (Weber, 1964). It occurs at the same stratigraphic level as the Upper orebody and B zone.
sulphides but is 500 ft. higher in elevation than the Upper orebody and 600 ft. higher than the B zone. Thus it is situated on top of a structural dome flanked to the south-west and south-east by the Silvermines Fault and sloping downwards towards the sulphide bodies to the north-west and north-east. Present day topography still reflects the dome and the surface level around the barite deposit is about 100-200 ft. higher than the ground surface above the sulphide bodies.

The barite deposit is contained in a succession very similar to that enclosing the sulphide bodies. It is underlain by chert and overlain by the Dolomite Breccia previously described. The underlying chert is exposed in parts of the open pit. It is well bedded with a slight nodular development. At the extreme east end of the deposit some pyritic chert occurs at the top of the chert sequence. Elsewhere the chert is succeeded by one or two feet of finely interbanded red chert, hematite and red-stained barite. This passes up into more massive red barite which at higher levels loses its red colour. The upper half of the barite body is grey and contains some disseminated pyrite. Near the upper contact pyrite becomes more common, occurring as thin bands parallel to stratification, and in the eastern part the barite is overlain by a layer of massive sulphide. The sulphide tends to occur in hollows or troughs in the upper surface of the barite so that its lower surface undulating and its thickness variable up to about 10 ft. The sulphide is mainly pyritic but contains a lesser amount of sphalerite and galena. The sulphide is overlain...
Dolomite Breccia, which contains some pyrite in its lower part.

The portion of the ore-bearing horizon between the barite and sulphide bodies has not been explored in any detail but evidence available suggests that it contains small amounts of barite and sulphide disseminated in the Dolomite Breccia.

The lowest part of the barite deposit contains finely laminated barite, chert and hematite and the laminae are usually folded or contorted (Fig. 170). The cryptocrystalline chert laminae contain some very small concentrically banded concretions or spherulites of red quartz (Fig. 171). In places the concretions occur in very finely banded zones and the bands tend to curve around them (Fig. 172). In other places concentrically banded colloform chert or quartz surrounds irregularly shaped patches of finer-grained, structureless chert (Fig. 173) or lines tiny vugs.

In the main part of the deposit, which is composed almost entirely of barite, fine laminations are absent and the barite occurs either as a compact fairly fine-grained aggregate or as coarse radial growths. The radial structures occur mainly in the central part of the deposit. They are up to several inches across and occur in the form of spherulites or as curving encrustations around patches of finer grained barite (Fig. 174). The radiating crystals are invariably pinkish. The more compact variety of barite occurs in the upper part of the deposit. It is grey and appears structureless in hand specimen. However, thin section
Fig. 170. Contorted, interlaminated chert and hematite (darker grey) from the base of the Magcobar barite body.

Fig. 171. Fine-grained chert (dark grey) containing small, concentrically banded chert concretions. Base of Magcobar barite body.

Fig. 172. Finely laminated chert deformed by a chert concretion. Base of Magcobar barite body.

Fig. 173. Chert with a colloform texture surrounding a patch of structureless chert. Base of Magcobar barite body.
Fig. 175. Laths and other grains of barite enclosed by pyrite (white). Magoobar barite body.

Fig. 176. Part of a barite-sulphide vein from the K zone. Wall-rock (dark) at bottom of photograph. Above, colloform galena, sphalerite and pyrite, concentrically banded and with convexity directed toward centre of vein. Top right, coarse grained barite representing central part of the vein.
and sphalerite is coarse grained and it contains a small amount of chalcopyrite, features typical of the Lower ore-body.

The veins vary considerably in their attitude, continuity and thickness and in the grain-size and zoning of contained minerals. Some vein fillings are coarse grained and unzoned whereas others are composed of finely banded to colloform layers displaying a variety of zonal features. Usually sulphides form the outer parts of the veins and barite occurs in their centres. The sulphide zone may contain pyrite, sphalerite and galena which themselves show no consistent ordering but show intricate and variable interbanding.

Part of a vein containing colloform sulphides and barite is shown in Figure 176. The gross zoning in the vein is from sphalerite in the outer part through pyrite to barite at the centre. However microscopic examination reveals thin intricate bands of pyrite in the sphalerite and also discontinuous bands of galena in sphalerite close to its contact with pyrite. The sphalerite is concentrically banded, light to dark brown in colour, and with a strong radial crystal development. It also contains traces of chalcopyrite. The pyrite is also concentrically banded. Some bands are extremely thin and fine grained whereas others are thicker and possess coarse radially directed crystals. The coarser pyrite is fractured and veined by galena and sphalerite, and marcasite is common in the pyrite along fracture margins. Barite at the centre is in the form of a matte of large lath shaped crystals. Another example of a similarly banded ve
filling consists in part of interbanded sphalerite and pyrite. The sphalerite has a radial texture and in each band the radiating crystals nucleate on the adjacent outer pyrite band surface and diverge towards the centre of the vein.

The curving colloform surfaces are always directed towards the centre of the vein. Other examples of finely interbanded vein fillings have either flat or only slightly curved interfaces and thus there appears to be a gradation from planar to colloform banding.
CHAPTER 7
ANALYTICAL AND RELATED RESULTS

Cobalt and nickel content of pyrite.

Cobalt and nickel values in pyrite have been quoted for many types of deposits and it appears that the absolute values and ratios give some indication of the origin of the deposits. For this reason some pyrite samples from the Upper and Lower orebodies were analysed to give some idea of the order of magnitude of the Co and Ni contents. One frambooidal aggregate and eleven samples of colloform pyrite from different parts of the Upper orebody, and two samples of bedded pyrite in black shale and one of veined pyrite from the Lower orebody were analysed. The results are given in Table 3.

The Co and Ni content in the colloform pyrite from different parts of the orebody does not show much variation and the Co: Ni ratio is consistently less than one. The frambooidal pyrite has slightly more Ni and a very low Co: Ni ratio. The pyrite samples from shale bands in the Lower Dolomite are similar to those from the Upper orebody but the vein pyrite from the Lower orebody has a Co: Ni ratio greater than one.

Silver and antimony content of galena.

The silver and antimony contents were determined in three
CHAPTER 7
ANALYTICAL AND RELATED RESULTS

Cobalt and nickel content of pyrite.

Cobalt and nickel values in pyrite have been quoted for many types of deposits and it appears that the absolute values and ratios give some indication of the origin of the deposits. For this reason some pyrite samples from the Upper and Lower orebodies were analysed to give some idea of the order of magnitude of the Co and Ni contents. One framboidal aggregate and eleven samples of colloform pyrite from different parts of the Upper orebody, and two samples of bedded pyrite in black shale and one of veined pyrite from the Lower orebody were analysed. The results are given in Table 3.

The Co and Ni content in the colloform pyrite from different parts of the orebody does not show much variation and the Co:Ni ratio is consistently less than one. The framboidal pyrite has slightly more Ni and a very low Co:Ni ratio. The pyrite samples from shale bands in the Lower Dolomite are similar to those from the Upper orebody but the vein pyrite from the Lower orebody has a Co:Ni ratio greater than one.

Silver and antimony content of galena.

The silver and antimony contents were determined in three
TABLE 3

Cobalt and nickel content of pyrite

<table>
<thead>
<tr>
<th>Sample</th>
<th>Ni. p.p.m.</th>
<th>Co. p.p.m.</th>
<th>Co:Ni</th>
</tr>
</thead>
<tbody>
<tr>
<td>Colloform pyrite from Upper orebody</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1) north-central part</td>
<td>32</td>
<td>41</td>
<td>1.28</td>
</tr>
<tr>
<td>2) central part</td>
<td>132</td>
<td>44</td>
<td>0.32</td>
</tr>
<tr>
<td>3) south-central</td>
<td>75</td>
<td>54</td>
<td>0.72</td>
</tr>
<tr>
<td>4) extreme south</td>
<td>46</td>
<td>44</td>
<td>0.96</td>
</tr>
<tr>
<td>5) extreme north</td>
<td>72</td>
<td>39</td>
<td>0.54</td>
</tr>
<tr>
<td>6) north-east</td>
<td>75</td>
<td>43</td>
<td>0.57</td>
</tr>
<tr>
<td>7) south-central</td>
<td>125</td>
<td>43</td>
<td>0.34</td>
</tr>
<tr>
<td>8) south</td>
<td>790</td>
<td>158</td>
<td>0.20</td>
</tr>
<tr>
<td>9) south</td>
<td>50</td>
<td>44</td>
<td>0.88</td>
</tr>
<tr>
<td>10) south</td>
<td>94</td>
<td>44</td>
<td>0.46</td>
</tr>
<tr>
<td>11) Unknown location</td>
<td>58</td>
<td>44</td>
<td>0.76</td>
</tr>
</tbody>
</table>

Framboidal pyrite

<table>
<thead>
<tr>
<th>Ni. p.p.m.</th>
<th>Co. p.p.m.</th>
<th>Co:Ni</th>
</tr>
</thead>
<tbody>
<tr>
<td>190</td>
<td>58</td>
<td>0.31</td>
</tr>
</tbody>
</table>

Pyrite in shale in Lower Dolomite (1)

<table>
<thead>
<tr>
<th>Ni. p.p.m.</th>
<th>Co. p.p.m.</th>
<th>Co:Ni</th>
</tr>
</thead>
<tbody>
<tr>
<td>35</td>
<td>33</td>
<td>0.94</td>
</tr>
</tbody>
</table>

(2)

<table>
<thead>
<tr>
<th>Ni. p.p.m.</th>
<th>Co. p.p.m.</th>
<th>Co:Ni</th>
</tr>
</thead>
<tbody>
<tr>
<td>48</td>
<td>35</td>
<td>0.73</td>
</tr>
</tbody>
</table>

Pyrite in vein, Lower Orebody

<table>
<thead>
<tr>
<th>Ni. p.p.m.</th>
<th>Co. p.p.m.</th>
<th>Co:Ni</th>
</tr>
</thead>
<tbody>
<tr>
<td>50</td>
<td>73</td>
<td>1.45</td>
</tr>
</tbody>
</table>
galena samples from the Upper orebody, three from the Lower orebody and one from the B zone (Table 4). With one exception the values for silver lie in the range 300 to 700 p.p.m. and there is no great difference between galenas from the Upper and Lower orebodies. Using the mine assay values for lead and silver the silver content was recalculated as a fraction of the galena at each sample location. These results are given for comparison and they are of the same order as those determined by analysis of the galena samples.

The amount of antimony in galena in the Upper and Lower orebodies was found to be in the range 50-250 p.p.m. with one exception of 605 p.p.m. The one sample from the B zone contains a significantly higher amount, 892 p.p.m.

**Vickers hardness of galena.**

A determination of the Vickers hardness of galena was made on a series of samples from the Upper orebody, the Lower orebody and the B zone. A Leitz Miniload hardness tester was used. For each polished section twelve diamond impressions were made and each diagonal was measured three times. The corresponding range of hardness and average hardness for each specimen were calculated (Fig. 177).

The hardness of galena in the Upper orebody shows a wider spread in each sample than it does in each sample in the Lower orebody but the range of average values for different samples in each orebody is about the same. The overall average hardness in the Lower orebody is very close to that in the Upper orebody, 71.4 and 71.8 respectively. Three samples
**TABLE 4**

Silver and antimony content of galena

<table>
<thead>
<tr>
<th>Sample</th>
<th>Ag. p.p.m.</th>
<th>Sb. p.p.m.</th>
<th>Ag. content calculated from assay figures. p.p.m.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper orebody</td>
<td>483</td>
<td>176</td>
<td>760</td>
</tr>
<tr>
<td></td>
<td>703</td>
<td>605</td>
<td>593</td>
</tr>
<tr>
<td></td>
<td>69</td>
<td>52</td>
<td>267</td>
</tr>
<tr>
<td>Lower orebody</td>
<td>337</td>
<td>240</td>
<td>777</td>
</tr>
<tr>
<td></td>
<td>520</td>
<td>120</td>
<td>432</td>
</tr>
<tr>
<td></td>
<td>511</td>
<td>240</td>
<td>656</td>
</tr>
<tr>
<td>B zone</td>
<td>649</td>
<td>892</td>
<td>figures not available</td>
</tr>
</tbody>
</table>

* Using Averages of 5-foot core samples and assuming all silver is in galena.
Fig. 177. Vickers hardness values for galena in the Upper, Lower and B zone orebodies. The horizontal line represents the range and the vertical dash the average for each sample. Each sample consists of 12 grains in one polished section.
from the B zone which were tested show a very wide range of hardness in each sample and also a much higher average value then in the other orebodies. The overall average of the three samples is 85.8.

