Genesis of the Rosebery massive sulphide deposit, western Tasmania, Australia.

by

Neil K. Martin (BSc. Hons.)

Submitted in fulfilment of the requirements for the degree of Doctor of Philosophy

UNIVERSITY OF TASMANIA

September, 2004
Declaration

This thesis contains no material that has been accepted for a degree or diploma by the University or any other institution, except by way of background information and duly acknowledged in the thesis, and to the best of the candidate’s knowledge and belief, contains no material previously published or written by another person, except where due acknowledgement is made in the text of the thesis.

Neil K. Martin

Date:

Confidentiality

This thesis is not to be made available for loan or copying for two years following the date this statement was signed. Following that time, the thesis may be made available for loan and limited copying in accordance with the Copyright Act 1969.

Neil K. Martin

Date:
The Rosebery massive sulphide deposit, with historical production and current resources of ~32.7Mt @ 14.5% Zn, 4.4% Pb, 0.58% Cu, 145ppm Ag and 2.2ppm Au, is located in western Tasmania. It is hosted within a Middle Cambrian post-collisional succession of predominantly submarine, calc-alkaline volcanic and non-volcanic sedimentary rocks. The ore occurs as single or stacked, sulphide and barite-rich lenses with podiform to sheet-like morphologies, distributed over 3km of strike and >1.5km depth below the current surface.

The basal stratigraphic unit of the east-dipping mine sequence (the footwall volcanics) comprises a syn-eruptive succession of thick rhyolitic pumiceous mass flow units that were rapidly emplaced on the seafloor, and succeeded by rhyolitic to dacitic turbiditic mass flow and suspension settled volcanioclastic sediments of the transitional stratified volcanioclastics (TSV). The quartz and lithic content of the TSV increased with time as sediment provenance became more distal. The base of the Hangingwall Volcaniclastics was marked by a reduced volcanioclastic influx and deposition of a carbonaceous mudstone, followed by further volcanioclastic mass flow units with a distal clastic provenance. Intrusion of rhyolitic to dacitic peperite sills occurred prior and subsequent to the onset of mineralisation. The current study has shown that syn-depositional displacement of the sediments occurred as the result of basement faulting and sill emplacement, with cessation of significant fault movement prior to deposition of the Hangingwall Volcaniclastics. The mine sequence is truncated at the upper and lower margins by major Devonian reverse faults. Mineralisation is hosted within sediments at the top of the footwall volcanics and within the TSV.

Regional low-temperature diagenesis commenced upon deposition of the volcanioclastic sediments. Hydrothermal circulation through unconsolidated volcanioclastic sediments commenced prior to sediment compaction and lithification, locally preserving uncompacted vitriclastic textures. Hydrothermal alteration proximal to ore comprises a halo of quartz ± sericite ± Mn/Fe-carbonate assemblages, with a thin discontinuous chlorite assemblage located immediately beneath sulphide ore. The current study has identified a transgressive zone of intense quartz-rich alteration beneath P lens that now delineates what was a zone of fluid upflow along a syn-depositional fault. A broad halo of sericite-rich alteration envelops the ore and proximal alteration assemblages, but does not extend into the Hangingwall Volcaniclastics, as the hydrothermal system had waned prior to their emplacement.

Isotopic data indicates a hydrothermal fluid derived through the modification of seawater as it circulated through the Cambrian volcanic succession and underlying Precambrian basement, from which sulphur and metals were leached. At the mine scale, diffuse hydrothermal upflow was loosely focussed along syn-sedimentary faults defined during this study, resulting in the localisation of ore lenses along the margins of a fault-bounded basin. Significant lateral fluid flow within ~200m of the seafloor was promoted by sediment stratification, thick packages of
fine-grained sediment, peperitic sills and the development of stratiform alteration zones. Lateral movement of hydrothermal fluids was focussed along coarser, more permeable horizons within the sediments. This resulted in stratiform zones of alteration and mineralisation, and in the intense alteration of the lower margins of a peperitic quartz-feldspar-phryic sill and a thick siltstone package now overlying the northern ore lenses.

The current study has identified features within the northern part of the mine that indicate a sub-seafloor mode of ore formation, including: massive barite and sulphide lodes that are hosted within rapidly emplaced mass flow and turbidite units; ore lenses that locally transgress bedding; massive sulphide ore with the same immobile element signature as enclosing mass flow sediments; and a laterally extensive halo of alteration and disseminated mineralisation that extends for hundreds of metres laterally away from massive ore, for tens of metres into the footwall, and for several metres into the hanging wall. The siting of ore lenses stratigraphically below a peperitic quartz-feldspar phryic sill that was emplaced prior to the onset of significant hydrothermal circulation further supports a sub-seafloor mode of ore formation. The ore and the host succession do not display any textures to indicate that mineralisation occurred on the Cambrian seafloor.

