THE SALMO LEAD-ZINC DEPOSITS:
A STUDY OF THEIR DEFORMATION AND METAMORPHIC FEATURES

by

ALAN STRATTON MACDONALD
B.Sc. University of Glasgow, 1962

A THESIS SUBMITTED IN PARTIAL FULFILMENT OF
THE REQUIREMENTS FOR THE DEGREE OF
DOCTOR OF PHILOSOPHY

in the

Department of Geological Sciences

We accept this thesis as conforming to the
required standard

THE UNIVERSITY OF BRITISH COLUMBIA
June, 1973
In presenting this thesis in partial fulfilment of the requirements for an advanced degree at the University of British Columbia, I agree that the Library shall make it freely available for reference and study. I further agree that permission for extensive copying of this thesis for scholarly purposes may be granted by the Head of my Department or by his representatives. It is understood that copying or publication of this thesis for financial gain shall not be allowed without my written permission.

Department of Geological Sciences

The University of British Columbia
Vancouver 8, Canada

Date 8th June, 1973
ABSTRACT

The problematic history of the strata-bound Salmo lead-zinc deposits, which occur in highly deformed, low-grade meta-sediments of the southern Kootenay Arc of British Columbia, has been investigated via structural analyses, fabric studies, and minor element determinations in three of the deposits: Reeves MacDonald, Jersey and H. B. mines.

Three phases of folding have been distinguished: an early phase of near-isoclinal folding, overturned toward the west, produced the major structures; a second phase of upright, more open folding modified the first phase folds by near coaxial refolding, by further closure and local flattening, and ultimately by westward-directed thrusting; a third phase of conjugate monoclinal folds and kink bands was produced by north-south compression related to northward-directed thrusting. The essentially tabular sphalerite-pyrite-galena ore bodies are involved, on all scales, in the folding and the ores exhibit internal structures ascribed to the differential movement of sulphides and host dolomite and calcite marbles.

Regional metamorphism, to lower greenschist facies, was synchronous with Phase 1 folding. Contact metamorphism by granite stocks (K-Ar age: circa 100 m.y. B.P.) postdates all phases of folding, and affects Jersey and H. B. deposits, and possibly also Reeves MacDonald; estimated temperatures are in
the range 425-600°C (at 1.5 kb). Seventy-one analyses of minor elements in the sulphides show that pyrite has Co: Ni<1, and that sphalerite has increasingly high Fe, Mn and Cd contents with increasing grade of contact metamorphism, except for an anomalous enrichment in Fe, Cd, Cu and Ag with depth in Reeves MacDonald mine. Pyrite retains brittle deformation textures until hornblende hornfels facies is attained when recrystallization becomes increasingly important, and ultimate breakdown to pyrrhotite occurs. Sphalerite and galena typically have granoblastic-polygonal recrystallization textures but exhibit widespread slight lattice bending and subgrain development, believed to postdate annealing recrystallization; local deformation twinning and kinking, recognized only in sphalerite, may pre-date annealing recrystallization.

X-ray fabric analyses of sphalerite, mainly from Reeves MacDonald mine, show well developed preferred orientation of (111) parallel with composition layering, attributed to syntectonic recrystallization. Sphalerite from Jersey mine has more varied (111) subfabrics showing development toward small-circle patterns of [111] which probably represent annealing fabrics. Mylonitic sphalerite, from thin zones in Jersey mine, has a distinctive (111) subfabric with orthorhombic symmetry and a pattern approaching (110)[001]. Galena, which has variable mylonitic and recrystallized textures, has either a random fabric, or weak (200) subfabrics which may reflect plastic deformation on the system (001)[110]. Quartz c-axis subfabrics and orientations of deformation lamellae (determined optically) suggest that, in
quartzitic rocks, syntectonic recrystallization occurred during Phase 1 folding, whereas plastic deformation, and at least local recrystallization (in fine-grained material), was produced essentially by flattening during Phase 2, and possibly during Phase 3 folding. c-axis subfabrics in host dolomite marbles may also be indicative of recrystallization during flattening.

It is concluded that the sulphides exhibit structures, on all scales, equivalent to those recognized in the host rocks and that these indicate involvement in all phases of deformation, in regional metamorphism, and in contact metamorphism. The deposits are interpreted as being originally of Mississippi Valley type.
# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>ABSTRACT</td>
<td>ii</td>
</tr>
<tr>
<td>LIST OF TABLES</td>
<td>ix</td>
</tr>
<tr>
<td>LIST OF ILLUSTRATIONS</td>
<td>xi</td>
</tr>
<tr>
<td>ACKNOWLEDGEMENTS</td>
<td>xvi</td>
</tr>
<tr>
<td>I. INTRODUCTION</td>
<td>1</td>
</tr>
<tr>
<td>LOCATION AND PURPOSE OF THE STUDY</td>
<td>1</td>
</tr>
<tr>
<td>PREVIOUS WORK</td>
<td>4</td>
</tr>
<tr>
<td>II. REGIONAL GEOLOGICAL SETTING</td>
<td>7</td>
</tr>
<tr>
<td>STRATIGRAPHY</td>
<td>7</td>
</tr>
<tr>
<td>Proterozoic rocks</td>
<td>7</td>
</tr>
<tr>
<td>Palaeozoic rocks</td>
<td>10</td>
</tr>
<tr>
<td>Mesozoic rocks</td>
<td>15</td>
</tr>
<tr>
<td>IGNEOUS HISTORY</td>
<td>15</td>
</tr>
<tr>
<td>Basic dykes, sills (and volcanics?)</td>
<td>16</td>
</tr>
<tr>
<td>Felsite sills</td>
<td>16</td>
</tr>
<tr>
<td>Nelson-type plutonic intrusions</td>
<td>17</td>
</tr>
<tr>
<td>Coryell-type plutonic intrusions</td>
<td>19</td>
</tr>
<tr>
<td>Lamprophyre dykes and sills</td>
<td>22</td>
</tr>
<tr>
<td>STRATIGRAPHIC SETTING OF MINERALIZATION</td>
<td>23</td>
</tr>
<tr>
<td>Dolomite in the Reeves member</td>
<td>23</td>
</tr>
<tr>
<td>Dolomite in the Nelway/Metaline formation</td>
<td>25</td>
</tr>
<tr>
<td>SUMMARY</td>
<td>27</td>
</tr>
</tbody>
</table>
III. STRUCTURAL ENVIRONMENT

MINOR FOLDING

Phase 1 structures 32
Phase 2 structures 36
Phase 3 structures 37

EFFECTS OF SUPERIMPOSED FOLDING 43

Phase 2 superimposed on Phase 1 43
Phase 3 superimposed on Phases 1 and 2 46

SUMMARY OF STRUCTURAL STYLES AND RELATIVE AGES OF FOLDING 48

MAJOR FOLDING 48

H. B. mine area 50
Jersey mine area 52
Reeves MacDonald mine area 54

FAULTING 57

MINOR FRACTURES 61

STRUCTURAL SYNTHESIS 63

IV. METAMORPHIC ENVIRONMENT 68

INTRODUCTION 68

REGIONAL METAMORPHISM 70

Textural features of the regionally metamorphosed host rocks 70
Mineral assemblages, metamorphic facies, and timing of the regional metamorphic event 71
P-T conditions of regional metamorphism 72

CONTACT METAMORPHISM 73

Textural features of the contact metamorphosed host rocks 73
Nature and distribution of the contact aureoles 74
Reeves MacDonald mine area 75
H. B. mine area 75
Jersey mine area 77
Assemblages of the contact aureoles 80
P-T conditions of contact metamorphism 81
Fluid phase during contact metamorphism 86
### V. THE SULPHIDE DEPOSITS

#### INTRODUCTION

#### NATURE AND FORM OF THE SULPHIDE ORE BODIES

- Reeves MacDonald mine
- Jersey mine
- H. B. mine
- Summary

#### MINERALOGICAL VARIATION OF THE SULPHIDES

- Introduction
- Minor element content of sphalerite
- Minor element content of pyrite
- Minor element content of galena
- Occurrence of pyrrhotite
- Occurrence of a second generation of pyrite
- Summary

#### TEXTURAL FEATURES OF THE SULPHIDES

- Introduction
- Pyrite
- Sphalerite
- Galena
- Summary and interpretation

#### FABRIC OF THE ORES

- Introduction
- Sphalerite fabric
  - (111) Subfabric
    - (a) Specimens from Reeves MacDonald mine
    - (b) Specimens from Jersey mine
    - (c) Mylonitized specimens from Jersey mine
  - (220) Subfabric
    - Origin of sphalerite fabric
- Galena fabric
  - (200) Subfabric
- Dolomite fabric
- Quartz fabric
- Summary

#### ORIGIN OF THE ORES


VI. SUMMARY AND CONCLUSIONS 176

STRUCTURE 176
SULPHIDE DEPOSITS 177
DEFORMATION AND REGIONAL METAMORPHISM 178
CONTACT METAMORPHISM 180

BIBLIOGRAPHY 183

APPENDIX

A. MAGNESIAN CALCITE-DOLOMITE GEOTHERMOMETRY 197

B. X-RAY TEXTURE GONIOMETER ANALYSIS OF PREFERRED ORIENTATION 199

Sample preparation 199
Technique 200
Accuracy of the technique 204
Precision 207
Pole-figures to $\phi = 90^\circ$ 209

C. MINOR ELEMENT DETERMINATION BY ATOMIC ABSORPTION SPECTROPHOTOMETRY 212

Sample preparation 212
Analytical method 214
Analytical precision 214
Sphalerite coloration 218

D. ANALYSIS OF SPHALERITE BY ELECTRON MICROPROBE 222

E. MAPS OF STRUCTURAL DATA 225
LIST OF TABLES

| I. | Stratigraphic divisions of the southern Kootenay Arc | 8 |
| II. | Potassium-argon age determinations on three intrusive igneous rocks from Jersey mine area | 20 |
| III. | Structural elements and their nomenclature | 31 |
| IV. | Summary of structural styles and relative ages of folding | 49 |
| V. | Structural correlation along the southern Kootenay Arc | 65 |
| VI. | Metamorphic rock types and typical mineralogies of the exposed stratigraphic units in the mine areas: (i) outside contact aureoles, (ii) within outer contact aureoles, and (iii) within inner contact aureoles | 69 |
| VII. | Comparison of mean values of minor element contents of sphalerite samples from the three mines | 111 |
| VIII. | Comparison of mean values of minor element contents of different types of sphalerite from the Annex zone of Reeves MacDonald mine | 114 |
| IX. | Comparison of mean minor element contents of pyrite samples from the three mines | 116 |
| X. | Co and Ni contents of A. certain sedimentary rock types, and of B. pyrite from different types of ore deposit | 119 |
| XI. | d_{112} and estimated temperatures of crystallization of magnesian calcites from Reeves MacDonald and Jersey mines | 198 |
| XII. | Absorption coefficients, 2θ angles, and other relevant data for the mineral assemblages analyzed | 206 |
| XIII. | Numbers and locations of ore specimens used in X-ray fabric analysis | 211 |
XIV. Minor element contents of sphalerite samples from Reeves MacDonald mine 215

XV. Minor element contents of sphalerite samples from Jersey and H. B. mines 216

XVI. Duplicate analyses of minor elements in sphalerite samples from the three mines 217

XVII. Sphalerite colour notation, based on Munsell Rock Colour Chart 219

XVIII. Minor element contents of pyrite samples from the three mines 221

XIX. Analyses by electron microprobe of iron and zinc in sphalerite 224
LIST OF ILLUSTRATIONS

FIGURE

1. Generalized geology of the southern Kootenay Arc showing the location of economically important strata-bound lead-zinc sulphide deposits 2

2. Intrusive rocks in the Salmo area. Jersey mine area is enlarged to show location of specimens dated by potassium-argon method 21

3. Examples traced from photographs of minor fold styles from the three fold generations distinguished in the mine areas 34

4. Equal-area projections of Phase 1 minor structural elements from the three mine areas 35

5. Plots of thickness variations across three Phase 2 minor fold hinges, developed in quartzite, quartz phyllite 38

6. Equal-area projections of Phase 2 minor structural elements from the three mine areas 39

7. Equal-area projections of Phase 3 minor structural elements from the three mine areas 41

8. Examples of interference fold structures resulting from superimposition of Phase 2 folds on earlier formed Phase 1 folds 45

9. Diagram showing the effects of Phase 2 folding on Phase 1 structures: (a) by refolding and (b) by closure and differential flattening 47

10. Vertical cross-sections (subnormal to major fold axes) of H. B. mine area 51

11. Vertical cross-sections (subnormal to major fold axes) of Jersey mine area 53

12. Inclined cross-section (subnormal to major $F_1$ fold axes) of Reeves MacDonald mine area 55
### FIGURE

<table>
<thead>
<tr>
<th>Figure Number</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>13</td>
<td>Main fault and fold patterns in the southern Kootenay Arc</td>
<td>58</td>
</tr>
<tr>
<td>14</td>
<td>Equal-area projections of poles to joint surfaces from the three mine areas</td>
<td>62</td>
</tr>
<tr>
<td>15</td>
<td>Contact metamorphic effects in Reeves MacDonald mine area</td>
<td>76</td>
</tr>
<tr>
<td>16</td>
<td>Contact metamorphic effects in H. B. mine area</td>
<td>78</td>
</tr>
<tr>
<td>17</td>
<td>Contact metamorphic effects in Jersey mine area</td>
<td>79</td>
</tr>
<tr>
<td>18</td>
<td>Stability fields of grossularite and wollastonite in H$_2$O-CO$_2$ mixtures at 2 kb</td>
<td>87</td>
</tr>
<tr>
<td>19</td>
<td>Mode of occurrence of grossularite skarn within calcite marble of the Reeves member</td>
<td>87</td>
</tr>
<tr>
<td>20</td>
<td>Schematic representation of the range of contact metamorphic temperatures affecting the three sulphide deposits</td>
<td>89</td>
</tr>
<tr>
<td>21</td>
<td>Diagrammatic east-west longitudinal section of Reeves MacDonald mine showing the ore zones</td>
<td>93</td>
</tr>
<tr>
<td>22</td>
<td>Level plans of the Reeves ore zone illustrating the change in configuration down plunge</td>
<td>93</td>
</tr>
<tr>
<td>23</td>
<td>Plan view and representative cross-sections of Jersey mine workings, showing the main structural features</td>
<td>97</td>
</tr>
<tr>
<td>24</td>
<td>a. Isometric view of the H. B. ore zones, showing their configuration and spatial relations. b. Typical vertical section of the No. 1 zone, looking north</td>
<td>101</td>
</tr>
<tr>
<td>25</td>
<td>Examples of layered ores from the three mines</td>
<td>104</td>
</tr>
<tr>
<td>26</td>
<td>Examples of breccia ores from Reeves MacDonald and Jersey mines</td>
<td>105</td>
</tr>
<tr>
<td>27</td>
<td>Examples of fold structures involving sulphides from Reeves MacDonald and Jersey mines</td>
<td>106</td>
</tr>
<tr>
<td>28</td>
<td>Logarithmic plot of Cd vs. Fe contents of sphalerite samples from the three mines</td>
<td>109</td>
</tr>
<tr>
<td>29</td>
<td>Ternary plot of Mn, Cu and Ag contents of sphalerite samples from the three mines</td>
<td>110</td>
</tr>
</tbody>
</table>
FIGURE

30. Logarithmic plot of Co vs. Ni contents of pyrite samples from the three mines 118
31. FeS contents (mole %) of sphalerite samples from Jersey mine 124
32. Semi-quantitative microprobe determinations of iron variation across sphalerite grains 126
33. Grain size variation of pyrite and sphalerite between the three mine areas 130
34. Photographs and photomicrographs of microstructures and textures developed in pyrite 132
35. Photomicrographs of recrystallization textures developed in sphalerite, etched with thio-urea + HCl 135
36. Photomicrographs of deformation textures and microstructures developed in sphalerite, etched with thio-urea + HCl 136
37. Photomicrographs of textures and microstructures developed in galena, etched with thio-urea + HCl 140
38. (111) pole figures for sphalerite from Reeves MacDonald mine 148
39. (111) pole figures for sphalerite from Jersey mine 149
40. (111) pole figures for a. partially mylonitized, and b. mylonitized sphalerite from Jersey mine 151
41. (220) pole figures for sphalerite from a. Reeves MacDonald and b. Jersey mines 152
42. a. Translation directions in the (111) plane of sphalerite. b, c. (111) pole figures of naturally deformed sphalerite, showing inferred ideal subfabrics and possible translation directions 154
43. (200) pole figures for galena from Jersey mine 157
44. a. Translation directions in the (100) plane of galena. b. Translation directions in the (110) plane of galena. c. Example of (200) pole figure of naturally deformed "gneissic" galena 159
<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>45.</td>
<td>Preferred orientation of dolomite $c$-axes in four dolomite marbles</td>
<td>162</td>
</tr>
<tr>
<td>46.</td>
<td>Preferred orientation of quartz $e$-axes in four quartzites, a quartz phyllonite, and a quartz schist</td>
<td>165</td>
</tr>
<tr>
<td>47.</td>
<td>Top: Equal area projections of poles to 75 sets of deformation lamellae, and of $c$-axes of host grains in each of two quartzites. Centre: Histograms of angles between lamellae pole and $c$-axis in each grain. Bottom: Deduced maximum principal stress directions</td>
<td>167</td>
</tr>
<tr>
<td>48.</td>
<td>Geometry of the Phillips X-ray texture goniometer set-up for reflection and transmission modes</td>
<td>201</td>
</tr>
<tr>
<td>49.</td>
<td>Intensity correction curves established from reflection scans of random specimens</td>
<td>203</td>
</tr>
<tr>
<td>50.</td>
<td>Duplicate X-ray reflection scans for various sphalerite specimens</td>
<td>208</td>
</tr>
<tr>
<td>51.</td>
<td>Plot of minor element contents vs. colour for all sphalerites (49) analyzed by AA spectro-photometry</td>
<td>220</td>
</tr>
</tbody>
</table>
PLATE

I. Planar structures in the Reeves MacDonald mine area 226

II. Linear structures in the Reeves MacDonald mine area 227

III. Major structures in the Reeves MacDonald mine area 228

IV. Planar structures in the Jersey mine area 229

V. Linear structures in the Jersey mine area 230

VI. Major structures in the Jersey mine area 231

VII. Planar (A) and linear (B) structures in the H. B. mine area 232

VIII. Major structures in the H. B. mine area 233
ACKNOWLEDGEMENTS

This study was done under the supervision of Dr. J. V. Ross who originally suggested the Salmo area for a study of this kind. Technical assistance was provided by J. E. Harakal, N. Dillon, A. Bentzen, and many other members of the Department of Geological Sciences, and also by R. G. Butters of the Department of Metallurgy, University of British Columbia.

Dr. J. T. Fyles of the British Columbia Department of Mines is thanked for his friendly encouragement and interest in the study. The willing cooperation, in the field, of mine geologists and managers is gratefully acknowledged, in particular G. G. Addie, then of Reeves MacDonald Mines Ltd.

Field expenses were provided through Grant NRC-A2134 awarded to J. V. Ross, and accommodation was kindly made available by Reeves MacDonald Mines Ltd., Canadian Exploration Ltd., and Consolidated Mining and Smelting Co. Ltd. The author was a recipient of a University of British Columbia Summer Research Scholarship in 1972.
These rocks occur in one of the Cordilleran zones of maximum orogenic shearing and mashing, with complete recrystallization. Numberless crumplings, overturnings and faultings characterize the region, which, as already noted, has been the scene of repeated igneous injections in the form of dikes, sills, stocks, and batholiths. . . . To such principal difficulties in analyzing the complex assemblage of strata along the Pend D'Oreille river there was added that common disadvantage of the geologist on the Forty-ninth Parallel, the dense evergreen forest with its deep mat of brush and fallen timber.

—Daly, 1912
SECTION I

INTRODUCTION

LOCATION AND PURPOSE OF THE STUDY

The Kootenay Arc (Hedley, 1955) is a complex arcuate structural belt lying within the south-central part of the Eastern Cordilleran Fold Belt (Fig. 1). It comprises juxtaposed Proterozoic, Palaeozoic and Mesozoic rocks affected in varying degrees by polyphase deformation and metamorphism. Within this highly deformed belt, mainly Lower Palaeozoic miogeosynclinal rocks separate late Palaeozoic and Mesozoic eugeosynclinal rocks in the west from Precambrian rocks of the Belt-Purcell anticlinorium in the east. Origin of the Arc has been widely debated, over the last decade, in terms of both its geometry (details of which remain to be defined) and its evolution. Also problematic is the metallogenesis and subsequent history of numerous strata-bound lead-zinc sulphide deposits, occurring in Cambrian carbonate rocks along the length of the Arc.

The Salmo area (Fig. 1) was selected for study because it includes three major lead-zinc deposits—Reeves MacDonald, Jersey and H. B. mines—which are closely located along strike of a strongly curved section of the southern Kootenay Arc, near the Canada-U.S.A. border. It was considered that a detailed examination of the stratigraphy, structure and metamorphism of both country rocks and enclosed sulphides, together with further
Figure 1. Generalized geology of the southern Kootenay Arc showing the location of economically important strata-bound lead-zinc sulphide deposits (compiled from Fyles, 1970; Ross, 1970; and Yates, 1970).
laboratory studies, would (i) clarify details of the geometry of the southern Kootenay Arc and perhaps of its evolution (at least in this segment), (ii) elucidate the history of the lead-zinc sulphides, and (iii) provide information on the behaviour of sulphide assemblages under conditions of regional and/or thermal metamorphism.

Field work for this thesis occupied four months, during summers of 1969-71, spent on the properties of Reeves MacDonald, Canex Jersey and Cominco H. B. mines. Detailed structural mapping was carried out on 1 in to 400 ft base maps of surface areas immediate to the mines and numerous visits were made underground to collect oriented sulphide samples (except at H. B. mine which was closed down in 1966).

In the thesis, the stratigraphy of the region is reviewed, the various types and ages of intrusive igneous activity are considered, and the nature and stratigraphic position of the strata-bound sulphide deposits are discussed in relation to that framework. The polyphase nature of deformation in the three mine areas is then described together with analyses of the various fold geometries and fracture patterns. Major structural features of the mine areas are re-interpreted and an attempt made to integrate these into the evolution of this segment of the Kootenay Arc. P-T conditions of regional and of local contact metamorphism are estimated and their textural and mineralogical effects on both sulphides and host rocks determined. The fabrics of the sulphides and of the quartz- and dolomite-rich host rocks are studied by X-ray and optical methods and their
responses to deformation and metamorphism compared. Finally, the implications of the various deformation and metamorphic features are discussed in terms of the metallogenesis and subsequent history of the strata-bound sulphides.

PREVIOUS WORK

Gold lode-mining in Ymir and Sheep Creek camps between 1896 and 1916 first stimulated interest by the Geological Survey of Canada in the Salmo area, resulting in the West Kootenay map of McConnel and Brock (1904). In this, the Rossland volcanic group was distinguished from metasediments to the east which were assigned to the Nisconlith and Selkirk series which were supposed to be of Cambrian age and to young eastward. Daly (1912) also traversed and mapped the southern edge of the area for the International Boundary Commission. He renamed the Nisconlith series as the Pend d'Oreille group and tentatively equated them with the Cache Creek group, considered to be of Carboniferous age. He also noted their structural complexity.

Revival of gold mining activity after 1928 was followed by Walker's (1929, 1934) one inch to one mile geological map and description of the geology and mineral deposits of Salmo area. Walker established the basic stratigraphic framework, subdividing the lower, predominantly clastic succession into separate formations. All of these, together with the Pend d'Oreille series were assigned to the Precambrian. He also described lead-zinc occurrences in carbonate rocks near the base of the Pend d'Oreille series and classified them as structurally controlled replacement deposits.
Post-1940, the decline of gold mining and the concurrent development and exploitation of the large strata-bound lead-zinc sulphide occurrences in the area resulted in renewed mapping and reinterpretation of the geology by both the British Columbia Department of Mines and the Geological Survey of Canada. Little (1950, 1951, 1960) remapped the Salmo area (later incorporated into the Nelson map-area, west half) and on palaeontological grounds assigned Walker's Pend d'Oreille series to the Lower Palaeozoic, subdividing it into the Laib group (Lower Cambrian), Nelway group (Middle Cambrian) and Active formation (Ordovician). In 1953, Mathews made a detailed study of the declining Sheep Creek gold camp, refining the stratigraphy of the host Reno and Quartzite Range formations and detailing the structural controls on gold lode-mineralization. Fyles and Hewlett (1959) mapped (1 in to 2000 ft) the Lower Palaeozoic rocks east and south of Salmo and redefined the Laib group as a formation subdividing it into four members. This aided in interpreting the complex structure developed in these rocks which was shown to be the product of at least two phases of deformation. From detailed mapping of their surface geology and from examination of the lead-zinc sulphide deposits themselves it was concluded that mineralization was by replacement, structurally controlled by the second phase of folding.

Numerous descriptions of individual Salmo lead-zinc deposits have also been published by B. C. Department of Mines, company and other geologists and will be referred to in the body of the thesis.
There are too a number of regional studies which have direct local application to Salmo area. Hedley (1955) first defined the Kootenay Arc and described its structural and metallogenic features. These were further amplified and developed by Fyles (1966, 1967, 1970) and by Yates et al. (1966) and Yates (1970). The evolution of the Arc has been also the subject of wide debate by Wheeler (1966, 1970), Ross and Kellerhals (1968), and Ross (1970). In addition, the metallogenesis of the regionally distributed lead-zinc deposits has been discussed. Muraro (1966) ascribed the gross shapes of the sulphide deposits and their internal structures to penetrative deformation and suggested that mineralization pre-dated regional metamorphism. However, Fyles (1967) considered the evidence for the latter suggestion to be very uncertain. Sinclair (1964, 1966) and Reynolds and Sinclair (1971) determined lead isotope abundances for deposits in the southern Kootenay Arc and distinguished two populations of anomalous lead isotopes corresponding to strata-bound and to vein-type deposits. Both types were interpreted as having been emplaced at approximately the same time, circa 150 m.y. ago (i.e., pre-regional metamorphism). Sangster (1970a, b) determined sulphur isotope abundances in sulphides from the Salmo deposits which showed that these could have been derived from Cambrian sea water sulphate and hence could have been syngenetic-diagenetic in origin.
SECTION II

REGIONAL GEOLOGICAL SETTING

STRATIGRAPHY

The southern Kootenay Arc, although primarily a tectonic feature, is outlined by the distribution of Palaeozoic miogeosynclinal and transitional rocks. To the east, these are marginally infolded with clastic and minor volcanic rocks of Late Proterozoic age, while to the west they are in complex structural contact with Upper Palaeozoic and Mesozoic eugeosynclinal argillites and volcanics.

Table I shows the stratigraphic succession and correlations established for the adjacent Metaline, Salmo and Kootenay Lake areas of the southern Arc.

Proterozoic rocks

The Late Proterozoic succession forming the east side of the Arc was described and defined as the Windermere Series by Walker (1926).

The basal Toby conglomerate formation marks the major unconformity separating these Windermere rocks from the underlying Belt-Purcell Series (a mid-Proterozoic fine-grained clastic assemblage forming the Purcell anticlinorium). It is of strongly bimodal grain size and very variable in lithology and thickness but was considered as being too extensive to represent a true
Table 1. Stratigraphic divisions of the southern Kootenay Arc (compiled from Little, 1960; Fyles, 1970; Ross, 1970; and Yates, 1970).

<table>
<thead>
<tr>
<th>Period</th>
<th>Metaline District</th>
<th>Salmo District</th>
<th>Larderud District</th>
</tr>
</thead>
<tbody>
<tr>
<td>JURASSIC</td>
<td></td>
<td>JOYER GROUP</td>
<td></td>
</tr>
<tr>
<td>TRIASSIC</td>
<td>ROSSLAND GROUP</td>
<td>ROSSLAND GROUP</td>
<td>SLOCAN GROUP</td>
</tr>
<tr>
<td>PERMIAN</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PENNSYLVANIAN</td>
<td>MT. ROBERTS FN.</td>
<td>MT. ROBERTS FN.</td>
<td>MILFORD GROUP</td>
</tr>
<tr>
<td>MISSISSIPPIAN</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DEVONIAN</td>
<td>unnamed formation</td>
<td></td>
<td></td>
</tr>
<tr>
<td>SILURIAN</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>UPPER ORDOVICIAN</td>
<td>LEDBETTER SLATE FN.</td>
<td>ACTIVE FORMATION.</td>
<td>LARDEAU GROUP</td>
</tr>
<tr>
<td>MIDDLE ORDOVICIAN</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>LOWER ORDOVICIAN</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>UPPER CAMBRIAN</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MIDDLE CAMBRIAN</td>
<td>METALINE FORMATION</td>
<td>NELWAY FORMATION</td>
<td></td>
</tr>
<tr>
<td>LOWER CAMBRIAN</td>
<td>MAITLEN PHYLLITE FN.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>WINDERMERE</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

- Horizons containing Pb-Zn mineralization

Legend:
- LAF: Lower Laib member
- U: Upper Laib member
- E: Emerald member
- T: Reeves member
- R: Reeves member
- M: Reeves member
- Q: QUARTZITE RANGE FN.
- S: THREE SISTERS FN.
- M: Monk formation
- L: Leola volcanics
- I: Irene volcanics
- T: Toby conglomerate
- H: HORSETHIEF CREEK GP.
fanglomerate (Rice, 1941). A recent study (Aalto, 1971) suggests that it is actually a tillite although this possibility was also briefly entertained by earlier workers. Upwards it becomes interbedded with the Irene volcanic formation, greenstones of andesitic composition believed to represent a volcanic extrusive and pyroclastic assemblage (Rice, 1941). Farther south, in Washington, Miller (1969) has recognized pillows in the equivalent Leola volcanics which are of tholeitic composition. These volcanics thin both north and south away from the general vicinity of the International Boundary.

Conformably overlying the Irene volcanics is the Horsethief Creek group or its equivalents in the Salmo area, the Monk and the Three Sisters formations. The lower part, equivalent to the Monk formation, has been described as a "heterogeneous assemblage" by Walker (1934) and as a "waste-basket formation" by Park and Cannon (1943), being predominantly phyllite with intercalations of conglomerate, quartzite and arenaceous limestone. In the south, this becomes increasingly arenaceous upwards into massive coarse-grained quartzite and grit of the Three Sisters formation (or lower part of the Gypsy quartzite in the Metaline district). To the north, however, the lithological heterogeneity continues to the top of the undivided Horsethief Creek group, indicating a marked southward facies change corresponding to increasing proximity of source.

This variable coarse clastic and volcanic assemblage appears to represent deposition in a rapidly subsiding trough, marginal to the Purcell anticlinorium, in which movements were
probably epeirogenic and fault controlled (Yates, 1970). Formations generally thin northwards and, in the Metaline area at least, thin even more rapidly to the west and southwest. However their ultimate western limit is unknown (Park and Cannon, 1943).

The boundary between Precambrian and Cambrian has not been accurately defined in the southern Kootenay Arc due to the poor fossil record and the fact that coarse clastic sedimentation continued virtually unbroken into the Palaeozoic era. Little (1960) has arbitrarily placed the boundary in the Salmo area at the base of the Hamill group since the lower part correlates lithologically with the upper Gypsy quartzite formation in which Park and Cannon (1943) found trilobite fragments.

**Palaeozoic rocks**

The **Hamill group** are the Lower Cambrian representatives of continued clastic deposition. In the Salmo area, the group is subdivided into the **Quartzite Range** and overlying **Reno formations**. The former comprises a massive white quartzite of remarkable purity and lateral persistence which can be recognized as a distinctive lithofacies as far north as Duncan Lake (Fyles, 1964). Upwards it becomes increasingly impure and where phyllite intercalations occur the Reno formation is distinguished.

In the Salmo area, this Lower Cambrian clastic succession appears, in spite of increasing intensity of subsequent deformation, to thin markedly toward the west from the Sheep Creek anticline, to lose cross-bedding and to become finer grained and increasingly argillaceous (Fyles and Hewlett, 1959).
Appearance of carbonate rocks marks the transition from a hitherto predominantly clastic succession into the mixed carbonate and pelitic sequence of the Laib formation (Little, 1960) and its correlatives, the Mohican, Badshot and Index formations (Lardeau district) and the Maitlen phyllite (Metaline district). The transitional rocks, interbedded thin limestone marble and phyllite, have been distinguished in the Salmo area as a separate member (the Trueman) by Fyles and Hewlett (1959). In addition to their varied lithology they are extremely variable in thickness, apparently as a result of intense deformation and thus may not provide a very reliable stratigraphic marker.

They are overlain by grey and white, locally dolomitic, limestone/marble of the Reeves member/Badshot formation which provides perhaps the best stratigraphic marker of the Kootenay Arc, by virtue of its distinctive and persistent lithology and palaeontological control (the presence of Lower Cambrian archaeocyathids at widely separated localities). The origin of the dolomite facies, important as host to the Salmo-type lead-zinc deposits, has been the subject of some controversy which will be discussed in detail below.

This limestone is not without variation, east-west lithological changes having been demonstrated by Fyles and Hewlett (1959) in the Salmo area and by Fyles (1964) in the Duncan Lake area. Eastwards the limestone becomes impure, passing into argillaceous limestone and calcareous siltstone and the only dolomite is apparently the product of local alteration adjacent to veins and intrusions.
Telescoping of east-west facies changes and drastic thinning by deformation causes local correlation problems, since the Reeves/Badshot limestone may then be confused with limestone units of the Trueman member or its equivalent (Fyles and Hewlett, 1959; Fyles, 1964) and with units of the lower Lardeau group (Crosby, 1968).

An abrupt upward change to a predominantly pelitic succession follows everywhere along the southern Arc.

In the Salmo area, a black argillite-phyllite member, the **Emerald** (Fyles and Hewlett, 1959) is distinguished from overlying green phyllites and micaceous quartzites which comprise the upper part of the Laib formation. This member is restricted to the westernmost outcrops of Lower Cambrian strata which may again reflect facies telescoping. It does, however, lithologically very closely resemble the Ordovician Active formation and if equivalent may therefore be structurally emplaced either by pre-folding décollement thrusting, or by complex folding and associated thrusting.

In the Salmo-Metaline segment there is a gradual transition from Laib and Maitlen phyllite formations up into the predominantly carbonate **Nelway** and **Metaline formations**, respectively, indicating the persistence of a shelf environment in this segment of the Arc. To the north in the Lardeau segment, the pelitic Index formation continues unchanged, reflecting continued subsidence of the Lardeau trough.