Iron content of sphalerite.

About 140 grains of sphalerite from the Upper and Lower orebodies and the B zone were analysed for their iron content using an electron microprobe. The values determined are all low and in the range 0-4 wt.% Fe.

In the Lower orebody the iron content ranges from 0.5 - 2.5 wt% and is related to colour, i.e. the darker the sphalerite the higher its iron content. In vein-filling sphalerite there is, in places, a zonal tendency with the lowest iron content occurring in sphalerite at the outer margin in contact with dolomite and the highest content in the inner sphalerite.

In the Upper orebody the sphalerite near the south end has the higher iron content whereas that around the northern margins contains no detectable iron. The highest values, around 3.6 wt.% Fe, were found in coarse-grained sphalerite interbanded with shale at the base of the orebody. There is considerable variation in iron content between different grains within the one polished section and the variation is related to colour. Where blebs or lamellae of chalcopyrite occur in sphalerite its colour is darkest and iron content greatest.

Different bands in colloform sphalerite were analysed and, as above there is a strong relationship between colour
and iron content. Very light coloured bands surrounding galena (Figs. 101-106) or calcite contain almost no iron whereas darker bands, which usually occur in the outer parts of colloform structures, contain up to 1 wt% Fe.

Colloform sphalerite does not however always show a consistent increase in iron content from the centre outwards as is shown by results from that illustrated in Figure 115. From the centre outwards there are four distinct sphalerite bands: a core of a medium shade, a paler fibrous band, a thin, very dark band with a granular texture, and an outermost light fibrous band. The respective iron contents are: about $\frac{1}{2}$, about 1, 1-2 and $\frac{1}{4}$-3/4 wt%.

Sphalerite interbanded with colloform pyrite usually has an iron content of less than 1 wt%. Sphalerite in the fault zone above the Upper orebody is mostly dark in colour and has iron contents ranging from $\frac{1}{2}$ to 4 wt% and the variations follow the same pattern as for the Upper orebody. Sphalerite in the B zone is almost colourless and the samples analysed contain no detectable iron.

**Sulphur isotope ratios.**

The ratios of the sulphur isotopes $S^{34}$ and $S^{32}$ were determined for some sulphide minerals in the Upper and Lower orebodies and for barite from the Magcobar deposit and from galena-barite veins in Devonian sandstone at Shallee, half a mile west of the Mogul mine. The results, in terms of $\delta S^{34}$, are given in Table 5.
<table>
<thead>
<tr>
<th>Location</th>
<th>Mineral</th>
<th>$\delta^{34}\text{S}$ (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper orebody</td>
<td>Pyrite</td>
<td>-4.9</td>
</tr>
<tr>
<td></td>
<td></td>
<td>-10.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td>-16.8</td>
</tr>
<tr>
<td></td>
<td></td>
<td>-19.4</td>
</tr>
<tr>
<td></td>
<td></td>
<td>-29.7</td>
</tr>
<tr>
<td></td>
<td>Sphalerite</td>
<td>-14.2</td>
</tr>
<tr>
<td></td>
<td></td>
<td>-19.0</td>
</tr>
<tr>
<td></td>
<td></td>
<td>-20.7</td>
</tr>
<tr>
<td></td>
<td>Galena</td>
<td>-3.6</td>
</tr>
<tr>
<td>Lower orebody</td>
<td>Sphalerite</td>
<td>1.8</td>
</tr>
<tr>
<td></td>
<td></td>
<td>10.0</td>
</tr>
<tr>
<td></td>
<td>Galena</td>
<td>-1.4</td>
</tr>
<tr>
<td></td>
<td></td>
<td>-8.5</td>
</tr>
<tr>
<td>Magcobar</td>
<td>Barite</td>
<td>15.1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>15.1</td>
</tr>
<tr>
<td>Shallee</td>
<td>Barite</td>
<td>16.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>18.8</td>
</tr>
</tbody>
</table>
CHAPTER 8
FRAMBOIDS

Terminology.

The term "framboid" was first used by Rust (1935) to denote "clusters of tiny pyrite cubes and grains, the whole with a spheroidal outline" (op. cit., p. 407). The term was generally accepted and is still commonly used (Berner, 1969; Amstutz et al., 1967; Kallioski, 1966; Love and Amstutz, 1966). Other terms have been used by some authors for the same feature, e.g. "sphere" (Greensmith, 1962; Steinike, 1963) and "spherule" (Vallentine, 1963). In translations of Russian literature the term "globulite" is used (Lebedev, 1967; Skripchenko, 1968). Love and Amstutz (1969) have used "cluster" because the features they were describing deviated slightly in appearance from those defined by Rust as frambooids. Because "framboid" appears to be the most widely accepted term it will be retained in this discussion.

Distribution of frambooids.

In the Mogul mine frambooids are abundant in the Upper orebody and are also fairly common in the Lower orebody and associated pyritic, shaly beds. They also occur to a lesser extent in the Dolomite Breccia above the Upper orebody but are conspicuously absent or very uncommon in the
non-ore parts of the black, pyritic shale and Muddy Reef horizon immediately below it where pyrite occurs mainly in subhedral to euhedral grains of greatly varying size. Thus framboids occur in three different lithologies: massive sulphide, shale or shaly limestone and pure dolomite.

In the massive sulphide their distribution is erratic. Locally they occur in great concentration with an estimated 60,000 per cu. m.m. whereas in other parts they are uncommon or absent. Most commonly bands of very fine grained sphalerite form the matrix of the framboids but in some instances dolomite or clay or both occur with sphalerite. Locally fine granular pyrite forms the entire matrix.

In the Dolomite Breccia the framboids are sparsely scattered. They occur in granular dolomite either enclosed by dolomite grains or along grain boundaries and also in interbedded clay-rich laminae and in stylolitic seams containing clay. They commonly occur in association with euhedral pyrite grains of the same size order (Fig. 178).

Textural Features.

A considerable range of textural variation is displayed by the framboids especially in respect of outline, size distribution (2-34 microns) and the size, shape and packing density of individual pyrite grains. Except for size distribution the differences do not appear to depend on host-rock lithology and framboids of almost identical appearance can be found in any of the three lithologies previously named.
Fig. 178. Framboid and euhedral pyrite crystals in close association in dolomite.

Fig. 179. Framboids with non-spherical outlines.
Fig. 1: Fractoids with well defined margins in a clay-erite matrix. Etched.
Fractoids with well-defined margins in a nanocrystalline matrix. Etched.
Fig. 182. Rounded area of separated individual framboild grains probably representing a partially disintegrated framboild.

Fig. 183. Scattered, minute pyrite grains similar to those in fig. 182 and probably representing a completely disintegrated framboild.

Fig. 184. Framboilds in a matrix containing uniformly dispersed, minute pyrite grains.

Fig. 185. Framboilds and pod-shaped aggregates of minute pyrite grains in shale.
framboids (Fig. 185) so there is probably a relation between the two forms.

(b) Individual grains.

Individual grains in the framboids vary considerably in size, shape and packing arrangement. Within any one framboid the grain size remains constant and this holds almost without exception. The same feature has also been pointed out by Love and Amstutz (1966) for framboids in the Chattanooga Shale. Although there are many exceptions there is a noticeable tendency for the size of the grains to increase as the size of the framboid increases. Because of this relationship the number of grains across the diameter of a framboid, and consequently the total number of grains in it, have a tendency to remain constant. A count of grains per diameter was made on 80 randomly chosen framboids and the results were plotted in the form of a histogram of number of grains per diameter against the number of framboids in each class (Fig. 186). From the 7-8 grains per diameter class to the 15-16 class the frequency increases gradually and then drops off sharply for groups with 17 or more grains per diameter. Elimination of the effect of sectioning would probably further increase the 15-16 grain class and considering this as the most commonly occurring number of grains per diameter the most common number of grains in a whole framboid would be in the neighbourhood of 16,000. Love and Amstutz (1969) have also reported a proportional relationship between grain diameter and framboid diameter from framboidal pyrite in Permian andesite
Fig. 186. Histogram showing the frequency distribution of the number of individual grains in a frambold in terms of the number of grains across a diameter of each frambold. Counts from 80 frambolds.
at Plotz, Germany. Examples of the above relationship can be seen in framboids in Figure 189. Exceptions to the rule, in this case large framboids with small grains, are shown in Figure 188.

The shapes and interrelationships of individual grains are usually not visible on an unetched polished surface (Fig. 189) but it was found that etching for about one minute with 1:1 HNO₃ was sufficient to reveal the grain relationships. The grains are always close-packed and cannot have more than a thin film of matrix material, if any, between them. They are always angular except possibly in very small framboids where they are too small to distinguish clearly optically. The grain outlines in section are cubic (Fig. 190), six-sided (Fig. 191) or irregular (Fig. 192). Granular aggregates of polygonal grains also occur (Figs. 193 and 194).

In most framboids the individual grains are arranged randomly. In a few there is a tendency for grains in the outer parts to be arranged in concentric rings (Fig. 195). This tendency occurs only if the grains are cubic and the regular arrangement extends over no more than the outer three rings. A very rare occurrence is a close packing of grains with hexagonal outlines as shown in parts of the large framboid in Fig. 191. This appears to correspond to what has been termed the "hexagonal pattern" by Love and Amstutz (1966, p. 286), formed by close packing of pyritohedral crystals. Love and Amstutz (op. cit.), in a study of framboids from the Chattanooga Shale, have recorded
Fig. 1c7. Two framboids of different sizes but containing about the same number of individual grains. Etched.

Fig. 1c8. Large framboids with unusually small individual grains. Etched.

Fig. 1c9. Framboids occurring as a group in a dolomite-clay matrix.
Fig. 190. Framboids in which some of the individual grains are cubic. Etched.

Fig. 191. Large framboid with some hexagonal grains. A very thin layer of pyrite encrusts the framboid (visible around lower margin). Etched.

Fig. 192. Framboids with irregularly shaped individual grains. Etched.

Fig. 193. Framboid consisting of a granular aggregate of pyrite grains. Etched.
Fig. 194. Framboid composed of a granular aggregate of pyrite grains. Etched.

Fig. 195. Framboid (centre) with outer grains arranged concentrically. Etched.

Fig. 196. Small frambooids in shale. Etched.

Fig. 197. Framboid with galena (darker grey) between individual grains. Etched with dilute HCl.

Fig. 198. Framboids enclosed in a large sphalerite grain and with sphalerite between individual grains. Etched.

Fig. 199. Framboid (centre) with pyrite (darker grey) between individual grains. Etched.
internal ordering in 28% of the framboïds observed and have recognised four different types of ordering. In the present study much fewer than 1% of the framboïds showed even a tendency toward ordering and the hexahedral pattern was the only one of the types defined by Love and Amstutz to be found.

(c) Size Analyses.

Measurements of framboïd diameters were made on specimens from three different localities, in all of which framboïds were present in large numbers. Polished sections cut perpendicular to the stratification were used for each location. Seven or eight photographs at high magnification were taken of each section and measurements made using a photograph of a micrometer under the same conditions of magnification. Between 200 and 260 measurements were made on each section. The results were plotted in the form of histograms with a class interval of two microns. Some measurements are slightly inaccurate because of framboïds with ragged outlines or others which departed from the spherical. In both cases an average value was taken. Those which departed greatly from the spherical were neglected. It is also possible that a few framboïds in the 0-4 micron range were missed because in this size range it was sometimes difficult to distinguish between framboïds and small grains of non-framboïdal pyrite.

Measurements made on framboïds in massive sulphide from the north-east margin of the Upper orebody are shown in Figures 200 and 201, which represent measurements from two
Fig. 200. Histogram showing the size distribution of framboids in massive sulphide in the north-east part of the Upper orebody. Based on 236 measurements from one polished section.

Fig. 201. As fig. 200 but measurements made on a specimen from one foot higher in the section. Based on 213 measurements from one polished section.
Fig. 202. Histogram showing the size distribution of framboids in massive sulphide in the south-central part of the Upper orebody. Based on 206 measurements.