Initial primary sphalerite-galena-rich mineralisation formed under relatively low-temperature (~200-250°C) and near neutral conditions, as a hot hydrothermal fluid mixed with cooler seawater and reacted with host sediments through which it passed. This mineralisation was characterised by primitive sulphide textures, including spongiform, atoll, colloform and skeletal morphologies. Utilising the laser ablation-ICP-MS analytical technique, this phase of mineralisation was found to be characterised by a low temperature trace element signature within the sulphide minerals that includes: elevated levels of Mn-Ni-As-Ag-Sb-Au-Tl-Pb and low levels of Co-Bi in pyrite; low Co:Ni and Bi:Pb ratio values in pyrite; low levels of Cu-Fe and elevated levels of Zn-Ag-Sb-Pb in sphalerite; and low levels of Sb-Ag in galena. The low temperature sulphide and barite mineralisation locally preserved an albitised volcaniclastic plagioclase component. The low temperature sphalerite-galena mineralisation was locally overprinted by and interspersed with mineralisation formed at higher temperatures, as hot hydrothermal fluids passed through permeable horizons within the ore and adjacent sediments. This phase of mineralisation produced more coarsely crystalline pyrite textures, including anhedral to euhedral grains, aggregates, and overgrowths on earlier sulphide phases. This phase of mineralisation was also characterised by a high temperature trace element signature within sulphide minerals that included: elevated levels of Bi-Sn, moderate levels of Ni-Ag-Sb-Tl-Pb-Co, and low levels of Mn-As-Au in pyrite; elevated Bi:Pb ratio values and moderate Co:Ni ratio values in pyrite; an increased Fe content in sphalerite; and galena with elevated levels of Sb-Ag. The lateral flow of high temperature (~300°C) hydrothermal fluid within the immediate footwall to sphalerite-galena ore produced discontinuous disseminated to massive pyrite-chalcopyrite mineralisation and chlorite alteration. The footwall mineralisation
was characterised by euhedral pyrite morphologies, and a high temperature sulphide mineral trace element signature comprising: elevated Co-Bi in pyrite; elevated Co:Ni and Bi:Pb ratio values in pyrite; and high levels of Fe-Mn-Cu-Sb-Bi in sphalerite.

Sphalerite-galena ore lenses are vertically and laterally zoned, with Fe-Cu enrichment at the base and proximal to interpreted zones of hydrothermal upflow, and Au-Ag-Pb-barite enrichment at the upper and lateral margins. The zonation was the due to the influence of physicochemical gradients present at the time of initial ore mineral precipitation, in combination with subsequent zone refinement processes. The vertical zonation is locally inverted.

In comparison with the northern ore lenses, a cluster of sulphide ore lenses at the southern end of the mine exhibits evidence for higher temperatures of ore formation that includes: significantly elevated average Fe and Cu contents; low average Au and Ag contents; low average δ³⁴S values (<12‰); and elevated Co, Ni and Co:Ni ratio values within footwall pyrites. This suggests that the southern ore lenses were located proximal to a major site of hydrothermal upflow at Rosebery, at the intersection of two northeast and northwest trending Cambrian faults. The peripheral ore lenses, including those at the northern end of the mine, formed at lower temperatures after mixing with a greater volume of locally circulating seawater.

Devonian deformation and metamorphism produced a pervasive cleavage, brittle-ductile faulting and folding, and syn-tectonic quartz-carbonate veining containing locally remobilised sulphides variably enriched in Au±Ag. Sulphide ore exhibits deformation fabrics at the meso- and micro-scale. Late syn- to post-deformation Devonian granite intrusion produced minor faulting and veining, and metasomatic fluids that resulted in replacement of sphalerite-galena ore with massive pyrite-pyrhhotite and magnetite-biotite ± chalcopyrite assemblages. Sulphide annealing occurred syn- to post-metasomatism. The host sequence was later subjected to brittle-ductile faulting and minor dolerite intrusion.
Acknowledgements

The first person I must thank is my wife Joanna, for her love and support during the birth of this thesis. Now I can get a haircut and a real job.

Thanks to Ross Large and Stuart Bull for their excellent supervision and guidance over the last few years. Hopefully the trace element work will be of use down the track. The other academic staff at CODES and the School of Earth Sciences were also a great help when my principal supervisors were hiding in foreign locations. Particular thanks to Bruce Gemmell, Dave Cooke, Mike Solomon, Jocelyn McPhie, Ron Berry, Garry Davidson, Marc Norman, Leonid Danyushevsky, Wally Herrmann, Cathryn Gifkins and Tony Crawford. Thanks also to Simon Stephens and Peter Cornish for their very practical assistance. To Dave Steele, thanks for keeping the beast alive against all odds.

I must thank Pasminco for their technical and financial support, and for access to a challenging ore body. I am especially grateful to Peter Edwards and Andrew McNeill for making most things possible. To Richard Fare, Susan King, Bronwyn Turner, Charles Carnie, Cam Graves, Hans Georgi and Rob Willis, all I can say is thanks for the technical support and for welcoming me into your homes. Some even invited me twice. Thanks also to Snake, Rowie, Gun and Bill for helping me find the drill core I needed, and for regularly checking to make sure I hadn’t frozen solid.

To Skakalec and the Juggernauts, namely Andrew Wurst, Darryl Clarke, Andrew Stewart, Mike Agnew and Tim Ireland, thanks for not getting me locked up in a Turkish jail. Thanks also to Alan Wilson and Glen Masterman for helping to keep me sane most of the time. A great vote of thanks to Kim Denwer for his timely assistance and for letting me occasionally win at cards.
1. Introduction
   1.1. Location
   1.2. Exploration History and Mining Status
   1.3. Previous Studies and Genetic Models
   1.4. Study Objectives
   1.5. Methodology
   1.6. Thesis Structure

2. Regional Geology
   2.1. Introduction
   2.2. Mount Read Volcanics
      2.2.1. The Sticht Range Beds
      2.2.2. The Eastern quartz-phyric sequence
      2.2.3. The Central Volcanic Complex
      2.2.4. The Western volcano-sedimentary sequences
      2.2.5. The Tyndall Range Group
      2.2.6. Andesitic to basaltic volcanics
      2.2.7. Tholeiitic mafic rocks
   2.3. Geochronology of the Mount Read Volcanics
   2.4. Granites, Deformation and Metamorphism
      2.4.1. Cambrian granites and associated porphyries
      2.4.2. The Cambrian Delamerian deformation
      2.4.3. The Devonian Tabberabberan deformation
      2.4.4. Metamorphism and Devonian granites
   2.5. Immobile Element Geochemistry of the Mount Read Volcanics
   2.6. Tectonic Setting of Western Tasmania
   2.7. Summary