The Nelway/Metaline formations are divided into lower limestone, middle dolostone and upper limestone members, described
in detail by Park and Cannon (1943) and by Dings and Whitebread (1965). The lower member is composed of dark grey limestones and calcareous phyllites and contains Middle Cambrian trilobites. The middle member consists of medium to thick bedded dolostone of various kinds which locally is host to productive lead-zinc mineralization in the Northport district, west of Metaline, and to minor mineralization in the Salmo area near Nelway (Fyles and Hewlett, 1959). Although typically a massive grey limestone sequence, the upper member is also dolomitized but only in its uppermost 200 ft which is host to economically important lead-zinc mineralization at Metaline Falls.

Algal structures in limestone and shaly limestone have been described from several stratigraphic levels in the Metaline formation (Dings and Whitebread, 1965). No Upper Cambrian fossils have been found and the contact with the overlying Ledbetter slate formation is very abrupt. Since the latter contains Lower and Middle Ordovician graptolites not far above its base, a disconformity may separate the two formations (Park and Cannon, 1943) or even a pre-folding décollement, but no regional discordance has been recognized.

The Ledbetter slate is correlated with the Active formation of the Salmo area which contains Lower Ordovician graptolites at one locality northeast of Salmo village (Little, 1960). They are lithologically very similar, comprising black argillite, slate and phyllite with minor calcareous and siliceous intercalations, not unlike the Lardeau group to the north.
The upper limits of both formations are very poorly defined due to poor exposure and structural complexity. In the Metaline area, the Ledbetter is succeeded by unnamed formations composed of slate with minor exotic, bioclastic limestone and pebble conglomerate, containing Silurian and Devonian fossils (Dings and Whitebread, 1965). This predominantly black shale assemblage of Ordovician-Devonian age reflects the extension southwards and subsequent persistence of a restricted basin (?) environment along the southern segment of the Arc.

These unnamed formations are assumed to be the correlatives of the upper Lardeau group to the north, represented by the volcanic Jowett formation and the overlying clastic Broadview formation which are however of unknown age. The provenance of the latter is problematic since it differs so radically from the underlying thick shale sequence. Alternatively, the Broadview formation has been equated with the Horsethief Creek group (Read, 1966).

Above this, the stratigraphic record is fragmentary and incomplete for the remainder of the Upper Palaeozoic era. This apparently reflects the culmination of local tectonism and the development of a eugeosynclinal environment to the west of the Arc.

In the Salmo-Metaline sector the Pennsylvanian is represented in limited outcrops west of the Arc by the Mt. Roberts formation which comprises slate, limestone, andesite and banded tuff with a Pennsylvanian (and possible Permian) fauna and flora (Little, 1960). The same stratigraphic break is recognized
in the Lardeau trough in the unconformity separating the
Milford group from older rocks below.

**Mesozoic rocks**

In the Salmo area, Triassic-Jurassic rocks are structurally juxtaposed against the lower and middle Palaeozoic rocks of the Kootenay Arc. These are distinguished as the Ymir and Rossland groups but their relative ages are obscure, the former being predominantly sedimentary (argillite, quartzite, and limestone) and the latter mixed volcanic (andesitic volcanics and pyroclastics) and sedimentary (argillite, impure sandstone and conglomerate). They are correlated to the north with the Slocan and Kaslo groups which crop out widely to the west of Kootenay Lake. The Slocan group consists of argillite, impure sandstone and limestone of Triassic age in part (Little, 1960), whereas the Kaslo group is a greenstone, breccia and tuff assemblage of uncertain age, which is assumed to stratigraphically underlie the Slocan group (Little, 1960).

**IGNEOUS HISTORY**

Igneous activity has been widespread, in both time and space, throughout the southern segment of the Arc. Types of activity are extremely varied but each is apparently widely distributed along some 150 miles of the Arc between the Lardeau and Metaline districts. Radiometric age determinations on these various types are relatively few but on the basis of cross-cutting relationships, alteration and internal deformation their relative ages appear to be (in order of decreasing age):
(a) Basic dykes, sills (and volcanics?)
(b) Felsite sills
(c) Nelson-type granitic batholiths and satellite stocks
(d) Coryell-type alkaline stocks
(e) Lamprophyre dykes

**Basic dykes, sills (and volcanics?)**

A number of older basic, minor intrusive and possible volcanic rocks have been described from different localities in the southern Arc. Their relative ages are obscure but they are compositionally different from the younger lamprophyres and apparently older than all other intrusive types, being highly altered and internally deformed.

In the Duncan Lake area, Fyles (1964) has described sill-like amphibolite masses in the upper Hamill group and Mohican formation which vaguely resemble feldspar-chlorite schists and greenstones of definite volcanic origin occurring in the Lardeau group, much higher in the succession.

Older basic dykes in the Sheep Creek gold camp (Mathews, 1953) are foliated unlike the younger lamprophyre suite, and distinctly more silica-rich and soda-poor.

Northeast of Reeves MacDonald mine, a thick sill-like mass of intensely deformed, augen-textured greenstone with schistose margins occurs within the Quartzite Range formation in the core of the Salmo River anticline.

**Felsite sills**

Felsite sills, commonly as narrow multiple swarms of extraordinary strike length paralleling the main foliation, are
a distinctive regional intrusive feature. They frequently display a weak foliation and, in the Duncan Lake and Ainsworth-Kaslo areas (Fyles, 1964, 1967) may be boudinaged and even folded.

In the Sheep Creek area, a sill swarm closely follows the Reno-Laib formational contact, except where this is parasitically folded on the east limb of the Central syncline (Mathews, 1953).

To the west, in the vicinity of Salmo, they intrude the upper Laib and Active formations, lying west and east respectively of the main Salmo River-Jersey anticlinal structure. Locally at least (e.g., on the Salmo River northeast of the Reeves MacDonald mine) they show an incipient foliation, apparently coplanar with the axial planes of Phase 2 folds, suggesting that they may predate or in part by synchronous with the latter.

**Nelson-type plutonic intrusions**

On the regional scale, granitic rocks of the Nelson and Kuskanax batholiths (Fig. 1) dominate the western concave side of the Arc and appear to emphasize its curvature. Numerous satellite plutons and stocks are also widely distributed along and across the Arc, causing local complications due to contact thermal metamorphism and marginal deformation. Their distribution appears largely irregular except for a possible weak east-southeast alignment. Marginally, the intrusives may show brittle internal deformation and local intense folding of the country rocks, suggestive of forceful intrusion.
The main intrusive phase is composed of porphyritic granite with subordinate phases of nonporphyritic granite, granodiorite, quartz monzonite, diorite and locally syenite.

Intrusion everywhere appears to postdate the main phases of folding. Potassium-argon age determination of 170 m.y. on biotite (Leech et al., 1963) and of 158 m.y. on biotite and 164 m.y. on hornblende (Nguyen et al., 1968) indicate intrusion in the Middle Jurassic which is consistent with other geological evidence (Little, 1960).

Stocks are particularly numerous in the Salmo and Sheep Creek areas and have been described in a general way by Fyles and Hewlett (1959) and Mathews (1953) as steep-sided irregular masses displaying varied contact effects, with margins varying from diffuse to xenolithic and cross-cutting. Thermal aureoles are up to several thousand feet wide and may overlap so that little of the country rock can have escaped reheating. The stocks are composed usually of biotite granite or quartz monzonite with only minor amounts, if any, of other phases.

All three areas mapped are in some degree affected by contact thermal metamorphism but only near the Jersey mine do intrusions actually occur close to the lead-zinc ore bodies. There, three small stock-like bodies of biotite granite, elongate parallel with the local foliation, intrude the main fold structures and locally cut the ore zones. Skarn-type tungsten mineralization occurs where the intrusions are in contact with either of the calcareous Trueman or Reeves members, resulting in complex superimposed mineralization.
Biotite from a medium-grained facies of the Dodger stock gave a potassium-argon age of 100 \pm 3.0 \text{ m.y.} (Table II, Fig. 2). This is very similar to circa 100 \text{ m.y.} concordant ages determined from hornblende, muscovite and biotite from the Spirit pluton and Kaniksu batholith in Stevens County, Washington (Yates and Engels, 1968). However other stocks in the Salmo district which have been dated (also by potassium-argon method) are the Lost Creek quartz monzonite stock (119 m.y. biotite age, Leech et al., 1963) and the Porcupine Creek granodiorite stock (128 m.y. biotite age and 152 m.y. hornblende age, Wanless et al., 1967) (Fig. 2). The latter pair of ages suggest that these stocks are related to the Nelson suite. Although the Dodger and affiliated small stocks are possibly related to the Cretaceous plutonic event represented by the Spirit pluton-Kaniksu batholith and perhaps also by the Bayonne and Fry Creek batholiths (100 m.y. biotite and muscovite ages, Leech et al., 1963), it seems more likely that they belong rather to the Nelson suite and have suffered argon loss, possibly by reheating during the subsequent Eocene igneous event.

Coryell-type plutonic intrusions

A number of very small pipe-like stocks of alkaline affinities, occurring in the Salmo area, are correlated with the syenitic rocks of Rossland and farther west, which have been described as the Coryell-type by Little (1960). The only one relevant to this study, the Salmo River stock, lies a short distance to the west of the Canex Emerald tungsten mine (Fig. 2). It was described by Daly (1912) as a porphyritic augite monzonite.
Table II. Potassium-argon age determinations on three intrusive igneous rocks from Jersey mine area (specimen locations are shown in Fig. 2).

<table>
<thead>
<tr>
<th>Specimen number</th>
<th>Unit</th>
<th>Rock type</th>
<th>Mineral</th>
<th>%K ± S*</th>
<th>$^{40}$A radiogenic $^{40}$A total</th>
<th>$^{40}$A radiogenic $^{40}$A radiogenic (10$^{-5}$ cc STP/g)</th>
<th>$^{40}$A radiogenic $^{40}$A radiogenic (10$^{-3}$)</th>
<th>Apparent age (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CX70-19</td>
<td>Dodger stock</td>
<td>Biotite granite</td>
<td>Biotite</td>
<td>6.08 ± 0.04</td>
<td>0.92</td>
<td>2.467</td>
<td>5.995</td>
<td>100 ± 3.0</td>
</tr>
<tr>
<td>CX70-17</td>
<td>Salmo River stock</td>
<td>Augite monzonite</td>
<td>Biotite</td>
<td>7.23 ± 0.02</td>
<td>0.87</td>
<td>1.467</td>
<td>2.998</td>
<td>50.6 ± 1.5</td>
</tr>
<tr>
<td>CX70-21</td>
<td>Tertiary dyke</td>
<td>Biotite lamprophyre</td>
<td>Biotite</td>
<td>7.04 ± 0.01</td>
<td>0.88</td>
<td>1.398</td>
<td>2.934</td>
<td>49.5 ± 1.5</td>
</tr>
</tbody>
</table>

*S = Standard deviation of quadruplicate analyses

Potassium analyses by J. E. Harakal and V. Bobik using Baird Atomic KY and KY3 flame photometers
Argon analyses by J. E. Harakal using AE1 MS-10 mass spectrometer
Figure 2. Intrusive rocks in the Salmo area (after Little, 1960). Jersey mine area is enlarged to show location of specimens dated by potassium-argon method. Ages from the Lost Creek stock and from the Porcupine Creek stock are by Leech et al. (1963) and Wanless et al. (1967).
Biotite from this stock gave a potassium-argon age of 50.6 ± 1.5 m.y. (Table II). This is in close agreement with potassium-argon ages determined from other alkaline intrusive rocks assigned to the Coryell suite in Stevens County, Washington (Yates and Engels, 1968) and in the Rossland district (Fyles et al., 1973).

**Lamprophyre dykes and sills**

Biotite- and olivine-lamprophyre dykes and sills are found throughout the southern segment of the Arc. They display little internal deformation and apparently postdate all other intrusive and major tectonic events.

In the Salmo and Sheep Creek areas, the lamprophyre dykes are not generally well exposed but are often encountered in underground workings where they cut mineralization of all types. They may contain locally derived xenoliths of quartzite, phyllite, sulphide, etc., but more interesting exotic xenoliths of granite have also been noted in the Sheep Creek area (Mathews, 1953), along the Pend Oreille River near the Reeves MacDonald mine, and within zones in the H. B. mine (Warning, 1960).

Biotite from a lamprophyre dyke in the Jersey mine gave a potassium-argon age of 49.5 ± 1.5 m.y. (Table II, Fig. 2). This is essentially the same age as that determined for the alkaline Salmo River stock. Yates and Engels (1968) and Fyles et al. (1973) have already shown that this mid-Eocene event was widespread and that intrusive and extrusive activity were essentially coeval.
STRATIGRAPHIC SETTING OF MINERALIZATION

Strata-bound lead-zinc deposits in the southern Arc are restricted to carbonate units of Lower and Middle Cambrian age which invariably have been dolomitized.

The Reeves (Badshot) limestone/marble, which forms the Lower Cambrian host, is dolomitized only in relatively small lenticular zones to which the sulphides, if present, are largely confined.

In contrast, the Middle Cambrian Metaline (Nelway) formation is much more extensively dolomitized and indeed its middle member is almost wholly dolomite and forms a productive sulphide host in the Northport district. Its upper grey limestone member is also dolomitized but only in its uppermost 200 ft in the Metaline district, where it is host to economically important sulphide mineralization.

There has been considerable argument concerning the origin of these dolomite-sulphide associations. In particular the dolomite has been variously considered as being of epigenetic, diagenetic, or syngenetic origin.

**Dolomite in the Reeves member**

Within the calcite marble of the Reeves member, dolomite forms flat tabular zones essentially conformable with the host; although not confined to one single horizon and therefore not stratiform the zones do seem to be strata-bound. They are located near the base of the Reeves member, locally even in contact with the underlying Trueman member, and in general are less than 100 ft thick. However there is a notable exception at the
Reeves MacDonald mine where an easterly structural repetition of the Reeves member has been virtually completely dolomitized throughout its total thickness of 350-400 ft (Plate III). The dolomite zones are elongate parallel with the major structural trends and appear to have been involved in all phases of deformation (see Section III). Contacts between dolomite and host calcite marble are usually interlayered vertically over short distances and finely and intricately interfingered laterally. The dolomite marble is distinct lithologically since it is composed usually of more than 80% dolomite in contrast with calcite marble which is more than 80% calcite (Green, 1954). Quartz, probably recrystallized after chert, is ubiquitous, usually in minor amounts, as small lenses and thin layers. However the dolomite marble may locally be highly siliceous in its upper part as in the structurally repeated eastern outcrop of the Reeves member at Reeves MacDonald mine (Plate III).

Several varieties of dolomite marble can be recognized:

(i) **massive**—pale grey to white homogeneous dolomite
(ii) **banded**—alternating pale grey/black layered dolomite
(iii) **sheared**—highly sheared pale grey/black dolomite in which S-planes are illustrated by graphitic laminae

The dolomite marble may be brecciated also but this is usually in association with sulphides where the latter form the matrix to blocks, fragments and lenses (some intricately folded) of dolomite marble.

Fyles and Hewlett (1959) and Fyles (1964) believe the dolomite and its siliceous varieties were formed by epigenetic
replacement of host Reeves limestone/marble and that the replace-
ment was structurally controlled. An alternative primary or
diagenetic origin is implied by Sangster (1970a) who believes
the carbonates formed, in association with bedded chert, in a
deep-water euxinic environment.

Dolomite in the Nelway/Metaline formation

The middle dolomite member of the Metaline formation is
primarily composed of light grey, fine- to medium-grained, thick
bedded dolostone but several subordinate varieties are also
distinguished. These range from very fine-grained grey to medium-
grained black massive dolostone, together with distinctive mottled
and striped varieties which may be of algal origin. Because of
its widespread distribution and great thickness (more than 3000
ft), Dings and Whitebread (1965) regard the middle dolomite unit
as being of diagenetic origin.

The equivalent Nelway middle member is composed of light
to dark grey massive dolostone commonly with chert nodules.
However it also includes a unit of finely interbanded light grey/
black dolostone in which the black layers contain abundant
irregular small lenses infilled with white dolomite and chert
resembling "bird's eye" structures, possibly of dessication
origin.

Little (1960) regarded the dolomite as of primary origin.
Fyles and Hewlett (1959), who noted that the limestone-dolostone
contact at the base of the middle member was irregularly trans-
gressive on a broad scale and often complexly interbanded,
concluded that "the complex pattern results from the vagaries of dolomitization" but did not speculate on the origin of the dolomite.

The upper limestone member of the Metaline formation is dolomitized in its uppermost 200 ft in the Metaline Falls area. This unit, known as the Josephine horizon, consists of irregularly but pervasively brecciated dolostone with a matrix of fine-grained black carbonaceous dolomite and jasperoid. The brecciated dolostone is texturally very variable and includes a striped variety which has been interpreted as being of algal origin (McConnel and Anderson, 1968). The latter authors propose that this Josephine horizon is stratigraphic and hence of sedimentary and diagenetic origin whilst Park and Cannon (1943) and Dings and Whitebread (1965) consider the dolomite to be of epigenetic origin.

These various dolomite occurrences are from different structural, as well as stratigraphic, levels and thus display widely differing intensities of deformation. A comparison of the occurrences, allowing for structural variation, suggests that they are all strata-bound although their limits are not always clearly or simply defined. There is no evidence that the dolomitization process was structurally controlled or localized as proposed by those favouring an epigenetic origin although of course the dolomite zones are themselves deformed. In general, too, the dolomite zones are frequently more or less siliceous, slightly carbonaceous and argillaceous. Striped and mottled varieties, distinguished in the less highly deformed Middle
Cambrian units have been interpreted to be of algal and/or diagenetic origin. However, Zeuger (1972) has pointed out that finely laminated dolomite and dessication structures developed in dolomite are not necessarily diagnostic of a supratidal or intertidal origin and may be found even in subtidal dolomite bioherms.

It is suggested that the general features of these various dolomite occurrences are more compatible with an early replacement origin (possibly diagenetic) in a bank environment rather than with a later, structurally controlled, epigenetic origin.

SUMMARY

Accumulation of a clastic wedge (initiated by epeirogenic movements) marginal to the Purcell Arch extended from Late Proterozoic to Lower Cambrian times. Overstep from the northeast onto the Arch occurred by earliest Cambrian period via the structurally controlled northeast-trending Eager trough and presumably continued with the progressive denudation of the Arch which is reflected in the increasingly fine grain size of the clastics upwards in the sedimentary pile.

A widespread shelf-edge bank carbonate (Reeves/Badshot limestone) marked the culmination of this prograding accumulation in the middle Lower Cambrian. Carbonate deposition persisted in the Salmo-Metaline sector at least until the end of the Middle Cambrian. Most of the dolomite occurrences in the Lower Cambrian Reeves member/Badshot formation and in the Middle Cambrian
Nelway-Metaline formation are considered to be of a syngenic-diagenetic origin as the occurrences are essentially strata-bound in form, exhibit slight facies differences from their "host" limestones, and contain problematic structures possibly of algal and of dessication origins. To the north, in the Lardeau district, the shelf environment was replaced toward the end of the Lower Cambrian period by deeper, more restricted slope or basin environment which, by the Lower Ordovician, had developed farther south along the Arc and persisted there until the Silurian and perhaps Devonian period when there is some evidence of influx of clastic material from a presumed western source.

The unconformity beneath the Mount Roberts formation (and the Milford group) apparently marks mid-Palaeozoic tectonism in the Kootenay Arc region. The nature of this deformation is obscure although it has been postulated as reflecting the development of eastward-verging nappe structures (Ross and Kellerhals, 1968). The subsequent depositional record is of mixed andesitic volcanic/sedimentary eugeosynclinal character with material being derived largely from the west. However this record is incomplete, there being little evidence of deposition during the Permian and Lower Triassic periods.

The Upper Triassic-Jurassic Ymir and Rossland groups unconformably overlie the Pennsylvanian Mount Roberts formation as apparently the Kaslo and Slocan groups similarly overlie the Milford group, reflecting the continuation (or re-initiation) of tectonism in the early Mesozoic era. These Triassic-Jurassic
rocks are themselves folded but plutonic intrusion by the Nelson batholith and its satellites post-dates this deformation which must therefore be pre-Middle Jurassic in age.

Many of these satellites have radiometric apparent ages much younger than those of the Nelson batholith itself. Several small granite stocks in the vicinity of Jersey mine are probably examples of such satellites. One of them, the Dodger stock, which contact metamorphoses and locally intrudes lead-zinc mineralization, was found to have an apparent potassium-argon age from biotite of 100 m.y. Other potassium-argon age determinations on biotites from the Salmo River alkaline stock and from a lamprophyre dyke confirm the occurrence in this area of the Eocene igneous event which is found to be widespread in the southern Kootenay Arc.
SECTION III

STRUCTURAL ENVIRONMENT

The three mine areas are located within a westerly verging, folded and thrust Lower Palaeozoic sequence referred to as the "Mine Belt" by Fyles and Hewlett (1959) and as the "Thrust Belt" by Yates (1970). Economic mineralization is restricted to the Reeves member where it is exposed on the upper limb of the major complex anticlinal structure which dominates the belt. Structural mapping included most of the latter structure within the mine areas.

The same sequence of structural events was recognized in all three areas although there are marked differences in the relative intensities of these events and lesser differences in both geometry and style of deformation.

Deformation is polyphase, with three phases of folding and various fracture systems having been recognized.

MINOR FOLDING

The various structural elements associated with folding were distinguished in the field on the basis of their morphology and orientation. These elements with the nomenclature used are listed in Table III and their orientations and locations shown in Plates I, II, IV, V and VII.

Relative ages of the different phases of folding were established mainly from parallel or cross-cutting relationships"
Table III. Structural elements and their nomenclature.

<table>
<thead>
<tr>
<th>Surface element</th>
<th>Linear element</th>
</tr>
</thead>
<tbody>
<tr>
<td>$F_0$</td>
<td>bedding</td>
</tr>
<tr>
<td><strong>Phase 1</strong></td>
<td></td>
</tr>
<tr>
<td>$F_1$</td>
<td>axial plane micaceous foliation</td>
</tr>
<tr>
<td>$F_0$</td>
<td>$F_0/F_1$ intersections</td>
</tr>
<tr>
<td>$AP_1$</td>
<td>fold axial surface</td>
</tr>
<tr>
<td>$F_1$</td>
<td>foliation</td>
</tr>
<tr>
<td>$F_2$</td>
<td>crenulation cleavage</td>
</tr>
<tr>
<td>$L_2$</td>
<td>crenulation of $F_1$ or $F_2$ trace on $F_{0-1}$</td>
</tr>
<tr>
<td>$AP_2$</td>
<td>fold axial surface</td>
</tr>
<tr>
<td>$F_3$</td>
<td>strain-slip cleavage</td>
</tr>
<tr>
<td>$L_3$</td>
<td>kink bands in $F_1$ or $F_3$ trace on $F_{0-1}$</td>
</tr>
<tr>
<td>$AP_3$</td>
<td>fold axial surface</td>
</tr>
<tr>
<td>$F_0$</td>
<td>fold axis</td>
</tr>
</tbody>
</table>
between cleavages and fold axial surfaces, and from refolding of earlier cleavages and related linear structures. Slight differences in fold style of successive phases may also be recognized but these are much complicated by the varied response of different lithologies during any particular phase of folding.

Following a convention established by other workers in the Kootenay Arc, these phases of folding have been designated Phase 1, Phase 2, and Phase 3. However it should be understood that these are not necessarily equivalent to phases distinguished by other workers beyond the immediate Salmo area, as will be considered below.

**Phase 1 structures**

Although gross lithological units are easily distinguished, recognition of internal bedding surfaces ($F_0$) is frequently obscured by the $F_1$ foliation, except where the bedding is cut at high angles by the latter as in Phase 1 fold closures. Very commonly the earliest recognized foliation ($F_1$) is near coincident with bedding so the two surfaces could be distinguished only with difficulty. Thus in many places, particularly in quartz phyllites and in carbonate units, only a composition layering (designated $F_{0-1}$) was mapped. The major penetrative foliation comprises a slightly micaceous, platy cleavage in quartzite, an intense micaceous-chloritic foliation or "schistosity" in phyllites, and a cleavage in carbonate rocks illustrated only by sparse micas, or by graphitic partings.

The earliest recognized lineations ($L_1$) are generally not very obvious structures involving intersections between bedding
and the F₁ foliation. They include bedding traces, usually illus-
trated by colour bands on the F₁ schistosity, mica-edge traces on bedding surfaces, and cleavage mullions developed at
to fold hinges in quartzitic rocks.

The earliest recognized folds deform only bedding and
have axial planes parallel with the F₁ foliation. They are
generally tight to isoclinal structures, either asymmetric with
single hinges or symmetric with multiple hinges and long para-
sitically folded limbs. Where strong axial plane cleavage is
developed, as in more competent rock types, hinges may be more
or less serrate or even locally obliterated by transposition.

All these folds, irrespective of lithology involved,
display marked limb attenuation (Fig. 3), with the most spec-
tacular attenuations being displayed by thin siliceous layers
within the Trueman and Reeves carbonate members. Rootless intra-
folial folds may also be developed within these latter units.
The folds are considered to approximate those of "similar" type
(Ramsay, 1962a) on the basis of the weak ductility contrast
displayed by the different lithologies involved and on the basis
of cross-sectional shape of the folded layers.

Phase 1 linear structures (fold axes and lineations)
have variable low to moderate plunges generally south to south-
west or less frequently north to northeast or even northwest
(Fig. 4). Axial surfaces too have variable orientations but most
commonly are moderately inclined easterly to southerly. Very
similar orientations are exhibited by bedding and F₁ foliation
surfaces implying that Phase 1 folding is essentially isoclinal.
Figure 3. Examples traced from photographs of minor fold styles from the three fold generations distinguished in the mine areas.
Figure 4. Equal-area projections (lower hemisphere) of Phase 1 minor structural elements (nomenclature as in Table III) from the three mine areas: a-c from H. B.; d-f from Jersey; g-i from Reeves MacDonald. Data from Plates I-VIII. Inset map (after Fyles and Hewlett, 1959) shows location of mine areas in relation to the host Reeves member (black) and to intrusive stocks.
That the variations are largely a consequence of refolding is shown by the dispersion of poles to these surfaces along great circles centred essentially on the modal Phase 2 fold axes (Fig. 4a, d, g).

Phase 2 structures

Phase 2 minor folds are distinguished as open to tight asymmetric folds, generally upright to slightly overturned in attitude (Fig. 3). Asymmetry viewed downplunge (usually southward) is almost invariably dextral. Hinge profiles vary from rounded to subangular, tending toward the former in more massive rocks and toward the latter in finely interbedded and/or strongly cleaved rocks. Attenuation of the limbs is apparently slight.

Axial plane cleavage ($F_2$) is weakly developed in massive rocks and is represented by a crenulation cleavage in phyllitic rocks. Mineral alignment along these planes is limited to sparse micas, possibly of mimetic origin. The crenulation cleavage is widely developed and the crenulation itself, or microfolding of the $F_1$ foliation, forms a particularly prominent lineation ($L_2$) throughout the three areas. Locally a lineation may also be developed on the fold limbs perpendicular to the fold hinges but the age of this relative to the lineation $L_2$ is not known.

These Phase 2 folds are considered to be "flexural-slip" folds modified by flattening (Ramsay, 1962a) because of the contrast in response between competent and incompetent units, and the generally slight attenuation of the fold limbs. The latter was investigated by measuring thickness variations across fold
hinges in several mesoscopic folds and it was found that the orthogonal thickness \( t \) tended to maximum values at the hinges whereas the thickness \( T \) measured parallel to the axial surface tended to minimum values at the hinges (Fig. 5). This geometry is characteristic of flattened "flexural-slip" folds (Ramsay, 1962a).

Phase 2 linear structures plunge at low to moderate angles generally south to south-southwest or less frequently north to north-northeast (Fig. 6). They are approximately coaxial with Phase 1 structures in the H. B. and Jersey areas where modal orientations differ by only 5-10°. Structural trends diverge more markedly in the Reeves MacDonald area where modal orientations differ by approximately 20°. Axial planes exhibit considerably less variation than do Phase 1 axial planes and are usually steeply inclined east-southeast to southeast so that Phase 2 folds can be distinguished geometrically from those of the earlier phase.

Phase 3 structures

Phase 3 folds are shallow to close asymmetric, monoclinal structures, plunging at moderate to steep angles. The larger folds in all lithologies tend to have gently rounded hinges whereas smaller representatives, especially in well-foliated phyllites, have distinctly angular hinges and are really small kink folds (Fig. 3).

Where deformation is more intense, closely spaced kink bands and, locally, crenulations may be developed. However axial plane cleavage is generally very weak or absent. More commonly
Figure 5. Plots of thickness variations across three Phase 2 minor fold hinges, developed in quartzite, quartz phyllite. T, the layer thickness measured parallel to axial planes, is plotted against perpendicular distance from the axial planes. t, the orthogonal thickness of the layer(s), is plotted against distance along the composition layering.
Figure 6. Equal-area projections (lower hemisphere) of Phase 2 minor structural elements (nomenclature as in Table III) from the three mine areas: a-b from H. B.; c-d from Jersey; e-f from Reeves MacDonald. Data from Plates I-VIII. Inset map (after Fyles and Hewlett, 1959) shows location of mine areas in relation to the host Reeves member (black) and to intrusive stocks.
a discrete strain-slip cleavage in phyllitic rocks or a fracture cleavage in more massive rocks is developed. Traces of these planar structures on the $F_1$ foliation or on composition layering produce a weak lineation ($L_3$) with the same orientation as the kink fold and larger monoclinal fold axes.

Phase 3 structures are non-penetrative. They are however widely developed, generally on a mesoscopic scale but locally also on a macroscopic scale, as in H. B. mine area.

The folds can be divided, on the basis of their asymmetry and orientation, into two sets (Fig. 7):

(i) Sinistral folds ($F_3$) plunging at moderate to steep angles east to northeast
(ii) Dextral folds ($F_3$) plunging at moderate to steep angles east-southeast to southeast

Axial planes (or kink planes) are generally subvertical to vertical. The two sets are believed to be coeval and to form a conjugate system, although only rarely are true conjugate pairs recognized. Both sets may however be recognized in all areas although their relative intensities vary markedly.

The orientation of principal stress axes can be determined from the mutual attitudes of the kink planes of conjugate fold systems (Johnson, 1956; Ramsay, 1962c). In this study, modal orientations of kink planes from the two singular sets of folds (sinistral and dextral) were used to determine the stress field orientation; a procedure which Clifford (1968) has shown to be valid where truly conjugate arrays are lacking. This shows in the maximum principal stress bisecting the obtuse angle
Figure 7. Equal-area projections (lower hemisphere) of Phase 3 minor structural elements (nomenclature as in Table III) from the three mine areas: a-c from H. B.; d-f from Jersey; g-i from Reeves MacDonald. Data from Plates I-VIII. Inset at lower right shows geometry and monoclinic symmetry of the Phase 3 conjugate system. Inferred positions of principal stress axes are shown in c, f, and i; 1, 2, 3 refer to subscripts on $\sigma_1 > \sigma_2 > \sigma_3$. 
between the conjugate kink planes (Fig. 7). This is a not uncom-
mon situation in naturally deformed rocks (Roberts, 1971) which
is in accord with the experimental data of Paterson and Weiss
(1966) but not with that of Donath (1964, 1968) or Borg and
Handin (1966). Herein lies a controversy concerning the rela-
tionship between kink bands and principal stress orientations;
in particular, whether or not kink planes are generated parallel
with planes of maximum shearing stress. Paterson and Weiss
(1966) and Weiss (1968) maintain that no causal relationship
exists and propose that kink bands nucleate at a point or line
source, grow by kink plane migration (continuously increasing in
width) and that this is achieved by slip within and along the
foliation. On the other hand, Donath (1964, 1968) and Dewey
(1965, 1969) regard kink bands as foliation segments of constant
length, externally rotated between kink planes which parallel
surfaces of maximum shearing stress. Their positions are con-
sidered to be already defined at infinitesimal strain and there-
after remain fixed.

Whatever the mechanism for their formation, the Phase 3
fold systems in the study areas consistently have monoclinic
symmetry and the deduced maximum principal stress directions are
directed approximately north-south (Fig. 7c, f, i). In Jersey
and H. B. areas, the deduced maximum principal stress direction
is sub-parallel with the modal orientation of the composition
layering whereas at Reeves MacDonald it is more inclined to the
layering. This variation in attitude of the layering relative
to the maximum principal stress may account for variation in
the relative intensities of the two sets of the conjugate fold system from one area to another, as has been illustrated by Roberts (1971).

EFFECTS OF SUPERIMPOSED FOLDING

Phase 2 superimposed on Phase 1

Phase 2 folding has produced widespread reorientation and hence modification to the geometry of earlier structures, both linear and planar.

Farther north in the Kootenay Arc, Ross (1970) has shown that the earliest linear structures (L1) plunge both north and south and are frequently curvilinear. He has proposed that this is produced by variable slip within the foliation parallel with the a2-direction of Phase 2 folding and that culminations/depressions of L1 therefore coincide with axial surfaces of Phase 2 folds. Within the mine areas, Phase 1 linear structures display similar plunge variations and curvilinear lineations have been observed locally. However, Phase 2 linear structures also exhibit plunge culminations/depressions on a broad scale so that, unless there has been differential flattening of Phase 2 folds, all the variation in L1 orientation cannot be the product of slip during Phase 2 folding. That it is in part can be demonstrated on the Phase 2 culminations where L2 is uniformly horizontal but L1 has variable plunge (±10°) as at the northern end of the Jersey mine area (Plate IV).

In spite of their variable orientation, no clearly defined dispersion path (in stereographic projection) of L1 linear
structures about the Phase 2 axes of folding can be determined, presumably because of their near coaxial trends; the difference in modal orientations is less than 5° at Jersey, 10° at H. B. and 20° at Reeves MacDonald (compare Fig. 4 and 6). It is thus not possible because of the ill-defined \( L_1 \) dispersion paths to deduce the specific deforming process operative during Phase 2 folding. Phase 1 planar elements, both the micaceous-chloritic foliation and fold axial surfaces, are clearly folded about the Phase 2 axes, as illustrated by the dispersion of poles to both sets of surfaces along great circles centred essentially on the modal Phase 2 axes (Fig. 4a, d, g). This is best illustrated in Jersey mine area (Fig. 4d) where macroscopic Phase 2 folding is correspondingly best developed.

Interference fold structures of two distinct types have been recognized (Fig. 8). The basin/dome structures (Type 1 of Ramsay, 1962b) are less common and have been recognized only within calcite marble units. They are produced by the same mechanism as the curvilinear \( L_1 \) lineations already described. More common, and recognized in all lithologies, are approximately coaxially refolded folds, corresponding to Type 3 of Ramsay (1962b), in which the Phase 1 axes were oriented close to the Phase 2 axial surfaces (Fig. 8c-f).