Fig. 203. Histogram showing size distribution of framboids in a sulphide layer in the Muddy Reef. Based on 253 measurements.
Fig. 204. Histogram showing the size distribution of frambolds in a sulphide layer in the Muddy Reef. Based on 217 measurements from one polished section.
specimens separate vertically by one foot. Both specimens contain about 85% sulphide and in each the framboids occur in a matrix of fine-grained sphalerite or sphalerite and dolomite. In both the framboids show a relatively high frequency over a wide size range: 6-24 microns in one and 6-18 in the other. One has a fairly well defined maximum in the 14-16 micron class and the other has two poorly defined maxima in the 6-8 and 14-16 micron classes. The maximum diameter measured was 34 microns.

In a specimen taken from the south-central part of the Upper orebody the range of higher frequencies is greatly reduced and the distribution approaches the unimodal with a positive skew (Fig. 202). The maximum is in the 6-8 micron range with a subsidiary maximum in the 10-12 micron class. This specimen contains about 90-95% sulphide and the framboids occur in a fine-grained sphalerite matrix with some clay impurity.

The third locality chosen was in the portion of the Upper orebody which occurs within the Muddy Reef. Counts were made on two specimens separated vertically by two feet. In one specimen (the upper one) the sulphide content is about 60% and the framboids are confined to interbedded clay-rich layers (Fig. 203). In the other, which is just below the limit of massive sulphide, framboids are uniformly distributed in a matrix of dolomite and clay with a little sphalerite (Fig. 204). Both show a positively skewed distribution. The upper one has a maximum in the 12-14 micron class and members up to 30 microns. The lower one has an
extremely well defined maximum with 32.5% of the framboids in the 8-10 micron class and members of more than 14 microns are very uncommon.

In summary, size analyses suggest that framboids in the massive sulphide have a wide size range and poorly defined and variable modes but those outside it in a non-sulphide matrix have a much narrower size range and do not attain the large dimensions of the former.

Visual observations confirm the above tendencies. Throughout the massive sulphide framboids were observed to vary greatly in size and to reach large dimensions. The largest found has a diameter of 70 microns (Fig. 191). In shaly beds with a small sulphide content the framboids are smaller and fairly uniformly sized (Fig. 196) but do not occur with a sufficiently high frequency to make grain-size analyses by photographic means feasible.

Some measurements were also made of individual grains in the framboids. The most commonly occurring size is about one micron and the range extends from under 0.2 to 4.6 microns.

Mineralogy of framboids.

All individual grains in framboids appear from optical examination to be pyrite. The black matrix material in the framboids is present in only very small amounts and was not identified. It may be organic material as reported by Vallen-tyne (1963) and others or "melnikovite pyrite" with a lower reflecting power than pyrite as stated by P. Ramdohr in a
personal communication to Love and Amstutz (1966). In some cases where the frambooidal grains are detached (Fig. 190), the cementing material, at least in the outer parts, appears to be the same as the host rock of the framboid. Galena (Fig. 197), sphalerite (Fig. 198) and pyrite (Fig. 199) have also been found locally forming the matrix between individual grains. Where galena or pyrite form the matrix it appears that each was deposited in the interstices between individual grains because the grains retain their assumed original shape indicating a loosely bound framboid. Where sphalerite forms the matrix of the group of frambooids (Fig. 198) individual pyrite grains show a tendency to become very small and widely separated indicating their partial replacement by sphalerite.

**Grouping of frambooids.**

Love and Amstutz (1966) have reported groups of 2 to 7 or more frambooids in contact in shales. In the present study a few such groups were found in different environments. Figures 205 and 206 show groups of small frambooids occurring in massive sulphide, in both cases surrounded by fine granular pyrite. Figure 184 shows part of an isolated pod-shaped group of frambooids and similar completely isolated groups have been found in the Dolomite Breccia. The effect of sectioning makes it difficult to determine the number of frambooids in contact but in the examples illustrated there are considerably more than seven in each group. It should be pointed out that usually where frambooids are in contact
they barely touch, i.e. the spherical outline of one is not interfered with by the other. There are, however, exceptions in which an undeformed framboid occurs partly within a concave indentation in an adjoining one (one example in each of Figures 205 and 206). In no case is there an outer ring or rings of individual grains common to both framboids (compare with pyrite overgrowths: below).

**Overgrowths on framboids.**

The majority of framboids are free of encrusting minerals but a few possess this feature. In all examples encountered the encrusting mineral is pyrite. It forms rims varying in thickness from a very thin film (Fig. 191) to greater than the diameter of the framboid (Fig. 207). The rims are usually composed of well-zoned subhedral pyrite crystals radiating outwards (Fig. 208). Commonly the rimming of neighbouring framboids coalesces and outer rims form enclosing several framboids (Fig. 209). This is in contrast to individual framboidal grains which never form a structure common to two or more framboids. A final stage in the encrusting process can lead to a large group of framboids becoming enclosed by the rimming pyrite to form a large area of solid pyrite (Fig. 210). Hundreds of framboids have been observed so enclosed in one group.

**Discussion.**

Excellent accounts of the literature on framboids have been given by Vallentyne (1963) and Love and Amstutz (1966)
Figs. 205, 206. Group of framboids surrounded by granular pyrite. In fig. 205 one of the framboids (right) appears to indent an adjacent one. Etched.

Fig. 207. Two framboids with a thick rim of later pyrite. Etched.

Fig. 208. Framboid with a well zoned pyrite rim. Etched.
Fig. 209. Small groups of framboïds enclosed by a single zoned rim of later pyrite. Etched.

Fig. 210. Several framboïds enclosed by granular pyrite. Etched.
and only a brief summary of the work need be given here. Framboids have been found in many environments, often not associated with economic deposits. They are common in fresh water and marine muds and peats and were known in these environments at the end of the 19th century. The results of recent detailed investigations on these types have been published by Vallentyne (1963) and Love (1964). Framboids are also common in shales, especially where carbonaceous matter is present, and recent publications on these include Love (1957), Love and Amstutz (1966) and Skripchenko (1968). They have also been reported from carbonate rocks (Honjo et al., 1965; Greensmith, 1962). Deposits of economic value in which framboids occur include disseminated base metal deposits in andesites (Steinike 1963; Love and Amstutz, 1969), silver-bearing vein deposits in andesitic volcanics, Idaho (Anderson 1963), tin-bearing vein deposits in Cornwall, England (Schouten, C., 1946, p. 521), massive sulphides in a volcanic environment (Kanehira and Bachinski, 1967), Mississippi Valley type deposits (Rust, 1935) and bedded sedimentary sulphide deposits such as the Mt. Isa deposit (Solomon, 1965).

Early investigators, among them Rust (1935), suggested that framboids formed by simultaneous crystallization from many points within a globule of pyrite or melnikovite gel. Later Schneiderhohn (see Love and Amstutz, 1966, p. 278) proposed the theory that they formed by fossilization of sulphur bacteria, i.e. that the sulphur particles, which form part of the bacteria, combined with metals after death.
of the organism to produce sulphides and retain the original structure. This idea was strongly contested by C. Schouten, (1946) who, after a detailed study of sulphur bacteria, concluded that, on the basis of size, shape and packing arrangement, the structure of a framboïd was not comparable with that of any sulphur bacterium. Schouten did agree however that sulphate-reducing bacteria may have provided much of the \( \text{H}_2\text{S} \) necessary for the formation of pyrite framboïds and that the framboïds were essentially syngentic. Although Love (1957) discovered an organic matrix in framboïds of Carboniferous age, which backed up Schneiderhohn's conclusions, he (1964) and others (e.g. Vallentyne 1963) subsequently proved that framboïds from Recent muds did not have an organic matrix and thus could not represent fossilised bacteria. Love and Amstutz in 1966 left the question of origin rather open because of the lack of knowledge about the behaviour of unstable sulphides during diagenesis but made a generalised suggestion that a sulphide gel was first produced from bacterially reduced sulphide and iron. This gel passed relatively quickly through an intermediate crystalline phase before crystallizing completely as pyrite.

Since 1966 conflicting views on the origin of framboïds have again appeared. Love and Amstutz (1969) from a study of framboïds in andesites reiterated their view that they originated as globules of gel. However Skripchenko (1968) again proposed a biogenic origin after a study of a coaly and bituminous shale associated with the Blyava and Kom- somol'skoye deposits. He made a detailed optical study of
framboids and found a correlation between their development and the proximity of organic remains (graptolites). Those within the organic debris are composed of a large number of tightly packed, small pyrite grains with a very small amount of matrix (presumed organic) and they are surrounded by a shell of "pyrite gel", whereas those more removed from the organic remains have fewer, larger individual grains, more matrix material and lack the gel encrustment. From these observations Skripchenko concludes that the framboids represent the bodies of sulphate reducing bacteria in which pyrite was deposited during their lifetime. The closer the bacteria were to rotting organic material the greater was their activity and hence the greater the amount of $\text{H}_2\text{S}$ and consequently pyrite they produced. Skripchenko does not mention the presence of an intermediate crystalline phase and seems to infer that pyrite was the original precipitate. Of interest in connection with the ideas of Skripchenko is a study, to which he does not refer, by Neves and Sullivan (1964). These authors studied minute pyrite grains in spore assemblages from Upper Palaeozoic coals and shales and found that the organic material between pyrite grains formed part of the surrounding spore material and did not represent the tissue of a separate bacterial organism. They concluded that the pyrite grain clusters formed in the exine at points which had been attacked by sulphate-reducing bacteria and that the exine became moulded around the growing pyrite crystals.

Kato (1967) also emphasises the importance of organic material and suggests that framboids are formed by the
reaction of Fe$^{2+}$ with sulphur which has been ionized to S$^{2-}$ in a colony of bacteria. The resultant FeS$_2$, initially in the form of a gel, crystallizes out in the framboidal form.

Papunen (1966) studied framboids occurring in a peat bog in Finland. He suggested that they formed by a reaction between H$_2$S liberated by bacteria in the bog water and colloidal droplets of humus with adsorbed iron hydroxide. Kalliokoski and Cathles (1969) agree with this suggestion and have added a few more observations in support of it.

The first successful attempt to produce framboids in the laboratory has recently been reported by Berner (1969). They were produced at a temperature of 65$^\circ$C and neutral pH by the reaction of FeS with elemental sulphur in saturated H$_2$S solution. The object of the experiment was to show that micro-organisms are not necessary in the production of the characteristic framboidal texture and prove what has been suspected by many authors.

Despite a lengthy history of research on framboids most of their features are still not satisfactorily explained and some of these can be discussed with reference to the framboids in the Mogul mine.

As already mentioned framboids in shale are smaller (mode 8-10 microns) and have a narrower size range than those in massive sulphide (mode 12-16 microns). Published size analyses are few but some of those available show some correlation with the above results. Kanehira and Bachinski (1967), who studied framboids from a Cu-Zn massive sulphide orebody at Tilt Cove, Newfoundland, recorded a modal size of
12-14 microns, similar to that at Mogul, but the maximum size measured, 20-22 microns, is much smaller. No further measurements from this type of orebody were found in the literature. Measured diameters recorded from frambooids in shales or recent muds are for the most part much smaller. Love and Amstutz (1966) studied frambooids in the Chattanooga Shale (Devonian) from Tennessee and in shales of the Rammelsberg orebody (Devonian). In both a well defined maximum occurred in the range 5-6 microns. Love (1957) found a modal diameter of about 7.5 microns for frambooids from Carboniferous shale in Scotland. Skripchenko reports an average grain size of 10-20 microns for frambooids in a carbonaceous shale of Lower Palaeozoic age. No modal size range was given. Size measurements on frambooids reported from Recent muds have been made by Vallentyne (1963) and Love (1964). Vallentyne reported a modal size of between 11 and 13 microns which would be the equivalent of about 9-11 microns had they been measured in polished section (effect of sectioning reduces diameters by a factor of 0.8165, Jackson et al. (1963), quoted by Love and Amstutz (1966, p. 281). Love on the other hand found that the mode varied in different samples and ranged between 4 and 20 microns, which in polished section, would be between 3 and 16 microns. Some other published diameter dimensions are an average of 8-10 microns (Park, 1967) for frambooids in a bioclastic limestone, and a modal value of 10-12 microns (Love and Amstutz, 1969) for those in andesites associated with disseminated base metal sulphides.
Although the number of size analyses available is far from adequate to show a distinct trend in the relationship between size and environment there is a slight indication that framboïds isolated in sediments and representing the only sulphide present are smaller than those enclosed in a massive or near massive sulphide. This would suggest that ready availability of the necessary elements may lead to the growth of larger framboïds. Also significant in this respect is Skripchenko's (1968) observation that framboïds in organic remains (where bacteria can produce a greater amount of $\text{H}_2\text{S}$) are both larger and contain a much greater percentage of pyrite and less matrix than those farther from the organic material. This illustrates that the amount of $\text{H}_2\text{S}$ present is more critical than the amount of iron and its availability influences the size of the framboïds.