3. Local geology
   3.1. Introduction
### Contents

3.2. **Geology of the Northern Central Volcanic Complex** 27  
3.2.1. The Sterling Valley and Mount Black Volcanics 29  
3.2.2. The White Spur Formation 30  
3.2.3. Intrusives 30  
3.3. **Geology of the Rosebery Mine** 31  
3.3.1. Previous work 31  
3.3.2. The footwall volcanics 33  
3.3.3. Transitional stratified volcaniclastics (TSV) 42  
3.3.4. The hanging wall volcaniclastics 46  
3.3.5. Intrusives 50  
3.4. **Structure of the Rosebery Mine** 55  
3.4.1. Bedding ($S_0$) and early compaction cleavage ($S_1$) 55  
3.4.2. Syn-depositional faulting 55  
3.4.3. Devonian deformation 57  
3.4.3.1. Cleavage 57  
3.4.3.2. Faulting and veining 58  
3.4.4. Post-Devonian deformation 61  
3.4.5. The Rosebery and Mount Black Faults 62  
3.5. **Summary** 63  

4. **Alteration: - Assemblages, Distribution Timing and Mineralogy** 65  
4.1. **Introduction** 65  
4.2. **Previous Work** 65  
4.3. **Diagenetic Alteration** 66  
4.3.1. Relict smectitic, zeolitic and feldspathic assemblages 66  
4.3.2. Chlorite and sericite pseudofiamme 70  
4.4. **Mineralisation-Related Hydrothermal Alteration** 70  
4.4.1. Sericite dominant alteration 71  
4.4.2. Quartz ± sericite dominant alteration 71  
4.4.3. Chlorite alteration 80  
4.4.4. Carbonate alteration 81  
4.5. **Devonian Alteration** 86  
4.5.1. Metamorphic and deformation-related alteration 86  
4.5.2. Granite-related metasomatic alteration 89  
4.6. **Mineral Chemistry** 95  
4.6.1. Carbonate 95  
4.6.2. Chlorite 97  
4.6.3. Sericite 100  
4.7. **Alteration Halo Geochemistry** 102
4.8. Discussion 104
4.9. Summary 104

5. Mineralisation
5.1. Introduction 107
5.2. Previous Work 107
5.3. Ore Lens Morphology and Distribution 109
  5.3.1. K lens and P lens 112
5.4. Ore Mineral Zonation 113
5.5. Ore Textures and fabrics 116
  5.5.1. Disseminated mineralisation 116
  5.5.2. Massive sulphide ore 116
  5.5.3. Lithic fragments and barite clots 119
  5.5.4. Vein-style mineralisation 120
  5.5.5. Other primary textures 125
5.6. Sulphide Mineralogy and Textures 125
  5.6.1. Pyrite 126
  5.6.2. Arsenopyrite 137
  5.6.3. Sphalerite, galena and chalcopyrite 138
  5.6.4. Tennantite, tetrahedrite and other sulpho-salts 141
  5.6.5. Gold and electrum 141
  5.6.6. Gangue mineralogy 143
5.7. Deformation-related ore remobilisation 146
5.8. Devonian Metasomatism 149
5.9. Discussion and Summary 150

6. Metal Distribution Within the Rosebery Ore
6.1. Introduction 155
6.2. Previous Work 155
6.3. Methodology 156
  6.3.1. Results 156
6.4. Discussion 163
  6.4.1. Metal associations across the ore lenses 163
  6.4.2. Application of the zinc ratio 165
  6.4.3. The southern ore lenses 168
6.5. Summary 168
7. Sulphide Trace Element Chemistry

7.1. Introduction 170
7.2. Previous Work 170
7.3. Analytical Methods 171
7.4. Comments on LA-ICP-MS Analysis 173
7.5. Results 175
  7.5.1. Pyrite 176
  7.5.2. Sphalerite 180
  7.5.3. Tetrahedrite and tennantite 183
  7.5.4. Galena 183
  7.5.5. Arsenopyrite, chalcopyrite and pyrrhotite 185
  7.5.6. Discussion 185
  7.5.7. Co:Ni in pyrite 186
  7.5.8. Bi:Pb in pyrite and sphalerite 192
  7.5.9. Tl in pyrite 194
  7.5.10. Au in pyrite 195
  7.5.11. Ag in galena, chalcopyrite and sphalerite 196
  7.5.12. Cd, Mn and Fe in sphalerite 197
  7.5.13. Trends in the tennantite-tetrahedrite series 202
7.6. Summary 203

8. Isotopes

8.1. Introduction 205
8.2. Sulphur Isotopes 205
  8.2.1. Previous studies 205
  8.2.2. Current study 207
    8.2.2.1. Aim 207
    8.2.2.2. Analytical methods 207
    8.2.2.3. Results 208
  8.2.3. Discussion and interpretation 209
    8.2.3.1. Rosebery δ34S signatures 209
    8.2.3.2. Rosebery δ34S – spatial trends 211
    8.2.3.3. Rosebery δ34S – sulphur isotope geothermometry 213
    8.2.3.4. Regional δ34S trends 215
  8.2.4. Sulphur isotope model – discussion 218
  8.2.5. Summary 222
8.3. Carbon and Oxygen Isotopes 223
  8.3.1. Whole Rock Oxygen Isotopes 226
Contents