Flattening of Phase 1 folds has presumably also occurred during Phase 2 folding. This is believed to be reflected in the occurrence of intense shearing/transposition in fold closures and perhaps also in the occurrence of boudinage structures within carbonate units on fold limbs, although the latter structures
Figure 8. Examples of interference fold structures resulting from superimposition of Phase 2 folds on earlier formed Phase 1 folds. a-d are developed in calcite marble; e-f are developed in quartz phyllite and quartzite.
may have developed wholly during Phase 1. Shearing/transposition appear to be particularly intensely developed where Phase 2 macroscopic folds are either absent or weakly developed, as in the H. B. and Reeves MacDonald mine areas. It is also perhaps significant that the nearest approach to coplanarity between modal $F_1$ and $F_2$ axial surfaces occurs in the H. B. mine area (compare 4c and 6b). This may suggest that, where Phase 1 structures were already steeply inclined, these early folds were tightened up and flattened rather than refolded (s.s.) during Phase 2 deformation (Fig. 9).

**Phase 3 superimposed on Phases 1 and 2**

Phase 3 folding is a non-penetrative event so that its effects are localized. These are most clearly illustrated in part of H. B. mine area where several macroscopic sinistral folds refold earlier structures (Plate VIII). There, Phase 1 linear structures have very variable orientations but are dispersed (in stereographic projection), with considerable spread, along an 80° small circle locus centred on the Phase 3 sinistral axis (Fig. 4c). Such constant angles between $L_3S$ and $L_1$ would imply, according to Ramsay (1962), that Phase 3 folds are "flexural-slip" in type. (These folds have been referred to, above, as kink folds and the controversy surrounding the mechanism of formation of such folds has been referred to.) Poles to the related planar structures, $F_1$ foliation and axial surfaces, are dispersed along a very similar small circle locus (Fig. 4a, b).

Insufficient Phase 2 structures were recognized within this particular sub-domain to illustrate their response to Phase
Figure 9. Diagram showing the effects of Phase 2 folding on Phase 1 structures: (a) by refolding and (b) by closure and differential flattening.
3 sinistral folding. Elsewhere in H. B. mine area and throughout Jersey and Reeves MacDonald mine areas no equivalent clearly defined dispersion of either Phase 1 or Phase 2 linear structures can be distinguished. Both, however, do show culminations/depressions within the composition layering \((F_{0-1})\), as has been referred to above, but the development of the broad culminations/depressions exhibited by Phase 2 linear structures is not clearly understood. The most likely explanation is that they developed as a result of Phase 3 folding since the trend of the culminations/depressions in plunge in the Jersey and H. B. mine areas parallel the traces of Phase 3 axial surfaces (Plates VI, VIII).

SUMMARY OF STRUCTURAL STYLES AND RELATIVE AGES OF FOLDING

Three phases of folding, all at least of local significance, have been recognized. Their styles, average orientations and age relations are summarized in Table IV.

Using local evidence, only an upper age limit can be placed on these since they all pre-date granitic intrusion in the area, dated at 100 m.y. (Lower Cretaceous).

The relative timing of the phases is as indicated but the continuity or otherwise cannot be fully established within the small areas studied.

MAJOR FOLDING

The three strata-bound sulphide deposits are all located on the east limb of complex major anticlinal structures referred to as the Salmo River anticline and the Jersey anticline in
### Table IV. Summary of structural styles and relative ages of folding.

<table>
<thead>
<tr>
<th></th>
<th>Phase 1</th>
<th>Phase 2</th>
<th>Phase 3</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Distribution</strong></td>
<td>Minor folds: common</td>
<td>Minor folds: common but locally restricted</td>
<td>Minor folds: widespread but variably developed</td>
</tr>
<tr>
<td></td>
<td>Major folds: dominant</td>
<td>Major folds: common but locally restricted</td>
<td>Major folds: locally developed</td>
</tr>
<tr>
<td><strong>Surface folded</strong></td>
<td>$F_0$</td>
<td>$F_0, F_1$</td>
<td>$F_0, F_1, F_2$</td>
</tr>
<tr>
<td><strong>Morphology</strong></td>
<td>Tight to isoclinal, similar folds with attenuated limbs</td>
<td>Open to close, asymmetric flattened flexural-slip folds</td>
<td>Shallow to open, asymmetric monoclinal folds and kink bands (conjugate)</td>
</tr>
<tr>
<td><strong>Axial surface foliation</strong></td>
<td>$F_1$ schistosity in phyllite, axial plane cleavage in quartzite and limestone</td>
<td>$F_2$ crenulation cleavage in phyllite, weak axial plane cleavage in massive rocks</td>
<td>$F_3$ strain-slip to crenulation cleavage in phyllite, weak fracture cleavage in massive rocks</td>
</tr>
<tr>
<td><strong>Lineation</strong></td>
<td>$L_1$ intersections of $F_0/F_1$</td>
<td>$L_2$ crenulation of mica and $F_2/F_0$ intersection</td>
<td>$L_3$ intersection of $F_3/F_0$</td>
</tr>
<tr>
<td><strong>Attitude</strong></td>
<td>Variably inclined axial surfaces, variable axial trends, low to moderate plunge</td>
<td>Steep axial surfaces, mainly S to SSW axial trends, low to moderate plunge</td>
<td>Steep axial surfaces, NW and SE axial trends, moderate to steep plunge</td>
</tr>
</tbody>
</table>
Reeves MacDonald and Jersey-H. B. mine areas respectively (Fyles and Hewlett, 1959). Although they are not continuous these appear to be essentially the same structure exposed at different levels. It is proposed that the name—Salmo River anticline—be extended to include the whole structure. Overall plunge is southward but plunge culminations contribute to persistence of the structure along strike. Structural complexity diminishes with increasing depth in the anticline which is reflected in degree of exposure of the quartzitic core rocks. Complications at higher levels are due to multiple hinges, to tectonic thickening/thinning particularly of the Trueman member, and to the effects of tectonic slides (Fleuty, 1964) and thrusts (see descriptions on p. 57).

H. B. mine area (Plate VIII, Fig. 10)

Major structure of H. B. mine area is the most complex of the three areas. Additional structural complications are due to Phase 3 major folding, to obscure stratigraphic relations between dolomite marble and calcite marble of the Reeves member, and to the uncertain nature of phyllite intercalations within the Reeves member.

Phase 1 major folds plunge southward and have axial surfaces inclined steeply eastward. These are most clearly illustrated by the anticlines cored by quartz phyllite/schist of the Reno formation (Fig. 10). Internal structure of the Reeves member is obscure due to widespread intense transposition. Dolomite marble within the member forms huge pod-like masses which are difficult to interpret structurally because their
Figure 10. Vertical cross-sections (subnormal to major fold axes) of H. B. mine area. Locations and general geology are shown on Plate VIII.
limits cannot be assumed to have been, even originally, stratigraphic. Very thin intercalations of phyllite also within the Reeves member are interpreted as stratigraphic and not as extremely attenuated isoclines as envisaged by Pyles and Hewlett (1959).

Generally weak development of Phase 2 folding and widespread development of transposition structure is interpreted as indicating tightening and hence possible flattening of Phase 1 folds. Low to high angle thrusting which further complicates the structure is believed to have occurred toward the end of Phase 2 deformation (Fig. 10).

Phase 3 sinistral macroscopic folds are well developed in the area. These plunge steeply northeast and have subvertical axial surfaces. Effects of this folding contribute to subsurface complexities in structure of the dolomite marble "pods" and hence in configuration of the ore zones southward downplunge along the main host Phase 1 structure (Fig. 10). The latter is inferred to be a tight synclinally folded dolomite marble lens occupying what has been referred to as the H. B. syncline.

**Jersey mine area (Plate VI, Fig. 11)**

The major anticlinal structure in Jersey mine area is in part recumbent, with the predominantly phyllite-marble limbs dipping more gently eastward than the more upright quartzitic core rocks (Fig. 11).

The limbs are folded into relatively open, upright slightly asymmetric Phase 2 folds. These have steep, generally eastward-dipping axial surfaces. Core rocks of the anticline
Figure 11. Vertical cross-sections (subnormal to major fold axes) of Jersey mine area. Locations and general geology are shown on Plate VI.
contain only minor Phase 2 folds superimposed on the Phase 1 structure. The latter is complicated by tectonic thickening and repetition in multiple hinges of the Trueman member, and by sliding at the base of that member. The slight departure in axial trends between the two fold generations can be appreciated from the distribution of major axial traces illustrated in Plate VI.

Phase 3 folding is represented by sparsely developed mesoscopic kink folds; but at the northern end of the area a broad culmination causes northward plunge reversals of both Phase 1 and 2 linear structures. Its trace parallels axial traces of Phase 3 minor sinistral folds and so the culmination is considered to be genetically related to Phase 3 deformation.

Reeves MacDonald mine area (Plate III, Fig. 12)

The most distinctive structural feature in Reeves MacDonald mine area is the nearly east-west strike and more southwesterly fold axial trends. According to Yates (1970) this abrupt change in strike is produced by aggregate displacement of northward-directed thrusts and associated tear faults. Appropriately oriented thrusts exist in Reeves MacDonald area but cross-cutting faults exhibit only weakly oblique displacements. Development of the east-west strike could also be partly explained by invoking large scale dextral kink folding (Phase 3) accompanying thrusting, with the northern antiformal hinge located east-northeast of the mine area (Fig. 1) and the middle limb striking east-west through the mine area. A complementary
Figure 12. Inclined cross-section (subnormal to major $F_1$ fold axes) of Reeves MacDonald mine area. Location and general geology are shown on Plate III.
synformal hinge cannot be located since the structure farther west is concealed by thrusting. However there is little real evidence that Phase 3 folding occurred on such a large scale.

Within the mine area, structure is dominated by the Salmo River anticline. Its axial surface dips steeply south-southeast and its plunge (up to 50°) is anomalously steep. South of the anticline lies the Reeves syncline which conformably encloses dolomite marble and associated sulphide mineralization. This structure was considered by Fyles and Hewlett (1959) to be a secondary (Phase 2) fold but was reinterpreted by Macdonald (1970), on the basis of minor fold geometry, to be a Phase 1 fold. It is a steeply plunging, extremely attenuated syncline and the relatively competent dolomite marble in its core is highly transposed. The structure is believed to have been flattened during Phase 2. The Reeves member is repeated south of the Reeves syncline and appears to occupy the overturned limb of another syncline, perhaps originally a structurally higher part of the Reeves syncline now juxtaposed along a lag surface.

Whereas Phase 2 minor structures are ubiquitous and intensely developed (except in carbonate units) the major folds are relatively open structures of limited dimensions. Axial traces transect Phase 1 traces at 60-70° reflecting a divergence in trends between the two phases (Plate III) which probably contributes to the southward developing curvature of this segment of the Arc.
FAULTING

Major faults developed within and adjacent to the so-called "Mine Belt" or "Thrust Belt" south of Salmo are shown in Figure 13 (after Little, 1960; Frebold and Little, 1962; Yates, 1970). The faulting includes tectonic slides (not shown in the Figure) and thrust faults cut by a variety of transverse faults.

Tectonic sliding, albeit on a limited scale, has contributed significantly to the development of the complex Salmo River anticline. The main locus of sliding has been along the Trueman member located between the Reeves marble member and the underlying thick quartzitic sequences, resulting in varied and locally drastic attenuation of that member and bringing Reeves marble into contact with the quartzitic rocks (Plates III, VI, VIII). From their distribution in relation to the major fold structures, slide movements are presumed to have been concurrent with Phase 1 folding perhaps also being reactivated by flattening during Phase 2.

The thrust faults, in contrast, are discrete structural breaks which cut across the stratigraphic section and juxtapose widely disparate units. They are represented by the Waneta, Argillite, Black Bluff and other lesser faults which dip steeply eastward to southward (Fig. 13). The Waneta fault marks the structural contact between Lower Palaeozoic and Mesozoic rocks along the western margin of the "Thrust Belt" whereas the Black Bluff fault marks the eastern limit of the belt. This latter fault was considered by Ross (1970) to be the southern expression of a major sole thrust separating westerly verging parautochthonous
Figure 13. Main fault and fold patterns in the southern Kootenay Arc (after Little, 1960; Frebold and Little, 1962; Yates, 1970).
rocks from easterly verging allochthonous rocks. However within this area no difference in vergence was recognized across this fault. The Argillite fault lies between these faults in the north but merges southwestwards with, and apparently supercedes the Black Bluff fault. It everywhere separates Lower Cambrian rocks from the Ordovician Active Formation to the east and may represent a pre-folding décollement thrust since the Active Formation is everywhere apparently in fault contact with other units.

The age of thrust faulting is problematic. Fyles and Hewlett (1959) believe it to have been initiated during the first phase of folding and that movement continued during the second phase of folding. They also consider thrusting to have been directed essentially westwards.

Certainly the north-trending thrust segment, as illustrated by the Argillite fault in H. B. mine area, has been refolded by, and hence predates Phase 3 event. The southwest to west-southwest trending segment extends into the Metaline-Colville quadrangles of northeast Washington where thrusting is directed essentially northward and is associated with large scale east-west kink folding (Yates, 1970). Similarly in the Reeves MacDonald area, thrusting appears to be related to kink folding, especially since principal stress directions derived from assumed conjugate Phase 3 sets is compatible with northward-directed movement.
Thus thrusting may be of two distinct ages comprising (a) pre-Phase 3 (Phase 2?) westward-directed movements, and (b) Phase 3 northward-directed movements.

Transverse faults include a variety of northeast- to northwest-trending faults which cut across folds and thrusts alike. Some of these may be strike-slip faults but there is conflicting opinion with respect to this.

In the Salmo map area, Walker (1934) and Frebold and Little (1962) have described right-lateral strike-slip movements along northwest to west-northwest-trending faults, such as the Ripple Creek fault cutting the southern end of the Sheep Creek anticline and the Mt. Erie faults west of Salmo village. A conjugate northeast-trending set is developed but only weakly so. Fyles and Hewlett (1959), however, found no evidence for strike-slip movement on the Ripple Creek fault.

A contrasting sense of movement along early, similarly oriented system of strike-slip faults in the Sheep Creek area is described by Mathews (1953). There, left-lateral slip occurs along northwest-trending fractures and right-lateral along northeast-trending fractures, with the latter being the dominant set.

No evidence for strike-slip movements was found within the mine areas; north-northwest striking faults in the Reeves MacDonald area are either normal dip-slip or slightly oblique-slip faults.

A group of north to north-northeast striking, normal (?) faults is also recognized which appear to both pre- and post-date
granitic intrusion both in the Salmo district (Fyles and Hewlett, 1959) and in the Metaline-Colville quadrangles (Yates, 1970). Some are occupied by sheared lamprophyre dykes suggesting that movements continued at least until the Eocene (see Section II, p. 22).

MINOR FRACTURES

Joint orientations measured in the course of mapping are illustrated, via equal-area projections of poles to joint surfaces and via rose diagrams in Figure 14.

Cross joints are the dominant fractures in all three areas and may comprise paired, closely oriented sets. They are subvertical to vertical and strike normal to the main structural trends. If the paired maxima (Fig. 14) are real they may reflect different stress regimes associated with Phase 1 and Phase 2 folding. In many places, these fractures have opened to form quartz-carbonate-filled gash veins which in Jersey mine area may be fringed by grossularite-actinolite skarn implying that opening of these fractures either preceded or accompanied intrusion of granitic stocks in the district. It is suggested that they were formed by slight extension accompanying intrusion.

A set of steep joints, striking north to north-northeast, is also developed but it is not clear whether these represent true longitudinal joints with respect to Phase 1 and 2 fold structures, or are related to the high-angle, north to north-northeast striking faults developed throughout the district.

Several other weakly developed joint sets occur. These are generally moderately to steeply inclined and strike either
Figure 14. Equal-area projections of poles to joint surfaces, and rose diagrams of joint directions from the three mine areas: (a) from H. B.; (b) from Jersey; and (c) from Reeves MacDonald. Inset map (after Fyles and Hewlett, 1959) shows location of mine areas in relation to the host Reeves member (black) and to intrusive stocks.
approximately northeast or northwest, oblique to Phase 1 and 2 fold trends. They are subparallel with Phase 3 trends and could represent conjugate shear joints related to the inferred Phase 3 north-south compression.

STRUCTURAL SYNTHESIS

Phase 1 and Phase 2 axial trends are near coaxial in H. B. and Jersey mine areas but diverge more widely in Reeves MacDonald area. Both phases also have westward vergence. There is, however, no local evidence to determine whether they both result from a single protracted period of deformation or whether they result from two essentially distinct deformation periods.

Phase 1 folding was near isoclinal and was synmetamorphic (lower greenschist facies) whereas Phase 2 folding was more open and developed at submetamorphic temperatures. The effect of Phase 2 folding superimposed on earlier Phase 1 structures has been to produce refolding and in some cases tightening up of Phase 1 folds resulting in flattening which has led to increased tectonic thinning of incompetent units, shearing of more competent units and ultimately thrusting.

Phase 3 deformation is non-penetrative and is represented by kink folding (believed to be conjugate) and by weak strain-slip cleavage, both cutting at a high angle across earlier structural trends except toward the south in Reeves MacDonald area. The geometry of this phase is interpreted as resulting from an approximately north-south compression, i.e., the product of a different stress system than that which produced the two earlier phases. This Phase 3 event would appear to be the direct
equivalent of thrusting and kink folding about east-west axes described by Yates (1970) from Metaline-Colville districts. The effect of superimposition of Phase 3 has been to produce effective, if limited, north-south shortening possibly related to northward-directed thrust faults developed near and south of the International Boundary. These north-south directed structures could be the result of accommodation movements following east-west stacking of Phase 1 and Phase 2 folds. Some of the conjugate strike-slip faults recognized by other workers in Salmo district may be related to these movements (Mathews, 1953; Frebold and Little, 1962).

Brittle-type extension preceded and/or accompanied granitic intrusion, dated at 100 m.y. age (see Section II, p. 19), in the vicinity of the mines, producing closely spaced east-west jointing and gash veining.

North to north-northeast normal faulting and fracturing in part post-dates granitic intrusion. Lamprophyre dyke intrusion dated at 49 m.y. (Yates and Engels, 1968; this study: Section II, p. 22) is partly controlled by this same fracture system.

When a regional structural synthesis is attempted (Table V) it becomes clear that Phase 3, unlike Phases 1 and 2, cannot be correlated from north to south along the southern Kootenay Arc although there are general similarities in the character of deformation with open folds and strain-slip or fracture cleavage having been developed. Farther north, these folds tend to be coaxial with the two earlier phases and have been
Table V. Structural correlation along the southern Kootenay Arc.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>PHASE</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Open to tight folds</td>
<td>Isoclinal similar folds</td>
<td>Isoclinal, intra-foial similar folds</td>
<td>Isoclinal, attenuated similar folds</td>
</tr>
<tr>
<td>F₁: A.P. cleavage dipping SE</td>
<td>F₁: A.P. cleavage to schistosity</td>
<td>F₁: A.P. cleavage to schistosity</td>
<td>F₁: A.P. cleavage to schistosity</td>
<td></td>
</tr>
<tr>
<td>L₁: 20-30 plunge SW</td>
<td>L₁: variable plunge</td>
<td>L₁: low plunge N or S</td>
<td>L₁: variable low plunge</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Open to close, asymmetric, flattened flexural-slip folds</td>
<td>Open to isoclinal, asymmetric, flattened flexural-slip or similar folds</td>
<td>Open, asymmetric folds</td>
<td>Open to tight, asymmetric flexural-slip or similar folds</td>
</tr>
<tr>
<td>F₂: crenulation cleavage dipping steeply E</td>
<td>F₂: A.P. to crenulation cleavage</td>
<td>F₂: crenulation cleavage dips SW</td>
<td>F₂: A.P. cleavage to schistosity, steep dip E or W</td>
<td></td>
</tr>
<tr>
<td>L₂: low plunge S-SSW</td>
<td>L₂: low plunge N or S</td>
<td>L₂: low plunge S to SE</td>
<td>L₂: low plunge N to SE</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>Major kink fold</td>
<td>Shallow to open flexural folds, kinks</td>
<td>F₃: strain-slip or fracture cleavage dips NE</td>
<td>Asymmetric minor and chevron folds</td>
</tr>
<tr>
<td>L₃: axis plunges v. steep N80W</td>
<td>L₃: steep plunges NE and SE</td>
<td>L₃: steep plunge W</td>
<td>Orthorhomic folds</td>
<td>L₇: steep plunge W to NW</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>F₇: conjugate shears dip NE and E</td>
<td>Local intense deformation marginal to granitic intrusion</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>L₇: plunge ESE and S</td>
<td>Variably oriented kinks and fractures, related to emplacement of Nelson batholith</td>
</tr>
</tbody>
</table>

Local intense deformation marginal to granitic intrusion.
considered as resulting from a single protracted deformation (Ross, 1970). Whereas in the south, Phase 3 structures transect earlier structures except near the Canada-U.S.A. border where there is the abrupt westward swing in strike of the Arc. Yates (1970) has proposed that this abrupt swing or bend is produced by aggregate displacement of northward-directed thrusts and related tear faults. This bend is, however, imposed on a less marked regional curvature which is related to southward-developing divergence of Phase 1 away from Phase 2 structural trends, possibly accompanied by a reduction in intensity of the latter.

In the Salmo district, both Phase 1 and Phase 2 structures display westward vergence at the levels observed. Ross (1970) has suggested that such westerly verging Lower Palaeozoic rocks along the east side of the Kootenay Arc, represent a parautochthonous cover displaced westward off the Purcell Arch, possibly as a consequence of uplift. Further, Phase 2 folding in these cover rocks developed as a result of reaction with allochthonous nappe structures verging easterly towards the Purcell Arch.

Farther south, in the vicinity of the International Boundary, Phase 1 folds trend more southwest, gradually become more open structures, and are less (if at all) affected by Phase 2 folding, suggesting that there eastward crowding was less marked.

Timing of the deformation in the southern segment of the Arc is problematic. Yates (1970), for example, believes
deformation in Metaline-Colville districts, although perhaps initiated in the Triassic, was at a maximum producing the main folding in the late Jurassic and was completed before 100 m.y. ago (mid-Cretaceous).

In Salmo district, at least two sets of macroscopic folds have been recognized in the Jurassic Rossland group west of the mine areas (Little, 1960; Frebold and Little, 1962). They constitute upright, asymmetric folds plunging at low angles south to southwest (equated by Frebold and Little with Phase 1 folds in the mine areas) which locally are warped by northeast-trending asymmetric folding overturned to the southeast (Fig. 14). From their geometry, these folds might more plausibly be equated with Phase 2 and Phase 3 structures developed in the mine areas, but even this correlation is very uncertain. However, it is suggested that Phase 1 folding at least is pre-Jurassic in age.

The chronology of deformation farther north along the Arc is also a matter of some debate. Ross and Kellerhals (1968) and Ross (1970) have proposed that the Slocan group of Triassic age is affected only by Phase 3 folding and hence that Phase 1 and Phase 2 folding is entirely of Palaeozoic age. Read (1966) and Wheeler (1970), however, believe that (i) Phase 1 folding, although Palaeozoic in age, pre-dates the Milford group and is therefore pre-early Mississippian in age, and also that (ii) Phase 2 folding may have affected rocks of Mesozoic age.

Whilst the meagre chronological evidence from Salmo district contributes little to this debate, it does at least permit structural correlations to be made with the more southerly segment of the Kootenay Arc in Washington.
SECTION IV

METAMORPHIC ENVIRONMENT

INTRODUCTION

The metamorphic character of the Lower Palaeozoic rocks in the Salmo district has been briefly described by Fyles and Hewlett (1959) who distinguished a contact metamorphic event superimposed on low-grade regional metamorphism.

In this study, the limits of the contact aureoles in the mine areas have been defined and outer/inner contact "zones" distinguished where possible. Characteristic textural and mineralogical features of the host rocks are described and thermal metamorphic features are separated from those of regional metamorphism.

The pressure-temperature conditions accompanying both metamorphic events have been estimated from experimentally determined mineral reactions and from X-ray diffraction determinations of calcite compositions from coexisting calcite-dolomite assemblages. Such estimates applied to the lead-zinc sulphides at least allow a comparison to be made of their mineralogical and textural responses to contact metamorphism (see Section V).

A summary of the relevant stratigraphy, metamorphic rock types and typical mineral constituents of the mine areas is presented in Table VI.
Table VI. Metamorphic rock types and typical mineralogies of the exposed stratigraphic units in the mine areas: (i) outside contact aureoles, (ii) within outer contact aureoles, and (iii) within inner contact aureoles.

<table>
<thead>
<tr>
<th>Stratigraphic Unit</th>
<th>OUTSIDE THE CONTACT AUREOLES ROCK TYPE and typical mineralogy</th>
<th>WITHIN THE OUTER AUREOLES ROCK TYPE and typical mineralogy</th>
<th>WITHIN THE INNER AUREOLES ROCK TYPE and typical mineralogy</th>
</tr>
</thead>
<tbody>
<tr>
<td>ACTIVE FORMATION</td>
<td>SLATE, PHYLITEx: graphite-quartz-albite-calcite-muscovite</td>
<td>SLATE, PHYLITEx: graphite-tremolite-albite-calcite-muscovite</td>
<td>(not determined)</td>
</tr>
<tr>
<td></td>
<td>LIMESTONE: graphite-quartz-muscovite</td>
<td>MARBLE: graphite-tremolite-muscovite-calcite</td>
<td></td>
</tr>
<tr>
<td>UPPERC LAIB MR.</td>
<td>PHYLITEx: albite-quartz-chlorite-muscovite</td>
<td>PHYLITEx: albite-chlorite-biotite-quartz-muscovite</td>
<td>SCHIST: plagioclase-cordierite-biotite-quartz-muscovite</td>
</tr>
<tr>
<td></td>
<td>PHYLONITE: quartz and perthite porphyroclasts in phylite matrix</td>
<td></td>
<td></td>
</tr>
<tr>
<td>EMERALD MR.</td>
<td>PHYLITEx: graphite-quartz-albite-clinozoisite-muscovite</td>
<td>PHYLITEx: graphite-tremolite-albite-clinozoisite-muscovite</td>
<td>(not present)</td>
</tr>
<tr>
<td></td>
<td>LIMESTONE: graphite-quartz-muscovite</td>
<td>MARBLE: graphite-tremolite-muscovite-calcite</td>
<td></td>
</tr>
<tr>
<td>REEVES MR.</td>
<td>CALCITE MARBLE: quartz-muscovite-dolomite-calcite</td>
<td>CALCITE MARBLE: tremolite-dolomite-calcite and quartz-tremolite-calcite</td>
<td>CALCITE MARBLE: diopside-dolomite-calcite</td>
</tr>
<tr>
<td></td>
<td>DOLOMITE MARBLE: quartz-phlogopite-calcite-dolomite</td>
<td>DOLOMITE MARBLE: tremolite-calcite-dolomite</td>
<td>DOLOMITE MARBLE: forsterite calcite-dolomite</td>
</tr>
<tr>
<td>TRUEMAN MR.</td>
<td>CALCITE PHYLITEx: quartz-muscovite-dolomite-calcite</td>
<td>CALC-SILICATE HORNFELS: plagioclase-actinolite-(diopside)-calcite-quartz</td>
<td>CALC-SILICATE HORNFELS: grossularite-idocrase-scapolite-wollastonite-diopside-plagioclase, etc.</td>
</tr>
<tr>
<td></td>
<td>PHYLITEx, QUARTZ PHYLITEx: (chloritoid)-chlorite-albite-quartz-muscovite</td>
<td>PHYLITEx, QUARTZ PHYLITEx: albite-chlorite-biotite-muscovite-quartz</td>
<td>SCHIST, QUARTZ SCHIST: albite-biotite-muscovite-quartz</td>
</tr>
<tr>
<td></td>
<td>QUARTZITE: chlorite-muscovite-quartz</td>
<td>QUARTZITE: muscovite-biotite-quartz</td>
<td>QUARTZITE: muscovite-biotite-quartz</td>
</tr>
<tr>
<td>QUARTZITE RANGE FN.</td>
<td>PHYLITEx, QUARTZ PHYLITEx: (chloritoid)-albite-clinozoisite-muscovite-quartz</td>
<td>PHYLITEx, QUARTZ PHYLITEx: albite-chlorite-biotite-muscovite-quartz</td>
<td>SCHIST, QUARTZ SCHIST: plagioclase-cordierite-biotite-andalusite-quartz</td>
</tr>
<tr>
<td></td>
<td>QUARTZITE: chlorite-muscovite-quartz</td>
<td>QUARTZITE: muscovite-biotite-quartz</td>
<td>QUARTZITE: muscovite-biotite-quartz</td>
</tr>
</tbody>
</table>
REGIONAL METAMORPHISM

Textural features of the regionally metamorphosed host rocks

Chlorite and/or white mica (mainly muscovite (2M), or phlogopite (2M) in dolomite marble) are aligned parallel with both $F_0$ and $F_1$ surfaces in all rock types. In phyllites, the muscovite is commonly bent around $L_2$ crenulations where it locally exhibits subgrain development. It also occurs locally as random flakes and as sparse developments along $F_2$ cleavage planes; the latter possibly as mimetic alignments. Chloritoid, abundant in some chloritic phyllites as elongate metacrysts, displays a marked preferred orientation parallel with $F_1$ surfaces.

Quartz is generally strained. In dolomite and calcite marbles, quartz phyllites and phyllonites, it is intensely so and larger grains display distinct dimensional preferred orientation, strongly undulose extinction, subgrain development and serrate grain boundaries. These larger grains occur commonly as flattened augen or lenses locally in a matrix of polygonal-shaped, strain-free quartz of fine grain size (15-30 μm). Such fine-grained polygonal textures are typical of quartz occurring in the various fine-grained phyllites. In quartzites, the quartz exhibits only slight dimensional preferred orientation, weak strain extinction and typically granoblastic textures. However, crystallographic preferred orientation is strongly developed, as shown by stereographic plots of $c$-axes orientations (Fig. 46a,b). In places, sub-basal deformation lamellae are intensely developed (Fig. 47).
The carbonates, calcite and dolomite, form recrystallized equigranular aggregates within which the $F_1$ foliation is outlined by graphitic partings and aligned white mica. Dolomite exhibits fine-grained (30-80 μm) granoblastic textures and calcite fine- to medium-grained (less than 300 μm) subgrano- blastic textures.

Mineral assemblages, metamorphic facies, and timing of the regional metamorphic event

Mineral assemblages from pelitic rocks, in parts of Reeves MacDonald and H. B. mine areas which have not been modified by contact metamorphism, indicate that regional metamorphism was Barrovian in type and restricted to lower greenschist facies. These assemblages include:

- chloritoid-chlorite-muscovite-quartz
- albite-epidote-chlorite-muscovite-quartz
- calcite-chlorite-muscovite-quartz

The presence of chloritoid, chlorite and albite together with epidote and the general absence of coeval biotite define the subfacies to be quartz-albite-muscovite-chlorite subfacies (Turner and Verhoogen, 1960, p. 534).

Timing of this metamorphism was apparently essentially synchronous with Phase 1 deformation, as evidenced by the preferred orientation of phyllosilicates and chloritoid. To what extent, if any, metamorphism outlasted this phase of deformation is not clear as textures were subsequently affected by Phase 2 deformation which appears to have been a purely tectonic event, unaccompanied by mineralogical change. It is true that in places
white mica is sparsely aligned parallel with Phase 2 axial plane cleavage but, like biotite, it is probably of later mimetic origin (i.e., post-Phase 3).

**P-T conditions of regional metamorphism**

The occurrence of chloritoid allows a lower P-T limit to be estimated for the regional metamorphic event in this district. Ganguly (1968) and Hoschek (1969) have investigated experimentally the stability of chloritoid and the latter has proposed the following reaction for its formation:

\[
\text{chlorite} + \text{kaolinite (or pyrophyllite)} = \text{chloritoid} + \text{quartz}
\]

Based on equilibrium conditions for the synthesis of chloritoid, this would imply a temperature of formation of approximately 425°C between 4-6 kb total pressure. The general absence of biotite would also suggest that temperatures were not much in excess of 400°C.

The temperature of crystallization of magnesian calcite, from calcite-dolomite assemblages, was also determined by X-ray diffraction using the method of Graf and Goldsmith (1955, 1958) (see Appendix A). Samples, from the Reeves member at the Reeves MacDonald mine, contained a little over 3 mole % MgCO₃, corresponding to a temperature of crystallization of 450°C ± 25°C. This provides only a minimum temperature since later annealing may have caused exsolution. Goldsmith and Newton (1969) have shown that the effect of total pressure on the solubility of MgCO₃ in the calcite structure is slight, amounting to only 0.12 mole % per kb (measured in the range 500-800°C). Although these samples are from outside the limit of the contact aureole, as indicated by
the absence of random biotite, it is not possible to be certain that the indicated temperature of crystallization is that of regional metamorphism unaffected by later contact metamorphism, temperature estimates for the two events being very similar (cf. p. 82).

CONTACT METAMORPHISM

Textural features of the contact metamorphosed host rocks

Chlorite, occurring with biotite in outer aureoles, is generally absent from inner aureoles except where it is retrogressive, and hence mimetic, after biotite. The F₁ foliation is therefore illustrated mainly by white mica and mimetic biotite. Both of the latter also occur as random flakes and mimetic alignments parallel with F₂ cleavage. In addition, biotite forms random poikiloblasts and clusters nucleated around magnetite grains. Other post-tectonic porphyroblasts and poikiloblasts are formed by andalusite and cordierite respectively.

Quartz, where present, is generally less intensely strained than outside the contact aureole. In marbles and calcite phyllites, quartz has reacted to form various calc-silicates. In both schists and quartz schists, the quartz is typically strain-free and has well-developed polygonal-shaped grain boundaries. In quartzites, textures vary from granoblastic to interlocking; the latter apparently being achieved by selective grain growth. Strain extinction is generally weak but locally may be intense and associated with sub-basal deformation lamellae.
Dimensional preferred orientation is lacking but crystallographic preferred orientation persists as illustrated by $c$-axes orientations (Fig. 46e-g).

Dolomite has a characteristic medium-grained (80-250 μm) granoblastic to polygonal texture. In contrast, calcite exhibits wide variation in grain size from medium to coarse (0.2-5.0 mm), typically allotriomorphic textures and relatively intense twinning development.

The calc-silicates--tremolite, diopside and forsterite--occur as sparse, randomly oriented laths and idioblasts within massive dolomite and calcite marbles. In layered calc-silicate hornfels, grossularite-rich layers and/or actinolite- and diopside-rich layers alternate with quartz-calcite-zoisite schist and biotite schist layers. Associated calc-silicate minerals include idocrase, scapolite and wollastonite. Closely spaced fractures filled with quartz-calcite cut through these calc-silicate assemblages, especially the more competent grossularite layers.

Nature and distribution of the contact aureoles

The distribution of granitic stocks within and around the mine areas is shown in Figure 2, from which it is obvious that of the contact metamorphic effects described, only those in Jersey mine area show a close spatial relationship to exposed intrusions.