A few measurements of individual grain sizes have also been published. The most commonly occurring grain size appears to be about one micron or slightly less but values as high as 7 microns have been found (Skripchenko 1968). Thus the grain size in framboïds presently under study falls within the range so far reported.

The proportional relationship between framboïd diameter and individual grain diameter, found to hold in the present study and also apparent in other cases (Love and Amstutz 1966 and 1969), has also been mentioned indirectly by Honjo et. al. (1965). They made grain size measurements under an electron microscope on two framboïds. The framboïds had diameters of 6 and 9 microns and individual grain sizes of 0.5-0.8
and 1.2-1.8 microns respectively. In contrast to the above observations Skripchenko (1968) found no relationship between framboid size and grain size in the material he studied. The proportional relationship suggests that increase in overall size of framboids is caused by growth of individual grains. This would be possible if individual grains were only loosely bound during the formational stage. Then growth of each grain would cause an overall expansion of the framboid. However, confirmation of this process is lacking because etching of some larger framboids did not reveal growth zoning in any of the grains. Whether their homogeneous nature is an original feature or due to recrystallization is not possible to say but fine zoning of an order narrower than individual grains can be retained as is illustrated by pyrite encrusting the framboids (Figs. 191 and 208).

Further information regarding the growth of framboids is gained from their relationship where in contact. In such cases the lack of zones of grains surrounding the touching framboids clearly shows that they did not form by outward additions of concentrically precipitated pyrite grains, but rather that they grew by expansion originating internally and when two growing framboids impinged the more rapidly growing one caused an indentation in the other. From a study of the morphology of individual grains Kallickoski and Cathles (1969) also suggested that framboids grew by a process of expansion originating internally.

The structure of overgrowths, which are always composed of pyrite, makes it clear that they were precipitated
layer by layer on the framboids to form concentric rings around each one and eventually around groups of them if they are closely spaced. This secondary pyrite does not normally occur inside the framboid margins but its banding follows closely irregularities on the framboid surface caused by projecting individual grains. Therefore the framboid must have been in its final consolidated form before deposition of the secondary pyrite. If the framboidal pyrite formed first as another iron sulphide then there must have been a time gap between deposition of this and the secondary pyrite. Thus the framboid is a distinct early feature developed by a different mechanism and independant of the secondary pyrite. The secondary pyrite is probably that which gives rise to the larger colloform concretions and which uses framboids as nucleation centres. This view contradicts that of Love and Amstutz (1969), who describe relatively coarse grained pyrite with a radiating structure surrounding framboids and consider it to be part of the framboids. They thus interpret this as a link between framboids and colloform structures. However one of their photographs (fig. 1 no. 6) clearly shows the coarser grained rim surrounding at least three framboids and therefore it appears equivalent to the secondary pyrite of the present study and later than the framboids.

For the most part pyrite is the only mineral found in the individual grains in framboids. However there are a few exceptions. Hydrous sulphides have been reported from framboids in Recent muds (Love and Amstutz, 1969, p. 276).
Lebedev (1967) has reported framboïds composed of sphalerite but the one he illustrates (p. 17) does not appear to have a true framboïdal structure and in fact the photograph appears to have been touched up by pen. Lebedev also reports cassiterite framboïds but provides no illustrations. Framboïds consisting of bornite, chalcocite, chalcopyrite, galena and sphalerite have been reported from the Rammelsberg deposit and other bedded sulphides. These have true framboïd morphology but have been proved by A. Schouten, (1946) to be pseudomorphs after pyrite framboïds. In the Mogul deposit, galena, which occurs in only a few framboïds, appears to replace only the matrix material. On the other hand where sphalerite is the infilling, which occurs even more rarely, the pyrite grains appear to have been partly or wholly replaced but not pseudomorphed by sphalerite. More usually however galena and sphalerite do not enter the framboïds even though the framboïds are embedded in these minerals.

The occurrence of framboïds in pure dolomite of the Dolomite Breccia is probably due to their formation in anaerobic micro-environments within dead organisms. Because of the effects of resedimentation and recrystallization most of the fossils have now been obliterated. Also, in the Dolomite Breccia especially, framboïds occur very close to euhedral pyrite grains of about the same size. This could possibly indicate recrystallization of framboïds to euhedral pyrite but because there are no recognisable intermediate stages it is more probable that the two forms are independant.

The mechanism of formation of framboïds is still unknown
if it is concluded that they are not formed by replacement of sulphur globules in sulphur bacteria. Their occurrence in Recent muds has proved that they can be syn-sedimentary or very early diagenetic. There is good evidence (Berner, 1964, p. 830; also Berner, 1967 and 1969; Kaplan et al., 1963) that pyrite in these anaerobic environments forms in two stages, firstly through formation of amorphous or poorly crystalline FeS from $\text{H}_2\text{S}$ released by sulphate-reducing bacteria and iron from the sediments, and secondly by reaction of the FeS with elemental sulphur to produce pyrite. If pyrite in framboids has gone through these stages the spherical shape may have been brought about by globular aggregation of amorphous or gel-like FeS with later diffusion of sulphur into the globules causing nucleation of pyrite to start at many centres. An excess of sulphur would cause the whole globule to convert to pyrite possibly with an equigranular texture, whereas a sulphur deficiency would result in the formation of a framboid with separated subhedral or euhedral grains. Thus the idea of Rust (1935) and others who postulated crystallization starting at many points in a gel may be at least partly correct. For the above mechanism to operate the sulphur need not necessarily be biogenic in origin as has been shown experimentally by Berner (1969).
Sedimentary features.

(a) Features in Mogul and Magcobar deposits.

A variety of features in the Upper orebody and its host rocks has been described and illustrated. Almost all of these have analogies to features found in orebodies of a similar type in other parts of the world and can be attributed to processes of sedimentation. They provide the strongest single line of evidence relating to the actual mechanism of emplacement of the orebody.

Of particular note is the change near the base of the orebody from a well-bedded, quiet water sequence below to a brecciated sequence above and a lateral confinement of the latter to the area containing the stratiform deposits. The coinciding of an abrupt change in conditions of sedimentation with the base of the orebody is the first point of evidence indicating that the orebody is closely related to the sedimentary sequence.

The rounded pyrite nodules occurring in the Muddy Reef are probably concretions formed during diagenesis in undisturbed sediments. In the shale immediately below the initial brecciated beds the fine laminae of sphalerite (Figs. 26 and 27) are apparently deposited alternately with argillaceous
layers because they are displaced by small pre-consolidational faults and are involved in small-scale contortions associated with the faults. The nebulous structures in thin galena bands (Fig. 28) in the same vicinity also appear to be due to soft-sediment deformation. The inter-lamination of platy galena and clay minerals and their bending around larger pyrite grains (Fig. 29) further suggests a sedimentary origin for the galena (and pyrite). Finely bedded pyrite (Fig. 30) was also involved in soft-sediment deformation. In other places larger aggregates of pyrite (Fig. 31) apparently were fragments interbedded with finely laminated shale-sphalerite layers and caused bowing of the laminations. The occurrence of fossil fragments in pyrite beds below the orebody (Fig. 35) indicate firstly a non-replacement mechanism of pyrite formation and secondly the onset of disturbed water conditions bringing in larger fragments to a zone undergoing fine-sediment deposition.

The overlying layers of unsorted sulphide and other fragments embedded in a clay matrix mark the base of the highly disturbed sequence. The origin of brecciataion is most clearly revealed by features in the Dolomite Breccia and there appear to be two related mechanisms of formation. Firstly the poorly developed stratification planes (Fig. 49), even in the more massive parts, indicate that it is a sedimentary breccia and this is confirmed by the presence of the few fairly thin and continuous beds of shale between coarse breccia layers. The presence of graded bedding in the fine-grained layers (Fig. 55) and also in a coarser layer
Fig. 56) show that turbidity currents played some part in deposition. However as much of the formation is fairly massive and completely unsorted it is unlikely that turbidity current activity was the major means of deposition. The unsorted nature and the occurrence of angular matching fragments (Fig. 52) are better explained by gravity sliding of semi-consolidated sediments. The occurrence of doubly brecciated fragments (Figs. 57 and 58) shows that reseparation occurred at least twice and the lithology of the fragments shows that they are all of local origin and at least some are Waulsortian limestone.

Similar interpretations can be drawn from features within the orebody itself. The basal layers of unsorted sulphide and dolostone fragments embedded in clay minerals (Figs. 32-34) indicate fracturing of sulphide and dolostone layers, their mixing in unconsolidated muddy layers and the gravity sliding of the whole. There is a considerable density difference between the various particles but it has been shown by Evans (1966) that as the fluid pressure approaches the lithostatic pressure corresponding to flotation of the overburden, the shear stress required to move a block approaches zero and thus fragments of high density could be moved with relative ease under these conditions.

In the overlying more massive sulphide the presence of variably sized and shaped fragments of sulphide indicates continuation of a similar gravity sliding process through a period of sulphide deposition on the sea bed. The few intercalated thin beds of shale and finely laminated sulphide
indicate periods of cessation of the disturbed conditions. In the sulphide succession the clay mineral lubricant is usually not present and appears to be replaced by a sphalerite mud represented by the fine-grained structureless sphalerite now found between the mainly pyrite fragments. Further evidence that a sphalerite mud acted as a matrix is given by the deformation of finely banded pyrite-sphalerite-dolomite layers by pyrite fragments (Fig. 44) and also by the fact that, below the orebody, it occurs interbedded with shales and has yielded in the same way as shale to soft-sediment deformation. There is evidence of double brecciation in the sulphide (Fig. 41) identical to that in the Dolomite Breccia but no good evidence of graded bedding was found.

The banding of fragments occurring in the uppermost few inches of the massive sulphide indicates that water currents were the dominant transporting agent at that point and the sharp upper contact which occurs in places indicates at least partial consolidation of the sulphide before deposition of overlying brecciated sediments. However in other places the highly contorted nature of the upper sulphide contact indicates that the sulphide was still unconsolidated when overlying sediments were deposited. The upward-projecting tongues of sulphide could have been caused by weight of overburden on the soft sulphide. Further evidence that the sulphide, including pyrite, was in an unconsolidated state is that pyrite forms a matrix to dolostone fragments above the orebody (Figs. 47 and 48) and may have been squeezed upwards from the orebody.
The mode of occurrence of pyrite in the Dolomite Breccia above the barite deposit is also indicative of its unconsolidated nature at the time of deposition. The small, unbroken, pyrite spheres with a radiating structure (Fig. 62) and the radiating growths of pyrite around dolostone fragments (Fig. 63) show fairly clearly that pyrite crystallized after breccia deposition. The irregularly shaped pod of pyrite with rounded upper surfaces (Fig. 61) seems to represent an accumulation of unconsolidated pyrite in a hollow with the development of globular upper surfaces in contact with mud above. Lower in the same picture however, the thin, fractured band of pyrite parallel to bedding appears to have been solidified before brecciation. The very thin veins of pyrite (Fig. 59) connecting stratiform bands may be due to the squeezing of unconsolidated pyrite along fractures, and the slightly displaced angular fragments (Fig. 60) appear to have been pulled apart in a band of unconsolidated pyrite.

The succession in the vicinity of the B zone contains similar features and the same interpretations can be applied. The only differences are in the greater amount of finely laminated sulphide indicating slightly less disturbed conditions, and the occurrence of Waulsortian limestone above the thinner Dolomite Breccia indicating an earlier cessation of disturbing influences in that area.

The small-scale interbanding of hematite, chert and barite at the base of the Magcobar barite deposit is evidence pointing towards its sedimentary deposition. The small
folds (Fig. 170) and bending of bands around chert concretions (Fig. 172) are indications of soft-sediment deformation there also. The rest of the barite does not show any sedimentary features but the ubiquitous development of spherulitic structures indicates recrystallization which could have been responsible for obliterating earlier features. Because of the uniformity of composition, however, sedimentary features may never have been clearly marked. The trough-like depressions filled by sulphide on the upper surface of the barite could represent washouts in which sulphide was deposited or may simply indicate sinking of a heavy sulphide layer into unconsolidated barite.