8.4. Strontium Isotopes 227
8.5. Lead Isotopes 227
8.6. Summary 229

9. Discussion and Synthesis

9.1. Previous Genetic Models 232
9.2. Devonian of Cambrian? 232
9.3. At or Below the Seafloor? 237
9.4. Regional Scale Hydrothermal Circulation 242
9.5. Local Hydrothermal Circulation 243
9.6 Mineralisation 246
9.7. The Hydrothermal Fluid 249
9.8. Other Sub-Seaﬂoor Replacement Deposits 252
9.9. Synthesis 254

References 259
iii Figures

Figure 1.1 Rosebery location plan. 1
Figure 1.2 Genetic models - Cambrian. 4
Figure 1.3 Genetic models - Devonian. 4
Figure 2.1 Bedrock geological map of the Mount Read Volcanics. 11
Figure 2.2 Time-space diagram. 12
Figure 2.3 Location of Devonian granites beneath Rosebery mine. 20
Figure 2.4 Representative geochemical fields - Suites I to V. 21
Figure 2.5 Hypothetical tectonic development of the Tasmanian section of south-eastern Australia shown as crustal cross-sections between 600 Ma and 480 Ma. 24

Figure 2.6 Alteration cross section 1440mN. 37
Figure 2.7 Geological cross section 1320mN. 38
Figure 2.8 Geological cross section 1200mN. 39
Figure 2.9 Geological cross section 1080mN. 40
Figure 2.10 Geological cross section 960mN. 41
Figure 2.11 Transitional Stratified Volcaniclastics. 43-44
Figure 2.12 Quartz-feldspar-phyric peperite within the TSV. 51-52
Figure 2.13 Schematic of peperitic intrusive facies. 53
Figure 2.14 Cambrian syn-sedimentary fault traces. 56
Figure 2.15 Devonian deformation. 59
Figure 2.16 Macro-scale folding interpretations. 61
Figure 2.17 Post-Devonian deformation. 62
Figure 2.18 Schematic of peperitic sill emplacement. 53
Figure 2.19 Geological cross section through F lens showing distribution of granite-related replacement zones. 90

Figure 2.20 Alteration in the Rosebery mine sequence: Devonian metamatism. 91-93
Figure 2.21 Alteration in the Rosebery mine sequence: Mineralisation-related. 72-74
Figure 2.22 Alteration cross section 1440mN. 75
Figure 2.23 Alteration cross section 1320mN. 76
Figure 2.24 Alteration cross section 1200mN. 77
Figure 2.25 Alteration cross section 1080mN. 78
Figure 2.26 Alteration cross section 960mN. 79
Figure 2.27 Alteration in the Rosebery mine sequence: mineralisation-related. 82-84
Figure 2.28 Alteration in the Rosebery mine sequence: deformation-related. 87
Figure 2.29 Geological cross section through F lens showing distribution of granite-related replacement zones. 90

Figure 2.30 Alteration schematic. 105
Figure 2.31 Ore lens distribution and Cambrian fault traces. 108
Figure 2.32 Immobile element ratios - K lens ore samples. 111
Figure 2.33 Cross Section of Rosebery ore zone at ~150mN 114
Figure 2.34 Copper and zinc distribution - 15 level. 114
Figure 2.35 Disseminated mineralisation. 117-118
Figure 2.36 Banded barite and sulphide ore. 118
Figure 2.37 Banding in sulphide ore. 121
Figure 2.38 Lithic material in sulphide ore. 122-123
Figure 2.39 Pre-deformation veins. 124
Figure 2.40 Bladed mineral textures. 124
Figure 2.41 Annealed sulphide textures. 126
Figure 2.42 Pyrite textures -framboids and overgrowths. 129
Figure 2.43 Framboid size distribution. 130
Figure 2.44 Pyrite textures -spongiform and atolls. 131-132
iii Figures