The outermost limit of contact metamorphism that could be detected mineralogically was taken at the first appearance
of biotite in pelitic rocks and of actinolite/tremolite in carbonate rocks. An inner contact aureole was defined by the presence of cordierite in pelitic rocks and of forsterite + calcite, or of diopside or grossularite in carbonate rocks.

Higher metamorphic grade may be attained locally, close to intrusive contacts, as evidenced by the presence of fibrolite and andalusite. Such occurrences are rare and no additional contact zone was defined.

Reeves MacDonald mine area. Figure 15 illustrates the occurrence of post-tectonic biotite in pelitic rocks and of occasional actinolite in carbonate rocks in Reeves MacDonald mine area. Andalusite is not developed. The sporadic distribution of these indicator minerals outlines an outer contact aureole, limited to the northern overturned limb of the Salmo River anticline. Contact metamorphic effects fall off southward, disappearing near the axial trace of the anticline, approximately 1000 ft north of the outcrop of the Reeves MacDonald ore zone. This fall-off in contact effects may in part be due to topography since there is a 1200 ft increase in elevation southward from the Salmo River to the axial trace. The nearest exposed intrusive body is located 8000 ft north of the Salmo River anticline. Conceivably, closer subsurface intrusions, other than dykes, may exist but none has been encountered in underground workings. Granitic xenoliths have, however, been found in lamprophyre dykes.

H. B. mine area. Figure 16 shows the distribution of contact metamorphic minerals—biotite in pelitic rocks and
Figure 15. Contact metamorphic effects in Reeves MacDonald mine area. Shown are occurrences of (i) post-tectonic biotite in pelitic and semi-pelitic rocks, and (ii) actinolite skarn.
tremolite, actinolite and diopside in carbonate and calc-silicate rocks—which indicate the existence of a contact aureole in H. B. mine area. The occurrences, especially of diopside, are concentrated on the lower north and south slopes of Sheep Creek, suggesting an increase in intensity of the metamorphic aureole with depth.

As a result of the absence of cordierite from pelitic rocks and the sporadic distribution of diopside a distinct inner aureole could not be defined. The localized occurrences of diopside skarn and diopside-bearing hornfels are attributed to the local influx of magmatic water causing reductions in CO₂ pressures.

The aureole is developed in the core and on the upper limb of the complex, major anticline which dominates the structure at H. B. mine. The ore zones lie at least within the outer contact aureole as tremolite is commonly developed in the dolomite-sulphide assemblages of the Garnet Zone and talc is reported (Warning, 1960) as abundant in the H. B. ore zones.

Possible intrusive sources for thermal metamorphism are spatially far removed; the Dodger and Annie Rooney Creek granitic stocks lie 1.25 miles to the south and the Nevada Mountain granitic intrusion lies approximately the same distance to the east (Fig. 2).

Jersey mine area. Figure 17 illustrates outer and inner aureoles distinguished in Jersey mine area which in this case are developed "peripheral" to exposed intrusions. The pattern, however, is much complicated by lithological variation and by
Figure 16. Contact metamorphic effects in H. B. mine area. Shown are occurrences of (i) post-tectonic biotite in pelitic and semi-pelitic rocks, (ii) actinolite and tremolite in carbonate rocks, and (iii) diopside in calc-silicate hornfels and in skarn.
Figure 17. Contact metamorphic effects in Jersey mine area. Shown are (i) a diffuse outer aureole defined by occurrences of post-tectonic biotite in pelitic and semi-pelitic rocks, and of actinolite and tremolite in carbonate rocks; (ii) an inner contact aureole defined by occurrences of cordierite in pelitic and semi-pelitic rocks, and of diopside, forsterite and grossularite in carbonate rocks.
disruption due to later faulting. Nor are the Emerald and Dodger stocks simple, steep-sided plugs. They have considerable subsurface extensions (moderately well known from mining activities) so the subsurface configuration of the contact metamorphic aureoles departs considerably from that observed at the surface.

The lead-zinc sulphide ore zones lie mainly within the inner contact aureole and are themselves locally cut by granitic off-shoots.

Assemblages of the contact aureoles

Equilibrium mineral assemblages established for pelitic and carbonate lithologies in the contact aureoles of the three areas are summarized below:

| Outer Aureole          | Pelitic rocks: | quartz-muscovite-chlorite-biotite-albite |
|                       | Carbonate rocks: | calcite-dolomite-talc calcite-dolomite-tremolite calcite-actinolite-plagioclase-quartz |
| Inner Aureole          | Pelitic rocks: | quartz-muscovite-biotite-cordierite(-andalusite) |
|                       | Carbonate rocks: | dolomite-calcite-forsterite calcite-dolomite-diopside (grossularite-diopside-|
|                       |                | ( wollastone-calcite |
|                       |                | ( grossularite-diopside-plagioclase |
|                       | Calc-silicate and skarn: | ( grossularite-idocrase-diopside- |
|                       |                | ( scapolite |
|                       |                | ( grossularite-zoisite-quartz |
|                       |                | (diopside-quartz-calcite |

Absence of chlorite (except for retrogressive material) and presence of cordierite with or without andalusite in pelitic rocks, plus the various assemblages from calc-silicate and
carbonate rocks define the metamorphic facies of the inner contact aureole as hornblende-hornfels facies (Turner, 1968, p. 193).

Mineral assemblages from the outer zone of the contact aureole, in particular the association of chlorite and biotite in pelitic rocks and the occurrence of talc or more commonly tremolite with dolomite and calcite in carbonate rocks, define the metamorphic facies there as corresponding to albite-epidote-hornfels (Turner, 1968, p. 190).

P-T conditions of contact metamorphism

The prevailing lithostatic pressure at the time of intrusion (Lower Cretaceous) has been estimated at 2 kb ± 500 bars by Greenwood (1967) from an area 1.5 miles east-southeast of H. B. mine. This estimate was based on stratigraphic work of Little (1950). In the light of stratigraphic revisions by Fyles and Hewlett (1959) and Little (1960), a slightly more conservative estimate of 1.5 kb, corresponding to 20,000 ft of overlying rock, is proposed.

Assuming fluid pressure was equal to total pressure, estimates of the range of temperatures existing during contact metamorphism can be arrived at by considering the experimentally determined stability relations of certain minerals from the assemblages characterizing the contact aureoles established in both pelitic and carbonate rocks.

The lowermost temperature limits are difficult to determine as the reactions which produce biotite and tremolite are
poorly defined experimentally. These minerals probably formed according to the reactions (Winkler, 1967, p. 25, 99):

\[ \text{muscovite} + \text{prochlorite} = \text{biotite} + \text{Al-chlorite} + \text{quartz} + \text{H}_2\text{O} \]
\[ \text{dolomite} + \text{quartz} + \text{H}_2\text{O} = \text{tremolite} + \text{calcite} + \text{CO}_2 \]

Equilibrium data, calculated by Skippen (1971), indicate that tremolite would form either directly according to the above reaction, or indirectly from talc + calcite at temperatures in the range 415-500°C for partial pressures of CO\(_2\) greater than 0.2 at a total fluid pressure of 2 kb. This is in fair agreement with estimates for the lower temperature limit of albite-epidote hornfels facies of metamorphism at the outer edge of contact aureole which is generally taken to be at approximately 400°C at 1.5 kb (Winkler, 1967, p. 70; Turner, 1968, p. 256).

Talc reported in the H. B. ore zones presumably formed according to the reaction (Winkler, 1967, p. 23):

\[ \text{dolomite} + \text{quartz} + \text{H}_2\text{O} = \text{talc} + \text{calcite} + \text{CO}_2 \]

Its localized and concentrated occurrence is attributed to introduction (via fractures?) of water and silica to dolomite marble; according to Winkler (1967, p. 25) high H\(_2\)O:CO\(_2\) ratios would favour the formation of talc over tremolite. This is confirmed by the experimental data of Skippen (1971) which gives equilibrium temperatures for the above reaction of 400-460°C for partial pressures of CO\(_2\) between 0.2-0.5 at a total fluid pressure of 2 kb. At higher partial pressures of CO\(_2\) tremolite forms in place of talc. Elsewhere, even at H. B. mine, tremolite is the more common hydrous silicate presumably because X\(_{\text{CO}_2}\) values were higher, although Gordon and Greenwood (1970) have suggested
that phlogopite (a common minor constituent of the dolomite marble) forms in place of talc where potassium and aluminum are available, as in muscovite, thus allowing the direct formation of tremolite over a wider range of CO$_2$ partial pressures.

The temperatures developed within the inner contact aureole in Jersey mine area, which is marked by the presence of cordierite in pelitic rocks and of forsterite and diopside in calcareous rocks, can only be approximately calibrated since the reactions have either been only partly investigated experimentally, or they are divariant.

Cordierite, which apparently develops simultaneously with andalusite, possibly forms according to the reaction (Winkler, 1967, p. 71):

$$\text{chlorite} + \text{muscovite} + \text{quartz} = \text{cordierite} + \text{biotite} + \text{andalusite} + H_2O$$

In quartz-, muscovite-bearing pelitic rocks such as these, it probably becomes stable at temperatures near 500°C for pressures between 1 and 2 kb (Hirschberg and Winkler, 1968; Seifert and Schreyer, 1970).
The experimentally determined, divariant reactions for the appearance of the relevant calc-silicate minerals are:

(a) tremolite + calcite + quartz = diopside + CO₂ + H₂O

for which equilibrium temperatures have been determined in the range

\[ T = 510-540°C \text{ for } X_{CO₂} > 0.2 \text{ at } P_f = 1 \text{ kb} \] (Metz and Winkler, 1964)

and \[ T = 475-540°C \text{ for } X_{CO₂} > 0.2 \text{ at } P_f = 2 \text{ kb} \] (Skippen, 1971)

(b) tremolite + dolomite = forsterite + calcite + CO₂ + H₂O

for which equilibrium temperatures have been determined in the range

\[ T = 510-560°C \text{ for } X_{CO₂} > 0.2 \text{ at } P_f = 1 \text{ kb} \] (Metz, 1967)

and \[ T = 535-580°C \text{ for } X_{CO₂} > 0.2 \text{ at } P_f = 2 \text{ kb} \] (Skippen, 1971)

The different determinations are in reasonable agreement for reaction (b) but not for reaction (a) which is generally taken as marking the beginning of the hornblende hornfels facies of contact metamorphism. The data of Skippen (1971) suggest lower temperatures for this reaction (a) which are more closely in agreement with the estimate of Turner (1968, p. 258) for the beginning of hornblende hornfels facies.

Grossularite, occurring locally within the inner contact aureole, provides some additional information as to pressure-temperature conditions. Its stability field in H₂O-CO₂ mixtures is shown in Figure 18 (Gordon and Greenwood, 1971) which indicates that its formation is favoured by water-rich fluids. Within layered calc-silicate hornfels of the Trueman member,
grossularite probably formed according to the reactions:

\[
\text{zoisite} + \text{calcite} + \text{quartz} = \text{grossularite} = \text{CO}_2 + \text{H}_2\text{O}
\]

\[
\text{calcite} + \text{anorthite} + \text{quartz} = \text{grossularite} + \text{CO}_2
\]

for which equilibrium temperatures are in the range

\[
T = \text{c.}500-590^\circ\text{C for } X_{CO_2} < 0.15 \text{ at } P_f = 2\ \text{kb} \quad \text{(Gordon and Greenwood, 1971)}
\]

However its occurrence in massive skarn suggests a metasomatic origin (discussed below).

Local occurrence of fibrolite with andalusite in pelitic rocks close to intrusive contacts permits an estimate to be made of the highest temperatures attained during contact metamorphism. Holdaway (1971) has shown that fibrolite, due to its high surface energy, is more likely to have formed from muscovite than from andalusite. If so, then the experimentally defined breakdown of muscovite (Althaus et al., 1970) to aluminum silicate according to the reaction:

\[
\text{muscovite} + \text{quartz} = \text{Al}_2\text{SiO}_5 + \text{orthoclase}
\]

which occurs at

\[
T = 600^\circ\text{C for } P_{H_2O} = 1.5\ \text{kb} \quad \text{(Fig. 20)}
\]

can be used to calibrate the upper temperature limit.

A country rock temperature immediately adjacent to the granitic intrusions can also be arrived at, using Jaeger's (1957) ideal model of a dyke-like intrusive mass losing heat mainly by conduction. Assuming an intrusive temperature of 700°C, a load pressure of 1.5 kb and a geothermal gradient of 30°C/km gives an approximate contact temperature of 600°C. It should be emphasized that because of the assumptions involved this
represents a very approximate estimate. Thus it will be shown below that there was probably considerable flow of \( \text{H}_2\text{O} \) accompanying intrusion so that heat transfer was also, in part, by convection.

As a check on these estimates, the temperature of crystallization of magnesian calcite, from calcite-dolomite assemblages, was determined by X-ray diffraction, using the method of Graf and Goldsmith (1955, 1958). Samples from the Reeves member at Jersey mine, collected close to or within the inner contact aureole, contained approximately 4 mole\% \( \text{MgCO}_3 \), corresponding to a temperature of formation of 490°C ± 25 (see Appendix A).

**Fluid phase during contact metamorphism**

Local abundance of grossularite as skarn, mainly within calc-silicate hornfels but also as restricted occurrences within carbonate marble, in the inner contact aureole at Jersey mine suggests that metasomatism played an important role in its formation. Its occurrence and distribution are apparently controlled by lithological boundaries and fractures. The grossularite tends to be concentrated in zones parallel with composition layering near the top of the Trueman member along its gradational contact with the Reeves marble. Locally within the marble it may form fringes and masses adjacent to fractures (Fig. 19). Such locations are assumed to have been subject to high \( \text{H}_2\text{O} \) partial pressures which would favour the formation of grossularite; aqueous fluids expelled from the crystallizing granite intrusions moving through the layered calc-silicate Trueman
1. Zo + Ca + Q = Gr + CO₂ + H₂O
2. Ca + An + Q = Gr + CO₂
3. Ca + An + Wo = Gr + CO₂
4. Ca + Q = Wo + CO₂
5. Zo + CO₂ = Ca + An + H₂O

Figure 18. Stability fields of grossularite and wollastonite in H₂O-CO₂ mixtures at 2 kb (from Gordon and Greenwood, 1971).

Figure 19. Mode of occurrence of grossularite skarn within calcite marble of the Reeves member.
member and only affecting the overlying Reeves marble in the vicinity of fractures. The latter with its presumably higher CO$_2$ partial pressures developed diopside and forsterite as the main calc-silicate minerals. Idocrase and scapolite are commonly associated with grossularite and together they suggest that Si, Al and Cl (at least) were also introduced in the aqueous fluids to contribute to skarn formation.

SUMMARY

Regional metamorphism in the three mine areas was restricted to lower greenschist facies and was apparently synchronous with Phase 1 folding. Temperatures accompanying Phase 2 and Phase 3 folding were, in contrast, submetamorphic.

Evidence for post-tectonic overprinting by contact metamorphism has been found in all three mine areas: in Reeves MacDonald and H. B. mine areas, diffuse outer aureoles (remote from any intrusive source) occur in which albite-epidote hornfels facies was not generally exceeded, whereas in Jersey mine area, hornblende hornfels facies was attained within an inner contact aureole developed adjacent to small granite stocks.

The various temperatures estimated for contact metamorphism can be approximated (making allowances for some of the small pressure differences involved) and summarized as follows:

Temperature range of outer aureole: 425 - 475°C ± 25°
Temperature range of inner aureole: 475 - c. 600°C

It is therefore possible to calibrate the temperatures to which the lead-zinc sulphide deposits have been elevated, as illustrated schematically in Figure 20.
Figure 20. Schematic representation of the range of contact metamorphic temperatures affecting the three sulphide deposits. Curve (a) indicates approximately the appearance of biotite (Turner, 1968), and curve (b) the appearance of cordierite (Hirschberg and Winkler, 1968). Curve (c) is the equilibrium curve for the breakdown of muscovite, in the presence of quartz, to aluminum silicate and orthoclase (Althaus et al., 1970), and (d) is the melting curve for granite (Luth et al., 1964).
SECTION V

THE SULPHIDE DEPOSITS

INTRODUCTION

Although it can be readily demonstrated that the Salmo sulphide deposits pre-date granitic intrusion in the area, there is controversy as to whether or not they have been involved in regional metamorphism. Consequently several different interpretations of their history have been proposed.

Fyles and Hewlett (1959) considered the sulphides to be replacement bodies structurally controlled by and only in part deformed by Phase 2 folding, i.e. post-metamorphic. In contrast, Muraro (1966) suggested that the sulphides pre-dated both folding and metamorphism and proposed, in particular, that the Duncan Lake deposit (a Salmo type deposit 85 miles farther north) was originally of Mississippi Valley type (Muraro, 1962).

Callahan (1964, 1967) classified the Salmo deposits as Appalachian type (i.e., deformed Mississippi Valley type) which were subsequently modified by granitic intrusion.

A pre-regional metamorphic history for the ores has also been implied by Sangster (1970a, b) who proposed a syngenetic-diagenetic origin for the sulphides "by concentration in deep-water carbonates, deposited in a 'shale-starved' basin environment."

90
Lead isotope abundances from the Kootenay Arc have been determined by Sinclair (1964, 1966), Reynolds and Sinclair (1971) and more recently by LeCouteur (1973). Two isotope populations are recognized, one corresponding to the Salmo strata-bound deposits and the other to the Slocan and Ainsworth vein-type deposits and rock-leads from the Nelson batholith. These are all anomalous (multistage) leads but they lie on separate anomalous lead lines which indicates that their parent common leads differed in composition. It was suggested by Sinclair and by Sinclair and Reynolds that their subsequent histories were similar and that both were emplaced at approximately the same time, 150 m.y. ago (i.e., post-regional metamorphism). However, it is possible to interpret the "emplacement" of the strata-bound deposits as having occurred as far back as the Lower Palaeozoic (LeCouteur, 1973).

NATURE AND FORM OF THE SULPHIDE ORE BODIES

The three deposits appear to be basically very similar, the major differences in form apparently being a function of their structural environment and the lesser differences, such as mineralogical and textural variation, a function of their varied, subsequent thermal histories.

Their basic mineralogy is simple, consisting of sphalerite-pyrite-galena. Pyrrhotite and minor chalcopyrite are rare additions possibly related to thermal metamorphism of the ores by lamprophyre dykes and granitic intrusions. Gangue consists mainly of dolomite, with lesser calcite, minor quartz
(or various calc-silicates within contact metamorphic zones), white mica and graphitized carbonaceous material.

The sulphides are present either as disseminations along the composition layering $(F_{0-1})$ or as more massive concentrations in which sphalerite and/or galena form the matrix to brecciated or disrupted and folded fragments of pyrite and of dolomite and calcite marbles. These two types of sulphide occurrence will be referred to as layer ore and breccia ore.

**Reeves MacDonald mine**

Previous descriptions of the geology of the Reeves MacDonald mine have been published by White (1949), Green (1954), Fyles and Hewlett (1959) and Addie (1970).

The ore body has the form of a steeply plunging, tight syncline in the core of the Reeves syncline which (as has been discussed in Section III) is interpreted as a Phase 1 fold. Oblique dip-slip faults cut the ore body creating separate ore zones (Fig. 21) which display a progressive change in orientation down plunge:

- Reeves ore zone Az. 215°/50°
- E. MacDonald ore zone Az. 235°/48°
- Annex ore zone Az. 252°/37°

The main hinge of the ore body is apparently much thickened and together with the hinges of parasitic minor folds is typically serrate and could be interpreted either as a highly attenuated isocline or as a lithological contact $(F_0)$ highly transposed along the $F_1$ foliation. The overturned, northern limb of the body has the greater strike length of the two limbs.
Figure 21. Diagrammatic east-west longitudinal section of Reeves MacDonald mine showing the ore zones: 1. No. 4; 2. O'Donnel; 3. B. L.; 4. Reeves; 5. East MacDonald; 6. West MacDonald; 7. Point; and 8. Annex (after Addie, 1970).

Figure 22. Level plans of the Reeves ore zone illustrating the change in configuration down plunge. The two levels are shown in their correct relative positions (from mine records).
and this becomes emphasized down plunge as the normal limb and eventually the hinge zone itself disappear (Fig. 22).

Sulphide mineralization is concentrated near the upper contact of the dolomite marble host with typical Reeves calcite marble. The lower limit is marked by a well-defined, footwall pyrite layer whereas the upper limit, which is much less distinct, approximates the interface between dolomite and calcite marbles.

Sulphides occur either as disseminations parallel with composition layering creating a "gneissic" layering or, in higher concentrations, as the matrix to variable breccias, containing rounded to markedly platy fragments of dolomite and calcite marble.

In the "gneissic" layered ore, contacts between sulphides and marble are not sharply defined (Fig. 25a, b) due to the disseminated nature of the sulphides. However massive pyrite layers occur in the footwall of the ore body and these are sharply inter-layered with graphitic layers and dolomite marble. These pyrite-graphite layers may also be tightly folded and more or less brecciated within intensely foliated ($F_1$) dolomite marble (Fig. 27a). It is suggested that this well-defined layering represents original bedding and that the folding is of Phase 1 age. Other mesoscopic folds are outlined by layered, disseminated ore but these have not been clearly identified as to age. Further complications result from the presence of numerous detached folds within breccia ore, discussed below.

Breccia ore appears to be irregularly distributed and may pass gradationally (vertically or laterally) either into
layered ore or into foliated or massive dolomite marble. The sulphide matrix may be predominantly either sphalerite or pyrite but not usually galena which is present in relatively minor amounts. The breccia fragments, mostly dolomite marble but also calcite marble and quartz, may be (i) apparently randomly oriented, or (ii) aligned parallel with the F1 foliation, or (iii) particularly the platy fragments, they may be folded and locally boudinaged (Fig. 26a, b, c, 27c). These folds display no systematic geometry and appear to result from the deformation of detached, relatively competent, platy fragments within a less competent, sulphide-rich matrix. These structures have been described by Addie (1970) as examples of thixotropic deformation but it seems unlikely that these would survive unmodified by the subsequent intense deformation which is reflected in the dolomite marble host (Fig. 25b). The variable nature of the breccia is puzzling but appears to be related in part to whether the host marbles were layered or massive. They very closely resemble the "milled" or durchbewegt ores commonly developed in sulphide-schist assemblages during regional dynamothermal metamorphism (Vokes, 1968, 1969). In these, progressive deformation outlasts metamorphism so that fragments of schistose host rocks are tectonically incorporated into still ductile sulphide layers.

Gash and irregular veining by coarse-grained "pegmatitic" sulphides and gangue occur in both the layered and the breccia ores (Fig. 25c). These are strongly developed downplunge in the major structure, particularly in the deepest ore zone--the
Annex. They are tentatively interpreted as representing local chemical remobilization, in the sense envisaged by Mookherji (1970), i.e., locally derived and transported some short (but unknown) distance in solution. They are apparently not deformed to any significant degree and, by analogy with the Jersey mine area (Section III, p. 61), their association with gash-veins suggests a relationship with granitic intrusion in the area.

**Jersey mine**

The geology of Jersey mine has been described by Whishaw (1953), Rennie and Smith (1957), Fyles and Hewlett (1959) and Bradley (1970).

The Jersey ore bodies consist of a series of elongate, tabular, folded masses plunging 10-15° southward and converging in the same direction. They comprise two, essentially separate, main zones subdivided (for mining purposes) into a series of parallel subzones, A to J (Fig. 23). Each of these main zones occurs within a dolomite marble host. The more western of these, containing the A-B zones, is located at the base of the Reeves member in contact with the Trueman member whereas the eastern host, containing the C-J zones, is located at a higher level within the Reeves member but converges downplunge with the lower host.

Two distinct styles of folding, developed on all scales, modify the form of the ore bodies: (i) recumbent, tight asymmetric folds (equated with Phase 1) and (ii) upright, more open symmetric or slightly asymmetric folds (equated with Phase 2). Both were classified by Fyles and Hewlett (1959) as
Figure 23. Plan view and representative cross-sections of Jersey mine workings, showing the main structural features (from mine records).
secondary (Phase 2) structures but, from their geometry, they are separable into distinct generations (Fig. 27d, e, f). Macroscopic examples of the former are represented by the A-zone and F-zone folds ("skarn rolls" in mine terminology) and of the latter by the D-zone anticline (Fig. 23). Both types of structures are commonly complicated by low-angle thrusts with displacements of up to several hundred feet. In addition, the tabular ore bodies interfinger with the dolomite marble hosts so that they are generally complex in detailed form. An interesting feature of the ore bodies is that they tend to be much thicker along the macroscopic Phase 2 synclinal troughs than on the adjacent Phase 2 anticlinal crests (Fig. 23, cross-sections 1 and 2). This led Fyles and Hewlett (1959) and subsequent mine geologists (e.g., Bradley, 1970) to postulate that emplacement of the ore was somehow structurally controlled by Phase 2 folding.

As in Reeves MacDonald mine, the sulphides form both layered and breccia ores. The layered ores are the dominant type and, due to the darker colour of the sphalerite and the generally coarser grain size of the sulphides, the layering tends to be more distinct than in the Reeves MacDonald ores. The sulphide layers may be predominantly sphalerite and/or galena or pyrite, forming disseminations parallel with composition layering ($F_{0-1}$) in white or grey/white dolomite marble (Fig. 25d, e). Massive pyrite/graphite bands, similar to those occurring in the footwall of Reeves MacDonald mine, have been
observed locally but they are intensely brecciated and partially obscured by recrystallization.

With increased sulphide concentration, the disseminated sulphide layers become increasingly massive and in many places sphalerite and/or galena layers enclose augen of coarse-grained dolomite, of pyrite, and of dolomite marble (Fig. 26d, e). Where sulphide concentrations exceed several inches in thickness, characteristic breccia structures are usually developed, as in the galena-rich bands of the A-, F- and G-zones. Of these, the most spectacular occurs in the lowermost A-zone where it is involved in Phase 1 folding. There the galena layer is up to six feet thick and contains abundant quartz and dolomite as rounded fragments (Fig. 26f).

Mesoscopic folds can be observed at numerous locations within the ore zones. Those involving layered ore have the same geometry as the macroscopic structures affecting the ore bodies as a whole and can be separated into Phase 1 and Phase 2 generations (Fig. 27d, e, f). Folds within the breccia ore are generally detached and hence tend to be more ptygmatmic.

Coarse-grained, "pegmatitic" ore is uncommon in Jersey mine but where observed (in the F- and G-zones) it is sphalerite-rich and associated with minor galena and dolomite. Cross-cutting veinlets of pyrite, paralleling late gash-veins, also are known but the genetic relations between these occurrences are unclear.
H. B. mine

The geology of the H. B. mine has been described in part by Green (1954) and more fully by Irvine (1957), Fyles and Hewlett (1959) and Warning (1960).

As production ceased in 1966, the writer was able to examine only surface exposures so that the following description of the ore bodies is largely based on the above previous work.

There are two distinct occurrences of productive ore, the H. B. and Garnet ore bodies (Plate VIII, Fig. 10). The H. B. ore bodies take the form of three elongate, crudely ellipsoidal zones dipping steeply toward the east and plunging 15-20° southward. These are connected by gently dipping tabular ore breccia zones which plunge southward at the same angle (Fig. 24). The system, with a plunge length of greater than 3000 ft, is located in the core of a complex major syncline, the H. B. syncline, which is considered (Section III, p. 52) to be a Phase 1 fold, modified by flattening during Phase 2 deformation.

The Garnet ore body is located southwest of the H. B. ore bodies across the Garnet fault (Fig. 10). It occurs as a single isolated zone, also steeply dipping to the east and plunging at a low angle southwards. It has a plunge length of approximately 2000 ft and is exposed at the surface in glory-hole workings.

The steeply dipping ore zones consist of numerous disseminated sulphide layers and lenses paralleling an intense cleavage foliation (F₁?) developed in the host dolomite marble (Fig. 25f). These have been described as mineralized shear
Figure 24. a. Isometric view of the H. B. ore zones, showing their configuration and spatial relations. b. Typical vertical section of the No. 1 zone, looking north.
zones which were developed parallel to an earlier axial plane cleavage related to major folding (Irvine, 1957; Warning, 1960).

The connecting flat breccia zones, which are apparently truncated by the steeply dipping ores, are interpreted by the same authors as zones of shearing and brecciation formed at a late stage of the main folding and subsequently mineralized. As these breccias also contain fragments of cleaved dolomite marble, Warning (1960) proposed the following sequence of structural events preceding mineralization: (i) folding and development of axial plane cleavage, (ii) low angle shearing and brecciation and (iii) steep shearing. Fyles and Hewlett (1959) related the gently dipping breccia zones to thrust zones which were considered to be secondary structures (Phase 2) and the control for mineralization.

The breccia ores have higher sulphide contents than the steep layered ores and from their descriptions closely resemble the breccia ores of Reeves MacDonald and Jersey mines which are believed to be post-mineralization breccias. If the breccias of H. B. mine also post-date mineralization, then their formation could have accompanied folding and the steep layered ores could have formed by transposition of breccia ore (or its precursor) along $F_1$ axial cleavage either during Phase 1 folding or during Phase 2 closure of Phase 1 folds.

Summary

It is concluded that no unique structural control of the emplacement of mineralization can be demonstrated as the ores are either located in, and apparently deformed by, Phase 1 major
structures (the Reeves and H. B. synclines at Reeves MacDonald and H. B. mines respectively), or are apparently deformed by both Phase 1 and Phase 2 fold structures of smaller dimensions (Jersey mine).

Internally, mesoscopic structures of the ores themselves suggest that the ores have indeed been deformed as:

(i) layered disseminated ores outline mesoscopic folds with the same geometry as the macroscopic structures;

(ii) layered, more massive ores show flaser/augen structures indicating differential response of the various sulphides to deformation;

(iii) breccia ores record intense and prolonged post-mineralization deformation.
Figure 25. Examples of layered ores from the three mines.

a. Typical layered sphalerite-pyrite/dolomite ore from Reeves MacDonald mine. Note transition into breccia ore at upper right. Reeves MacDonald mine, E. MacDonald zone, 610' level.


c. "Pegmatitic" sphalerite-dolomite-quartz in irregular/gash veining. Reeves MacDonald mine, Annex zone, 800' level.

d. Layered dark dolomite marble and sphalerite, with augen and lenses of white calcite. Hammer for scale at lower left. Jersey mine, F-zone.

e. Detail of typical layered dark sphalerite/white dolomite ore from Jersey mine. Jersey mine, G-zone. Specimen CX69-22.

f. Detail of dark sphalerite/white dolomite-calcite ore, transposed along F1 foliation (?). H. B. mine, Garnet zone. Specimen HB70-18.
Figure 25. Examples of layered ores from the three mines.
Figure 26. Examples of breccia ores from Reeves MacDonald and Jersey mines.

a. Contorted and brecciated dolomite marble within and adjacent to dark sphalerite-pyrite layer. Reeves MacDonald mine, E. MacDonald zone, 400' level.

b. Platy fragments of layered dolomite marble in sphalerite-rich matrix of breccia ore. Reeves MacDonald mine, Annex zone, 875' level. Specimen RM70-34.

c. Detail of relations between black graphitic dolomite marble and sphalerite in sphalerite-rich breccia ore. Same specimen as b. above.

d. White calcite and grey dolomite augen and lenses within streaky sphalerite-galena ore. Jersey mine, open pit west of 4050' adit.

e. White calcite and grey dolomite augen and lenses within galena-rich matrix (dark grey). Jersey mine, open pit west of 4050' adit. Specimen CX69-11.

f. Detail of galena and sphalerite-rich breccia ore containing rounded, white dolomite fragments. Jersey mine, D-zone. Specimen CX69-55.
Figure 26. Examples of breccia ores from Reeves MacDonald and Jersey mines.
Figure 27. Examples of fold structures involving sulphides from Reeves MacDonald and Jersey mines.

a. Interbedded and folded massive pyrite (white) and graphitic dolomite (grey). Reeves MacDonald mine, Reeves zone "glory-hole".

b. Disseminated pyrite (grey) outlining fold in calcite marble (oblique section). Reeves MacDonald mine, Reeves zone "glory-hole".

c. Breccia ore containing folded dolomite marble layers in pyrite-sphalerite matrix. Note rootless closure at lower right and lack of systematic fold geometry. Reeves MacDonald mine, Reeves zone "glory-hole".

d. Recumbent Phase 1 fold in layered white dolomite marble/sphalerite-galena-pyrite ore, cut by thrust faults. Jersey mine, E-zone.

e. Recumbent Phase 1 fold in grey/white dolomite marble with minor sphalerite-rich layers. Jersey mine, F-zone.

f. Upright Phase 2 fold in white dolomite marble and interlayered dark dolomite marble/sphalerite. Jersey mine, F-zone.
Figure 27. Examples of fold structures involving sulphides from Reeves MacDonald and Jersey mines.
MINERALOGICAL VARIATION OF THE SULPHIDES

Introduction

A superficial comparison of the ores from the three deposits reveals relatively obvious mineralogical, as well as textural, variations between them. The former comprise: (i) a general increase in the iron content of sphalerite from Reeves MacDonald to H. B. to Jersey mines, (ii) local association of pyrrhotite with pyrite in H. B. and Jersey mines, and (iii) superimposition of scheelite-molybdenite skarn mineralization on the older sphalerite-galena mineralization at Jersey mine.

Other more cryptic variations, which have also been noted in the past, include: (a) a general increase in the cadmium content of sphalerite from north to south implied by mine production figures (Sinclair, 1964); (b) increased cadmium and silver values in the Annex ore zone at Reeves MacDonald mine (G. G. Addie, 1971, pers. comm.); and (c) local change in the antimony and bismuth contents of galena said to be a result of contact metamorphism at Jersey mine (Sinclair, 1964). Many of these changes appear to be connected with contact metamorphism which is generally accepted as post-dating lead-zinc mineralization (Muraro, 1966).

The mineralogical effects of contact metamorphism on strata-bound sulphide ores in general have been reviewed by Vokes (1969) and may be summarized as including:

(a) partial melting of sulphides (Fe-Pb-Zn-S system)
(b) dissociation of primary pyrite to pyrrhotite and/or magnetite
(c) exsolution and concentration of chalcopyrite
(d) increased iron content of sphalerite

In this study, the mineralogical effects were investigated primarily via quantitative determinations of minor element contents of sulphides. Seventy-one determinations were carried out using atomic absorption spectrophotometry (49 sphalerite and 17 pyrite analyses) and electron microprobe (5 sphalerite analyses). Details of sample preparation, analytical methods and precision are given in Appendix C.