(b) Comparison with features in other areas.

A review of the literature reveals that some sedimentary features have previously been described from deposits in a sedimentary environment. Usually there is not a great range of features described in any one deposit but considered collectively a great number of the features found in this study have been found and similarly interpreted before.

The most conspicuous feature of the Mogul deposits, the resedimentation caused by gravity sliding and turbidity current activity, is also the feature most widely referred to in the literature and appears to be the most characteristic feature of this type of deposit. The brecciation has not always been described in the ore itself but occurs commonly in the host rocks either above or below.

What appears to be the most spectacular brecciation occurs in sediments immediately below the large Sullivan
orebody in British Columbia (Freeze, 1966). Occurring there are thick graded beds and intraformational conglomerates with large fragments indicating, according to Freeze, violent marine slides. All the fragmental material is derived from the local basin. In the H.Y.C. deposit, which is a large, stratiform, lead-zinc orebody in Proterozoic sediments in Northern Territory, Australia, both Cotton (1965) and Davis (1969) have described slump breccias and graded turbidites and according to Cotton the breccia beds occur between ore horizons but contain mineralised fragments.

Schneider (1964) and Maucher and Schneider (1967) have described some of the lead-zinc occurrences in the Eastern Alps. The deposits occur interbedded with calcareous sediments on top of a plateau reef and the authors draw attention to the change from a monotonous uniform carbonate succession to a turbulent water succession with graded beds and resedimentation breccias at the base of the ore-bearing beds. The sedimentary breccias include sulphide among other fragments. Schneider suggests that the brecciation is due to tremors associated with volcanic activity which has produced some tuffs.

Schulz (1964) in his theory of origin of lead-zinc deposits in the Calcareous Alps particularly mentions mechanical redistribution of chemically precipitated sulphides with the production of resedimented breccias and other sedimentary structures. He relates the resedimentation to earthquake shocks.

Garlick (1969) in a long article documents in detail
sedimentary structures in the ore-bearing strata of the Zambian Copperbelt. Although this is a dominantly copper orebody it contains features similar to the Mogul deposit including the development of sedimentary breccias. Other examples of sedimentary breccias have been mentioned by Amstutz et al (1961) in the layered lead deposits of Fredericktown, Missouri and by Zimmermann (1969) in the zinc-lead deposits of Shullsburg, Wisconsin.

Some of the above authors have also referred to deformation of fine shale or sulphide layers by sulphide clasts, similar to that shown in Figure 31, but the best illustrations are those of Schneider (1964) and Maucher and Schneider (1967). Other references to similar features are by El Baz and Amstutz (1964) and Amstutz and Bubenicek (1967) in the zinc-lead deposits of Fredericktown, Missouri; by Zimmermann and Amstutz (1964) who describe shale layers bent around spherical barite clasts in the Arkansas barite district; and by Zimmermann (1969) who illustrates sphalerite fragments depressing layering in limestone layers. All the above authors agree that this feature, usually referred to as a geopetal structure, is due to sulphide (or barite) clasts having been deposited mechanically with the enclosing fine-grained sediment.

The nebulously banded structure in fine-grained galena (Fig. 28) has also been described by Zuffardi (1967; 1969) in galena in Silurian lead-zinc deposits in Sardinia. He ascribes both it and features similar to the platy sulphides interlaminated with clay minerals (Fig. 29) to a sedimentary origin.
Sedimentary interlaminations of shale and sphalerite have been described by Davis (1969) in the H.Y.C. deposit in Australia. The laminations are displaced by small faults similar to those shown in Figure 26 and Davis considers the faults synchronous with deposition, thus the sphalerite layers must have been precipitated at the same time as the shale. The sphalerite in Figure 26 is extremely fine grained and provides evidence of cryptocrystalline syngenic layers with properties similar to those of clay layers.

(c) Summary.

In summary there appears to be good evidence that the brecciation and other features in the Mogul deposits and host rocks are of sedimentary origin and that the Upper orebody forms an integral part of the sedimentary sequence. The sulphides of the Upper orebody appear to have been precipitated on an unstable sea floor in the form of a cryptocrystalline mud which, before complete solidification was shaken, possibly by earthquake shocks, broken up and slid downslope under the influence of gravity. Clay mud probably formed the lubricating agent for the basal part but higher horizons, devoid of clay, appear to have slid in a muddy sulphide matrix composed mainly of sphalerite and possibly galena. The shaking and gravity sliding process continued after sulphide deposition ceased and in the less dense carbonate horizons turbidity currents played a small part.

The distance moved during gravity sliding was probably small and the depth of material removed in each slide was also small otherwise fragments of lower horizons would
appear at higher levels which is not the case. Thus the slope down which the material moved must also have been small.

Schneider (1964) and others have attributed instability to effects of nearby volcanic centres but at Silvermines it is more probably related to movement on the Silvermines fault than to the volcanic centre 20 miles distant.

Textures and mineralogy.
(a) Colloform textures.

(1) Previous work.
The literature on colloform texture is voluminous. Early writers suggested that these textures were indicative of the constituents involved having passed through a colloidal or gel stage early in their history. Since then other writers have supplied observational and experimental data to back this up but more often the theory has been adopted as an unquestioned self-evident truth. Because of the latter complacent attitude Roedder (1968) has recently reinvestigated the problem by means of fluid inclusion and extensive microscopic studies and has come the conclusion that gels or colloids play little if any part in the formation of colloform textures. His conclusions have since been essentially corroborated by Haranczyk (1969).

Some of those who have strongly advocated a colloidal origin for the textures are Boydell (1924; 1926), Lindgren (1925), Lasky (1930), Bastin (1950), Chukhrov (1966), Edwards (1965) and Lebedev (1967). The most extensive
experimental work has been done by Lebedev (1967) who then used the results as an aid in the interpretation of naturally occurring textures.

Literature prior to 1968 has been fairly comprehensively reviewed by Roedder (1968). He has compiled the previously published evidence in favour of a colloidal origin and then critically reviewed it by reference to his own work on colloform sphalerite and galena from Pine Point, North-West Territories as well as other parts of North America and Europe. Evidence in favour of a colloidal origin is, according to Roedder, based on four main physical characteristics, which are:

(a) the overall rounded outline supposedly caused by the effects of surface tension acting in a gel;
(b) shrinkage or syneresis cracks due to dehydration of a gel;
(c) internal radial and concentric structures generally assumed due to recrystallization of a gel and diffusion of ions through a gel respectively;
(d) the argument that the colloidal state is a necessary intermediate during condensation.

Roedder believes that colloform textures result from direct crystallization from a liquid and originally based his argument on the observation that large-scale crustification textures in veins, always considered crystallizational, are identical except in size to colloform textures and that there is a gradation between them. Regarding the effects of surface tension he points out that in a globule
the force of surface tension is inversely proportional to
the radius and thus the force should smooth out small sur-
face irregularities and there is a limiting radius above
which the force could not overcome the strength of the gel.
However, observations show that there is a great range of
radii of colloform structures and that small irregularities
are common. Roedder also points out that due to surface
tension small globules coming in contact should flow or
coalesce to form one but in nature touching globules with
sharp reentrant angles are common and this is inexplicable
if they were originally gels.

Roedder's main objection to the evidence supplied by
syneresis cracks is that gels originally contain 50-90% or
more water but the cracks account for a maximum shrinkage
of about 0.2%.

As regards concentric banding and radial crystalline
texture he can find no convincing evidence to preclude
origin through layer by layer precipitation from solution,
growth of radial crystals from solution, or recrystallization
to a radial aggregate from a fine-grained fibrous or banded
aggregate.

Roedder believes that the colloidal state is not neces-
sarily an essential intermediary during condensation except
during formation of the original nucleus on which the mi-
neral can subsequently precipitate directly from solution.
His own observations also show that in radially directed
sphalerite crystals, trace elements show a preference to
precipitate on particular faces, which could not happen if
the sphalerite had crystallized from a gel globule.

Roedder concludes that colloform textures are formed by rapid precipitation due to high degrees of supersaturation, which are readily attained for sulphides when even small amounts of sulphide ion are added to a solution containing metal ions. Under high supersaturation large numbers of sulphide nuclei may form on a flat surface but because of the instability of a flat growing front of crystals, any irregularities will become the centre of more rapid crystal growth with the result that the growing surface will become bulbous or colloform. He thinks that if gels did occur as an intermediate stage they are more likely to have given rise to the cryptocrystalline, almost structureless sphalerite which he has observed.

Haranczyk (1969) basically agrees with Roedder but has reservations about rare occurrences of sphalerite bands in which the grain size is as small as colloid particles (10^{-5} cm.). To explain this and the gradation to coarser grain-sizes he proposes that solutions intermediate between colloidal and true ones can exist and thus a range of textures from colloidal (structureless) to crystallizational (colloform) can form. To the intermediate textures which show traces of colloform structures he gives the term "hemicolloform".

Early experimental work on zinc sulphide and other compounds failed to produce truly amorphous gels (Roedder, 1968, p. 467). However Lebedev (1967) has produced a zinc sulphide which was amorphous to X-rays for up to 3 years
after its production. Lebedev produced the ZnS gel by adding a highly concentrated solution of sodium sulphide to weaker solutions of zinc sulphate. Aging of the gels led to the formation of globules about 0.3 microns in diameter, globulites (aggregates of globules) and oolites (concentrically banded spheres) up to 4 m.m. in diameter, all of which eventually became crystalline. Of particular note is the fact that gel globules in contact became only partly coalescent. Lebedev experimented with the addition of electrolytes such as NaCl and CaCl₂ to the gels and found in general that crystallization was faster and bigger oolites, up to 8 m.m. in diameter, were produced. None of the large oolites are illustrated but a cross-section of a smaller one shows a structureless centre surrounded by one fibrous layer. Experiments with mixed ZnS-PbS gels showed that sphalerite always formed as globules or globulites either as a distinct layer or embedded in galena. Galena on the other hand always crystallized rapidly into skeletal or dendritic crystals and did not form globular structures.

Roedder (1968) criticised Lebedev's work on the grounds that he did not explain why earlier experimental work failed to produce gels. Another criticism is that Lebedev attributed to a colloidal origin a great variety of forms which bear no relation to those he produced. Many of these forms are of a type which Roedder considers almost certainly formed from solution.

Further confusion on the origin of colloform textures arises when some of the work of physical chemists on gels
and colloids is considered. It is generally agreed (Glass-
tone and Lewis, 1966) that gels can be divided into two
types: elastic and non-elastic. The elastic gels are usu-
ally formed of organic substances and can occur in the form
of globules with elastic properties. Inorganic substances
usually possess non-elastic properties and do not globu-
late. They harden in the shape of their confinement as a
rigid mass of thin walled, sub-microscopic tubes. From this
evidence it would appear impossible to obtain globules in
a sphalerite gel or any other inorganic gel. However natu-
rally occurring silica gel, the existence of which is not
questioned (Roedder, 1968; Lindgren, 1933; Khopf, 1916),
has been found by Lebedev (1967) to occur in both the form
of minute tubes and globular structures.

There is also some confusion in the literature conser-
ring the formation of concentric bands by the Liesegang
ring process and this has been mentioned by Roedder (1968).
It should also be pointed out that those who favour formation
of banding by diffusion through a gel propose a reaction
between the assumed sulphide disperse phase and ions diffu-
sing through the gel. However the experimental work demon-
strating the formation of Liesegang rings (Hatschek, 1922)
showed that rings were formed in an organic gel due to a
solution diffusing through it and reacting with the disper-
sion medium (a different solution) in the gel. The disperse
phase, which may be gelatin or a similar substance, takes
no part in the reaction.

Roedder's (1968) work is a significant contribution to
the literature on colloform textures and it clearly indicates that they can form by crystallization from solution. However, small-scale colloform textures originating in a gel, for example silica gel, are known to occur, although they are probably rare, and the problem of distinguishing with certainty between the two types still remains.

(2) Discussion of evidence from the Upper orebody.

An examination of the wide variety of colloform textures contained in these deposits reveals much evidence bearing on their mode of formation. Many of the features are similar to ones described by Roedder (1968) and interpreted by him as indicating crystallization from solution.