Figure 5.15 Pyrite textures - colloform and skeletal. 133-134
Figure 5.16 Pyrite textures - other. 135-136
Figure 5.17 Arsenopyrite, sphalerite, galena, tetrahedrite, electrum. 139-140
Figure 5.18 Albite occurrences in sulphides and barite. 144-145
Figure 5.19 Ore remobilisation. 147-148
Figure 6.1 Major elemental distribution in K lens ore. 157-159
Figure 6.2 Major elemental distribution within P lens ore: P6 and P7-P9 sub-lenses. 160-162
Figure 6.3 Interpreted temperature gradients across northern ore lenses. 164
Figure 6.4 Zn ratio value (100Zn/(Zn+Pb)) - variation with temperature and salinity. 165
Figure 6.5 Zn ratio values - cross section view of traverse 1. 166
Figure 6.6 Zn ratio values "cut-off" values. 167
Figure 6.7 Lower Zinc ratio cut off values for northern ore lenses. 167
Figure 7.1 K lens geochemical traverses. 171
Figure 7.2 Electron microprobe backscatter images of pyrite. 173
Figure 7.3 Time-resolved LA-ICP-MS data profile. 175
Figure 7.4 Histogram of Ti values from LA-ICP-MS analysis of K lens pyrites. 175
Figure 7.5 LA-ICP-MS analysis scatter plots - pyrite. 178-179
Figure 7.6 Microprobe analysis scatter plots - sphalerite. 181
Figure 7.7 LA-ICP-MS analysis scatter plots - sphalerite. 182
Figure 7.8 Microprobe analysis scatter plots - tennantite-tetrahedrite. 184
Figure 7.9 Co vs Ni diagram for pyrite from Rosebery and Mt Lyell 187
Figure 7.10 LA-ICP-MS analysis scatter plots - Co and Ni in pyrite. 187
Figure 7.11 Southern ore lenses - ~300ppm Co in footwall pyrites. 188
Figure 7.12 Average Co vs Ni values in pyrite of sedimentary/diagenetic and submarine hydrothermal origin. 190
Figure 7.13 Average Co vs Ni values in pyrite of Devonian origin - Tasmania. 191
Figure 7.14 LA-ICP-MS analysis scatter plots - Bi and Pb in pyrite. 193
Figure 7.15 Fe vs Cu in sphalerite from various Tasmanian deposits. 198
Figure 7.16 Fe content of sphalerite vs elemental ratios and contents of pyrite. 200
Figure 7.17 Fe content of sphalerite vs Fe content of tennantite-tetrahedrite series. 200
Figure 7.18 Relationship between As-Sb and Ag-Cu in Rosebery tennantite-tetrahedrite series. 202
Figure 7.19 Fe content of sphalerite vs presence of tennantite-tetrahedrite. 202
Figure 8.1 Rosebery d34S data for principal stratigraphic units and ore mineralisation. 210
Figure 8.2 Average sulphide d34S vs metal content of major ore lenses. 212
Figure 8.3 Average sulphide d34S vs metal content of major ore lenses. 212
Figure 8.4 Sulphur isotope temperature estimates. 214
Figure 8.5 d34S data for Cambrian and Devonian deposits in western Tasmania. 216
Figure 8.6 Model of sulphur circulation in hydrothermal cell. 219
Figure 8.7 Rayleigh isotopic fractionation of Cambrian seawater. 220
Figure 8.8 Carbon and oxygen isotopes - Rosebery carbonate. 224
Figure 8.9 Carbon and oxygen isotopes - Cambrian vs Devonian systems. 224
Figure 8.10 Pb isotope data - Cambrian vs Devonian. 228
Figure 8.11 Pb isotope data - Cambrian vs Devonian. 228
Figure 9.1 Hydrothermal fluid pathways - P lens. 245
Figure 9.2 Hydrothermal fluid pathways - K lens. 245
Figure 9.3 Fluid pathways - P lens. 246
Figure 9.4 fO2 – pH diagrams for hydrothermal fluid at 250°C and 300°C. 251
Figure 9.5 Temperature vs aFe/aH2O diagram. 251
Figure 9.6 Development of the Rosebery mine stratigraphy and ore mineralisation. 255-256
iv Tables

Table 2.1  Age data for lithostratigraphic units of the Mount Read Volcanics.  17
Table 3.1  Stratigraphy of the Hangingwall Volcaniclastics for the Bastyan Dam-Rosebery-Dalmeny area.  47
Table 4.1  Alteration types and their relative timing, Rosebery-Hercules area  67
Table 4.2  Temperatures of diagenetic mineral development  68
Table 4.3  Micro-scale alteration textures  81
Table 5.1  Electrum and gold mineral associations.  142
Table 5.2  Rosebery mineralisation paragenesis.  152
Table 7.1  Electron microprobe and LA-ICP-MS results - pyrite.  176
Table 7.2  Elemental correlation trends - pyrite.  177
Table 7.3  Electron microprobe Ti analyses - pyrite.  177
Table 7.4  Electron microprobe and LA-ICP-MS analyses - sphalerite.  180
Table 7.5  Elemental correlation trends - sphalerite.  181
Table 7.6  Electron microprobe analyses - tennantite-tetrahedrite series.  183
Table 7.7  Electron microprobe analyses - galena  184
Table 7.8  Electron microprobe analyses - chalcopyrite.  185
Table 7.9  LA-ICP-MS analyses - chalcopyrite and arsenopyrite.  185
Table 7.10  Elemental ratios of Bi, Sb, Ag vs Pb - pyrite and sphalerite.  193
Table 7.11  Elemental ratios of Cd, Mn and Fe vs Zn in sphalerite - Rosebery and Devonian ores.  199
Table 8.1  Summary of sulphur isotope results from current study - K lens and P lens.  208
Table 8.2  Summary of sulphur isotope data from all units.  211
Table 8.3  Summary of sulphur isotope results from sulphide and barite lodes.  213
Table 8.4  Summary of carbon and oxygen isotope data in carbonate.  225
Table 9.1  Metal contents - Rosebery vs modern hydrothermal sulphides  249

v Appendices (volume 2)

Appendix 1.  Sample Catalogue
Appendix 2.  Whole Rock Analyses
Appendix 3.  Microprobe Analyses - Silicates
Appendix 4.  Microprobe Analyses - Sulphides
Appendix 5.  LA-ICP-MS Analyses - Sulphides
Appendix 6.  Sulphur Isotope Analyses
Appendix 7.  Geological Logs
Appendix 8.  Specific New Data Collected & Presented
Frequently Used Abbreviations etc.