Minor element content of sphalerite

Iron, cadmium, manganese, silver and copper values in 49 sphalerite samples from the three deposits were determined by A. A. spectrophotometry (Tables XII and XIII of Appendix C). Lead values determined as measure of sample contamination, varied from nil to 4.7 wt. % but showed no systematic correlation with other minor element values. Mean values of minor element contents of sphalerites from the three deposits are compared in Table VII. In this Table, the Reeves MacDonald mine samples have been subdivided into those from Reeves MacDonald zone vs. those from Annex zone as there are significant differences between them, as will be discussed below. Scatter diagrams of individual sample values are illustrated via a logarithmic plot of cadmium against iron (Fig. 28), and a ternary plot of logarithmically transformed manganese, silver and copper values (Fig. 29). These indicate differences between the four sample groups which were statistically tested using the
Figure 28. Logarithmic plot of Cd vs. Fe contents of sphalerite samples from the three mines.
Figure 29. Ternary plot of Mn, Cu and Ag contents (in ppm logarithmically transformed) of sphalerite samples from the three mines.
Table VII. Comparison of mean values of minor element contents of sphalerite samples from the three mines (G = geometric mean in ppm; S.D. = standard deviation expressed as logarithms).

<table>
<thead>
<tr>
<th>Minor element</th>
<th>Reeves MacDonald mine</th>
<th>H. B. mine</th>
<th>Jersey mine</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Annex zone (15 samples)</td>
<td>Reeves MacDonald zone (10 samples)</td>
<td>H. B. mine (9 samples)</td>
</tr>
<tr>
<td>Fe</td>
<td>14100 0.127 G</td>
<td>9900 0.138 G</td>
<td>27300 0.128 G</td>
</tr>
<tr>
<td>Cd</td>
<td>10400 0.078 G</td>
<td>4000 0.177 G</td>
<td>4100 0.071 G</td>
</tr>
<tr>
<td>Mn</td>
<td>60 0.275 G</td>
<td>75 0.423 G</td>
<td>385 0.239 G</td>
</tr>
<tr>
<td>Ag</td>
<td>71 0.551 G</td>
<td>11 0.313 G</td>
<td>7 0.402 G</td>
</tr>
<tr>
<td>Cu</td>
<td>51 0.177 G</td>
<td>21 0.236 G</td>
<td>17 0.205 G</td>
</tr>
</tbody>
</table>

Annex zone vs. Reeves MacDonald zone

<table>
<thead>
<tr>
<th>Minor element</th>
<th>Degrees of freedom</th>
<th>t-values</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fe</td>
<td>23</td>
<td>-2.837¹</td>
</tr>
<tr>
<td>Cd</td>
<td>11</td>
<td>-6.891¹</td>
</tr>
<tr>
<td>Mn</td>
<td>23</td>
<td>0.698</td>
</tr>
<tr>
<td>Ag</td>
<td>23</td>
<td>-4.196¹</td>
</tr>
<tr>
<td>Cu</td>
<td>23</td>
<td>-4.708¹</td>
</tr>
</tbody>
</table>

Reeves MacDonald zone vs. H. B. mine

<table>
<thead>
<tr>
<th>Minor element</th>
<th>Degrees of freedom</th>
<th>t-values</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fe</td>
<td>17</td>
<td>7.156¹</td>
</tr>
<tr>
<td>Cd</td>
<td>12</td>
<td>0.045</td>
</tr>
<tr>
<td>Mn</td>
<td>17</td>
<td>4.436¹</td>
</tr>
<tr>
<td>Ag</td>
<td>17</td>
<td>-1.047</td>
</tr>
<tr>
<td>Cu</td>
<td>17</td>
<td>-0.870</td>
</tr>
</tbody>
</table>

H. B. mine vs. Jersey mine

<table>
<thead>
<tr>
<th>Minor element</th>
<th>Degrees of freedom</th>
<th>t-values</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fe</td>
<td>22</td>
<td>3.252¹</td>
</tr>
<tr>
<td>Cd</td>
<td>22</td>
<td>6.592¹</td>
</tr>
<tr>
<td>Mn</td>
<td>22</td>
<td>1.256</td>
</tr>
<tr>
<td>Ag</td>
<td>22</td>
<td>0.345</td>
</tr>
<tr>
<td>Cu</td>
<td>22</td>
<td>2.407¹</td>
</tr>
</tbody>
</table>

¹Indicates mean values are significantly different at the 95% confidence level.
standard t-test of the mean values at the 5% level of significance. The t-values were computed using a FORTRAN program, STRIP (Seagraves, 1971) and the results are listed in Table VII. These suggest that there is a general increase in the iron, cadmium and manganese values of sphalerite in the order—Reeves MacDonald (exclusive of the Annex zone), H. B., Jersey. This is however clearly systematic only in the case of iron although moderate correlations do exist between iron and manganese \((r = 0.62)\) and between iron and cadmium values \((r = 0.56)\). Similar correlations between the iron and manganese contents of sphalerites have been noted in other occurrences by Fryklund and Fletcher (1961), Sims and Barton (1961) and Boyle and Jambor (1963) but according to Fryklund and Fletcher no correlation existed between iron and cadmium contents in their study.

The general trend of these minor element values corresponds with the observed increase in intensity of contact metamorphic effects, implying a relationship that is at least partly temperature dependent. Obviously, the other conditions attendant upon recrystallization (pressure, sulphur fugacity, etc.) and the pre-existing minor element concentrations are largely unknown so that their value as geological thermometers must remain at best qualitative. In this respect, the significance of the iron content (in particular) of sphalerite has been discussed by Kullerud (1953), Barton and Kullerud (1958) and Barton and Toulmin (1966) and has been shown, in pyrite + sphalerite assemblages, to be sensitive to not only temperature but also sulphur fugacity.
Differences in the minor element content of sphalerites are most marked between samples from the Reeves MacDonald ore zones and samples from the Annex ore zone at Reeves MacDonald mine. Those from the Annex zone are distinctly richer in iron, cadmium, silver and copper (Table VII and Fig. 28, 29). The reasons for this change down the plunge length of, what was prior to faulting, a continuous ore body are obscure; no increase in metamorphic grade has been recognized, although a contact metamorphic aureole could conceivably exist at this depth (see Section IV, p. 75). The most obvious physical difference down-plunge is an increase in very coarse-grained, sphalerite-rich "pegmatitic" ore veining in the Annex zone. This may be in either gash veins or in irregular veining which grades into typical recrystallized, fine-grained sphalerite of the layered and breccia ores (Fig. 25c).

When the minor element contents of the "pegmatitic" and the fine-grained sphalerites are compared, it is found that the only significant difference is in the iron contents (Table VIII). Thus it is not clear how the zonation in minor elements down-plunge is related, if at all, to the development of the coarse-grained, sphalerite-rich ore.

Many writers (Ramdohr, 1953; Juve, 1967; Lawrence, 1967; McDonald, 1967; Mookherjee, 1970; Vokes, 1971) suggest that coarse-grained, pegmatite- or vein-like sulphide-gangue bodies within strata-bound ores in metamorphic terrain are segregations mobilized during metamorphism. Such mobilization is believed, in general, to involve limited fluid transport (although the actual
Table VIII. Comparison of mean values of minor element contents of different types of sphalerite from the Annex zone of Reeves MacDonald mine (G= geometric mean in ppm; S.D. = standard deviation expressed as logarithms).

<table>
<thead>
<tr>
<th></th>
<th>Coarse-grained pegmatitic sphalerite</th>
<th>Fine-grained sphalerite</th>
<th>t-values</th>
<th>Degrees of freedom</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(7 samples)</td>
<td>(7 samples)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>G</td>
<td>S.D.</td>
<td>G</td>
<td>S.D.</td>
</tr>
<tr>
<td>Fe</td>
<td>16900</td>
<td>0.140</td>
<td>11900</td>
<td>0.068</td>
</tr>
<tr>
<td>Cd</td>
<td>10000</td>
<td>0.091</td>
<td>10700</td>
<td>0.072</td>
</tr>
<tr>
<td>Mn</td>
<td>77</td>
<td>0.358</td>
<td>45</td>
<td>0.120</td>
</tr>
<tr>
<td>Ag</td>
<td>47</td>
<td>0.521</td>
<td>67</td>
<td>0.233</td>
</tr>
<tr>
<td>Cu</td>
<td>43</td>
<td>0.177</td>
<td>61</td>
<td>0.166</td>
</tr>
</tbody>
</table>

\(^1\)Indicates mean values are significantly different at the 95% confidence level
physico-chemical nature of the transporting medium is rarely discussed), or in rarer cases to involve partial melting in the Fe-Pb-S, Fe-Pb-Zn-S, and Cu-Fe-Pb-S systems (Brett and Kullerud, 1967; Lawrence, 1967; Vokes, 1971). It is also suggested that the mobilization takes place selectively as relative enrichments of segregations, for example, in Pb, As, Sb, and Ag concentrations, have been described (Vokes, 1963, 1971).

In this study, the "pegmatitic" ore commonly occupies gash veins (which in the Jersey mine area are considered to have formed at a late stage of granite intrusion), appears to be essentially undeformed, and hence post-dates all three phases of folding and regional metamorphism. Therefore, if its mobilization is indeed of metamorphic origin, it can only be of contact metamorphic origin, perhaps exotically related to granite intrusion at depth in the area and resulting from upward migration of magmatic water through the ore body.

Minor element content of pyrite

The cobalt, nickel, copper and manganese contents of 17 pyrite samples from the three deposits were determined by A. A. spectrophotometry (Table XIX of Appendix C). Lead and zinc values, determined as a measure of sample contamination, are erratic and often relatively high (0.002 - 2.2%) but show no systematic correlation with the other minor element values. Means and standard deviations for the values are given in Table IX. The only significant difference (at the 95% confidence level) between the mean values of the three deposits is in the higher mean nickel content of the Reeves MacDonald pyrites.
Table IX. Comparison of mean minor element contents of pyrite samples from the three mines (G = geometric mean in ppm; S.D. = standard deviation in logarithms).

<table>
<thead>
<tr>
<th>Minor element</th>
<th>R. MacDonald mine (7 samples)</th>
<th>H. B. mine (5 samples)</th>
<th>Jersey mine (5 samples)</th>
<th>Synoptic (17 samples)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Co</td>
<td>11</td>
<td>0.148</td>
<td>4</td>
<td>0.355</td>
</tr>
<tr>
<td>Ni</td>
<td>149</td>
<td>0.163</td>
<td>16</td>
<td>0.351</td>
</tr>
<tr>
<td>Cu</td>
<td>10</td>
<td>0.131</td>
<td>8</td>
<td>0.507</td>
</tr>
<tr>
<td>Mn</td>
<td>2</td>
<td>0.147</td>
<td>21</td>
<td>0.506</td>
</tr>
<tr>
<td>Co:Ni</td>
<td>.07</td>
<td>.25</td>
<td>.13</td>
<td>.12</td>
</tr>
</tbody>
</table>

R. MacDonald mine vs. Jersey mine

<table>
<thead>
<tr>
<th>t-values</th>
<th>Degrees of freedom</th>
</tr>
</thead>
<tbody>
<tr>
<td>Co</td>
<td>-1.545</td>
</tr>
<tr>
<td>Ni</td>
<td>-3.634</td>
</tr>
<tr>
<td>Cu</td>
<td>-0.832</td>
</tr>
<tr>
<td>Mn</td>
<td>3.818¹</td>
</tr>
</tbody>
</table>

¹Indicates mean values are significantly different at the 95% confidence level
There is also a high correlation ($r = 0.81$) between cobalt and nickel values which are shown, plotted on a logarithmic scale, in Figure 30.

The genetic significance of Co:Ni ratios, in relation to strata-bound, disseminated copper deposits, was discussed by Davidson (1962) who noted their distinctively high Co:Ni ratios which more closely resembled those of acidic igneous rocks and affiliated skarn sulphides than those of either ancient or modern sediments. Cobalt and nickel contents of sedimentary rocks tend to be low and Co:Ni ratios are generally less than unity (Table X).

Loftus-Hills and Solomon (1967), from a study of pyritic ores in Lower Palaeozoic rocks of the Tasman geosyncline, found that sedimentary pyrite in shales is characterized by low Co:Ni ratios ($<0.35$) and by moderately high Ni contents ($>200$ ppm), although pyrite from granitic rocks (and associated ore) and from stratiform lead-zinc ores with volcanic associations had similar low ratios but much lower Ni contents.

A recent statistical study of analyses of minor elements in pyrite from many sources by Price (1972) shows that the mean (geometric) cobalt and nickel contents of sedimentary pyrite are significantly lower than those of pyrite from either massive sulphide-volcanic exhalative deposits or hydrothermal vein-type deposits, and that the mean Co:Ni ratio is less than one (Table X).

Very little information is available on the cobalt and nickel contents of pyrite from Mississippi Valley type deposits,
Figure 30. Logarithmic plot of Co vs. Ni contents of pyrite samples from the three mines.
Table X. Co and Ni contents (in ppm) of A. certain sedimentary rock types, and of B. pyrite from different types of ore deposit.

<table>
<thead>
<tr>
<th></th>
<th>Co</th>
<th>Ni</th>
<th>Co:Ni</th>
</tr>
</thead>
<tbody>
<tr>
<td>A.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Average shale</td>
<td>19</td>
<td>68</td>
<td>0.57</td>
</tr>
<tr>
<td>(Turekian and</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Wedepohl, 1961)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Average black shale</td>
<td>10</td>
<td>50</td>
<td>0.2</td>
</tr>
<tr>
<td>(Vine and Tourtelot,</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1970)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dolostones</td>
<td>-</td>
<td>126</td>
<td>?</td>
</tr>
<tr>
<td>primary</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Carbonates</td>
<td>4</td>
<td>11</td>
<td>0.36</td>
</tr>
<tr>
<td>(Graf, 1962)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>secondary</td>
<td></td>
<td>41</td>
<td>?</td>
</tr>
<tr>
<td>B.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sedimentary ores</td>
<td>41</td>
<td>65</td>
<td>0.63</td>
</tr>
<tr>
<td>(Price, 1972)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Massive sulphide</td>
<td>486</td>
<td>56</td>
<td>8.70</td>
</tr>
<tr>
<td>ores</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
few analyses having been done and these mostly semi-quantitative spectrographic analyses. A few analyses (20) from the stratiform zinc-lead ores of the Upper Mississippi Valley district indicate that pyrite and marcasite (common in these deposits) generally have low cobalt and nickel contents and Co:Ni ratios close to unity (Bradbury, 1961), although pyrite of different generations may have cobalt and nickel contents of up to 1% (Hall and Heyl, 1968). The available data, however, are insufficient to accurately characterize pyrite from these deposits.

Although cobalt and nickel contents of the Salmo pyrites most closely resemble those of syngenetic pyrite and of sedimentary rocks, they have almost certainly been modified by metamorphism (possibly both contact and regional), especially where a new phase, such as pyrrhotite, has been generated.

Cambel and Jarkovsky (1969), in a study of the pyritic deposits of the Western Carpathians, showed in coexisting pyrite and pyrrhotite that nickel was concentrated in the pyrrhotite lattice and cobalt in the pyrite lattice, under medium- to high-grade metamorphic conditions. Sangster (1967) has also proposed redistribution of cobalt and nickel between pyrrhotite and sphalerite as a consequence of the metamorphic conversion of pyrite to pyrrhotite in the New Calumet deposit, Quebec.

Although the Salmo pyrite data are too few, there is a suggestion that contact metamorphism and the appearance of pyrrhotite may be the cause of the significantly lower nickel values at Jersey and H. B. mines.
Even more uncertain are the effects, if any, of prior regional metamorphism, since pyrrhotite was not formed under these conditions. However, particularly at Reeves MacDonald mine (least affected by contact metamorphism), the bedded nature of some of the pyrite, its common association with graphitic dolomite and the generally low cobalt and nickel contents, with Co:Ni ratios <1, suggest that the pyrite is sedimentary in origin and that cobalt-nickel values are perhaps little fractionated under conditions of lower greenschist metamorphism.

**Minor element content of galena**

No determinations were made, in this study, of minor element contents of galena as suitable galena-rich material was available only from Jersey mine. It has already been shown (Sinclair, 1964) that galena (six analyses) from Jersey mine has high Sb:Bi ratios which are probably indicative of a low temperature origin (Malakhov, 1969). One additional sample, believed to be highly contact metamorphosed, had a very low Sb:Bi ratio which is more indicative of a high temperature origin as in skarn-type mineralization (Malakhov, 1969).

**Occurrence of pyrrhotite**

Pyrrhotite coexisting with sphalerite + pyrite occurs locally in all three deposits. At Reeves MacDonald mine, the occurrence of the three phase assemblage is very restricted and has been observed only in rare sulphide xenoliths from lamprophyre dykes cutting the ore body. In the H. B. mine, pyrrhotite occurs sparingly and then mostly as local concentrations in the
No. 1 ore zone (Warning, 1960). It is not known how its distribution relates, if at all, to lamprophyre dykes cross-cutting the ores.

Pyrrhotite occurrence is most abundant and widespread at Jersey mine where it is found near the base of the A-zone and sporadically throughout the eastern zones (F, G and H). Its distribution is apparently related to proximity of the Emerald and Dodger stocks (Fig. 17), but its mode of occurrence is apparently complex in that it not only is associated with sphalerite-pyrite-galena mineralization but possibly also occurs as a primary (?) constituent of skarn-type scheelite-molybdenite mineralization at the margins of the granite stocks.

The assemblage sphalerite + pyrrhotite + pyrite is of particular interest as the sulphur fugacity is fixed at given temperatures by the coexisting iron sulphides (Barton and Toulmin, 1966). It has been shown (Boorman, 1967; Scott and Barnes, 1971; Boorman, Sutherland and Chernyshev, 1971) that the sphalerite-pyrrhotite-pyrite solvus is vertical between 550° and 304°C so that the FeS content of sphalerite in equilibrium with hexagonal pyrrhotite + pyrite is fixed at approximately 21 mole % (at 1 bar) and is not therefore temperature sensitive. It is however pressure sensitive, as has been calculated by Barton and Toulmin (1966) and Scott and Barnes (1971), and appears to provide a useful geobarometer, providing the attainment of equilibrium can be established.

Five samples of sphalerite coexisting with pyrrhotite + pyrite were analyzed by electron microprobe for zinc and iron (see
Appendix D). Four of these samples were from Jersey mine and one from Reeves MacDonald mine. Apart from comparing FeS contents of these sphalerites with those from sphalerite + pyrite assemblages, a further purpose was to attempt to use the former assemblages as geobarometers to check estimates of total pressure during contact metamorphism. However, the lack of attainment of equilibrium in the ore during contact metamorphism (discussed below) has precluded their use as geobarometers.

The Reeves MacDonald sample was from a sulphide xenolith in a lamprophyre dyke of Eocene age. Sphalerite from it contained 17.82 ± 1.2 mole % FeS and was not, on the basis of this microprobe data, very homogeneous. Further, the presence of abundant blebs of pyrrhotite (possibly exsolved) in the sphalerite suggested that the FeS content might not be "original."

FeS contents of the Jersey sphalerites coexisting with pyrite + pyrrhotite were in the range 10.83-12.66 mole % (mean: 11.94 ± 0.64 mole % FeS). In comparison, the FeS contents of sphalerite coexisting with pyrite alone (analysed by A. A. spectrophotometry) were in the range 2.42-11.31 mole % (mean: 7.86 ± 1.92 mole % FeS). These results are inset as histograms in Figure 31 to illustrate their range in relation to the composition of sphalerite in equilibrium with pyrrhotite + pyrite between 550° and 304°C.

The localized distribution of pyrrhotite, together with the wide variation in sphalerite compositions (Fig. 31), indicate that equilibrium was not attained throughout the Jersey ores during contact metamorphism. Further, the coexisting pyrrhotite
Figure 31. FeS contents (mole %) of sphalerite samples from Jersey mine. Contents and locations of individual samples are shown on plan of mine workings. Inset shows projection (on the FeS-ZnS binary) of part of the system Fe-Zn-S at 1 bar (after Scott and Barnes, 1971). In relation to this, the histogram shows range in FeS contents of sphalerites from Jersey mine (white from sph + py and black from sph + py + po assemblages).
is mixed monoclinic and hexagonal, as determined directly by X-ray diffraction (Arnold, 1966), so that its association with pyrite suggests disequilibrium within the iron sulphide assemblage. The hexagonal pyrrhotite has a composition of approximately 47.5 atomic % Fe (determined from the curve for $d_{(102)}$ versus composition of Arnold, 1962) which is usual in ores containing both hexagonal and monoclinic pyrrhotite (Arnold, 1967). This has been interpreted as resulting from inversion below 304 C of hexagonal high-temperature pyrrhotite (composition: 46.5-47.5 atomic % Fe) to hexagonal low-temperature pyrrhotite and metastable monoclinic pyrrhotite (Arnold, 1969). This could be the result, in this case, of retrogressive contact metamorphism as indicated by the development of chlorite after biotite porphyroblasts in schists elsewhere in the contact aureole.

Sphalerite coexisting with pyrrhotite + pyrite locally contains minute blebs and lamellae (possibly exsolved) of chalcopyrite and/or pyrrhotite. Rare localized, minor concentrations of chalcopyrite have also been noted either in contact with both sphalerite and pyrrhotite, or with pyrrhotite alone. In general, even where no "exsolution" is observed, sphalerite is slightly less homogeneous (on the basis of semi-quantitative microprobe traverses for iron) than in sphalerite-pyrite assemblages, particularly when compared with specimens from H. B. and Reeves MacDonald mines (Fig. 32), which further suggests that equilibrium was not fully attained on the micro-scale.
Figure 32. Semi-quantitative microprobe determinations of iron variation across sphalerite grains in samples from (a, b) Jersey mine; (c) H. B. mine; (d) and Reeves MacDonald mine. (a) shows traverse of sphalerite adjacent to pyrrhotite.
Occurrence of a second generation of pyrite

A second generation of pyrite within sphalerite + pyrrhotite + pyrite assemblages in the Jersey ore is indicated by the local development of large idioblastic pyrite crystals (Fig. 34f) and of inclusion-bearing pyrite overgrowths on primary pyrite in contact with pyrrhotite. Late, cross-cutting pyrite veinlets elsewhere in the ore may also be of this generation although it has been suggested that these and other cross-cutting veins of arsenopyrite, pyrrhotite and chalcopyrite (not observed in this study) result from a separate episode of mineralization accompanying intrusion of the granite stocks and perhaps related to scheelite-molybdenite skarn mineralization (Sinclair, 1964).

The origin of this second pyrite generation is not known but may result from late-stage sulphurization of pyrrhotite. Such an increase in sulphur vapour pressure might also explain the occurrence of "exsolved" pyrrhotite in sphalerite from these assemblages (Barton and Toulmin, 1966).

Similar occurrences of secondary pyrite from dyke-contact aureoles in massive sulphides have been described by Johnson (1966), Graham (1968), and Mookherjee and Suffel (1968). These are ascribed to sulphurization of pyrrhotite and of ferromagnesian silicates near the intrusive contacts either by the sulphur released during the breakdown of the pyrrhotite (in the presence of $O_2$) to magnetite, or by the migration inwards towards the intrusion of sulphur released by the transformation of pyrite to pyrrhotite. Breakdown of pyrrhotite has not been observed,
however, in the Jersey ore nor does such a breakdown appear to have been likely if carbonaceous/graphitic material was associated with the ore at the time of contact metamorphism in similar abundance to that presently observed in Reeves MacDonald ore.

Summary

A number of mineralogical effects resulting from contact metamorphism of the sulphide assemblages have been determined. Some of these are tentative because of insufficient data. The effects include changes in the minor element contents of the sulphides, the appearance of new mineral phases, and perhaps also local mobilization of both sulphides and gangue. They may be summarized as follows:

(i) Changes in minor element contents:
   a. progressive increases in the Fe, Mn and Cd contents of sphalerite
   b. a possible decrease in the Ni content of pyrtie
   c. a possible decrease in Sb:Bi ratios in galena at the highest metamorphic grade (Sinclair, 1964)

(ii) Appearance of new minerals within the inner aureole of contact metamorphism:
   a. transformation of pyrite to pyrrhotite giving the disequilibrium assemblage sphalerite + pyrite + pyrrhotite
   b. possible exsolution of pyrrhotite and chalcopyrite from sphalerite
   c. crystallization of a second generation of pyrite
(iii) Formation of "pegmatitic" sphalerite-rich ore (in the Reeves MacDonald Annex zone) possibly by mobilization via a fluid phase of magmatic origin; enrichment of Fe, Cd, Ag and Cu is observed in both "pegmatitic" and host fine-grained ore.

Mineralogical evidence that the sulphides have undergone regional metamorphism is extremely slight. It is suggested by the cobalt-nickel contents of pyrite which appear to indicate a sedimentary, and hence pre-metamorphic, origin and also by the remarkable homogeneity, on the microscale, of sphalerite in the least contact metamorphosed specimens which may imply prior equilibration of the sulphides.

TEXTURAL FEATURES OF THE SULPHIDES

Introduction

The sulphide minerals were examined in both thin and polished sections for variations in grain size and shape, grain boundary configuration and deformation structure. Structure etching of the polished sections was used extensively to reveal grain and subgrain boundaries and intragranular microstructure. A thiourea + HCl solution was used for sphalerite and galena (Brebrick and Scanlon, 1957) and a KMnO$_4$ + H$_2$SO$_4$ solution for pyrite (Cameron, 1961, p. 210-211).

Pyrite

Pyrite occurs in a wide range of grain sizes, these being illustrated semi-quantitatively for the three areas in Figure 33. Its forms too are varied and include the following:
Figure 33. Grain size variation of pyrite and sphalerite between the three mine areas.
(i) folded/brecciated massive, prismatic layers associated with graphitic laminae
(ii) fractured porphyroclasts
(iii) subidioblastic grains and granoblastic-polygonal grain aggregates
(iv) large poikilitic idioblasts
(v) late cross-cutting veinlets

At Reeves MacDonald mine, (i) and (ii) are the common forms observed. Where massive layers occur, they are invariably folded and/or brecciated (Fig. 34a). More commonly, pyritic layers contain numerous, strongly fractured, angular fragments or porphyroclasts. The fracturing is usually irregular but regular conjugate fracturing has also been observed (Fig. 34b, c). Both these types may be healed with inclusion-free pyrite. Disseminated, fine-grained pyrite has a more widespread occurrence throughout the ores and tends to be either angular to subangular and slightly fractured or subidioblastic and in places embayed. Large idioblastic forms are rare and are apparently restricted to late, "pegmatite"-ore veining.

Average grain size tends to be higher at Jersey mine and granoblastic-polygonal and idioblastic pyrite are more common (Fig. 34e, f). In numerous places the configuration of granoblastic-polygonal aggregates within sphalerite-galena layers suggests that they represent recrystallized augen or porphyroclasts (Fig. 34d). In contrast the large poikilitic idioblasts are suggestive of neocrystallization (Fig. 34f). Perhaps related to the latter are late veinlets of massive pyrite which
Figure 34. Photographs and photomicrographs of microstructures and textures developed in pyrite.

a. Folded and cleaved pyrite (grey), interlayered with highly graphitic dolomite. Reeves MacDonald mine, Reeves zone "glory-hole". Specimen RM69-17.

b. Typical irregularly fractured and fragmented pyrite. Reeves MacDonald mine, Reeves zone "glory-hole". Specimen RM69-21. Magnification 80X.

c. Conjugate fracture system in pyrite porphyroclast. Reeves MacDonald mine, Annex zone. Specimen RM71-32. Magnification 100X.


Figure 34. Photographs and photomicrographs of microstructures and textures developed in pyrite.
cross-cut the composition layering and have the same orientation as the late fractures and gashes filled with quartz and calcite previously described (Section III, p. 61). Pyrite-rich layers are commonly partially recrystallized (granoblastic-polygonal texture) but within them survive brecciated, massive pyrite fragments, characteristically associated with graphitic laminae.

Pyrite from H. B. mine commonly occurs as layered, prismatic fragments associated with minor graphitic material. These closely resemble the form (i) described above but they also occur as nodular masses up to 1 cm in diameter which have an internal radiating, prismatic structure and may look coarsely colloform; X-ray diffraction studies did not reveal the presence of marcasite in any of this material. These forms are found both as relatively isolated masses within dolomite and calcite gangue and as a component of brecciated, highly pyritic layers. Disseminated, fine-grained pyrite varies from angular to rounded, subidioblastic grains, with the former occurring more typically in the brecciated pyrite layers and the latter typically in sphalerite-galena layers where the grains may be embayed.

**Sphalerite**

Sphalerite generally forms granoblastic aggregates composed of strongly twinned equidimensional grains. The twin lamellae are broad and mostly coherent and are thus considered to be mainly recrystallization or annealing twins (Klassen-Nekhudova, 1964; Stanton and Gorman, 1968) rather than deformation twins (compare Fig. 35d and 36f). Average grain size
increases progressively from Reeves MacDonald through H. B. to Jersey mines (Fig. 33), presumably reflecting increasing intensity of contact metamorphism. Numerous specimens from Jersey mine exhibit localized exaggerated grain growth as a result of this annealing recrystallization (Fig. 35d, upper right), possibly analogous to secondary recrystallization in metals. Grain size in the other deposits tends to be more uniform locally except in the deepest zone (the Annex) of Reeves MacDonald mine where patches of coarser grain size (circa 200 μm) occur locally and are possibly related to the development of late "pegmatitic" sulphides.

Etching reveals internal grain boundaries within many larger sphalerite grains (Fig. 36d). These outline a substructure of rounded "grains" which are distinctly bimodal in grain size (averaging 80 μm and 10 μm). Twin lamellae cut across these without deflection. They probably represent true subgrains since they are terminated at host grain boundaries.

Mylonitized sphalerite, forming thin flinty-looking layers (less than 1 cm thick) parallel with composition layering, has been recognized at several localities in Jersey mine (in the F-zone) (Fig. 36a-c). The only other recorded occurrence of similar material appears to be that described by Butler (1935) from the Friedensville Appalachian-type deposit. On the microscale the mylonitized sphalerite consists of ribbon-like or lensoid concentrations forming zones within sphalerite and carbonate gangue of normal grain size. When etched these concentrations are revealed as consisting of very fine-grained polygonal
Figure 35. Photomicrographs of recrystallization textures developed in sphalerite, etched with thio-urea + HCl.

a-c Examples of fine-grained granoblastic textured sphalerite from Reeves MacDonald mine. These textures are interpreted as essentially primary recrystallization textures. Sphalerite is light grey in colour and strongly twinned, dolomite is black (from etching), and pyrite is white. All magnifications 28X.

a. Annex zone, level unknown. Specimen RM71-34.
b. Annex zone, 800' level. Specimen RM71-52.

d-f. Examples of relatively coarse-grained subgranoblastic sphalerite from Jersey mine. These textures are interpreted as annealed textures resulting from grain growth. The coherent lamellar twinning is believed to be annealing or recrystallization twinning. Note exaggerated grain growth at top centre of Figure 35d, and also bending of twin lamellae in Figures 35e, f. Sphalerite is grey and strongly twinned, pyrite is white, dolomite is black, and galena is mottled grey (from etching). All magnifications 28X.

e. D-zone. Specimen CX69-44.
f. C-zone. Specimen CX69-45.
Figure 35. Photomicrographs of recrystallization textures developed in sphalerite, etched with thio-urea + HCl.
Figure 36. Photomicrographs of deformation textures and micro-structures developed in sphalerite, etched with thio-urea + HCl.

a. Mylonitic sphalerite. Section is normal to the lamination. Jersey mine, F-zone. Specimen CX69-25. Magnification 50X.

b. Detail of sphalerite "ribbons" from mylonitic layer, showing very fine-grained polygonal texture. Jersey mine, F-zone. Specimen CX69-25. Magnification 320X.

c. Extreme detail of polygonal texture developed in mylonitic sphalerite. Jersey mine, F-zone. Specimen CX69-25. Magnification 1000X.

d. Subgrain structure developed within a single twinned sphalerite grain. Jersey mine, D-zone. Specimen CX69-55. Magnification 1000X.

e. Kinking of recrystallization twins and the development of deformation twinning in sphalerite. Note also the serrate grain boundaries. Jersey mine, D-zone. Specimen CX69-56. Magnification 80X.

Figure 36. Photomicrographs of deformation textures and microstructures developed in sphalerite, etched with thio-urea + HCl.
aggregates with an average grain size of 5-10 \( \mu \text{m} \) (Fig. 36c) which shows them to be totally recrystallized.

Evidence of slight internal deformation in sphalerite aggregates is widespread. The most common is a weak to moderate bending of the annealing or recrystallization twin lamellae (Fig. 35e, f). Locally, more intense deformation is indicated by the development of kink bands, slip lines and serrate grain boundaries in host grains (Fig. 36e). The most intense deformation is shown by sphalerite porphyroclasts within mylonitic galena layers. These may have intensely developed deformation twinning (closely spaced, narrow, non-coherent lamellae), marked dimensional preferred orientation, and sparse matrix of fine-grained (5 \( \mu \text{m} \)), polygonal-shaped grains (Fig. 36f).

The less intense deformation effects are recognized in even the most highly annealed sphalerite aggregates from Jersey mine and their development is therefore believed to post-date, wholly or in part, contact metamorphism. It is not clear to what extent the intense strain shown by sphalerite porphyroclasts within galena layers also post-dates contact metamorphism; it may, in fact, represent plastic deformation which has survived annealing recrystallization.

Galena

The highest galena concentrations, in all three deposits, occur in breccia ores. Minor concentrations elsewhere in the ore bodies are localized in such tensioonal features as cross-fractures and boudinage "necks," suggesting that solid-state movement of the galena may have occurred.
There do not appear to be any widespread significant differences in texture or microstructure among the galenas from the three deposits, except that mylonitic lamination (see below) was not observed in the galena-poor ores of Reeves MacDonald mine. Unetched galena typically has xenoblastic form in relation to the other sulphides and to dolomite gangue. Etching reveals straight, or slightly curved, grain boundaries and uniformly granoblastic-polygonal textures (Fig. 37a). Although the average grain size is about 50 \( \mu \text{m} \), surprisingly wide variations in grain size occur locally (from less than 1 \( \mu \text{m} \) to 300 \( \mu \text{m} \)) even on the microscale (Fig. 37c). Grain shapes, whatever the grain size, are generally equidimensional except for some coarse-grained material from Reeves MacDonald mine which exhibits length/width ratios for the grains of up to 2:1 (Fig. 37b). Such dimensional preferred orientation could perhaps be attributed to slight flattening. However, similar textures have been induced experimentally by static recrystallization of deformed (kinked), coarse-grained galena (Stanton and Willey, 1972); continued heating of these reduced length/width ratios of the grains by sideways growth and indicated development toward equidimensional grain structure. Lyall and Paterson (1966) also noted development of elongate grain shapes resulting from static recrystallization of some single crystals of galena experimentally deformed by 001 slip.