Colloform pyrite was not described by Roedder but some features displayed by it here are similar to those he found in sphalerite and can be interpreted in the same way. Colloform structures of the dendritic type have a clearly defined elongation parallel to the main fibre direction (Figs. 72-76) and this combined with their branching pattern indicates that they grow from a nucleus at the point of convergence of the branching fibres. It is clear, especially in Figure 71 and 74 that there was a well defined preferential growth direction and, although it cannot be proved because of later brecciation, it is tempting to believe that they grew vertically upwards from a surface. Evidence of vertical growth of colloform, bedded sphalerite has been demonstrated by Zimmermann (1969, p. 256).

The gradation of the dendritic type of pyrite spheroid into the concentric type (Figs. 76-78) and the occurrence of
elongated concentric types (Fig. 79) shows that the two types are closely related and probably of the same origin. The elongation tendency is caused by a thicker development of bands along the axis of curvature of the bands (Fig. 80) and again the elongation suggests upward growth of the structures. Evidence that some colloform pyrite crystallized at the same time as surrounding granular pyrite is provided by the merging of the two types and the band of granular pyrite contained within the colloform pyrite in Figure 81. Such relationships would appear impossible from crystallization of a globule of gel.

Further evidence of direct crystallization is provided by interbanded coarse radial and very fine grained pyrite (Figs. 82-86). The coarse radial crystals (Fig. 84) end without crystal terminations against the very fine grained band. On the outer side of the fine grained band many groups of radiating crystals nucleate and grow outwards. The lack of crystal terminations suggests partial solution of the inner radiating crystals before deposition of the fine-grained band and if solution or leaching did occur it provides good evidence of the presence of true solutions because, according to Roedder (1968), colloidal solutions cannot remove by leaching material the same as the dispersed phase. Figures 85 and 86 demonstrate the deposition of a fine-grained layer without leaching of the underlying zone and again renewed nuclei of radiating crystals occur on the outer edge of the fine grained zone. Thus there is evidence of sequential deposition of the bands from the inside outwards.
and also evidence of the presence of true solutions.

Other features in concentrically banded pyrite structures which illustrate the sequential deposition of the bands, probably by direct precipitation, are as follows: a readily distinguishable fine-grained band occurring at the same location in several colloform structures (Fig. 89); crystal projections from an outer colloform surface (Fig. 90); bands of very variable character with a diffuse outer boundary (Figs. 91 and 92); euhedral crystals enveloped by concentrically banded layers and the reverse (Figs. 94 and 95 resp.).

One feature which appears to be related to recrystallization of a very fine-grained pyrite after deposition is the patchy development of a coarser-grained pyrite in a colloform structure (Fig. 87).

The concentric internal structure of the rod-shaped type shows that it probably originated in the same way as the more spherical concentrically banded structures.

The relatively large, radiating pyrite crystals (Fig. 99) are clearly due to direct crystallization as shown by their euhedral crystal zoning. The zoning is analogous with concentric banding in smaller colloform textures and thus the texture as a whole seems to be a link between these and coarser crustiform structures. This supports evidence cited by Roedder (1968) from his study of similar textures in sphalerite.

Features of more doubtful origin are the black irregularly shaped patches of black pyrite (Figs. 96 and 97). Their extremely fine grain-size, porous nature and irregular
shape could be evidence of origin by crystallization from a gel.

Turning to colloform textures in sphalerite and galena, the radial and concentric arrangement of these two minerals (Fig. 102) seems to indicate their contemporaneous deposition but with their relative proportions varying with time. A closer examination of the texture (Fig. 107) reveals formation of a sphalerite band by nucleation at, and radial growth from, several points on the inner galena band. Figure 106 shows sphalerite nucleating on an inner galena band but also reveals galena crystals growing radially outwards parallel to the sphalerite growth direction. Both the above features have also been described and similarly interpreted by Roedder (1968).

The skeletal appearance of galena in Figures 103-105 appears to be due to sectioning across a series of very delicate cubic hopper crystals. The galena seems to form the nucleus for later growth of radiating and concentrically banded sphalerite.

Globular texture which is found in both sulphides and carbonates provides a problem. In order for the spherical shape to develop the globules must have been suspended in a fluid medium and free to grow in all directions. The monomineralic globules resemble those produced experimentally by Lebedev (1967) from gels in that they are partly coalescent and in places form "chain globules" (Figs. 109, 111, 136 and 137). However globules composed of two or more minerals display delicate concentric banding (Figs. 1
113, 138 and 139) very similar to that in larger colloform structures indicating a similar origin. Particularly convincing with respect to layered deposition is the outer band of sphalerite surrounding two touching galena globules (Fig. 139). Another feature pointing to a crystallizational origin of globules is the tiny skeletal galena crystal which appears to form a nucleus for each of the pyrite globules in Figure 129. An indication regarding the suspension medium is provided by the sphalerite globules surrounded by structureless, fine-grained sphalerite (Fig. 115) which suggests that the globules may have formed within a crypto-crystalline sphalerite mud.

(3) Paragenetic sequences in the Upper orebody.

Concentric interbanding of different minerals is taken by Roedder (1968) to be sufficient proof of sequential precipitation from solution. Proof of this conclusion has already been illustrated here for galena-sphalerite banding and using the same argument the same conclusion can be drawn for sphalerite-pyrite banding (Figs. 119 and 120). If it is assumed that the banding represents the sequence of deposition then it becomes apparent that there is no well defined paragenetic sequence. Both sulphides and carbonates are interbanded in various orders and repetitions are common. Many of the discrete colloform pyrite structures probably formed early but there are exceptions to this (Figs. 136 and 139), shown by concentric bands of pyrite enclosing apparently earlier formed sphalerite - galena globules. Late-crystallizing pyrite is illustrated by its occurrence.
interstitial to globular sphalerite (Fig. 123) and interstitial to globular sphalerite-galema (Fig. 140). Two quite different stages of pyrite crystallization are illustrated by Figures 121 and 122 (122 an enlargement of part of 121 in oblique light). Concentrically banded sphalerite nucleated on and grew from the large dendritic pyrite structure. A later pyrite formed in the interstices between the colloform sphalerite spheroids and has taken on a scalloped appearance in vertical light (Fig. 121).

(4) Other colloform textures.

The most noticeable feature of colloform textures in the upper part of the Lower orebody is their lack of variety in comparison with those of the Upper orebody. Of particular significance is the absence of discrete dendritic and elongated concentric pyrite types, which, it was suggested, grew upwards from a horizontal surface in the Upper orebody. The irregularly banded colloform textures which do occur are assumed to have formed by sequential crystallization of various minerals on evidence derived from similar features in the Upper orebody.

The banded colloform textures in veins in the K zone are analogous with the concentrically banded type in the Upper orebody and, assuming the same growth mechanism, they must have grown from the vein walls inwards. The variation in the amount of curvature in the bands with local planar banding supports Roedder's (1968) statement that flat-growing crystal fronts are unstable and usually develop a curvature. As in the Upper orebody there is a varied paragenetic sequence
but barite, if present, always occupies the centre of the vein and therefore crystallized after the sulphides.

No detailed study was made of the colloform textures in the barite deposit but because in most places the barite occurs in interfering spherulites, with relatively coarse unzoned crystals, it is possible that it recrystallized from a cryptocrystalline barite mud or from a gel. Spherulites are indicative of recrystallization and commonly form in silica glass, a process which has been duplicated in the laboratory (White and Corwin; 1961). Barite gel can be formed in the laboratory (Lindgren, 1925) but is always crystalline in nature, so whether it started as a gel or cryptocrystalline mud is uncertain. However, its originally soft nature is indicated by depressions formed in its upper surface by overlying sulphides.

(5) Brecciation of colloform textures.

Fragmental pieces of colloform pyrite and sphalerite are common (Figs. 70, 80 and 103). In all cases the fractured edges are sharp and straight and there is no evidence of deformation by bending. This is further evidence that the colloform textured aggregates never were in a soft gel state.

(6) Recrystallization of colloform textures.

The granular texture revealed by etching colloform sphalerite appears to have developed secondarily without obliterating the earlier radial and concentric textures. This shows that recrystallization did not significantly change the original iron or trace element distribution
responsible for the early colour banding. Recrystallization of galena has also occurred without destruction of early structures as is shown in Figure 113 by coarse cleavage planes which cross-cut an area of galena containing fine, concentrically interbanded sphalerite. A similar recrystallization is not apparent in pyrite. An approach to a granular texture is displayed by some colloform aggregates but a radial tendency is also displayed (Fig. 92) so the texture may well be original.

(7) Summary.

Examination of the colloform textures in sulphides reveals much evidence that direct crystallization from solution was responsible for their formation. The most important points of evidence are: branching growth of pyrite; the preferred growth direction of discrete dendritic and concentric pyrite structures indicating upward growth from a surface; interbanding of bands of very different grain size and habit; possible solution before precipitation of a succeeding band; interbanding of different minerals; distinct evidence of nucleation of crystals of one band on the outer surface of the previous band; sharp angular fractures on brecciated colloform aggregates.

However the possible presence of some gel cannot be ruled out because some features such as the black, fine-grained, porous pyrite aggregates are difficult to explain and may be due to crystallization of a gel. Gel origin of the barite also is a possibility.

Colloform textures are evidence of relatively rapid
precipitation (Roedder, 1968) and their gradual disappearance downwards from the Upper to the Lower orebody indicates a decrease in the precipitation rate compatible with a change from a seabed to a sub-sediment environment of deposition.

Colloform textures can be used to determine paragenetic sequences which are inconsistent and often repetitive.

(b) Other textural features.

(1) Upper orebody.

In the Upper orebody some of the sulphide minerals appear to have crystallized diagenetically or later in the Dolomite Breccia. Evidence of this is supplied by the zoned subhedral to euhedral pyrite crystals (Fig. 141), and the interstitial nature of coarse-grained sphalerite with dolomite (Fig. 152). Other diagenetic features appear to be the exsolution of chalcopyrite from sphalerite, exsolution of boulangerite from galena (Fig. 155) and fracturing and veining of colloform pyrite aggregates by sphalerite (Fig. 148), and galena. The formation of marcasite in large irregular crystals in pyrite also appears to be a secondary feature and the intense fracturing (Fig. 151) commonly associated with it may be due to expansion because of its greater volume. Lastly the formation of fibrous quartz growths on some pyrite crystals may be secondary and due to shrinkage or redistribution of pyrite during recrystallization.

(2) No. 3 zone.

The occurrence of fragmented, planar banded and colloform pyrite and the delicate, dendritic pyrite-galena
aggregates indicates that the zone has a similar origin to the Upper orebody. The coarse-grained nature of much of the sphalerite may be due to its remobilization during brecciation.

(3) Lower orebody.

The sharp angular fracturing and very minor displacement of the interveining blocks (Figs. 158 and 159) in the Lower Dolomite indicate that the fracturing and thus mineralization occurred after the rock had reached the brittle stage. Earlier dolomitization of the formation may have been responsible for producing a porosity which could explain the fairly widespread sulphide occurrences interstitial to dolomite grains. The zoning of the veins and local inwardly directed crystal growth indicate filling of the veins from the walls inwards. However the inconsistent order of mineral zones shows that a universal paragenetic sequence cannot be applied. This implies that precipitation was a function of depth or some parameter other than time and may relate to the lead : zinc ratio which, when examined on a small scale, shows erratic variations with depth.

Features such as formation of galena and chalcopyrite along parallel fractures in pyrite and the irregular fracturing and veining of pyrite by sphalerite (Figs. 165 and 166) indicate solution and replacement of pyrite. However in other places there is evidence of secondary deposition of pyrite, shown by cross-cutting layering relationships (Fig. 163).

In the more shaly pyritic layers the bedded nature of
the pyrite indicates that it formed syngenetically but the pseudomorphing of euhedral dolomite grains in these layers by galena and chalcopyrite (Fig. 164) indicates a later origin for these minerals.

Thus the Lower orebody can be envisaged as a vein and interstitial filling deposit in a partly porous dolomitized formation. Sequential mineral zoning and solution and replacement of earlier minerals did occur but do not fall into a simple paragenetic sequence. Sedimentary sulphide in the orebody is minor and probably consists of thin pyrite layers only.

**Minor elements.**

(a) Iron content of sphalerite.

One significant feature of the iron content of sphalerite is that it is everywhere low (less than 4 wt. percent) and normal for sphalerite in carbonate sediments. Also significant is the greater spread of values in sphalerite of the Upper orebody over that in the Lower orebody. This may indicate more fluctuating conditions of precipitation in the Upper orebody.