**Mineral Abbreviations**
- ab - albite
- apy - arsenopyrite
- ba - barite
- sp - sphalerite
- ga - galena
- cpy - chalcopyrite
- py - pyrite
- po - pyrrhotite
- tet - tennantite-tetrahedrite
- el - electrum
- fp - feldspar
- qz - quartz
- se - sericite
- cl - chlorite
- ru - rutile
- to - tourmaline

**Thin Section Abbreviations**
- rl - reflected light
- tl - transmitted light
- xp - crossed polar light
- fov - field of view

**Mine Grid Datum**
- 0mE, 0mN, 0mRL = 378870.55mE, 5374181.69mN, -3049.49mRL (AMG)
- 0mE, 0mN, 0mRL = 0'E, 0'N, 0'RL (old grid) = 0'E, 0'N, 10,000'RL (old grid)
- Rosebery N = -11°18'10" TN
Chapter 1 : Introduction

1.1. Location

The Rosebery massive sulphide deposit has a total historical production and minable resource of ~32.7Mt @ 14.5% Zn, 4.4% Pb, 0.58% Cu, 145ppm Ag and 2.2ppm Au (P. Edwards pers. comm., 2002). It is located in western Tasmania (fig.1.1) and is hosted by the Cambrian Mount Read Volcanics, a succession that also hosts the Hercules, Hellyer, Que River and Mt Lyell sulphide deposits (Berry et al., 1998).

![Rosebery location plan](image)

Figure 1.1 Rosebery location plan.

1.2. Exploration history and mining status

Tom McDonald discovered the first indication of mineralisation in the area in 1893, in the form of alluvial gold along with boulders of zinc-lead sulphide. Steve Karlson discovered outcropping mineralisation on the Black P.A. shortly thereafter. Prospecting by the Rosebery Prospecting Association under the leadership of Tom McDonald continued for a further six months until it was concluded that an ore body was unlikely to exist. It was not until six months later that the main lode was discovered by deep trenching. The ore body was initially thought to be a Cu-Ag-Au deposit, and the Tasmanian Copper Mining Company and Primrose Mining Company were formed to work the northern and southern ends of the deposit. It was not until further development was undertaken that the ore was recognised as primarily Pb-Zn-rich. Between 1905 and 1913 lead-rich ore was railed to Zeehan for treatment, with the zinc content incurring a financial penalty by the smelters, as it was not extractable at that time (Hills, 1915; Easterbrook, 1962). From 1912 until 1915 the Tasmanian Metals Extraction Company unsuccessfully attempted metal extraction by the "bisulphate" process. The Electrolytic Zinc
Company of Australasia eventually acquired the Rosebery deposit in 1921, and full-scale production commenced in 1936 (Easterbrook, 1962). Pasminco acquired Rosebery through a corporate takeover in 1988. Pasminco refloated as Zinifex in early 2004, and are the current mine operators with a production level of ~750,000 t.p.a.

1.3. Previous studies and genetic models

More than 160 unpublished and published studies have been undertaken on some aspect of the Rosebery ore deposit and the immediate host rock succession. A summary of the relevant aspects of these studies will be presented within the introductory section of each chapter. Several key studies have however led to the development of different genetic models for the formation of primary ore at Rosebery.

Initial studies, predominantly of a regional nature, were completed by government geologists including Montgomery (1895), Harcourt-Smith (1898), Twelvetrees (1901), Waller (1902), Hills (1914; 1915; 1919), Finucane (1932a; 1932b; 1932c), and Stillwell (1934). Few geological observations were made with respect to the Rosebery mineralisation, primarily due to the lack of exposure during early mine development, and early genetic models for ore formation were based on a limited understanding of the geological context in which the ore bodies were located. Harcourt-Smith (1898), Twelvetrees (1901), Waller (1902), and Hills (1919) regarded the Rosebery mineralisation as the product of metasomatic replacement of the schistose host rocks by metalliferous solutions. Twelvetrees (1901) and Hills (1919) suggested that there was also some form of structural control on the distribution of ore lenses. Twelvetrees (1901) stated that “the alternative theory of the ore having been precipitated on a lake bottom is quite untenable”.

Finucane (1932a; 1932b; 1932c) considered the ores to have been emplaced along zones of structural weakness, with an initial phase of fluids altering the host rocks being followed by the intrusion of a viscous metalliferous fluid from which the ore crystallised. Stillwell (1934) considered the ore to have formed through replacement of quartz-sericite schist with the overlying black shale acting as an impermeable barrier to mineralising fluids. The proposed paragenetic sequence commenced with pyrite, followed by sphalerite and galena later in the ore forming process.

Changing ideas with respect to ore genesis began when Hall & Solomon (1962) noted that there was no obvious association of the ore with Devonian granites. The ores were instead interpreted to have possibly formed contemporaneously with volcanic lavas or sediments during the Cambrian, followed by a phase of redistribution and recrystallisation during the Devonian. Hall et al. (1965) were, however, not quite ready to replace the epigenetic replacement model of Stillwell (1934).
The study of Brathwaite (1969) was the most comprehensive compilation of geological observations made within the Rosebery mine until the later studies of Allen (1991; 1992; 1993a; 1993b; 1996). Brathwaite (1969; 1974) interpreted the ore to be contemporaneous with the Cambrian volcanic host sequence, having formed through chemical sedimentation of sulphides precipitating from metalliferous fluids venting under reduced conditions within a shallow marine basin. Disseminated and veinlet footwall mineralisation was interpreted to have formed through replacement of the footwall volcanics.

Solomon & Walshe (1979), incorporating the work of Sato (1972), proposed that the ore had precipitated from hot metalliferous fluids that upon venting had formed a buoyant plume in a shallow marine environment (fig.1.2). Chalcopyrite and pyrite, it was suggested, precipitated as a sulphide gel proximal to the vent as the fluids at this stage were effectively saturated with respect to Fe and Cu. Precipitation of sphalerite and galena occurred after a delay sufficient to allow entrainment of fine sulphide particles into the plume. The sphalerite and galena particles then rained down from the vented fluids as they spread laterally along the air/water interface.