At many places within the galena-rich breccia ores of H. B. and Jersey mines, mylonitic lamination (grain size from less than 1 to 5 \( \mu \text{m} \)) is developed which may enclose
porphyroclasts of pyrite and of polycrystalline sphalerite together with highly deformed fragments of dolomite marble and patches of coarser-grained galena (Fig. 37d, e). This mylonitic material, as far as can be determined (the finest grains after etching are not easily resolved with the microscope), generally forms equidimensional granoblastic-polygonal textures. The coarser-grained patches of galena, by analogy with the accompanying porphyroclasts of pyrite and sphalerite, may represent textural relicts which survived mylonitization and were subsequently annealed. Such annealing might be expected to be more effective, in terms of grain boundary adjustment and grain growth, than in the matrix as a result of higher stored strain energy in the relict material.

Subgrain boundaries were detected by etching in much of the coarser-grained galena, whether equidimensional or elongate in shape (Fig. 37b). These subgrains (grain size circa 5 μm) are produced by polygonization, presumably as a result of late minor deformation. Curvilinear cleavage traces attest to bend gliding also having occurred.

**Summary and interpretation**

Pyrite, of the three sulphides, undoubtedly provides the most obvious evidence that the ores have been deformed because of its failure by brittle fracture under a wide range of conditions (Graf and Skinner, 1970), and because the resulting fractured grains and porphyroclasts survive annealing at least until conditions of hornblende hornfels facies are reached.
Figure 37. Photomicrographs of textures and microstructures developed in galena, etched with thio-urea + HCl.


b. Galena aggregate showing dimensional preferred orientation. Equant subgrains are faintly visible within the large elongate grains. Reeves MacDonald mine, Annex zone, 950' level. Specimen RM71-47. Magnification 80X.

c. Granoblastic to polygonal textured galena, showing wide grain-size variation on the microscale. Subgrains are developed within the coarser-grained material on the right. Jersey mine, A-zone. Specimen CX69-73. Magnification 28X.


e. Very fine-grained, mylonitic galena containing porphyroclasts of pyrite and of polycrystalline sphalerite. H. B. mine, X-2 zone. Specimen HB71-3. Magnification 28X.
Figure 37. Photomicrographs of textures and microstructures developed in galena, etched with thio-urea + HCl.
Under these conditions increasing recrystallization, formation of pyrrhotite and of a second generation of pyrite, may all contribute to development of new textures.

Textures of the other sulphides, sphalerite and galena, are largely those of recrystallization—typically granoblastic to polygonal. In the H. B. and Jersey ores which are located within contact metamorphic aureoles these textures appear to result largely from annealing recrystallization. However, in the Reeves MacDonald ore, which is located approximately 1000 ft away from the outermost limit of a contact aureole detectable in pelitic rocks, it is not clear to what extent the recrystallization textures result from syntectonic recrystallization, from later annealing recrystallization, or from some combination of both.

Grain sizes, as illustrated by pyrite and sphalerite, show a progressive increase with increasing intensity of contact metamorphism from Reeves MacDonald to Jersey mines.

In general, galena has better equilibrated textures than does sphalerite; a contrast in behaviour which has been demonstrated experimentally in static recrystallization of deformed galena-sphalerite ore from Broken Hill (Stanton and Willey, 1970) and attributed to the very low annealing activation temperature (perhaps as low as 10-20°C) required for grain boundary readjustment in galena (Stanton and Gorman, 1968). Much higher temperatures (generally above 300°C) are required to produce appreciable grain growth during experimental annealing of galena aggregates (Siemes, 1964; Stanton and Gorman, 1968); growth rates
apparently being affected by the degree of plastic deformation and of preferred orientation previously developed in the aggregates (Stanton and Willey, 1972). The origin of some of the more varied textures developed in the sulphides of the study area can perhaps be explained by these experimental observations.

In the mylonitic galena layers, patches of coarser grain size are attributed to areas of higher stored strain energy prior to annealing. Both fine-grained matrix and coarse-grained lenses now exhibit granoblastic-polygonal textures and the only internal microstructures observed are bent cleavage planes and subgrains developed within larger individual grains, including well-equilibrated polygonal-shaped grains. Sphalerite porphyroclasts within galena aggregates (whether mylonitized or not) exhibit a wide range of internal effects, from intense plastic deformation to total recrystallization, suggesting that sphalerite was, in general, less responsive than galena to annealing.

Layered polycrystalline sphalerite aggregates, in comparison with galena, have only local, thin mylonitic layers developed within them. The latter have well-equilibrated polygonal textures whereas the unmylonitized aggregates have less well-equilibrated granoblastic textures which, in many places, show the effects of minor deformation—slight bending of annealing twin lamellae, and the development of subgrains post-dating grain growth and hence annealing. Localized more intense strain—shown by kinking of twin lamellae and the development of slip lines—may be of the same age, or more probably it is relict as in the sphalerite porphyroclasts already described.
The widespread slight deformation and recovery effects observed in sphalerite and galena could have developed at any time since contact metamorphism by granitic stocks in the area produced annealing; faulting continued at least until the Eocene thermal event represented by intrusion of lamprophyre dykes and alkaline stocks. Equivalent effects have been described from only slightly deformed vein-type sulphide deposits (Burn, 1971) and may be a feature of many sulphide ores as has been suggested by Stanton and Gorman (1968). In addition, it has been suggested that creep within mine pillars and in unsupported stopes might be sufficient to induce these features. However, considering sphalerite's somewhat higher annealing activation temperature (perhaps 60°C compared to 10-20°C for galena, according to Stanton and Gorman, 1968), temperatures within the mines are probably insufficient to promote recovery in sphalerite at least.

FABRIC OF THE ORES

Introduction

The optical determination of crystallographic preferred orientation in opaque minerals, such as sulphides, is difficult if not impossible, particularly where fine-grained polycrystalline aggregates are involved.

X-ray methods have, however, been used extensively. Bradshaw and Phillips (1970) have reviewed these methods and their application to petrofabric studies. They include:

(a) stationary and moving camera methods (Sander and Sachs, 1930; Wenk, 1963, 1965; Starkey, 1964)
(b) diffractometer and texture goniometer methods (Schulz, 1949a, b; Gehlen, 1960; Higgs et al., 1960; Davis, 1966; Baker et al., 1969)

(c) the X-ray universal stage (Paulitsch, 1963)

Applications to sulphide fabrics are relatively few. Gehlen (1960) first discussed the application of an X-ray texture goniometer method to the determination of preferred orientation in ore minerals, using the Schulz (1949a) reflection technique. Various error sources were also investigated and examples of preferred orientation were determined for pyrrhotite, hematite, pyrite, sphalerite and chromite. Siemes (1964) and Siemes and Schachner-Korn (1965), using the same method, determined the fabrics developed in "gneissic" galena (bleischweif) and the effects on them of annealing. Experimental deformation of galena at room temperature and its fabric development have also been studied by Siemes (1970). A similar study of sphalerite was conducted by Saynisch (1970) who also compared the experimentally produced fabrics with those of naturally deformed sphalerite. Chakrabarti (1969) briefly compared natural fabrics developed in both ore and gangue minerals from the Zawar (Rajasthan) lead-zinc deposits which occur in regional metamorphic terrain. An X-ray photographic method (Starkey, 1964) was used by Bertrand (1969) to test for preferred orientation in the regionally metamorphosed sulphide ore of the Normetal mine (N. W. Quebec) but subfabrics were found to be completely random and presumed annealed.
In this study, investigation of preferred orientation in the sulphide aggregates was undertaken to determine the nature and degree of any crystallographic preferred orientation developed in the ores, and to determine what fabric changes occurred as a consequence of increasing intensity of contact metamorphism. In metallurgical experience, at least, it is generally accepted that annealed fabrics may either resemble the pre-existing deformation fabric, or be completely different, or even be random (Barret and Massalski, 1966, p. 568-569; Byrne, 1965, p. 105). However, the deformation fabrics referred to are generally the result of cold working (plastic deformation at low temperatures) whereas in naturally occurring mineral assemblages the pre-annealing fabrics may result either from plastic deformation (cold working), from syntectonic recrystallization (hot working), or from both these processes.

To test for the existence of preferred crystallographic orientation, oriented slabs of sulphides, cut parallel to composition layering (Fo-i), were X-rayed using a standard Phillips texture goniometer. The Schulz (1949a) method was used to measure variations in reflection intensity of the Bragg angle for several crystallographic forms of sphalerite and galena. This involves simultaneous rotation about its own axis and tilting of the specimen slab in the incident X-ray beam which permits measurement of reflection intensities throughout hemispherical space above the slab surface whilst keeping the Bragg angle constant. The method suffers from the limitation that it provides limited coverage of the hemispherical reflection geometry
due to defocussing of the X-ray beam when the sample is tilted to high angles. Thus coverage was restricted to within 70° from the foliation pole. Details of instrument setting and possible error sources are given in Appendix B.

Measured reflection intensities were plotted in equal-area polar projection and contoured as multiples of a uniform distribution, taken to be the mean intensity expressed as unity (Baker et al., 1969). Such intensity plots are known as pole figures and are defined (Green et al., 1970) as the distribution on an equal-area projection of the poles to all planes of a crystal form with reference to a set of specimen coordinates.

**Sphalerite fabric**

Pole figures were constructed from reflection intensities obtained from (111) and (220) forms to determine the nature of preferred orientation developed in polycrystalline sphalerite. These are the two most intense reflections of sphalerite. In addition, (111) planes constitute the principal glide planes operative during plastic deformation (Buerger, 1928) and might therefore be expected to show preferred orientation.

**(111) Subfabric.** Various patterns of preferred orientation are illustrated by the (111) reflections. These are separable into several types which can be qualitatively related to observed textural variations.

(a) **Specimens from Reeves MacDonald mine** generally display primary recrystallization textures although these may be locally modified (particularly in the Annex zone) by areas of increased grain growth. Pole figures show a central
maximum of [111] falling within 20° of the normal to the specimen foliation, indicating a preferred orientation of (111) planes parallel to subparallel with the composition layering in the ore (Fig. 38). The symmetry is usually axial but local departures to near orthorhombic symmetry also occur (Fig. 38a, e). The latter may result from superimposition on the axial fabric of a later fabric element, or from modification to the axial fabric, for example by bend gliding—in this case about a southerly plunging axis (equivalent geometrically to the Phase 3 dextral axis recognized in the host rocks?).

The only specimen of polycrystalline sphalerite from H. B. mine which was analyzed was found to have a (111) subfabric very similar—in pattern, intensity and axial symmetry—to those of specimens from Reeves MacDonald mine.

(b) Specimens from Jersey mine, with average grain sizes two to three times larger than those of Reeves MacDonald specimens, are assumed to be highly annealed. In addition grain growth is exaggerated in many places as a result of secondary recrystallization. Pole figures show a range of patterns in which [111] sub-maxima are dispersed increasingly outwards to form a small circle annulus 40-45° from the foliation normal (Fig. 39). This implies the development toward a preferred orientation of (001) planes parallel with the layering in the ore. Such a subfabric (Fig. 39f) vaguely resembles the "cube texture" commonly developed by annealing in FCC metal
Figure 38. (111) pole figures for sphalerite from Reeves MacDonald mine. Sections are cut parallel with composition layering ($F_{0-1}$) and oriented north-south. The average lineation ($L_1$) direction is indicated. The broken circle represents a reference horizontal plane. Intensity contours are drawn at 0.55, 0.70, 0.85, 1.00, 1.15, 1.30, >1.45 X uniform distribution (maximum intensities are shown at lower right of each figure). Specimen numbers and locations are given in Table XIII (Appendix B).
Figure 39. (111) pole figures for sphalerite from Jersey mine. Sections are cut parallel with composition layering (F_0 - i) and oriented north-south. The average lineation (L_1) direction is indicated. The broken circle represents a reference horizontal plane. Intensity contours are at 0.70, 0.85, 1.00, 1.15, 1.30, > 1.45 X uniform distribution (maximum intensities are shown at lower right of each figure). Specimen numbers and locations are given in Table XIII (Appendix B).
sheet (Barret and Massalski, 1966, p. 570). It cannot, however, be clearly demonstrated that this represents a systematic trend related to increasing intensity of annealing.

(c) Mylonitized specimens from Jersey mine are very fine-grained and completely recrystallized with a well developed polygonal texture. Pole figures show a preferred orientation in which there is a tendency for two concentrations of [111] to develop (Fig. 40). This, in one of the specimens, approximates an ideal (110)[001] or (110)[110] subfabric in which there is a tendency for (110) planes to become aligned with the composition layering (Fig. 42c). (An ideal subfabric is defined as the pattern assumed by the pole figure maxima when superimposed on a standard projection (in this case cubic) of the form being analyzed.) The other specimen contained both mylonitized and unmylonitized sphalerite and is assumed, therefore, to show a hybrid fabric (Fig. 40a).

(220) Subfabric. Several pole figures of (220) reflections were also constructed (Fig. 41) to check that the (111) preferred orientations were expressed by other crystallographic forms. In particular, the (220) pole figure of the specimen from Reeves MacDonald mine (Fig. 41a) clearly confirms the existence of a central [111] maximum. Further, the distribution of [110] maxima suggests a weak directional fabric corresponding to the ideal (111) [112] pattern. The (220) pole figure of the specimen from Jersey mine is somewhat more complex, reflecting the
Figure 40. (111) pole figures for a. partially mylonitized, and b. mylonitized sphalerite from Jersey mine. Sections are cut parallel with the foliation and oriented north-south. Intensity contours are drawn at 0.85, 1.00, 1.15, 1.30, 1.45, >1.60 X uniform distribution (maximum intensities are shown at lower right of figures). Superimposed on Figure b is the ideal subfabric (110)[001]. Specimen numbers and locations are given in Table XIII (Appendix B).
Figure 41. (220) pole figures for sphalerite from a. Reeves MacDonald and b. Jersey mines. Sections are cut parallel with composition layering (F₀₋₁) and oriented north-south. The average lineation (L₁) direction is indicated. The broken circle represents a reference horizontal plane. Intensity contours are drawn at 0.8, 1.0, 1.2, 1.4, >1.6 X uniform distribution (maximum intensities are shown at lower right of each figure). Superimposed is the ideal subfabric (111)[112]. Specimen numbers and locations are given in Table XIII (Appendix B).
complexity of the (111) subfabric. However, a weak 3-fold symmetry can be detected which is also suggestive of the ideal directional fabric (111) [\( \overline{11\overline{2}} \)] (Fig. 41b).

**Origin of sphalerite fabric.** Buerger (1928) uniaxially compressed sphalerite and showed that it deformed by twin gliding on the system (111) [\( \overline{11\overline{2}} \)] (Fig. 42a). Saynisch (1970) found that [110] axes were aligned with the compression axis during axial deformation of polycrystalline sphalerite. He also demonstrated, by X-ray fabric analysis, that (111) [\( \overline{11\overline{2}} \)] and (110) [001] (Fig. 42b, c) were commonly developed subfabrics in naturally deformed sphalerite aggregates.

Assuming deformation to have been produced by simple shear, he showed, by applying the theory of Schmid and Boas (1950) to the 24 translation systems of \{111\} [\( \overline{11\overline{2}} \)], that both (111) [\( \overline{11\overline{2}} \)] and (110) [001] were stable "end positions." No details are given of the texture and microstructure of these naturally deformed sphalerite aggregates but it is implied that the fabrics result from plastic deformation.

Nearly all of the Salmo material has been recrystallized (prior to minor late straining) so that it is not clear from the textures whether the pre-annealing fabric was produced mainly by syntectonic recrystallization, or by plastic deformation. The very weak directional anisotropism of most of the fabrics, unlike those figured by Saynisch (1970) (Fig. 42b, c), suggests that plastic deformation *per se* was not the main orienting process; syntectonic recrystallization appears more likely. Certainly, syntectonic recrystallization would be expected under the
Figure 42. a. Translation directions in the (111) plane of sphalerite. b, c. (111) pole figures of naturally deformed sphalerite, showing inferred ideal subfabrics and possible translation directions (from Saynisch, 1970).
synmetamorphic conditions of Phase 1 folding and possibly also under the submetamorphic conditions of Phase 2 folding. Under these conditions, with estimated temperatures of approximately 425°C and 350°C respectively (±50°C), i.e., with $T/T_m = 0.5-0.6$ for sphalerite (where $T_m$ is the absolute M.P.), recovery would be expected to be an important process, producing polygonization and ultimately recrystallization. It is proposed, therefore, that the widespread preferred orientation of (111) planes parallel with composition layering in the ores results from syntectonic recrystallization.

The weak (111)[112] element within the more general (111) axially symmetric fabric is ascribed to minor, pervasive plastic deformation superimposed on the latter fabric. Age of this deformation is problematic; translation is indicated as having occurred within the plane of the composition layering in a direction normal to the main fold axes so that it could be related perhaps to Phase 2 folding but then would have to have survived annealing recrystallization. Indeed, there is some evidence that this may have occurred even in Jersey mine (p. 137).

It is not known to what extent the slight deformation and recovery effects, which post-date annealing recrystallization, contribute to the observed subfabric elements. Possibly this is limited to very minor modifications of the pre-existing subfabrics by bend gliding.

Mylonitic sphalerite, with its very fine-grained polygonal, strain-free grains, has a (111) subfabric which approaches an ideal (110)[001] pattern with orthorhombic symmetry. Such
mylonitic textures have been interpreted as forming in zones of restricted extensional flow at high strain rates (Ross, 1973). Flattening, as during Phase 2 folding, could produce these mylonite zones but there is little evidence for this within the ore zones of Jersey mine; there, refolding, rather than closure, of Phase 1 folds occurred. It is postulated, therefore, that the mylonite zones were formed by simple shear during the thrust movements which occurred at a late stage of Phase 2 folding. As observed by Tullis et al. (1973), in experimental deformation of quartzite, the magnitude of finite strain determines the microstructure, not whether the deformation is axial or rotational.

**Galena fabric**

The (200) form of galena was investigated because of its relatively high diffraction intensity. Measured intensities, however, were so weak for this form that analyses of other forms were not made. Only specimens from Jersey mine were used, other available material being either too low in galena content (Reeves MacDonald specimens) or of unknown orientation (H. B. specimens).

**200 Subfabric.** The (200) subfabric of variably mylonitized galena specimens from Jersey mine is either random or weakly developed as might be inferred from their generally well-equilibrated, granoblastic-polygonal textures. Their pole figures, where the subfabric is not random, show maxima restricted to within 15-20° of the normal to the specimen foliation and in some cases distinctly elongated north-northeast to south-southwest (Fig. 43). A preferred orientation of (001) planes parallel with the composition layering (P₀₋₁) is thus indicated.
Figure 43. (200) pole figures for galena from Jersey mine. Sections are cut parallel with composition layering ($F_{0-1}$) and oriented north-south. Broken circle represents reference horizontal plane. Average $L_1$ lineation direction in the foliation plane is indicated. Intensity contours are drawn at 0.75, 1.00, 1.25, >1.50 X uniform distribution (maximum intensities are shown at lower right of each figure). Specimen numbers and locations are shown in Table XIII (Appendix B).
The elongation of the central maxima, which changes the fabric symmetry from axial to orthorhombic, may result from bending of the (001) planes; such bend gliding is attested by the local development of curviplanar cleavage in these galena aggregates. The bend glide axes (from Fig. 43b, d) plunge at low angles southeast or northwest and therefore have somewhat similar orientations to Phase 3 dextral axes in the host rocks.

However, the origin of this preferred orientation of (001) planes is problematic. Mugge (1898) and Buerger (1928) observed that galena deformed plastically by slip on the system (001)[110] (Fig. 44a). This has been verified for deformation of polycrystalline galena by Lyall and Paterson (1966) who also offered evidence that an additional slip system (110)[110] operated at higher stresses when orientation was unfavourable for slip on (001)[110] (Fig. 44b). In addition, it has been shown by Siemes (1970) that axial deformation of polycrystalline galena at room temperature, between 0.3-5.0 kb confining pressure, results in uniform flow and the alignment of [110] poles parallel with the compression axis.

Conceivably, the observed (001) preferred orientation could result from plastic deformation but textures are recrystallized and deformation effects, as far as can be determined, are limited to lattice bending and subgrain development. Textures are mylonitic in many places but the subfabric does not resemble those developed in naturally deformed "gneissic" galena (bleischweif) which has various well developed (200) subfabrics approximating the ideal (110)[001] pattern (Fig. 44c). The
Figure 44. a. Translation directions in the (100) plane of galena. b. Translation directions in the (110) plane of galena. c. Example of (200) pole figure of naturally deformed "gneissic" galena (from Siemes, 1970), showing inferred ideal sub-fabric and possible translation directions.
latter are considered to have been produced by shearing deformation (Siemes and Schachner-Korn, 1965) but the axial deformation experiments of Siemes (1970) would suggest that preferred orientation of (110) planes, parallel with composition layering, could also be produced by flattening. However, as with mylonitic sphalerite from Jersey mine, these mylonite textures are considered as resulting from localized shear within the galena-rich layers, perhaps related to thrusting at a late stage of Phase 2 folding. The (200) subfabric, since it does not appear to be related to shearing, may result from annealing recrystallization, or from weak plastic deformation (either surviving annealing or imposed on an annealed random fabric).

Dolomite fabric

The orientations of c-axes of dolomite in four specimens of dolomite marble were determined, using a polarizing microscope and universal stage. Two of the specimens were from Reeves MacDonald mine and two from Jersey mine. Due to a general lack of twinning and the relatively low birefringence the c-axes were measured directly (Turner and Weiss, 1963, p. 239). For each specimen, 200 grains were measured in a pair of mutually perpendicular thin sections and the data combined in a single diagram normal to the Li lineation direction. The preferred orientations (Fig. 45) show relatively consistent, strong concentrations of c-axes subperpendicular to the composition layering (F0-1).

Such patterns are apparently typical of many dolomite tectonites (Ladurner, 1953; Christie, 1958), in particular the so-called S-tectonites. Neumann (1969) has produced similar
patterns by experimental recrystallization of Dover Plains and Knox dolomite under directed stress. Neuman also suggested, because of the common similarity between the preferred orientations of c-axes in recrystallized calcite and dolomite, that slip and twinning processes do not contribute to the development of the preferred orientations; plastic deformation, under experimental conditions at least, results in different c-axis sub-fabrics for the two minerals reflecting their different translation and twinning systems (Handin and Fairbairn, 1955; Higgs and Handin, 1959; Griggs et al., 1960). Recrystallization, involving the nucleation of new strain-free grains, is proposed by Neumann (1969) as the orienting process where plastic deformation cannot be demonstrated.

The dolomite specimens from the Salmo area, examined for preferred orientation, typically have uniform grain sizes, granoblastic to polygonal textures and exhibit only limited f-twinning (shown by less than 5% grains). Grain size differences, which exist between Reeves MacDonald (average 55 µm) and Jersey (average 125 µm) dolomite, and the distinctly more polygonal texture of the latter, imply that annealing recrystallization has affected at least the dolomite marble from Jersey mine area. There is also a slight suggestion that preferred orientation is more sharply developed in the Jersey dolomite (compare maxima in Fig. 45).

The concentrations of c-axes subnormal to composition layering in all the specimens are similar to those developed experimentally by axial deformation and are consistent with and
Figure 45. Preferred orientation of dolomite e-axes (lower hemisphere, equal-area projection) in four dolomite marbles. Sections are cut normal to the L₁ lineation. 200 grains were measured for each diagram and contours drawn at 1, 2, 3, 4, >5% per 1% area (maximum values are shown at lower right of each figure).

a. Reeves MacDonald mine, Reeves zone "glory-hole". Specimen RM69-18.

b. Reeves MacDonald mine, Reeves zone, 1900' level. Specimen RM69-60.

c. Jersey mine, D-zone. Specimen CX69-47.

d. Jersey mine, E-zone. Specimen CX69-43.
perhaps related to flattening perpendicular to the layering during recrystallization, as has been suggested for certain other dolomite tectonites (Neumann, 1969). These preferred orientations are interpreted as resulting from syntectonic recrystallization probably during Phase 1 deformation and possibly also during Phase 2 deformation. However, the dolomite may have been wholly, or partly, plastically deformed during Phase 2 with much of the strain being subsequently annealed out during contact metamorphism. Experimental recrystallization of dolomite suggests that early formed, well developed fabrics may be retained, or only slightly modified, during syntectonic recrystallization or during plastic deformation followed by annealing. Such modifications to the $c$-axes fabrics generally involve the development of small-circle girdles in place of concentrations close to the axis of compression; these appear to result from nucleation and growth of new strain-free grains with $c$-axes tending to be inclined 30-60° to those of old host grains (Neumann, 1969).

Quartz fabric

Quartz $c$-axes orientations in four quartzite and in two quartz phyllite/schist specimens were determined by the standard universal stage method. Relatively consistent preferred orientations were obtained (Fig. 46).

The subfabric diagrams for the four quartzites show peripheral "split-maximum" girdles with approximate orthorhombic symmetry but the symmetry axes are not symmetrically disposed with respect to specimen foliation and lineation (Fig. 46a,b,e,f). These patterns somewhat resemble, and are found in association
with, the "crossed-girdle" patterns commonly developed in other S-tectonites. The latter, however, have symmetry axes symmetrically disposed with respect to foliation and lineation and are interpreted as resulting from flattening perpendicular to the foliation, with the lineation approximating the direction of elongation (Sylvester and Christie, 1968). The patterns are thus correlated, in a general way, with strain. However, in situations of polyphase deformation, c-axes distributions probably also reflect effects of stress variation and fabric anisotropism in addition to a more general type of strain. Because of their over-all low symmetry and uniform patterns, the c-axes subfabrics of these Salmo quartzites are ascribed to more general strain than that resulting from flattening. The development of these subfabrics is presumed to result from syntectonic recrystallization during Phase 1 deformation, i.e. prior to flattening.

The quartz phyllonite from Reeves MacDonald mine area has an interesting fabric composed of two elements:

(a) elongate highly strained augen with a symmetrical "crossed-girdle" pattern of c-axes (Fig. 46c), and
(b) small polygonal strain-free grains with a small-circle girdle of c-axes normal to the foliation (Fig. 46d).

The quartz schist from Jersey mine area has a uniform well-equilibrated polygonal texture and a weak small-circle girdle of c-axes (Fig. 46g) resembling that of the small strain-free grains in the phyllonite from Reeves MacDonald mine area.

Similar small-circle girdles of c-axes have been produced in experimental syntectonic recrystallization of flint
Figure 46. Preferred orientation of quartz c-axes (lower hemisphere, equal-area projection) in four quartzites (a, b, e, f), a quartz phyllonite (c, d), and a quartz schist (g). Sections are cut normal to the L1 lineation. In the quartz phyllonite, 200 elongate strained grains (c), and 200 polygonal strain-free grains (d) were measured. In all other specimens 300 grains were measured. Contours drawn at 1.0, 2.0, 3.0, >4.0% per 1% area (maximum values shown at lower right of each figure).

a. Specimen RM69-23] Quartzite Range Fn., core of Salmo anticline
b. Specimen RM69-24] Quartzite Range Fn., core of Salmo anticline
c-d. Specimen RM69-13, U. Laib mamber, overturned limb of Salmo River anticline
e. Specimen CX69-3] Quartzite Range Fn., core of Jersey anticline
f. Specimen CX69-38] Quartzite Range Fn., core of Jersey anticline
g. Specimen CX69-4] Quartzite Range Fn., core of Jersey anticline
in the alpha quartz field during axial compression (Green et al., 1970), and in axial deformation experiments (under similar conditions) on quartzite (Tullis et al., 1973). There is, however, some doubt as to the relative roles of syntectonic recrystallization and of plastic deformation in the development of such preferred orientation. "Crossed-girdle" patterns, similar to those observed in quartzites naturally deformed by flattening (Sylvester and Christie, 1968), have also been produced by recrystallization during non-axially symmetric compression (Green et al., 1970) which suggests that these patterns are a more general strain equivalent of the small-circle patterns of preferred orientation.

Thus the preferred orientations in the two fine-grained quartz-rich rocks which have been examined are assumed to have been produced by flattening, presumably during Phase 2 folding. Deformation lamellae may or may not be developed in the quartzites. Two of the four specimens contained abundant well developed lamellae. The orientations of 75 sets of lamellae were measured in each together with the \(c\)-axes orientations of host grains. Histograms (Fig. 47) of the angular relations between these elements show the lamellae to be predominantly sub-basal, as commonly developed in quartzite tectonites (Friedman, 1964; Carter and Friedman, 1965). Avé Lallement and Carter (1971) have shown that identical sub-basal lamellae can be produced experimentally in quartz at moderate temperature and low pressure and propose that natural sub-basal lamellae should form below \(500^\circ\)C for strain rates \(<10^{-13}\)/sec. In Figure 47, the "arrow
Figure 47. Top: Equal area projections of poles to 75 sets of deformation lamellae (arrow heads), and of c-axes of host grains (arrow tails) in each of two quartzites (a. RM69-24 from Reeves MacDonald mine area; b. CX69-3 from Jersey mine area).  Centre: Histograms of angles between lamellae pole and c-axis in each grain.  Bottom: Deduced maximum principal stress directions (see text for explanation).
method" (joining \( c \)-axes to corresponding deformation lamellae poles) is employed to locate the orientation of the principal stress axes (Carter and Raleigh, 1969). The deduced positions indicate maximum compressive stresses oriented approximately normal to the specimen foliations implying that the plastic deformation was produced by flattening. For the specimen from Jersey area (Fig. 47b), this orientation appears symmetrically related to Phase 2 folding deformation whereas for the specimen from Reeves MacDonald area (Fig. 47a), the orientation most closely approximates that deduced from Phase 3 folding (Fig. 7i).

Annealing, as a result of contact metamorphism, appears to have had some effect on preferred orientation of \( c \)-axes; those of Jersey specimens are very similar to, but slightly weaker than, those of Reeves MacDonald specimens (Fig. 46). This agrees with annealing experiments in the \( \alpha \)-quartz field (Green et al., 1970) which have produced only weak to moderate changes in preferred orientation of \( c \)-axes (over limited time periods) particularly for small-circle girdle patterns.

Summary

X-ray fabric analyses show that the sulphide minerals possess varying degrees of preferred orientation. Galena, sampled from Jersey mine only, has a very weak fabric and corresponding, generally well-equilibrated polygonal texture. Sphalerite, in contrast, has well-developed, varied fabric, even where coexisting with galena.

Specimens of sphalerite from Reeves MacDonald mine have relatively systematic subfabric elements in which (111) planes
exhibit a preferred orientation parallel with composition layering in the ores. (220) subfabric patterns reflect this same preferred orientation and also indicate a very weak directional element for the combined fabric, corresponding to (111) [\{11\overline{2}\}] which is the twin gliding system operative in plastic deformation of sphalerite. This element is believed to post-date the axially symmetric (111) subfabric and may reflect plastic deformation features observed in the texture.

Sphalerite from Jersey mine shows more varied (111) subfabrics, ranging from Reeves MacDonald-type subfabrics to more complex, small-circle patterns of submaxima. The latter are interpreted as representing true annealing subfabrics, developed from the Reeves MacDonald patterns which are believed to represent relatively unmodified, pre-existing deformation fabrics produced by syntectonic recrystallization.

Mylonitic sphalerite from conformable thin zones within the Jersey ores has a distinctive orthorhombic (111) subfabric approaching an ideal (110) [001] pattern and is considered to result from simple shear within the composition layering, perhaps in association with thrusting at a late stage of Phase 2 folding.

The variably developed mylonite textures in galena from Jersey mine are believed to have formed in the same manner but the (200) subfabric, if developed, is very weak and does not resemble the patterns of preferred orientation observed in "gneissic" galena (Siemes and Schachner-Korn, 1965). The (200) subfabric is, therefore, considered to represent either a weak
annealing subfabric, or a weak subfabric produced by slip on the system (001)[110], either surviving annealing or superimposed on an annealed random fabric.

Quartzites have peripheral "split-maximum" girdle patterns of c-axes, resembling "crossed-girdle" patterns but they are not symmetrically disposed with respect to the foliation/lineation. Orientations of sub-basal deformation lamellae, measured in two of the quartzites, suggest that plastic deformation (assumed to post-date syntectonic recrystallization) was produced by flattening. Recrystallized grains in two fine-grained quartz-rich rocks were found to have small-circle girdle patterns of c-axes disposed symmetrically with respect to the foliation/lineation. These were interpreted as indicating recrystallization also produced during flattening.

Preferred orientations of c-axes in four dolomite marbles show single or small-circle maxima oriented approximately normal to composition layering in the marbles which may also be indicative of flattening.

Annealing recrystallization during contact metamorphism has generally weakened c-axis subfabrics in the quartz-rich rocks but appears to have had little effect on c-axis subfabrics in dolomite marble, although both groups show evidence of textural changes.

The main fabrics of the host rocks are ascribed, therefore, to syntectonic recrystallization with minor modifications, in the form of plastic deformation and/or recrystallization, being
produced, at least in part, by flattening during Phase 2 folding and possibly also, in the Reeves MacDonald area, during Phase 3 folding.

No subfabric elements recognized in the sulphides can be attributed directly to flattening, although mylonite textures in Jersey mine can be related to Phase 2 folding. It is by no means clear what the response of sphalerite in the Reeves MacDonald ores was to the flattening which occurred there during Phase 2 folding (see Section III, p. 56); the relative roles of syntectonic recrystallization and plastic deformation are not known. A preferred orientation of (110) planes parallel with the composition layering might be expected from axial deformation experiments (Saynisch, 1970) but is not observed. Syntectonic recrystallization, as during Phase 1 folding, may be the main orienting process. The mechanism for the development of the preferred orientation by recrystallization is not known but current experimental workers (e.g., Green et al., 1970) appear to favour the "oriented growth" theory, based on the mobility of high-energy grain boundaries (Barret and Massalski, 1966, p. 579). However, it should be noted that the thermodynamic theory of Kamb (1959) for selective grain growth, due to differences in strain energy under non-hydrostatic stress, predicts the development of preferred orientation of (111) or (001) planes in cubic crystals normal to the unique stress axis.
ORIGIN OF THE ORES

Evidence from this study as to the origin of the lead-zinc ores is largely inconclusive, or, to put it in the terminology of Snyder (1967), "permissive" (i.e., subject to dual interpretation).

The structural evidence shows that the ores have been penetratively deformed and are involved in both Phase 1 and Phase 2 fold structures. Mineralization, therefore, pre-dates Phase 1 deformation. The age of this deformation could not be defined within the study areas but, from regional stratigraphic considerations, it is believed to have occurred during the Devonian-Mississippian hiatus. If this interpretation is correct mineralization could have occurred at any time between the Lower Cambrian and Devonian periods, although similar mineralization in the Nelway/Metaline formation may further limit this interval to between Middle Cambrian and Devonian periods.