The relationship of iron content to the colour of the sphalerite seems fairly consistent in that all white and colourless sphalerite was found to have close to zero weight percent iron and darker varieties 2-4 wt. %Fe. The relationship of exsolved chalcopyrite to areas of dark and relatively high iron (3-4 wt. %) sphalerite is particularly consistent. Edwards (1965) has described similar relationships
but Roedder and Dwornik (1968) in a study of finely banded colloform sphalerite found that there was no relationship between colour and iron content. However, they were working in a slightly lower iron content range (0-2.9 wt%). The latter authors suggested, but were unable to prove, that a deviation from stoichiometry may be responsible for colour variation. Manning (1967) has shown that the amount of iron (Fe\(^{3+}\)) octahedrally bonded in interstitial positions in the sphalerite lattice can affect the colour. The ratio of iron (Fe\(^{2+}\)) in substitutional sites to that (Fe\(^{3+}\)) in interstitial sites is about 10 but may not remain constant. This uncertainty allows enough latitude for arguments both for and against the colour of sphalerite reflecting the total iron content. Possibly the ratio is variable and accounts for the conflicting observations by different authors.

(b) Silver and antimony content of galena.

The silver content of galena is normal in that the values fall within a commonly occurring range (Fleischer, 1955). The fact that the silver content predicted from assay figures is fairly close to that determined by analysis of the galena confirms the absence of silver minerals, but does not exclude the possibility that silver occurs in trace amounts in other minerals. The better defined, direct relationship between lead and silver in the Lower orebody (Fig. 13) indicates a more uniform distribution of silver in galena, and thus more uniform conditions of deposition there than in the Upper orebody.

The hardness of minerals has been related to trace
element content by Parnamaa (1963) and he states that for galena the hardness increases with increase in the silver content. The values of hardness obtained for galena from the Upper and Lower orebodies could be related to the silver content, and the wider variation occurring in the Upper orebody could, as above, relate to more variable conditions of deposition. However the effects of antimony on the hardness are probably also important and it is significant that a high antimony content was found in galena from the B zone where galena also has a much greater hardness. Further trace element and hardness determinations are needed to establish this correlation.

Also of importance regarding hardness and trace element distribution is the occurrence of boulangerite in galena. Boulangerite occurs only in the Upper orebody where it is erratically distributed. Its crystallographically controlled growth indicates its exsolution from galena and if so this could also be indicative of non-uniform distribution of trace elements and rapid original crystallization of galena in the Upper orebody.

In summary, trace element and hardness studies on galena point to more uniform conditions and probably a slower rate of deposition in the Lower than in the Upper orebody. (c) Nickel and Cobalt content of pyrite.

Literature referring to the nickel and cobalt content of pyrite has been comprehensively reviewed by Loftus-Hills and Solomon (1967). Carstens, Hegemann and Talluri (reported in Fleischer 1955) have all found that in pyrite of
sedimentary origin the cobalt content is less than the nickel content and usually less than 100 p.p.m. Davidson (1962) also has proposed that sedimentary pyrite can be distinguished from pyrite of magmatic-hydrothermal origin on the basis of the Co:Ni ratio which he said should be less than 1 in sedimentary pyrite. Loftus-Hills and Solomon (1967) however caution against careless use of this relationship because highly concentrated pyrite deposits in a sedimentary environment have probably accumulated under special conditions with the possible addition of elements from volcanic or other emanations which would obscure the original sedimentary Co:Ni ratio. The above authors then go on to show from their own extensive studies that pyrite of sedimentary origin (from shales) can be distinguished from other types of pyrite by its high Ni content and low Co:Ni ratios. Pyrite associated with granitic rocks has a similar Co:Ni ratio but values for both elements are much lower (average Ni about 16 p.p.m.).

The cobalt and nickel content of pyrite from the Mogul Upper orebody clearly corresponds to the set of values given by Loftus-Hills and Solomon (1967) for pyrite of sedimentary origin although the Mogul values for each element are slightly lower and the Co:Ni ratio slightly higher. The correspondence is especially significant because of the concentrated nature of pyrite in the Mogul deposit and may be taken as an indication that the deposit is uncontaminated by magmatic solutions, i.e. the nickel and cobalt have been derived entirely from the environment of sedimentation. Further
confirmation of this comes from a study by Tourtelot (1964) which shows that the Co:Ni ratios in marine and non-marine shales are similar to those in the Mogul pyrite but with slightly lower absolute values. Krauskopf (1956) shows similar ratios in sea-water. Krauskopf (1956) has also shown that in the sedimentary environment nickel and cobalt (also V and Mo) are concentrated by organic activity rather than by adsorption or sulphide precipitation. Thus the content of Ni and Co in pyrite in the Upper orebody is an indication that organic activity played a part in its formation. This agrees with the results of sulphur isotope work (see section on sulphur isotopes).

**Metal distribution.**

In the Lower orebody the occurrence of copper in association with quartz in the west and the occurrence of barite in the east suggests an east-west zoning with the west end nearer the source of mineralization if compared with zoning in other epigenetic deposits, for example the deposits of Cornwall, England (Park and MoDiarmand, 1964; pp. 164-172). In conjunction with the upward change in the zinc:lead ratio the overall direction of zoning appears to be upwards and eastwards.

The upward enrichment of zinc is unusual when compared with lead-zinc zoning in other epigenetic deposits in carbonate rocks, for example, in the Central Kentucky district (Jolly and Heyl, 1964) and the North Pennine ore deposits (Dunham, 1934) in both of which galena occurs farther from
the centre of mineralization. In these deposits however zoning is developed over several miles rather than over a few hundred feet as in the Mogul deposit. In the case of the Cornish deposits, where zoning is on a smaller scale, lead and zinc are closely associated although zinc may still have a tendency to occur nearer the centre. If chloride complexes play a part in the transport of metals, as fluid inclusions from many lead-zinc deposits suggest, then it is possible for reversals to occur in the solubilities of lead and zinc sulphides depending on the concentration of Cl (Krauskopf, 1967; p. 503). Thus it would be possible to explain a change from lead nearer the source to zinc farther from it. The change in the zinc:lead ratio, however, is not uniform as is shown by large fluctuations when the ratio is examined on a smaller scale. Thus, as well as an overall change in conditions upwards, there were small-scale fluctuations which produced large local reversals in the overall trend.

In the Upper orebody zinc is the predominant base metal as it commonly is in other deposits of sedimentary origin in carbonate rocks. The isopach map of the Upper orebody (Fig.9) suggests that its thickest part near the south end is at the source of mineralization and that there is some development of channels radiating northwards from the source. If this is so then here also lead tends to be concentrated near the source and zinc at a greater distance from it. Whether the zoning can be explained in the same way as it was for the Lower orebody is uncertain because if, as
sedimentary and other features indicate, the sulphides were deposited as sediments, then conditions of deposition are likely to be radically different from those in the Lower orebody. If they were deposited as sediments and subsequently reworked by gravity sliding and current activity then it might be expected that the denser galena would remain nearer the source than sphalerite. Metal zoning in this type of orebody has rarely been described in the literature but Freeze (1966) has described a similar pattern with lead nearer the source in the Sullivan orebody. This great orebody contains many features indicative of a sedimentary origin but Freeze considers it a replacement of sediments and has therefore had to invoke a complex chemical system to explain the zoning.

The continuity of change of the zinc:lead ratio from the Lower orebody to the southern end of the Upper orebody indicates that the two orebodies are closely connected and that the zinc and lead in both have a common source. The relationship of iron to lead and zinc is uncertain because of its great abundance in the Upper orebody and scarcity in the Lower orebody. Descriptions of numerous paragenetic relationships show that pyrite is precipitated early and work by Barnes (1962) has shown iron to be less soluble than zinc or lead in any anion complex, further confirming that it should be precipitated early. Thus if iron was brought up through the fracture system of the Lower orebody with zinc and lead, theoretically the Lower orebody should contain a much greater amount of pyrite. Its minor occurrence there
leads to the suggestion that iron in the form of pyrite in
the Upper orebody was derived from the sedimentary environment
and not brought up through the fracture system or at least
that a large amount was added from the sedimentary environ-
ment. Evidence in support of this suggestion comes from work
done by Schultz (1966) on ironstones close to the Tynagh
base metal deposit. He concluded that the iron and silica
in the ironstones were derived from weathering of uplifted
Lower Carboniferous and Devonian rocks rather than from
volcanic exhalations. Thus if the Lower Carboniferous sea
did contain a large amount of iron in solution the iron
could have been precipitated as a sulphide in a reducing
environment and thus explain the enrichment of iron in the
Upper orebody. As additional evidence White (1967) has fou-
nd that many thermal springs associated with mercury and
base metal deposits contain very little iron although a
few do contain appreciable amounts.

Sulphur isotopes.

Comprehensive accounts of sulphur isotope fraction-
ation processes have been given by Sangster (1968) and
Jensen (1967). In a reducing, submarine environment sul-
phate-reducing bacteria consume ocean water sulphate and
produce \( \text{H}_2\text{S} \), the sulphur of which has a reduced prop-
ortion of the isotope \( \text{S}^{34} \) or a relative enrichment in
\( \text{S}^{32} \) with respect to sea-water sulphate. If metal sulph-
ides are ultimately produced from the \( \text{S}^{32}\)-depleted \( \text{H}_2\text{S} \)
they will also have a reduced $S^{34}$ content, i.e. a lower $\delta S^{34}$
value where

$$\delta S^{34} = \frac{(S^{34}/S^{32}) \text{ Sample} - (S^{34}/S^{32}) \text{ standard}}{(S^{34}/S^{32}) \text{ Standard}} \times 1000$$

and where for the standard, which is sulphur from the Canon Diablo meteorite, $S^{32}/S^{34}=22.22$ (Sangster, 1968). The
reduction in $S^{34}$ depends on the metabolic rate of the bac-
teria and if this is low, fractionation is greater. Theo-
retically it has been predicted that the maximum reduction
in $S^{34}$ should be 73 permil (Jensen and Whittles, 1969) and
laboratory experiments have produced reductions of up to
40 permil. In nature reductions of up to 80 permil have
been found (Jensen and Whittles).

Corresponding to a reduction of $S^{34}$ in $H_2S$ there is
an increase in $S^{34}$ in the coexisting sulphate. However,
because the reservoir of sulphate ion in the surrounding
sea-water is so large, the amount of $S^{34}O_4$ produced from
bacterial activity is negligible and unless there is re-
stricted circulation the amount of $S^{34}$ in sea-water sulphate
remains constant. Sulphates which are precipitated from sea
water thus have a $\delta S^{34}$ value equal to that of sea-water of
that period.

The sulphides and sulphates of the Silvermines district
are strongly indicative of their having been deposited in a
marine environment. The $\delta S^{34}$ values of 15.1 permil in the
Magcobar barite are almost identical to the value of sea-
water sulphate (14.5%) for the Lower Carboniferous
(Sangster, 1968) showing that the sulphate ion was derived from sea-water. Sulphides from the Upper orebody have a wide range of δS34 values, which is typical of bacterially derived sulphur, and they also show extreme fractionation with respect to sea-water sulphate which indicates a slow rate of bacterial fractionation. The wide range and extreme fractionation are not typical of Mississippi Valley type lead-zinc deposits, which usually have a more restricted range and δS34 values above -10‰ (Sangster, 1968; Jensen, 1967). They do however resemble results obtained by Jensen and Whittles (1969) from bedded pyrite in shales in South Australia. They found a maximum fractionation there of about 48‰ which compares with the maximum of 44‰ found in the Mogul deposit. They also showed that the pyrite was purely sedimentary with no hydrothermal or magmatic associations.

Kaplan et al. (1963) and Vinogradov et al. (1962) have both reported wide δS34 values and extreme fractionation of sulphur in pyrite from recent marine sediments off southern California and in the Black Sea respectively.

Isotope ratios for both pyrite and sphalerite in the Upper orebody are comparable and show that both minerals crystallized under the same conditions. Further work needs to be done to determine if the same holds true for galena but textural relationships indicate contemporaneous deposition with sphalerite and pyrite and therefore all three minerals probably had the same source of sulphur.

Sulphur isotope ratios for galena and sphalerite in the Lower orebody, with an average δS34 value of close to zero,
indicate a lesser degree of fractionation. Although the
determinations are too few to draw certain comparisons they
bear a much closer similarity to results from Mississippi
Valley deposits (Jensen, 1967; p. 150) which are also con-
sidered to have formed from bacterially reduced sulphur.