Green (1983) supplemented the work of Brathwaite (1969) with further geological observations and isotopic and geochemical studies. After taking into account changing ideas on the geological context of the host sequence that suggested a deep marine setting, Green et al. (1981) and Green (1983), interpreted the ore to have formed in association with buoyancy reversal of a plume created from hot, highly saline mineralising fluids venting in a deep marine basin (fig.1.2). Initial hot fluids were thought to have led to the precipitation of chalcopyrite-pyrite rich ores proximal to the vent. Sphalerite and galena then precipitated as rising fluids cooled and mixed with seawater. The ore fluids gradually became denser than the surrounding seawater, resulting in a fluid buoyancy reversal, a commensurate collapse of the plume, and rapid deposition of the suspended sphalerite and galena particles. As no mineralised vent structure could be identified within the footwall, and as transgressive sulphide veins could rarely be identified in the ore, it was thought that hydrothermal fluids were forced to flow laterally under the depositing sulphide mound prior to venting at the lateral margins of the ore.

Sainty (1986) noted a zone of stringer pyrite-chalcopyrite mineralisation in the vicinity of F-lens, and interpreted this to be a primary vent structure. By comparison with genetic models proposed for some Kuroko ores (Sato, 1972; Ohmoto and Takahashi, 1983), the Rosebery ores were interpreted to have formed from mineralising fluids that vented at a seafloor location distal to the final ore position. The resultant dense metal-bearing fluids then migrated down slope into topographic lows and deposited the massive ore (fig.1.2).

Huston (1988), Huston and Large (1989), and Large (1990) proposed a coalesced mound model for Rosebery, based in part on genetic models developed by Eldridge et al (1983) and Campbell et al. (1984). Fluids were thought to have initially vented at several closely spaced
Figure 1.2 Genetic models - Cambrian.
Schematic summary of the genetic models proposed for Rosebery ore formation during the Cambrian. The common denominator is the deposition of sulphide mineralisation as the result of mixing a hot, upwelling metalliferous fluid with cooler seawater. (modified after Large, 1992)

Figure 1.3 Genetic models - Devonian.
Schematic summary of the deformation-related model proposed for Rosebery ore formation during the Devonian. Metalliferous fluids generated during metamorphism and deformation deposit sulphide mineralisation within zones of dilation resulting from the development of macro-scale foliation boudinage structures orientated along the X and Y axes. Image on right is a composite of mine cross sections and levels planes at the southern end of the mine. (modified after Aerden, 1992)
sites to form a series of small permeable coalesced mounds (fig.1.2). Subsequent mineralising fluids passing through the sulphide mound led to infill, recrystallisation and replacement of earlier mineralisation, with subsequent re-precipitation of remobilised ore components at the upper and lateral margins of the mound in a process termed “zone refinement”.

Allen (1991; 1992; 1993a; 1993b; 1996) mapped the volcanic and sedimentary stratigraphy and facies that comprise the Rosebery-Hercules host sequence at both a regional and mine scale. It was suggested that the Rosebery host sequence had been deposited in a deep marine setting prior to primary ore formation beneath the Cambrian seafloor. In this context, the ore was interpreted as having formed through a process of syn-volcanic, sub-seafloor replacement of permeable strata subsequent to deposition of the volcanioclastic host units (Allen, 1994a) (fig.1.2).

Aerden (1990; 1991; 1992; 1994) reverted to an epigenetic model for ore formation after a detailed microstructural study of J lens ore and the nearby Hercules deposit. Mineralising fluids were interpreted to have deposited the sulphide ores in dilational zones related to Devonian deformation, with development of macro-scale foliation boudinage structures as the major control on ore lens location and morphology (fig.1.3).

Solomon et al. (1990) and Solomon & Groves (1994), utilising the genetic model developed by McDougall (1984), interpreted the ore to have formed within a brine pool developed within a restricted basin environment (fig.1.2), with a low point at the basin margin acting as a weir for mineralising fluids. Hot, saline solutions venting within the basin mixed with less dense seawater, producing a negatively buoyant fluid which ponded within the basin. At initially low temperatures only Zn and Pb were carried in solution thereby leading to sphalerite-galena precipitation. As the fluid temperature increased a buoyant plume again formed carrying copper, lead, zinc and gold in solution. Chalcopyrite and pyrite precipitated from the fluid whilst Zn and Pb were carried out of the basin by the overflowing fluid. As the system cooled only Zn and Pb were carried in solution and subsequently precipitated from the ore fluids as they traversed the basin. Later ore fluids may have contained partially reduced seawater sulphate or mixed with seawater sulphate to precipitate barite.

1.4. Study objectives

The overall aim of this study was to evaluate the competing genetic models proposed by previous workers for the Rosebery deposit, and to develop a model for ore genesis that takes into consideration all of the evidence currently available. To this end several key questions were to be resolved during this study.

- What evidence exists to demonstrate a Cambrian or Devonian timing for primary ore mineralisation?
What controls on ore lens location were imposed by stratigraphic elements such as host lithology, grain size, and intrusive or volcano-sedimentary facies, or by other factors such as seafloor topography?

Can syn-sedimentary faults be recognised and if so, did they control ore lens location?

Did post-depositional structures control primary mineralisation? If not, what effects did post-mineralisation deformation and metamorphism have on the primary ore?

Does the textural and geochemical evidence indicate a replacement or exhalative mode of ore genesis?

What stratigraphic or structural controls were responsible for ore lens repetition?

What role did the quartz-porphyry sill at the northern end of the mine play in the formation of the primary ore?

What evidence exists, for example metal or isotope zonation, to demonstrate ore-fluid pathways and fluid chemistry controls on metal transport and ore deposition?