The local stratigraphic evidence shows that the ores are located in a dolomitic facies of the Lower Cambrian Reeves member and that they are essentially strata-bound. The Reeves member is considered to have been deposited in a shelf environment with the localized dolomitic facies having developed on shallow banks. Where the host dolomite marble has not been recrystallized during contact metamorphism, the sulphides (particularly pyrite) are commonly associated in the marble with minor amounts of graphitized carbonaceous material and with quartz.

Mineralogy of the ores is simple, comprising mainly sphalerite, pyrite and galena with only minor pyrrhotite and
and chalcopyrite locally in addition. Cobalt and nickel contents of pyrite are suggestive of a syngenetic-diagenetic origin for that mineral at least. Where least affected by contact metamorphism, sphalerite has uniformly low iron and manganese contents which are suggestive of a low temperature of formation.

The sulphur isotopic character of the Salmo ores (Sangster, 1970b) indicates derivation by biogenic reduction of sea water sulphate; $\delta^{34}S$ values are positive and restricted in range for the individual deposits. Sangster has interpreted these as indicating a syngenetic origin for the ores as the $\delta^{34}S$ values are somewhat isotopically lighter than those determined for sea water sulphate of Cambrian age (Thode and Monster, 1965). However, in practice it is not possible to differentiate between isotopic compositions of Cambrian and Ordovician sulphates so that, although the sulphur is derived from sea water, it need not be syngenetic with respect to the host rocks.

Lead isotope abundances in the Salmo deposits are anomalous (Sinclair, 1966; Reynolds and Sinclair, 1971). According to Snyder (1968), anomalous leads which are characteristic of the Mississippi Valley deposits are diagnostic of an epigenetic origin for the ores. However, numerous Mississippi Valley type deposits have non-radiogenic or single-stage leads, for example the Pine Point and Monarch deposits (Baadsgaard et al., 1965; LeCouteur, 1973) which suggest that the lead isotopic character of ores is not diagnostic as to origin.

In many of the above characteristics, stratigraphic, mineralogical and isotopic, the Salmo deposits resemble typical
Mississippi Valley deposits, as described by Ohle (1959) and by Snyder (1967, 1968). On similar grounds, Muraro (1962) considered the Duncan deposit, occurring in a very similar geological setting 80 miles north of Salmo (Fig. 1), to have been a Mississippi Valley type deposit prior to regional metamorphism. Some of the obvious differences in character, such as the fine grain size, relatively homogeneous nature of the sulphides and the lack of voids and open spaces, can be attributed to penetrative deformation and recrystallization.

Current theories as to the genesis of Mississippi Valley type deposits invoke various fluid transport and precipitation systems operative at the flanks of large sedimentary basins (Callahan, 1964; Jackson and Beales, 1967; White, 1968; Beales and Onasick, 1970; Hoagland, 1971; and others). The transporting fluids are generally considered, as a result of fluid inclusion studies (Hall and Friedman, 1963; Roedder, 1967), to be chloride brines of largely connate origin, with local contributions of meteoric or magmatic water being invoked in specific situations. Source of the brines and transported metals is assumed to be shales of the sedimentary basins undergoing diagenesis, compaction and eventual burial metamorphism. Two models for these metal-bearing brines are envisaged—one sulphur-rich and involving metal-sulphide complexes, and the other sulphur-deficient and involving metal-chloride complexes (White, 1968). It was to the second of these that Jackson and Beales (1967) subscribed. They proposed that sulphur-deficient brines escaping from deep sedimentary basins mix at shallow depths with H$_2$S-rich brines in
carbonate reservoirs, resulting in precipitation of ore minerals. A variation on their model is suggested for the formation of the Salmo ores. Black shales of the Active formation of Ordovician-Devonian age may have been the source for lead and zinc. The close association of pyrite (Co:Ni ratios <1) with graphitized carbonaceous material in the deposits suggests that a sulphide-rich environment may have been developed syngenetically or diagenetically in the host carbonates and hence been available to precipitate ores from zinc- and lead-bearing brines moving upward and eastward out of the shale basin. Such a model would satisfy both the sulphur and lead isotopic data.

The nature of the ore-bearing structures in the Salmo deposits prior to deformation is obscure. Breccias are extensively developed and are interpreted as being largely of tectonic origin but this does not preclude the prior existence of breccias of other origins--by sedimentary processes, or by solution and collapse. Indeed, in places dolomite breccia resembles crackle breccia but this resemblance may be superficial as a wide variety of transposition structures affect the dolomite marble.
SECTION VI

SUMMARY AND CONCLUSIONS

STRUCTURE

The structures of the three mine areas reveal a complex history involving at least three phases of folding:

Phase 1  overturned, near-isoclinal folds of "similar" type, with westward vergence.

Phase 2  upright, asymmetric folds of flattened "flexural-slip" type, with westward vergence. These are associated with westward-directed thrust faults.

Phase 3  monoclinal folds and kink bands forming a conjugate system with monoclinic symmetry. Deduced maximum principal stress directions are oriented approximately north-south, suggesting an association with northward-directed thrust faults.

Shearing and transposition in the cores of Phase 1 folds indicate that closure and flattening of these structures occurred, in addition to refolding, during Phase 2 deformation. Culminations and depressions in the attitude of Phase 1 linear structures result from interference between Phase 1 and Phase 2 fold structures whereas broad culminations and depressions affecting both Phase 1 and Phase 2 linear structures result from Phase 3 folding.

176
The marked curvature of the southern Kootenay Arc near 49°N is believed to be a composite effect produced by (i) southward developing divergence between Phase 1 and Phase 2 structural trends, and (ii) northward thrusting and associated steeply plunging monoclinal and kink folding of Phase 3 age.

The chronology of these phases of deformation cannot be determined within the mine areas although it can be shown that the last phase must pre-date granitic intrusion for which a minimum age of 100 m.y. B.P. was determined.

SULPHIDE DEPOSITS

The tabular sulphide bodies are shown to have been involved in both Phase 1 and Phase 2 folding. Layering within the sulphides and in the ores parallels composition layering in the host dolomite marbles and some places outlines mesoscopic folds of both ages. In other places, the sulphides may be transposed along the F1 foliation. Internal deformation is indicated by the development of augen/flaser structures involving both sulphides and marble, and by the development of mylonitic layers within the sulphides, sphalerite and galena. In many places too, the sulphides form the matrix to extensive tabular bodies of breccia which are essentially conformable with the composition layering. In the past, these were interpreted as mineralized breccias but are now interpreted as being largely of post-mineralization, tectonic origin resulting from differential movement between dolomite marble and more highly ductile sulphide, and also between sulphides of different ductilities. Detached folds within these breccia bodies display no systematic geometry and
and cannot be assigned to any particular phase of folding. Microstructural evidence for internal deformation is given mainly by pyrite which typically forms fractured grains and augen within equidimensional, recrystallized sphalerite and/or galena.

The deposits are considered to be ultimately of epi-genetic origin and were probably of Mississippi Valley type prior to deformation and metamorphism. They are thus not believed to differ fundamentally from the Metaline deposits of Middle Cambrian age; the differences that do exist probably resulting from (i) slight facies differences in the host carbonates, (ii) more intense deformation, and (iii) the more complex metamorphic history of the Salmo deposits.

DEFORMATION AND REGIONAL METAMORPHISM

The effects of deformation and regional metamorphism on the sulphides are relatively obscure, largely because they involve recrystallization which in many places is overprinted by the effects of contact metamorphism; of the three deposits, the Reeves MacDonald is least affected by the overprinting. Sphalerite from it shows remarkable homogeneity of composition on the microscale, as determined semi-quantitatively by electron microprobe. Textures also tend to be more uniform, with pyrite generally exhibiting the effects of brittle fracturing, and sphalerite and galena displaying typically granoblastic-polygonal textures and generally equant grain shapes.

X-ray fabric determinations show that sphalerite aggregates, mainly from Reeves MacDonald but also locally in H. B. and Jersey mines, have well developed preferred orientation of
(111) planes parallel with composition layering in the ore. The origin of this widely developed fabric is ascribed to syntectonic recrystallization. Conformable, thin mylonite layers within the sphalerite aggregates, at Jersey mine, have a quite different subfabric which approaches an ideal (110) [001] pattern and has orthorhombic as opposed to axial symmetry. The mylonite textures, also developed in galena from Jersey mine, are considered to be the result of intense movement caused by simple shear during Phase 2 folding. A weak (111)[112] element within the more general (111) sphalerite fabric probably results from plastic deformation with slip being indicated within the layering and acting in a direction approximately normal to the Phase 1/Phase 2 fold axial trends. There is widespread textural evidence of minor plastic deformation in the form of slight lattice bending and subgrain development (observed in both sphalerite and galena); much of this apparently post-dates grain growth and hence annealing. However, local more intense deformation shown by kinking and deformation twinning in sphalerite alone may predate annealing, and may contribute to the observed weak (111)[112] subfabric element.

Optical determinations were also made of quartz and dolomite fabrics. Quartz c-axes display "split maximum" peripheral girdles which are slightly asymmetric with respect to foliation/lineation and are believed to reflect a general type of strain resulting from syntectonic recrystallization during Phase 1 deformation. Quartz deformation lamellae were used to derive principal stress directions which indicate that the lamellae were produced
during flattening; the orientations of the maximum principal stress axes suggesting that flattening occurred during Phase 2 deformation (Jersey mine area) and possibly during Phase 3 deformation (Reeves MacDonald mine area). Recrystallized grains from a quartz phyllonite and a quartz schist (contact metamorphosed) have c-axis subfabrics in which small-circle girdles are developed normal to the foliation which indicates that recrystallization, at least locally, was also produced by flattening. Dolomite displays c-axes patterns strongly concentrated subnormal or normal to composition layering which may also be indicative of flattening although it is not clear what the relative roles of syntectonic recrystallization and plastic deformation (with subsequent annealing) were in the development of this subfabric.

CONTACT METAMORPHISM

The conditions of contact metamorphism and some, at least, of its effects on the sulphide deposits have been investigated. The Reeves MacDonald deposit is located close to the detectable limit of a contact aureole, the H. B. deposit lies within an outer aureole (albite-epidote hornfels facies) and the Jersey deposit is located mainly within an inner aureole (hornblende hornfels facies). Only in the last case can the contact aureole be spatially related to an intrusive igneous source. Estimated temperatures attained during contact metamorphism are in the range 425-600°C (±25°) for an assumed total pressure of 1.5 kb.

Grain sizes of sphalerite and pyrite show a relatively systematic increase in diameter with increasing metamorphism. Grain shapes are typically equant (although not exclusively),
with sphalerite tending to have granoblastic texture and galena a well developed polygonal texture. Pyrite, however, develops polygonal grain boundaries only under conditions of hornblende hornfels metamorphism, whereas sphalerite, under these conditions, develops more irregular boundaries as a consequence of exaggerated grain growth. Definite changes to the (111) preferred orientation of sphalerite aggregates also begin to occur under these conditions, perhaps as a result of this exaggerated grain growth. There is some suggestion that the new annealing subfabric ultimately approaches a pattern in which (001) planes are developed parallel with the composition layering. Galena, recrystallized under these conditions, has either a random fabric or a preferred orientation of (001) planes parallel with composition layering but the origin of this is uncertain as a comparison could not be made with less highly annealed material; possibly it is related to plastic deformation on the system (001)[110].

Seventy-one analyses of minor element contents of sphalerite and pyrite illustrate certain mineralogical changes also occurring with increasing metamorphism. Thus sphalerite shows a progressive, if not very systematic, increase in iron, manganese and cadmium contents. At the highest metamorphic grade, chalcopyrite and pyrrhotite may locally be exsolved from sphalerite which in general becomes much less homogeneous in composition. Pyrite shows slight changes in nickel content with increasing metamorphism which may or may not be significant. Under the higher conditions of hornblende hornfels facies, it begins to break down to pyrrhotite and, in the Jersey deposit,
forms a sporadically developed disequilibrium assemblage: sphalerite + pyrite + pyrrhotite in which the sphalerite contains less than 13 mole % FeS.

The central conclusion arising from this study is that the sulphide deposits have indeed been deformed and that they have been involved in all phases of deformation, in regional metamorphism and in contact metamorphism. The value of applying petrofabric techniques to opaque minerals has hopefully been demonstrated; syntectonic recrystallization being an even more cryptic process in sulphides than in other rocks. Textural studies, by themselves, serve to distinguish between the effects of plastic deformation and recrystallization but can neither demonstrate the existence of, nor define the nature of, crystallographic preferred orientation, both of which are essential to the interpretation of recrystallization fabrics.
BIBLIOGRAPHY


Arnold, R. G. 1962. Equilibrium relations between pyrrhotite and pyrite from 325° to 743°C. Econ. Geol. 57, 72-90.


———. 1969. Pyrrhotite phase relations below 304 ± 6°C at less than 1 atmosphere total pressure. Econ. Geol. 64, 405-419.


Hedley, M. S. 1955. Lead-zinc deposits of the Kootenay Arc. Western Miner and Oil Review 54, 31-35.


Ohle, E. L. 1959. Some considerations in determining the origins of ore deposits of the Mississippi Valley type. Econ. Geol. 54, 769-789.


Ramsay, J. G. 1962a. The geometry and mechanics of formation of "similar" type folds. J. Geol. 70, 309-327.
Ramsay, J. G. 1962b. Interference patterns produced by the superposition of folds of similar type. J. Geol. 70, 466-481.


Whishaw, Q. G. 1954. The Jersey lead-zinc deposit, Salmo, B. C. Econ. Geol. 49, 521-529.


APPENDIX A

MAGNESIAN CALCITE-DOLOMITE GEOTHERMOMETRY

The X-ray diffraction method of Graf and Goldsmith (1955, 1958) was used to determine equilibrium temperatures of coexisting calcite and dolomite. The d-spacing (112) of magnesium calcite in the samples was measured relative to the (111) peak of an internal silicon standard (2θ = 28.466°), each sample being oscillated at 1/2° 2θ/min three times across the relevant peaks. \( \Delta d_{(112)} \) was then found for each sample by subtracting the mean \( d_{(112)} \) from \( d_{(112)} \) for pure calcite (3.036Å). From this the mole % MgCO₃ in the calcite of each sample was determined using the \( \Delta d_{(112)} \) vs. mole % MgCO₃ graph of Graf and Goldsmith (1955). The equilibrium temperature was then determined from the calcite-dolomite solvus curve of Goldsmith and Newton (1969) which extends down to 400°C (equivalent to approximately 2 mole % MgCO₃ in calcite).

The 2θ angle of the relevant peaks cannot be measured with greater accuracy than ±0.025 2θ which is approximately equivalent to ±0.8 mole % MgCO₃ or ±30°C.
Table XI. $d_{112}$ and estimated temperatures of crystallization of magnesian calcites from Reeves MacDonald and Jersey mines.

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Location</th>
<th>$d_{112}$ (Å)</th>
<th>$\Delta d_{112}$ (Å)</th>
<th>Mean $\Delta d_{112}$ (Å) ± Std. Dev.</th>
<th>Mole % MgCO₃</th>
<th>Minimum temp. (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>REEVES MACDONALD MINE</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>RM69-65</td>
<td>1900' level, Reeves zone</td>
<td>3.0297</td>
<td>0.0060</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>RM69-75</td>
<td>550' level, E. MacDonald zone</td>
<td>3.0307</td>
<td>0.0050</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>RM69-39</td>
<td>460' level, E. MacDonald zone</td>
<td>3.0269</td>
<td>0.0088</td>
<td>0.0066±0.0014</td>
<td>3.2±0.5</td>
<td>450±25</td>
</tr>
<tr>
<td>RM69-81</td>
<td>420' level, E. MacDonald zone</td>
<td>3.0292</td>
<td>0.0065</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>JERSEY MINE</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CX69-6</td>
<td>SW of Jersey town-site</td>
<td>3.0248</td>
<td>0.0109</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CX69-24</td>
<td>70G haulage drift, G-zone</td>
<td>3.0257</td>
<td>0.0100</td>
<td>0.0091±0.0019</td>
<td>4.0±0.7</td>
<td>490±25</td>
</tr>
<tr>
<td>CX69-74</td>
<td>74J rib pillar, J-zone</td>
<td>3.0261</td>
<td>0.0096</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CX69-92</td>
<td>671F pillar, F-zone</td>
<td>3.0299</td>
<td>0.0058</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

$\Delta d_{112} = d_{112}$ (pure calcite) - $d_{112}$ (sample) where $d_{112}$ (pure calcite) = 3.0357 Å
APPENDIX B

X-RAY TEXTURE GONIOMETER ANALYSIS
OF PREFERRED ORIENTATION

An X-ray analysis of preferred orientation in polycrystalline sphalerite and galena samples was carried out using a Norelco diffractometer unit with standard Phillips texture goniometer attachment belonging to the Department of Metallurgy, University of British Columbia.

The Schulz (1949a) reflection technique was used. Its application to sulphides has been explored by Gehlen (1960) and to deformed quartz aggregates by Baker et al. (1969). More general descriptions of the method are given by Cullity (1956, p. 285-292) and Barret and Massalski (1966, p. 200-203).

Sample preparation

Oriented ore samples were cut parallel with composition layering. A reference direction, usually the average L1 direction for that locality, was marked on the cut plane. Polycrystalline samples of sphalerite or galena, as homogeneous as possible in the plane, were selected and oriented, and small plane parallel slabs (approximately 15-20 mm square by 2 mm thick) prepared. The upper surfaces were then ground and finely polished. The numbers and locations of all specimens used are given in Table XIII.
Technique

The geometry of the goniometer set-up is illustrated in Figure 48. The specimen slab was oriented at an angle $\theta$ to the incident beam (so that the Bragg angle, $2\theta$, for the material being analyzed is satisfied). The incident beam (Ni filtered Cu Kα radiation generated at 35 kv and 13 mA) irradiates the specimen over a narrow rectangular area. The slab is rotated in its own plane (\(\alpha\) rotation) through 360° in 8 minutes and simultaneously tilted (\(\phi\) rotation) 5° sideways in the incident beam for each successive \(\alpha\) rotation so that the reflection geometry assumes a spiral configuration. During these rotations the slab is also oscillated in its own plane perpendicular to the incident beam direction so that a larger surface area of the specimen is irradiated. The diffracted beam is received at a geiger counter and intensities are recorded on a strip chart with a speed of 0.5 inch per minute. Scale factor and multiplier settings on the rate meter were adjusted so that the full range of the chart recorder was used. The shortest time constants, generally 8-16 seconds, which yielded relatively smooth profiles were used, as recommended by Baker et al. (1969).

Tilting (\(\phi\) rotation) causes progressive reduction in intensity of the diffracted beam due to defocussing of the primary beam. Reflection scans were made only up to tilt angles of 70° which is the normal limit for this technique because of severe reduction in intensity. In this study the specimen was irradiated over a narrow rectangular area approximately 5 mm x 1 mm, as indicated by a fluorescent target. The width of this area is believed to have caused increased defocussing and hence
Figure 48. Geometry of the Phillips X-ray texture goniometer set-up for reflection and transmission modes.
absorption of the incident beam with increasing angles of tilt. In different goniometer set-ups described by Barrett and Massalski (1966, p. 202) and Gehlen (1960), specimens were irradiated over narrower rectangular areas (e.g., 8 mm x 0.5 mm and 3.7 mm x 0.6 mm respectively).

To correct for this effect, empirical curves were established by scanning a prepared specimen of randomly oriented sphalerite and a near-random specimen of naturally occurring galena. These curves, corrected for background, illustrate the decrease in intensity of the diffracted beam with increasing tilt (Fig. 49).

Using these curves, the intensity (or pole density $q$) in the deformed specimen at any point on the spiral scan path (defined by the coordinates $\alpha, \phi$) is given by:

$$q(\alpha, \phi) = \frac{I_{\text{obs}} - I_{\text{bkg}}}{I_{\text{random}}}$$

(Baker et al., 1969)

The background corrections ($I_{\text{bkg}}$) were made by setting the specimen off the Bragg angle and recording intensities at several values of $\phi$.

Intensities are read off the chart record at $5^\circ$ intervals of $\alpha$, giving 504 data points for each reflection scan to $\phi = 70^\circ$. The values are then corrected and plotted in spiral configuration on an equal-area polar projection. These are contoured, as multiples of a uniform distribution (taken to be the mean intensity expressed as unity), to produce a pole figure.
Figure 49. Intensity correction curves established from reflection scans of random specimens.
Accuracy of the technique

Possible sources of inaccuracy in the technique have been explored by Gehlen (1960) and include those resulting from:

(a) specimen shape and misalignment
(b) grain size effects
(c) effects of different minerals in the specimen
(d) "built-in" effects—α rotation at constant angular velocity; collimator shape

These problems were treated during this study as described below.

(a) Specimen shape and misalignment

A fluorescent target was used for aligning specimens and checking that the incident beam did not leave the specimen during rotation and oscillatory translation. Alignment was also checked by testing at high tilt angles (φ) that the specimen was still on the appropriate Bragg angle.

(b) Grain size effects

i. Oscillation in the plane of the specimen helps reduce the effect of diffraction by large individual grains. In this study fine-grained sulphides (e.g., circa 100 μm sphalerite from Reeves MacDonald mine) were oscillated ±4.5 mm whereas coarser-grained sulphides (e.g., circa 300 μm sphalerite from Jersey mine) were oscillated ±7.5 mm.

ii. Haessner (1958) has shown that ideally 10⁶ to 10⁵ grains should be irradiated for representative and hence reproducible pole-figures to be obtained. This condition appears to have been satisfied for specimens of Reeves MacDonald sphalerite and Jersey galena analyzed, but not in general for specimens of
Jersey sphalerite in each of which approximately $10^3$ grains were irradiated.

(c) Effects of different minerals in the specimen

None of the specimens were monomineralic. All contained a proportion of dolomite and/or pyrite in addition to the main sulphide of interest. A few contained both sphalerite and galena. Two possible effects of these associations are: (i) interfering reflections and (ii) reduction in intensities due to relatively high absorbances of associated minerals (Table XII).

The first effect was restricted to associations of sphalerite and pyrite and was minimized as far as possible by selecting specimen slabs with the lowest pyrite concentrations (0-5% by volume).

The second effect was most serious where accessory minerals had higher linear absorption coefficients than the mineral being analyzed, e.g., galena or pyrite occurring with sphalerite. This effect has been quantified by Gehlen (1960) but its application is valid only if there are no overlapping peaks, component grains are randomly distributed, and grain size is very small in comparison with mean depth of penetration ($1/\mu$ for the material being analyzed). Only the second of these conditions could be satisfied so that in the few cases where the specimen contained both sphalerite and galena in near equal proportions, analyses were made only for the more highly absorbing galena fractions.
Table XII. Absorption coefficients, 2θ angles (after Gehlen, 1960), and other relevant data for the mineral assemblages analyzed.

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Density (g/cm³)</th>
<th>Cu Kα μ*(cm²/g)</th>
<th>Cu Kα 2θ (cm⁻¹)</th>
<th>hkl</th>
<th>Interfering reflections</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sphalerite</td>
<td>4.09</td>
<td>70</td>
<td>285</td>
<td>28.55</td>
<td>111</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>47.60</td>
<td>220</td>
<td>Pyrite</td>
</tr>
<tr>
<td>Galena</td>
<td>7.58</td>
<td>221</td>
<td>1674</td>
<td>30.08</td>
<td>200</td>
</tr>
<tr>
<td>Pyrite</td>
<td>5.02</td>
<td>200</td>
<td>1002</td>
<td>28.53</td>
<td>111</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>47.46</td>
<td>220</td>
<td>-</td>
</tr>
<tr>
<td>Dolomite</td>
<td>2.85</td>
<td>50</td>
<td>143</td>
<td>30.96</td>
<td>104</td>
</tr>
</tbody>
</table>

μ* = mass absorption coefficient  
μ = linear absorption coefficient
Dolomite occurs in many of the specimens but due to its low absorbance and lack of interfering (hkl) reflections (Table XII) its effect was ignored.

(d) Built-in effects

i. Accuracy of any single scan is reduced at high tilt angles. This is because a rotation proceeds with constant angular velocity so that the scanning rate along the spiral path increases with increasing $\phi$ tilt. Theoretically this could be rectified by modifying the goniometer so that the angular velocity of a rotation is reduced with increasing $\phi$. In practice, however, this loss in accuracy is generally accepted because there is also loss in accuracy of detail at low tilt angles due to the smoothing effect of contouring central data plotted on pole-figures.

ii. Collimator design, including port line focus slit and aperture slit dimensions, directly control the shape and size of irradiated area of specimens and hence, as has been described above, the quality of intensity profiles.

**Precision**

Reproducibility was checked by periodically re-scanning selected specimen slabs (Fig. 50b).

To check reproducibility on the microscale, a selected specimen slab which had already been scanned was deeply etched and re-scanned. The resulting intensity profiles are essentially the same (Fig. 50c). This also serves to demonstrate that no surface "working" of the specimen had occurred during polishing, a problem encountered in metallurgical studies.
Figure 50. Duplicate X-ray reflection scans for various sphalerite specimens. Shown for reference is the reflection profile obtained from randomly oriented sphalerite.
To check whether pole-figures were representative of the subfabric developed on the hand-specimen scale, several pairs of similarly oriented slabs from different layers in the same hand specimen were scanned and compared. It was found that pole-figures from any one hand specimen were approximately similar at least in the distribution of intensity maxima (Fig. 50d).

Pole-figures to $\phi = 90^\circ$

Gehlen (1960) pointed out that a single reflection scan usually suffices for the investigation of preferred orientation in ore minerals. However, attempts were made by two different techniques to produce complete pole-figures because maxima may be developed close to $\phi = 70^\circ$ for certain preferred orientations of sphalerite (Saynisch, 1970; see also Section V, Fig. 42b). Definition of such maxima would be improved in a complete pole-figure.

i. Transmission method

A thin section made from a specimen slab (previously scanned by reflection) was mounted on Scotch tape as advised by Baker et al. (1969) and set up vertically in the incident beam so that the Bragg angle was satisfied (Fig. 48). The specimen was rotated in this plane through $360^\circ$ in 16 minutes, and simultaneously translated in the incident beam. With this geometry the incident beam is diffracted by (hkl) planes perpendicular to the specimen surface--analogous ideally to reflection scanning with tilt angle $\phi = 90^\circ$.
The specimen was then tilted manually 5° for each rotation through 360°, so that a series of concentric circular scans was obtained covering the pole-figure in the range φ = 70°-90° which is inaccessible to reflection scanning. In the Phillips goniometer used for this work, a 70° tilt was the minimum mechanically possible in transmission mode. Hence, there was lack of overlap between reflection and transmission scans and it was not possible to effectively normalize and then match transmission data with the reflection data to obtain a complete pole-figure.

ii. Mutually perpendicular specimens

Attempts to produce complete pole-figures were also made using three mutually perpendicular slabs. These were analyzed by reflection scanning and a composite pole-figure produced by rotation of the data into one plane. In practise, this approach was unrewarding due to the inhomogeneous distribution (layering) of the sulphides in sections normal to composition layering, and to the increased time involved, both in X-ray scanning and in data reduction.
Table XIII. Numbers and locations of ore specimens used in X-ray fabric analysis.

<table>
<thead>
<tr>
<th>Text</th>
<th>Specimen figure</th>
<th>No.</th>
<th>Location</th>
<th>Pole figure pattern</th>
</tr>
</thead>
<tbody>
<tr>
<td>(111) Sphalerite</td>
<td>REEVES MACDONALD MINE</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>38b</td>
<td>RM69-37</td>
<td>460' level</td>
<td>central maximum</td>
<td></td>
</tr>
<tr>
<td>-</td>
<td>RM69-76</td>
<td>500' level</td>
<td>&quot; &quot;</td>
<td></td>
</tr>
<tr>
<td>38a</td>
<td>RM70-1</td>
<td>Annex, 800' level</td>
<td>&quot; &quot;</td>
<td></td>
</tr>
<tr>
<td>-</td>
<td>RM70-2</td>
<td>Annex, 800' level</td>
<td>&quot; &quot;</td>
<td></td>
</tr>
<tr>
<td>38d</td>
<td>RM70-3</td>
<td>Annex, 800' level</td>
<td>&quot; &quot;</td>
<td></td>
</tr>
<tr>
<td>38c</td>
<td>RM71-32</td>
<td>Annex, dump</td>
<td>&quot; &quot;</td>
<td></td>
</tr>
<tr>
<td>38f</td>
<td>RM71-43</td>
<td>Annex, 1000' level</td>
<td>&quot; &quot;</td>
<td></td>
</tr>
<tr>
<td>38e</td>
<td>RM71-44</td>
<td>Annex, 975' level</td>
<td>&quot; &quot;</td>
<td></td>
</tr>
<tr>
<td>H. B. MINE</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>-</td>
<td>HB70-19</td>
<td>Garnet zone, pit</td>
<td>central maximum</td>
<td></td>
</tr>
<tr>
<td>JERSEY MINE</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>40b</td>
<td>CX69-25</td>
<td>71F heading</td>
<td>two maxima</td>
<td></td>
</tr>
<tr>
<td>-</td>
<td>CX69-44</td>
<td>49D stope</td>
<td>central maximum</td>
<td></td>
</tr>
<tr>
<td>39d</td>
<td>CX69-47</td>
<td>E-zone, 5500N</td>
<td>central maxima</td>
<td></td>
</tr>
<tr>
<td>39e</td>
<td>CX69-53</td>
<td>D-zone exit</td>
<td>small-circle maxima</td>
<td></td>
</tr>
<tr>
<td>39f</td>
<td>CX69-55</td>
<td>D-zone exit</td>
<td>small-circle maxima</td>
<td></td>
</tr>
<tr>
<td>39c</td>
<td>CX69-56</td>
<td>D-zone exit</td>
<td>small-circle maxima</td>
<td></td>
</tr>
<tr>
<td>39b</td>
<td>CX69-64</td>
<td>E-zone, 5200N</td>
<td>Central maxima</td>
<td></td>
</tr>
<tr>
<td>40a</td>
<td>CX69-73</td>
<td>74J pillar</td>
<td>two maxima</td>
<td></td>
</tr>
<tr>
<td>39a</td>
<td>CX69-89</td>
<td>659F pillar</td>
<td>central maximum</td>
<td></td>
</tr>
<tr>
<td>(220) Sphalerite</td>
<td>REEVES MACDONALD MINE</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>-</td>
<td>RM71-43</td>
<td>Annex, 1000' level</td>
<td>small-circle maxima</td>
<td></td>
</tr>
<tr>
<td>41a</td>
<td>RM71-44</td>
<td>Annex, 975' level</td>
<td>small-circle maxima</td>
<td></td>
</tr>
<tr>
<td>JERSEY MINE</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>41b</td>
<td>CX69-47</td>
<td>E-zone, 5500N</td>
<td>small-circle maxima</td>
<td></td>
</tr>
<tr>
<td>(200) Galena</td>
<td>JERSEY MINE</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>-</td>
<td>CX69-45</td>
<td>C-zone</td>
<td>random</td>
<td></td>
</tr>
<tr>
<td>43d</td>
<td>CX69-53</td>
<td>D-zone exit</td>
<td>elongate central maximum</td>
<td></td>
</tr>
<tr>
<td>43c</td>
<td>CX69-69</td>
<td>537A pillar</td>
<td>weak central maximum</td>
<td></td>
</tr>
<tr>
<td>43b</td>
<td>CX69-73</td>
<td>74J pillar</td>
<td>elongate central maximum</td>
<td></td>
</tr>
<tr>
<td>43a</td>
<td>CX69-78</td>
<td>70A stope</td>
<td>near random</td>
<td></td>
</tr>
</tbody>
</table>
Sample preparation

A. Sphalerite

Samples were crushed by hand, first roughly using an iron mortar and pestle and then more finely using a refractory mortar and pestle. After sieving to -50+80 mesh (or -80+200 mesh for some fine-grained samples from Reeves MacDonald mine) the samples were panned to wash and concentrate them. Separate galena grains were efficiently removed in this way. Further concentration was effected for many of the samples by the use of Bromoform in a separatory funnel to separate dolomite, calcite and quartz. Final concentration was achieved by running samples two or three times through a Frantz Isodynamic magnetic separator. Low iron sphalerites required a cross-slope of 10-11° and an operating current of 1 amp for efficient separation. With increasing iron content the cross-slope was increased to 15-16° and the current reduced to 0.8 amps. Diopside occurring in some of the Jersey mine samples had first to be removed at a current of 0.4-0.5 amps. Finally the sphalerite concentrates were ground for 2-3 minutes in an agate mortar. Preparation of samples thereafter was organized in separate batches of 12, made up of 10 samples with one duplicate and one blank.

212
200 mg of each sample was weighed out into a beaker, dissolved in 10 ml concentrated HCl and evaporated to dryness. The residue was dissolved in 1.5 M HCl, made up to 25 ml in a volumetric flask and then transferred to a polyethylene bottle ready for analysis.

B. Pyrite

Initial preparation of pyrite concentrates was similar to that for sphalerite. After crushing (generally to -80 mesh), sieving and panning, concentrates were passed through the magnetic separator to remove sphalerite. Then leaching with hot 6 M HCl helped remove galena, dolomite, calcite, and remaining sphalerite. For a number of samples, further concentration using Bromoform in a separatory funnel was required to remove quartz and remaining dolomite. The final concentrates were ground in an agate mortar for 2-3 minutes.

Marked interferences from iron are encountered in the analysis of pyrite for cobalt and nickel (Loftus-Hills and Solomon, 1967). This was overcome by the extraction of iron from the system, using a method described by Price (1972).

25 mg of sample was weighed into a beaker and then roasted in a muffle furnace for 2-3 hours at 550°C to remove sulphur. The resulting oxide was dissolved in several ml of 6 M HCl and evaporated to dryness. The residue was redissolved in 6 M HCl and made up to 25 ml in a volumetric flask. This solution was shaken up with methyl iso-butyl ketone (MIBK) in a separatory funnel so that the iron, as Fe$^{3+}$, is taken up in the MIBK which is then separated. The sample solution is then ready for analysis.
Analytical method

Sphalerite sample solutions were analyzed for Fe, Cd, Mn, Ag, Cu and Pb. Pyrite sample solutions were analyzed for Co, Ni, Mn, Cu, Zn and Pb. Analyses were done on a Techtron AA-4 Spectrophotometer; operating procedure was as outlined by Fletcher (1970). The most sensitive absorbance line was used for each of the metals determined. Standard solutions were used to establish calibration curves for absorbance vs. concentration from which the sample concentrations were read off. All standard solutions were in 1.5 M HCl except for the silver standard which was in 3 M HCl. Corrections for background absorption (measured using a hydrogen lamp) were necessary for determinations of cobalt, nickel and lead, and were tried but found unnecessary for cadmium and silver determinations.

Silver and cobalt concentrations were very low in many samples, requiring the use of expanded scale on the indicator unit; under this condition meter stability was improved by increasing the lamp current.