The $\delta^{34}S$ values (16-19‰) for barite in galena-barite
veins in Devonian sandstone at Shallee (Fig. 2) are slightly
higher than for Lower Carboniferous sea-water sulphate but
considerably lower than for Upper Devonian sea-water (22.3‰).
This could be explained by the precipitation of barite in
fractures saturated by overlying Lower Carboniferous sea-wa-
ter, and in such an environment restricted circulation could
permit some enrichment in the $S^{34}$ isotope.

In summary the sulphur isotope results indicate that
sulphides of the Upper orebody accumulated slowly in a basin
with reducing bottom conditions in which bacterial activi-
ty was prolific. Nearby, at the same time, barite was being
deposited by sea-water sulphate in a freely circulating
environment. Underlying the seabed sulphides and sulphates
were also being precipitated in sea-water saturated bedrock
where bacterial activity was less intense than on the surface.

Precipitation, transport and source of constituents.

(a) Precipitation.

The problem of the formation of pyrite in the sedimen-
tary environment has already been mentioned in discussion of
the origin of framboids. Because of evidence already pre-
sented it appears that pyrite in the Upper orebody is
sedimentary in origin and may have a similar mechanism of formation to that in framboids.

From work on recent sediments there is abundant evidence that pyrite in other forms as well as in framboids is commonly formed under reducing conditions in fine-grained clastic sediments. For example Kaplan et al. (1963) found that pyrite formed most rapidly in the uppermost layers of sediments in basins off the coast of southern California and Harmsen (1954) has reported pyrite in surface mud layers off the coast of Holland.

Although there is no doubt that pyrite can form in a sedimentary environment there is controversy over the process of its formation. There is considerable support for the reaction between an intermediate product, FeS and elemental sulphur to produce pyrite but Harmsen (1954) thought this unlikely because there was little evidence of free sulphur in recent sediments. He thought that pyrite was more likely to be formed by a reaction between original hydrated iron oxide and H₂S. Kaplan et al. (1963) favour the reaction of FeS and elemental sulphur because, although they too found only a very small amount of free sulphur in recent sediments, they thought it was being continuously formed in the sediment and used to form pyrite.

Experimental work has helped in the solution of the problem. Berner (1967) showed that pyrite could be formed by a reaction between hydrated iron oxide and H₂S but only at a pH of 4 or lower which is lower than in any known basin of pyrite formation. He thought that addition of sulphur
to FeS was a more probable process and showed experimentally that it did occur and that elemental sulphur could be produced by oxidation of $H_2S$ by ferric iron.

Roberts et al. (1969) have shown that pyrite can form much more rapidly from ferrous iron and the soluble ion $S_2^-\text{O}_2$ than it can from FeS and solid sulphur and suggest the former as a more probable process which occurs at a pH of less than 6. One of the objections to the latter process is that it occurs rapidly at room temperature only at low pH values. However, Berner (1969) has shown that by increasing the temperature to $65^\circ C$ the reaction can take place easily at a neutral pH.

Whether sulphur is in ionic or element form there appears to be sufficient evidence to show that pyrite is formed by the combination of sulphur with FeS which in turn was formed by a reaction between hydrated iron oxide and bacterially produced $H_2S$.

The above process could have produced at least some of the pyrite in the Upper orebody and the intermediate non-crystalline FeS stage could explain some of the flow or injection features in the pyrite which could not have formed after final crystallization of pyrite.

The reason for variation in texture from place to place is still unknown. Not all pyrite need necessarily have formed by the same process. Concentrically banded and radial types of colloform structures could have developed by precipitation from solution on to pyrite nuclei, such framboids, and textural features in fact suggest that...
is the more likely process.

The sedimentary precipitation of lead and zinc has rarely been mentioned in the literature but is probably not so complex as the formation of pyrite because the sulphur ion $S^{2-}$ is capable of precipitating many heavy metals as sulphides (Krauskopf, 1967, p. 271). Baas-Becking and Moore (1961) have shown experimentally that sphalerite and galena can be formed from the metal carbonates in solution and bacterially reduced sulphate.

(b) Transport.

Little can be said about the mode of transport of metals to their sites of deposition without investigations of fluid inclusions which would reveal the composition of the fluid from which the minerals crystallized. The only indication from work done is that the metals were unlikely to have been carried as sulphides or sulphide complexes because sulphur appears to have been provided at the site of deposition by bacterial activity. Considering abundant evidence from thermal water and fluid inclusion studies done elsewhere and summarized by White (1967; 1967a), it appears probable that chloride complexes may have played a major part in metal transport.

(c) Source.

The source of sulphur for all the orebodies is apparently sea-water and there is also evidence to show that at least some of the iron may have come from the sedimentary environment. However indications are that base metals, barium and minor elements were introduced to the veins and
basin of sedimentation from the sub-surface, but without the help of information provided by lead isotopes their source is uncertain. There are at least three possibilities: Lower Carboniferous shaly rocks down dip from the orebodies, Lower Palaeozoic shales and greywackes, or Lower Carboniferous volcanic rocks.

Sources of metal-bearing fluids have been discussed by Krauskopf (1967a) and he shows that shales could provide the quantities of metals required. Jackson and Beales (1967) have proposed that metals in the Pine Point orebodies were derived from laterally equivalent shales by connate water circulation, but more recent lead isotope work (Cumming and Robertson, 1969) does not support the theory. In the Mogul orebodies there is no evidence bearing on whether or not metals were derived from laterally equivalent rocks.

The thick sequence of Lower Palaeozoic shales and greywackes underlying the ore-bearing strata provide a more likely source of the metals as is indicated by the occurrence of scattered sphalerite-galena veins in the cores of the Silurian inliers. Russel (1968) has shown from a series of analyses that the Lower Palaeozoic rocks in Ireland contain on the average 20, 100, 45 and 700 p.p.m. respectively of Pb, Zn, Cu and Ba and thus he thinks they could be the source of the ores.

A tenuous connection with igneous rocks is provided by the fact that mineralization at the Gortdrum mine is associated with feldspar-porphyry dykes (Thompson, 1966). A slightly more likely possibility is that heat provided by ingeous
intrusion started a circulation of connate water which leached and carried metals to the surface. However, at present evidence of intrusion in the neighbourhood of the Mogul mine is lacking.

Russel (1968) has suggested that igneous intrusion and consequent leaching of Lower Palaeozoic rocks by connate water is the process leading to the deposition of the Mogul orebodies. He attributes igneous intrusion to major crustal fractures extending to the mantle and associated with continental drift. However evidence for his major north-south fracture zones is extremely scanty and there is no need to appeal to such large-scale features as a source of magma and cause of ore deposition.

**Summary of ore formation processes.**

From the foregoing evidence it appears that the stratiform sulphide and sulphate bodies were deposited on the sea floor in an area of hot spring activity and that the vein-type sulphides of the Lower orebody represent fracture filling deposits along the paths of ascent of thermal waters. The constituent metals may have been derived from the thick sequence of Lower Palaeozoic argillaceous sediments underlying the Carboniferous rocks and carried upwards by connate waters possibly set in circulation by heat associated with igneous intrusion. The major and minor fault planes in the Silvermines Fault system may have acted as channelways for the solutions.

Deposition of sulphides began when the rising solutions
came into contact with Lower Carboniferous rocks saturated with sea-water containing sulphate-reducing bacteria and this gave rise to the vein fillings.

The thermal waters reached the surface in a minor basin down-dropped along the Silvermines Fault. Circulation in the basin was restricted and bottom conditions were reducing. The metals, which were transported possibly as chloride complexes, were precipitated in the basin as sulphides on coming in contact with $H_2S$ produced by sulphate-reducing bacteria. At the same time some iron carried by sea-water and a small amount of nickel and cobalt concentrated by organic material were added to the precipitating sulphides. Barium migrated outwards from the basin and was deposited as barite in more freely circulating sea-water at a higher level. Deposition of sulphides and barite continued over a relatively long period of time but was interrupted on occasions by sudden sea-floor movements, probably associated with displacement on the fault, which caused brecciation, slumping and resedimentation of the deposits.

Formation of the basin of deposition may be attributable to the rising thermal waters which were probably slightly acidic and could have dissolved some of the underlying limestones leading to subsidence in the sea-floor. Repeated subsidences during deposition could have added to instability and contributed to the brecciation of the ores.

Instability of the sea-floor persisted for a time after ore deposition ended and caused slumping and resedimentation
of overlying limestones. After consolidation there appears to have been a "settling" or contraction in the sulphide body which is marked by vertical slip planes in the ore and coarse brecciation in the Dolomite Breccia above.

Whether the sulphides in the B zone are due to migration of metals along a trough in the sea-floor or to a second hot spring system immediately underlying the zone is uncertain.

Temperatures of deposition are unknown but the fact that even the vein fillings were deposited at shallow depth in rocks soaked with sea-water indicates that all deposition took place at very low temperatures.
CHAPTER 10
CONCLUSIONS

The stratiform sulphide and sulphate deposits in the Silvermines area were deposited as chemical sediments in association with hot spring activity in an overall shallow-water carbonate shelf environment and lie at the edge of a fault-bounded positive block, not necessarily emerged above sea-level. By comparison with other Irish Carboniferous base metal deposits, which occur in similar positions, the relationship with fault-bounded positive features is significant.

Abundant sedimentary features such as syn-sedimentary slump breccia, graded-bedding, interbedding of sulphide and shale layers, and geopetal structures confirm the sedimentary origin of the stratiform sulphide deposits. Study of colloform textures in the sulphides show that most of the minerals formed by direct precipitation from solution and that an extensive intermediate gel stage did not exist. However soft sediment features in sphalerite show that it was in a cryptocrystalline, muddy state at one stage and may have behaved like a gel. Local flowage features in pyrite can be explained by some of this mineral having passed through a muddy cryptocrystalline or gel FeS stage.

Framboidal pyrite occurs in both vein-type and stratabound
deposits but is by far the commonest in the latter where it occurs as the earliest pyrite type and forms nuclei for later pyrite deposition. Framboids may have formed by pyrite crystallization at points within minute FeS gel globules.

Sulphur isotope results show that sulphides in the Upper orebody formed from bacterially reduced sea-water sulphate. Fractionation was slow and extreme showing that conditions of circulation on the sea floor were very restricted and that bacterial activity was intense. On the other hand barite was deposited in an area of circulating sea-water at a higher level.

The relatively high nickel content with respect to cobalt in pyrite of the Upper orebody further confirms accumulation of the sulphides in a sedimentary environment in the presence of organic matter.

Vein-fillings of the Lower orebody represent deposition from circulating, metal-carrying solutions rather than re-mobilized sedimentary deposits. Sulphur isotope results show that the rocks were saturated by sea-water at the time of deposition.
REFERENCES


Roedder, E. 1968. The noncolloidal origin of "colloform" textures in sphalerite ores. Econ. Geol. V. 63, pp. 451-471.


Schouten, Cg. 1946. The role of sulfur bacteria in the formation of the so-called sedimentary copper ores and pyritic orebodies. Econ. Geol. V. 41, pp. 517-538.


Schouten, C.F. 1946. The role of sulfur bacteria in the formation of the so-called sedimentary copper ores and pyritic orebodies. Econ. Geol. V. 41, pp. 517-538.


Weber, W. W. 1968. A brief dissertation of the metallogene-
metric significance of the ore deposits at Silvermines,
Co. Tipperary, Ireland as a metallogenetic province.

White, D. E. 1967. Mercury and base metal deposits with asso-
ciated thermal and mineral waters. Geochemistry of hy-
drothermal ore deposits. Barnes, H. L. Ed., pp. 575-


Wynne, A. B. 1861. Memoirs of the Geological Survey (of Ire-
land). Explanations of accompany Sheet 134 of the map
of the Geological Survey of Ireland illustrating parts
of the counties of Clare, Tipperary and Limerick. (Sil-
vermines deposits described on pp. 31-46).

Zimmermann, R. A. 1969. Sediment-ore-structure relations in
barite and associated ores and sediments in the upper Mis-
issippi Valley zinc-lead district near Shullsburg, Wis-

Zimmermann, R. A. and Amstutz, G. O. A small-scale sedimentary
features in the Arkansas barite district. Developments
in sedimentology V. 2. Sedimentology and ore genesis.

Zuffardi, T. 1967. The genesis of stratiform deposits of
lead-zinc and barite in Sardinia. Genesis of lead-
zinc-barite-fluorite deposits. Brown, J. S. Ed.

——— 1969. Sedimentary ores of the Middle Silurian
207-309.