The results are tabulated in Tables XIV, XV and XVIII.

Analytical precision

Duplicate analyses (Tables XVI, XVIII) indicate that analytical precision was satisfactory. Differences can probably be attributed to sample inhomogeneities (inclusions of other minerals) and sample weighing and dilution errors.
Table XIV. Minor element contents of sphalerite samples from Reeves MacDonald mine.

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Colour</th>
<th>Fe %</th>
<th>Cd %</th>
<th>Mn ppm</th>
<th>Ag ppm</th>
<th>Cu ppm</th>
<th>Pb ppm</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>RM69-17a</td>
<td>PB</td>
<td>1.21</td>
<td>0.37</td>
<td>134</td>
<td>9</td>
<td>21</td>
<td>0</td>
<td>Glory hole, Reeves zone</td>
</tr>
<tr>
<td>RM69-35</td>
<td>DY</td>
<td>0.58</td>
<td>0.35</td>
<td>38</td>
<td>7</td>
<td>23</td>
<td>2250</td>
<td>460' level, E. MacDonald zone</td>
</tr>
<tr>
<td>RM69-37</td>
<td>MYB</td>
<td>1.04</td>
<td>0.30</td>
<td>35</td>
<td>19</td>
<td>21</td>
<td>17800</td>
<td>460' level, E. MacDonald zone</td>
</tr>
<tr>
<td>RM69-60</td>
<td>MYB</td>
<td>1.38</td>
<td>0.78</td>
<td>19</td>
<td>12</td>
<td>40</td>
<td>860</td>
<td>1900' level, Reeves zone</td>
</tr>
<tr>
<td>RM69-66a</td>
<td>MYB</td>
<td>1.17</td>
<td>0.51</td>
<td>28</td>
<td>21</td>
<td>14</td>
<td>4120</td>
<td>1900' level, Reeves zone</td>
</tr>
<tr>
<td>RM69-66b</td>
<td>MYB</td>
<td>1.27</td>
<td>0.46</td>
<td>53</td>
<td>21</td>
<td>16</td>
<td>1050</td>
<td>1900' level, Reeves zone</td>
</tr>
<tr>
<td>RM69-69a</td>
<td>MYB</td>
<td>1.07</td>
<td>0.45</td>
<td>113</td>
<td>18</td>
<td>30</td>
<td>7450</td>
<td>1900' level, Reeves zone</td>
</tr>
<tr>
<td>RM69-69b</td>
<td>PB</td>
<td>1.10</td>
<td>0.46</td>
<td>93</td>
<td>14</td>
<td>31</td>
<td>6190</td>
<td>1900' level, Reeves zone</td>
</tr>
<tr>
<td>RM69-74</td>
<td>DY</td>
<td>0.91</td>
<td>0.48</td>
<td>315</td>
<td>8</td>
<td>31</td>
<td>1510</td>
<td>1900' level, Reeves zone</td>
</tr>
<tr>
<td>RM71-17</td>
<td>DY</td>
<td>0.55</td>
<td>0.16</td>
<td>278</td>
<td>2</td>
<td>6</td>
<td>15400</td>
<td>2650' level, Reeves zone</td>
</tr>
<tr>
<td>RM70-1</td>
<td>MYB</td>
<td>1.27</td>
<td>1.11</td>
<td>74</td>
<td>1940</td>
<td>51</td>
<td>2000</td>
<td>800' level, Annex zone</td>
</tr>
<tr>
<td>RM70-2</td>
<td>MYB</td>
<td>1.37</td>
<td>1.09</td>
<td>35</td>
<td>36</td>
<td>61</td>
<td>138</td>
<td>800' level, Annex zone</td>
</tr>
<tr>
<td>RM70-3</td>
<td>MYB</td>
<td>1.13</td>
<td>0.96</td>
<td>54</td>
<td>96</td>
<td>63</td>
<td>1420</td>
<td>800' level, Annex zone</td>
</tr>
<tr>
<td>RM70-4a</td>
<td>MYB</td>
<td>1.27</td>
<td>1.01</td>
<td>58</td>
<td>126</td>
<td>38</td>
<td>1300</td>
<td>875' level, Annex zone</td>
</tr>
<tr>
<td>RM70-4b</td>
<td>PB</td>
<td>1.40</td>
<td>1.11</td>
<td>39</td>
<td>46</td>
<td>63</td>
<td>69</td>
<td>875' level, Annex zone</td>
</tr>
<tr>
<td>RM70-5</td>
<td>MYB</td>
<td>0.93</td>
<td>1.25</td>
<td>123</td>
<td>109</td>
<td>41</td>
<td>410</td>
<td>875' level, Annex zone</td>
</tr>
<tr>
<td>RM70-6</td>
<td>MB</td>
<td>2.21</td>
<td>1.12</td>
<td>131</td>
<td>13</td>
<td>44</td>
<td>6</td>
<td>875' level, Annex zone</td>
</tr>
<tr>
<td>RM71-34</td>
<td>MYB</td>
<td>0.97</td>
<td>1.00</td>
<td>43</td>
<td>76</td>
<td>93</td>
<td>1140</td>
<td>Annex dump</td>
</tr>
<tr>
<td>RM71-42</td>
<td>PB</td>
<td>1.66</td>
<td>0.98</td>
<td>40</td>
<td>164</td>
<td>49</td>
<td>1500</td>
<td>1000' level, Annex zone</td>
</tr>
<tr>
<td>RM71-44</td>
<td>DY</td>
<td>0.97</td>
<td>0.91</td>
<td>31</td>
<td>112</td>
<td>100</td>
<td>2430</td>
<td>975' level, Annex zone</td>
</tr>
<tr>
<td>RM71-45</td>
<td>MYB</td>
<td>1.31</td>
<td>1.51</td>
<td>61</td>
<td>36</td>
<td>38</td>
<td>62</td>
<td>950' level, Annex zone</td>
</tr>
<tr>
<td>RM71-46</td>
<td>MB</td>
<td>1.37</td>
<td>1.16</td>
<td>59</td>
<td>34</td>
<td>40</td>
<td>120</td>
<td>950' level, Annex zone</td>
</tr>
<tr>
<td>RM71-49</td>
<td>MB</td>
<td>2.44</td>
<td>0.94</td>
<td>105</td>
<td>107</td>
<td>45</td>
<td>25</td>
<td>925' level, Annex zone</td>
</tr>
<tr>
<td>RM71-53'</td>
<td>MB</td>
<td>1.86</td>
<td>0.99</td>
<td>212</td>
<td>81</td>
<td>20</td>
<td>125</td>
<td>800' level, Annex zone</td>
</tr>
<tr>
<td>RM71-54</td>
<td>GB</td>
<td>1.84</td>
<td>0.66</td>
<td>19</td>
<td>7</td>
<td>80</td>
<td>6</td>
<td>&quot;satellite&quot; o.z., 1000' level, Annex zone</td>
</tr>
</tbody>
</table>

1Analyses by A. S. Macdonald using Techtron AA-4 Spectrophotometer
2See Table XV. for explanation of colour notation
Table XV. Minor element contents of sphalerite samples from Jersey and H. B. mines.

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Colour</th>
<th>Fe %</th>
<th>Cd %</th>
<th>Mn ppm</th>
<th>Ag ppm</th>
<th>Cu ppm</th>
<th>Pb ppm</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>CX69-22</td>
<td>DB</td>
<td>3.10</td>
<td>0.63</td>
<td>1620</td>
<td>7</td>
<td>18</td>
<td>87</td>
<td>90G bench, G-zone</td>
</tr>
<tr>
<td>CX69-25</td>
<td>BR</td>
<td>4.75</td>
<td>0.49</td>
<td>411</td>
<td>29</td>
<td>38</td>
<td>4750</td>
<td>71F heading, F-zone</td>
</tr>
<tr>
<td>CX69-42</td>
<td>BR</td>
<td>4.52</td>
<td>0.68</td>
<td>994</td>
<td>4</td>
<td>31</td>
<td>2750</td>
<td>514E pillar, E-zone</td>
</tr>
<tr>
<td>CX69-43</td>
<td>DB</td>
<td>4.34</td>
<td>0.51</td>
<td>1390</td>
<td>21</td>
<td>35</td>
<td>10600</td>
<td>512E pillar, E-zone</td>
</tr>
<tr>
<td>CX69-44</td>
<td>DB</td>
<td>5.58</td>
<td>0.50</td>
<td>536</td>
<td>0</td>
<td>100</td>
<td>1590</td>
<td>49D stope, D-zone</td>
</tr>
<tr>
<td>CX69-47</td>
<td>BR</td>
<td>4.13</td>
<td>0.70</td>
<td>558</td>
<td>0</td>
<td>24</td>
<td>680</td>
<td>5500N, E-zone</td>
</tr>
<tr>
<td>CX69-50</td>
<td>BR</td>
<td>4.47</td>
<td>0.69</td>
<td>241</td>
<td>1</td>
<td>34</td>
<td>6</td>
<td>Anticline E. limb, D-zone</td>
</tr>
<tr>
<td>CX69-53</td>
<td>BR</td>
<td>6.59</td>
<td>0.51</td>
<td>344</td>
<td>48</td>
<td>38</td>
<td>47500</td>
<td>W. wall of exit, D-zone</td>
</tr>
<tr>
<td>CX69-55</td>
<td>BR</td>
<td>4.78</td>
<td>0.69</td>
<td>204</td>
<td>13</td>
<td>7</td>
<td>2500</td>
<td>E. wall of exit, D-zone</td>
</tr>
<tr>
<td>CX69-56</td>
<td>BR</td>
<td>4.94</td>
<td>0.58</td>
<td>181</td>
<td>36</td>
<td>50</td>
<td>106</td>
<td>E. wall of exit, D-zone</td>
</tr>
<tr>
<td>CX69-64</td>
<td>MB</td>
<td>1.40</td>
<td>0.59</td>
<td>356</td>
<td>3</td>
<td>25</td>
<td>94</td>
<td>5200N crosscut, E-zone</td>
</tr>
<tr>
<td>CX69-71</td>
<td>DB</td>
<td>3.80</td>
<td>0.59</td>
<td>1110</td>
<td>45</td>
<td>20</td>
<td>13400</td>
<td>556A pillar, A-zone</td>
</tr>
<tr>
<td>CX69-89</td>
<td>BR</td>
<td>5.75</td>
<td>0.64</td>
<td>352</td>
<td>13</td>
<td>38</td>
<td>4500</td>
<td>659F pillar, F-zone</td>
</tr>
<tr>
<td>CX69-93</td>
<td>BR</td>
<td>5.84</td>
<td>0.65</td>
<td>1080</td>
<td>8</td>
<td>20</td>
<td>62</td>
<td>70G stope, G-zone</td>
</tr>
<tr>
<td>CX69-95</td>
<td>BR</td>
<td>4.53</td>
<td>0.68</td>
<td>860</td>
<td>23</td>
<td>38</td>
<td>7620</td>
<td>654F pillar, F-zone</td>
</tr>
</tbody>
</table>

H. B. MINE

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Colour</th>
<th>Fe %</th>
<th>Cd %</th>
<th>Mn ppm</th>
<th>Ag ppm</th>
<th>Cu ppm</th>
<th>Pb ppm</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>HB70-2</td>
<td>VDR</td>
<td>4.09</td>
<td>0.34</td>
<td>180</td>
<td>7</td>
<td>14</td>
<td>38</td>
<td>3200' adit dump, Garnet zone</td>
</tr>
<tr>
<td>HB70-6</td>
<td>GB</td>
<td>2.38</td>
<td>0.53</td>
<td>560</td>
<td>18</td>
<td>21</td>
<td>100</td>
<td>Open pit, Garnet zone</td>
</tr>
<tr>
<td>HB70-8</td>
<td>MB</td>
<td>2.38</td>
<td>0.35</td>
<td>615</td>
<td>10</td>
<td>16</td>
<td>25</td>
<td>Open pit, Garnet zone</td>
</tr>
<tr>
<td>HB70-9</td>
<td>PB</td>
<td>2.40</td>
<td>0.45</td>
<td>432</td>
<td>8</td>
<td>7</td>
<td>0</td>
<td>Open pit, Garnet zone</td>
</tr>
<tr>
<td>HB70-11</td>
<td>BR</td>
<td>4.45</td>
<td>0.46</td>
<td>677</td>
<td>6</td>
<td>18</td>
<td>6</td>
<td>Open pit, Garnet zone</td>
</tr>
<tr>
<td>HB70-14</td>
<td>GB</td>
<td>2.21</td>
<td>0.38</td>
<td>465</td>
<td>3</td>
<td>34</td>
<td>25</td>
<td>2800' adit dump, S side Sheep Creek</td>
</tr>
<tr>
<td>HB70-15</td>
<td>GR</td>
<td>2.81</td>
<td>0.33</td>
<td>556</td>
<td>1</td>
<td>20</td>
<td>31</td>
<td>2800' adit dump, S side Sheep Creek</td>
</tr>
<tr>
<td>HB70-19</td>
<td>PB</td>
<td>1.76</td>
<td>0.40</td>
<td>162</td>
<td>29</td>
<td>11</td>
<td>62</td>
<td>Open pit, Garnet zone</td>
</tr>
<tr>
<td>HB70-21</td>
<td>VDR</td>
<td>2.95</td>
<td>0.46</td>
<td>238</td>
<td>10</td>
<td>27</td>
<td>8500</td>
<td>Open pit, Garnet zone</td>
</tr>
</tbody>
</table>

1Analyses by A. S. Macdonald using Techtron AA-4 Spectrophotometer.
2See Table XV for explanation of colour notation
Table XVI. Duplicate analyses of minor elements in sphalerite samples from the three mines.

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Fe %</th>
<th>Cd %</th>
<th>Mn ppm</th>
<th>Ag ppm</th>
<th>Cu ppm</th>
<th>Pb ppm</th>
</tr>
</thead>
<tbody>
<tr>
<td>RM71-42</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>a</td>
<td>1.66</td>
<td>0.99</td>
<td>40</td>
<td>164</td>
<td>49</td>
<td>1500</td>
</tr>
<tr>
<td>b</td>
<td>1.63</td>
<td>1.02</td>
<td>40</td>
<td>170</td>
<td>51</td>
<td>1630</td>
</tr>
<tr>
<td>RM69-74</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>a</td>
<td>0.91</td>
<td>0.49</td>
<td>315</td>
<td>8</td>
<td>31</td>
<td>1510</td>
</tr>
<tr>
<td>b</td>
<td>0.81</td>
<td>0.49</td>
<td>313</td>
<td>9</td>
<td>34</td>
<td>1550</td>
</tr>
<tr>
<td>HB70-14</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>a</td>
<td>2.21</td>
<td>0.37</td>
<td>465</td>
<td>3</td>
<td>34</td>
<td>25</td>
</tr>
<tr>
<td>b</td>
<td>2.15</td>
<td>0.39</td>
<td>452</td>
<td>3</td>
<td>34</td>
<td>25</td>
</tr>
<tr>
<td>CX69-89</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>a</td>
<td>0.57</td>
<td>0.64</td>
<td>352</td>
<td>13</td>
<td>38</td>
<td>4500</td>
</tr>
<tr>
<td>b</td>
<td>0.58</td>
<td>0.61</td>
<td>354</td>
<td>13</td>
<td>38</td>
<td>4690</td>
</tr>
<tr>
<td>RM71-45</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>a</td>
<td>1.31</td>
<td>1.15</td>
<td>61</td>
<td>36</td>
<td>38</td>
<td>62</td>
</tr>
<tr>
<td>b</td>
<td>1.31</td>
<td>1.16</td>
<td>59</td>
<td>34</td>
<td>38</td>
<td>62</td>
</tr>
</tbody>
</table>
Sphalerite coloration

Coloration in sphalerite specimens from the three deposits shows a progressive change from dusky yellow to blackish red, apparently with increasing grade of contact metamorphism (Tables XIV, XV, XVII). Such colour variation in sphalerite is commonly attributed to differences in iron content. However, Roedder and Dwornik (1968), in an electron microprobe study of colour banding in sphalerite from the Pine Point deposit, found no correlation between colour and iron content and could not satisfactorily explain the colour banding. Graeser (1969) suggested from a study of minor elements in sphalerite of relatively low iron content, from the Binnatal deposit, that colour was strongly influenced by small variations in manganese content which might not be detectable by electron microprobe. More recently, Scott and Barnes (1972) have shown that nonstoichiometry affects sphalerite coloration and they suggest that metal deficiency, due to formation under highly sulphidizing conditions, may cause dark coloration in sphalerite rather than increases in iron content.

A plot of minor element contents of sphalerite specimens from this study against colour variation suggests a correlation between iron, and also manganese (to a lesser extent) contents and coloration (Fig. 51). Such a correlation does not preclude metal deficiency especially since increasingly high sulphidizing conditions were probably attained during contact metamorphism in H. B. and Jersey mine areas. Hence this study suggests that sphalerite coloration is possibly a function both of the nature of the metals involved in the lattice as well as of an overall metal-deficiency.
Table XVII. Sphalerite colour notation, based on Munsell Rock Colour Chart (Goddard, et al., 1963).

<table>
<thead>
<tr>
<th>Notation</th>
<th>Colour</th>
<th>Colour code</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>REEVES MACDONALD MINE</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DY</td>
<td>dusky yellow</td>
<td>5Y 6/4</td>
</tr>
<tr>
<td>MYB</td>
<td>moderate yellowish brown</td>
<td>10YR 5/4</td>
</tr>
<tr>
<td>PB</td>
<td>pale brown</td>
<td>5YR 5/2</td>
</tr>
<tr>
<td><strong>REEVES MACDONALD ANNEX</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DY</td>
<td>dusky yellow</td>
<td>5Y 6/4</td>
</tr>
<tr>
<td>MYB</td>
<td>moderate yellowish brown</td>
<td>10YR 5/4</td>
</tr>
<tr>
<td>PB</td>
<td>pale brown</td>
<td>5YR 5/2</td>
</tr>
<tr>
<td>MB</td>
<td>moderate brown</td>
<td>5YR 3/4</td>
</tr>
<tr>
<td>GB</td>
<td>greyish brown</td>
<td>5YR 3/2</td>
</tr>
<tr>
<td><strong>H. B. MINE</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PB</td>
<td>pale brown</td>
<td>5YR 5/2</td>
</tr>
<tr>
<td>MB</td>
<td>moderate brown</td>
<td>5YR 3/4</td>
</tr>
<tr>
<td>GB</td>
<td>greyish brown</td>
<td>5YR 3/2</td>
</tr>
<tr>
<td>VDR</td>
<td>very dusky red</td>
<td>10R 2/2</td>
</tr>
<tr>
<td>GR</td>
<td>greyish red</td>
<td>5R 4/2</td>
</tr>
<tr>
<td>BR</td>
<td>blackish red</td>
<td>5R 2/2</td>
</tr>
<tr>
<td><strong>JERSEY MINE</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MB</td>
<td>moderate brown</td>
<td>5YR 3/4</td>
</tr>
<tr>
<td>DB</td>
<td>dusky brown</td>
<td>5YR 2/2</td>
</tr>
<tr>
<td>BR</td>
<td>blackish red</td>
<td>5R 2/2</td>
</tr>
</tbody>
</table>
Figure 51. Plot of minor element contents vs. colour for all sphalerites (49) analyzed by AA spectrophotometry. Colour becomes darker toward the right. Colour notation is as defined in Table XVII.
Table XVIII. Minor element contents of pyrite samples from the three mines.

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Form</th>
<th>Co ppm</th>
<th>Ni ppm</th>
<th>Mn ppm</th>
<th>Cu ppm</th>
<th>Zn ppm</th>
<th>Pb %</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>REEVES MACDONALD MINE</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>RM69-18</td>
<td>a massive</td>
<td>11</td>
<td>119</td>
<td>1</td>
<td>17</td>
<td>2</td>
<td>2.18</td>
<td>S. side of Reeves glory hole</td>
</tr>
<tr>
<td></td>
<td>b massive</td>
<td>14</td>
<td>146</td>
<td>3</td>
<td>18</td>
<td>2</td>
<td>2.89</td>
<td></td>
</tr>
<tr>
<td>RM69-20</td>
<td>a massive</td>
<td>8</td>
<td>92</td>
<td>2</td>
<td>6</td>
<td>219</td>
<td>0.11</td>
<td>S. side of Reeves glory hole</td>
</tr>
<tr>
<td></td>
<td>b massive</td>
<td>12</td>
<td>94</td>
<td>2</td>
<td>7</td>
<td>247</td>
<td>0.12</td>
<td></td>
</tr>
<tr>
<td>RM69-21</td>
<td>a disseminated</td>
<td>.9</td>
<td>147</td>
<td>2</td>
<td>10</td>
<td>159</td>
<td>0.16</td>
<td>N. side of Reeves glory hole</td>
</tr>
<tr>
<td></td>
<td>b disseminated</td>
<td>.8</td>
<td>147</td>
<td>2</td>
<td>12</td>
<td>184</td>
<td>0.18</td>
<td></td>
</tr>
<tr>
<td>RM70-26</td>
<td>a disseminated</td>
<td>11</td>
<td>135</td>
<td>2</td>
<td>10</td>
<td>322</td>
<td>0.43</td>
<td>Reeves glory hole</td>
</tr>
<tr>
<td></td>
<td>b disseminated</td>
<td>11</td>
<td>142</td>
<td>3</td>
<td>11</td>
<td>322</td>
<td>0.43</td>
<td></td>
</tr>
<tr>
<td>RM69-36</td>
<td>a massive</td>
<td>12</td>
<td>217</td>
<td>2</td>
<td>10</td>
<td>77</td>
<td>0.20</td>
<td>460' level, E. MacDonald zone</td>
</tr>
<tr>
<td></td>
<td>b disseminated</td>
<td>12</td>
<td>204</td>
<td>2</td>
<td>5</td>
<td>76</td>
<td>0.19</td>
<td>1900' level, Reeves zone</td>
</tr>
<tr>
<td>RM69-77</td>
<td>a disseminated</td>
<td>9</td>
<td>265</td>
<td>2</td>
<td>10</td>
<td>269</td>
<td>0.19</td>
<td>500' level, E. MacDonald zone</td>
</tr>
<tr>
<td></td>
<td>b disseminated</td>
<td>9</td>
<td>265</td>
<td>0</td>
<td>10</td>
<td>269</td>
<td>0.20</td>
<td></td>
</tr>
<tr>
<td>JERSEY MINE</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CX69-20</td>
<td>a disseminated</td>
<td>0</td>
<td>12</td>
<td>2</td>
<td>2</td>
<td>1030</td>
<td>0.03</td>
<td>65J stope, J-zone</td>
</tr>
<tr>
<td></td>
<td>b disseminated</td>
<td>1</td>
<td>15</td>
<td>2</td>
<td>4</td>
<td>1015</td>
<td>0.03</td>
<td></td>
</tr>
<tr>
<td>CX69-23</td>
<td>a disseminated</td>
<td>11</td>
<td>57</td>
<td>3</td>
<td>38</td>
<td>937</td>
<td>0.09</td>
<td>90G2 bench, G-zone</td>
</tr>
<tr>
<td></td>
<td>b disseminated</td>
<td>8</td>
<td>72</td>
<td>4</td>
<td>40</td>
<td>1000</td>
<td>0.09</td>
<td></td>
</tr>
<tr>
<td>CX69-26</td>
<td>a massive</td>
<td>14</td>
<td>96</td>
<td>6</td>
<td>6</td>
<td>24</td>
<td>0.10</td>
<td>44C stope, C-zone</td>
</tr>
<tr>
<td></td>
<td>b massive</td>
<td>14</td>
<td>100</td>
<td>6</td>
<td>8</td>
<td>30</td>
<td>0.12</td>
<td></td>
</tr>
<tr>
<td>CX69-44</td>
<td>a massive</td>
<td>8</td>
<td>68</td>
<td>5</td>
<td>9</td>
<td>275</td>
<td>0.20</td>
<td>49D stope, D-zone</td>
</tr>
<tr>
<td></td>
<td>b massive</td>
<td>9</td>
<td>68</td>
<td>5</td>
<td>11</td>
<td>306</td>
<td>0.21</td>
<td></td>
</tr>
<tr>
<td>CX69-75</td>
<td>a disseminated</td>
<td>3</td>
<td>17</td>
<td>6</td>
<td>3</td>
<td>3060</td>
<td>0.69</td>
<td>70G stope, G-zone</td>
</tr>
<tr>
<td></td>
<td>b disseminated</td>
<td>3</td>
<td>15</td>
<td>7</td>
<td>2</td>
<td>2970</td>
<td>0.67</td>
<td></td>
</tr>
<tr>
<td>H. B. MINE</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>HB70-1</td>
<td>a massive</td>
<td>6</td>
<td>16</td>
<td>23</td>
<td>9</td>
<td>1860</td>
<td>0.51</td>
<td>3200' adit dump, Garnet zone</td>
</tr>
<tr>
<td></td>
<td>b massive</td>
<td>3</td>
<td>15</td>
<td>23</td>
<td>9</td>
<td>1860</td>
<td>0.52</td>
<td></td>
</tr>
<tr>
<td>HB70-4</td>
<td>a disseminated</td>
<td>8</td>
<td>15</td>
<td>6</td>
<td>7</td>
<td>52</td>
<td>0.02</td>
<td>3500' adit dump, Garnet zone</td>
</tr>
<tr>
<td></td>
<td>b disseminated</td>
<td>3</td>
<td>16</td>
<td>6</td>
<td>9</td>
<td>61</td>
<td>0.02</td>
<td></td>
</tr>
<tr>
<td>HB70-10</td>
<td>a massive</td>
<td>0</td>
<td>11</td>
<td>51</td>
<td>3</td>
<td>3</td>
<td>0.20</td>
<td>S. Side of open pit, Garnet zone</td>
</tr>
<tr>
<td></td>
<td>b massive</td>
<td>0</td>
<td>9</td>
<td>53</td>
<td>3</td>
<td>2</td>
<td>0.19</td>
<td></td>
</tr>
<tr>
<td>HB70-12</td>
<td>a disseminated</td>
<td>2</td>
<td>7</td>
<td>82</td>
<td>4</td>
<td>3</td>
<td>0.11</td>
<td>Old pit N. of Garnet zone</td>
</tr>
<tr>
<td></td>
<td>b disseminated</td>
<td>2</td>
<td>7</td>
<td>83</td>
<td>3</td>
<td>2</td>
<td>0.11</td>
<td></td>
</tr>
<tr>
<td>HB70-17</td>
<td>a disseminated</td>
<td>1</td>
<td>61</td>
<td>7</td>
<td>59</td>
<td>3</td>
<td>1.79</td>
<td>Old adits, W. Edge of mine area</td>
</tr>
<tr>
<td></td>
<td>b disseminated</td>
<td>3</td>
<td>61</td>
<td>8</td>
<td>139</td>
<td>2</td>
<td>1.87</td>
<td></td>
</tr>
</tbody>
</table>

1Analyses by A. Bentzen using Techtron AA-4 Spectrophotometer
2indicates zinc values >3000 ppm
APPENDIX D

ANALYSIS OF SPHALERITE BY ELECTRON MICROPROBE

Five polished specimens of sphalerite co-existing with pyrrhotite and pyrite were selected for analysis. Analyses were carried out principally by J. E. Harakal on a J.E.O.L. JXA-3A electron microprobe belonging to the Department of Metallurgy, University of British Columbia. Specimens were analyzed for zinc and iron by comparing intensities of characteristic radiations (Kα) with those of pure metal standards. Operating voltage was 25 kv. Five grains in each specimen were analyzed. Duplicate analyses were made at several different points within the grains, a minimum of 10 analyses being made on each grain. Counting time for each analysis was 10 seconds. Standards were analyzed before and after each specimen run and background determinations were made for both standards and specimen at the end of each run.

Williams (1967) has described simple matrix correction parameters which can be applied to observed intensity ratios in the microanalysis of sphalerite. These, however, apply to measurements made at 20 kv on an instrument of different geometry than the above. In this study, the data was processed for instrumental corrections (deadtime, background) and matrix corrections (back scatter, ionization-penetration, absorption and fluorescence) using a FORTRAN program, "MAGIC" (made
available by L. C. Brown, Department of Metallurgy). Sulphur concentration was determined by difference so that all analyses totalled 100%. Assuming stoichiometric distribution of sulphur, the mole % FeS and ZnS present were calculated from each analysis and totalled to obtain some measure of the accuracy of the analyses (Table XIX). These results indicate that accuracy is adequate for the purposes of this study (see Section V, p. 122). CdS and MnS contents were not determined but on the basis of mean analyses of sphalerite from Jersey mine by AA spectrophotometry, these were assumed to total 0.85 mole %.

A number of semiquantitative analyses of sphalerite specimens from all three deposits were also carried out, usually via stepwise traverses, to check their homogeneity or otherwise (see Section V, p. 126). The analyses were made for Zn and Fe, and for Fe and Cd. Fe values are generally within 10% of values determined by AA spectrophotometry on samples from the same hand specimens.

\begin{verbatim}
RM69-39 RM69-60 RM70-1 CX69-55 HB70-19
AA:  1.04  1.38  1.27  4.78  1.76  wt. % Fe
EMP:  1.08  1.33  1.18  4.75  2.00
\end{verbatim}

Similar comparisons of Cd values indicate that the semiquantitative microprobe analyses are unreliable, perhaps because of relatively low Cd concentrations. Zn was not determined by the AA method.
Table XIX. Analyses by electron microprobe of iron and zinc in sphalerite (sulphur contents determined by difference, assuming stoichiometry).

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Weight %</th>
<th>Mole %</th>
<th>Total corrected for CdS + MnS*</th>
</tr>
</thead>
<tbody>
<tr>
<td>CX69-9</td>
<td>Fe 6.89 ± 0.97</td>
<td>FeS 10.83 ± 1.52</td>
<td>101.92 102.77</td>
</tr>
<tr>
<td></td>
<td>Zn 61.04 ± 0.90</td>
<td>ZnS 91.09 ± 1.34</td>
<td></td>
</tr>
<tr>
<td></td>
<td>S 32.07 ± 0.15</td>
<td></td>
<td></td>
</tr>
<tr>
<td>CX69-77</td>
<td>Fe 8.06 ± 0.30</td>
<td>FeS 12.67 ± 0.47</td>
<td>95.97  96.82</td>
</tr>
<tr>
<td></td>
<td>Zn 55.82 ± 0.68</td>
<td>ZnS 83.30 ± 1.01</td>
<td></td>
</tr>
<tr>
<td></td>
<td>S 36.12 ± 0.72</td>
<td></td>
<td></td>
</tr>
<tr>
<td>CX69-76</td>
<td>Fe 7.63 ± 0.14</td>
<td>FeS 11.99 ± 0.22</td>
<td>98.84  99.69</td>
</tr>
<tr>
<td></td>
<td>Zn 58.20 ± 0.34</td>
<td>ZnS 86.85 ± 0.50</td>
<td></td>
</tr>
<tr>
<td></td>
<td>S 34.17 ± 0.45</td>
<td></td>
<td></td>
</tr>
<tr>
<td>CX69-53</td>
<td>Fe 7.82 ± 0.21</td>
<td>FeS 12.29 ± 0.33</td>
<td>99.32  100.17</td>
</tr>
<tr>
<td></td>
<td>Zn 58.32 ± 0.25</td>
<td>ZnS 87.03 ± 0.37</td>
<td></td>
</tr>
<tr>
<td></td>
<td>S 33.86 ± 0.11</td>
<td></td>
<td></td>
</tr>
<tr>
<td>RM69-54B</td>
<td>Fe 11.34 ± 0.77</td>
<td>FeS 17.82 ± 1.21</td>
<td>96.92  -</td>
</tr>
<tr>
<td></td>
<td>Zn 52.87 ± 0.81</td>
<td>ZnS 78.90 ± 1.21</td>
<td></td>
</tr>
<tr>
<td></td>
<td>S 35.79 ± 0.30</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*Mean CdS + MnS content of Jersey sphalerites = 0.85 mole % (by AA spectrophotometry)
APPENDIX E

MAPS OF STRUCTURAL DATA
PLANAR STRUCTURES IN THE REEVES MACDONALD MINE AREA.

LEGEND

<table>
<thead>
<tr>
<th>Feature</th>
<th>Phase 1</th>
<th>Phase 2</th>
<th>Phase 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Foliation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fold Axial Surface</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lithological contact</td>
<td>Approx./Inferred</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Slide</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Thrust</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fault</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

No exposure

Adit

Units as in Plate II.

Structural mapping by A.S. MacDonald.
Main geological boundaries slightly modified after Fyles and Hewlett (1959).

SCALE: 500 1000 1000 ft
LINEAR STRUCTURES IN THE REEVES MACDONALD MINE AREA.

LEGEND

<table>
<thead>
<tr>
<th>Lineation</th>
<th>Phase 1</th>
<th>Phase 2</th>
<th>Phase 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Minor fold axis with vergence</td>
<td>Approx./Inferred</td>
<td>Approx./Inferred</td>
<td>Approx./Inferred</td>
</tr>
<tr>
<td>Lithological contact</td>
<td>Approx./Inferred</td>
<td>Approx./Inferred</td>
<td>Approx./Inferred</td>
</tr>
<tr>
<td>Slide</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Thrust</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fault</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Units as in Plate III.

No exposure ( ), Pit ( ), Adit ( )

Structural mapping by A.S. Macdonald.
Main geological boundaries slightly modified after Fyles and Hewlett (1959).

SCALE: 0 500 1000 ft
MAJOR STRUCTURES IN THE REEVES MACDONALD MINE AREA

MAP COMPiled FROM PLATES I & II

LEGEND

ACTIVE MN. graphitic phyllite, slate, limestone.
U. LAID MN. chlorite phyllite, phyllonite.
EMERALD MN. graphitic phyllite, slate, limestone.
REEVES MN. calcite marble.
TRUSMAN MN. quartz calcite phyllite marble.
RENO MN. quartz phyllite, quartzite.
QUARTZITE RANGE MN. quartzite.
PHASE I AXIAL TRACE
PHASE 2 AXIAL TRACE
SLIDE
THRUST
FAULT
NO EXPOSURE
ABIT
PIT

SCALE

0 500 1000 ft

PLATE III.
**LEGEND**

**A. PLANAR ELEMENTS**

- Sedimentation
- Foliation
- Fold Axial Surface

**B. LINEAR ELEMENTS**

- Lineation
- Minor Fold Axis

Lithological contact
Slide
Thrust
Fault
No exposure
Pit
Adit

Units as in Plate VIII.

Structural mapping by A.S. MacDonald.
Main geological boundaries slightly modified after Fyles and Hewlett (1959).

**SCALE**

- 1 cm = 500 ft
- 2 cm = 1000 ft

PLATE VII. PLANAR (A) AND LINEAR (B) STRUCTURES IN THE N.W. MINING AREA.