What was the likely source of the primary ore-fluids, and what effects did Devonian metasomatism have on the northern ore lenses?

1.5. Methodology

The current study was focussed on ore lenses at the northern end of the Rosebery mine that have been gradually developed and mined since initial discovery in 1991. This part of the mine was selected, as the ore lenses were less affected by Devonian metasomatism and have not been subjected to previous detailed study.

Initial work required the logging of ~15,000m of diamond drill core for the purpose of recording primary volcanic and sedimentary features, post-depositional alteration, textural and mineralogical features associated with ore mineralisation, and structural elements evident within the ore and the host rocks. The broad aim was to establish a stratigraphic framework for the northern end of the mine that could then be utilised to place the various phases of mineralisation, alteration and deformation into their correct context. Underground and surface exposures were utilised to place into context many of the features observed within the diamond core.

To aid primary rock identification a series of rock and ore samples were subjected to whole rock major and trace element analysis. A series of thin sections were also prepared and examined, with electron microprobe analysis undertaken on some specimens to aid in determining silicate mineral compositions.

Polished thin sections and resin mounts were prepared from mineralised specimens and
microscopically examined, primarily with the aim of determining a paragenetic sequence for sulphide formation. Many of the same sulphide sections and mounts were also utilised in a geochemical study that examined the trace element signature of individual sulphide minerals. A range of sulphide minerals and barite were collected for sulphur isotope analysis.

Data from the mine grade control and development drill datasets were examined and collated with the aim of defining the distribution of geological and structural features in long section across the whole mine. Ore grade control assay data was also compiled, in part for the determination of mineral zonation within individual ore lenses.

Geochemical datasets were compiled from published and unpublished sources and integrated with new data generated during the current study. This included whole rock major and trace element analyses, silicate and carbonate mineral microprobe analyses, sulphide mineral microprobe analyses, sulphur isotope analyses, and carbon and oxygen isotope analyses.

Previous published and unpublished studies were reviewed in the light of observations made during the current study.

1.6. Thesis structure

The first chapter of this thesis is aimed at introducing the Rosebery deposit, the exploration and mining history, the extent of previous studies, and the approach taken during this study.

The second chapter of this thesis comprises a summary description of the Cambrian Mount Read Volcanics, the succession that hosts the Rosebery deposit, and a tectonic context for the emplacement of this succession.

The third chapter is a more detailed examination of the northern part of the Mount Read Volcanics, and the lithologies, stratigraphy and volcanic facies that constitute the host succession across the northern end of the mine. A detailed stratigraphic framework is constructed for the first time across the northern end of the mine, which in turn establishes a geological context for the alteration, mineralisation and structural elements that are examined later. In particular, the timing of peperitic sill emplacement within the hanging-wall to the northern ore lenses is clearly established, placing an important constraint on the timing and location of sulphide ore formation. Identified for the first time, are a series of large-scale growth faults that appear to have controlled local basin development, host sedimentation, peperitic sill emplacement and the site of ore development during the Cambrian.

Chapter four examines alteration assemblages across the northern end of the mine, with a focus on timing, mineralogy and spatial extent. Placing these alteration assemblages into a
geological context has facilitated the establishment of a paragenetic sequence for the alteration assemblages related to primary mineralisation, again placing important constraints on the timing of ore formation. Identified for the first time is a zone of transgressive alteration beneath the northern ore lenses that is interpreted to be a conduit for hydrothermal fluid upflow directly related to barite and sulphide ore formation. The elemental chemistry of some gangue minerals is also examined.

Chapter five examines the ore mineralisation in detail, with a focus on micro- and meso-scale textural evidence within the ore as an indicator of ore genesis. The northern ore lenses were the focus of most work, as they are significantly less affected by Devonian deformation or subsequent granite-related metasomatism than the southern ore lenses. This has permitted a detailed examination of relict primitive sulphide textures to be undertaken for the first time at Rosebery. Controls on mineralogical zonation and ore lens distribution are examined at the mine scale, and are related to the distribution of mine-scale growth faults interpreted as part of this study. The immobile element composition of K lens ore and surrounding sediments is examined, with important implications for primary ore genesis. Ore gangue mineralogy is examined with a particular emphasis on albite, with the aim of placing constraints on the physicochemical factors controlling ore formation. The effects of deformation and metasomatism on primary ore are also examined.

Chapter six examines ore mineralogy and chemistry in more detail, with the aim of establishing elemental zonation at the ore lens scale. The mineral zonation within K lens and P lens ores is examined in the context of interpreted Cambrian growth faults, with a view to outlining hydrothermal fluid pathways active during ore formation.

Chapter seven examines the major and trace elements within different sulphide minerals. Previous workers have struggled to achieve sufficiently detailed spatial resolution and low detection limits required for this type of study at Rosebery. The use of laser ablation - inductively coupled - mass spectrometry (LA-ICP-MS) has for the first time permitted a comprehensive study of the trace element characteristics of sulphide minerals at Rosebery. This important study has aided in establishing some of the physicochemical factors that played a role in producing different sulphide textures, compositions and distributions.

Chapter eight summarises the isotopic signature of mineralisation and associated alteration at Rosebery, with a particular emphasis on the sulphur isotope data. Variation in sulphur isotope values across the entire mine is presented for the first time, with important implications for the nature of hydrothermal fluid circulation and ore formation.

Chapter nine examines the various genetic models proposed for the Rosebery deposit in the context of observations made during both the current and previous studies. Modern and
ancient analogues are utilised to indicate the likely nature of hydrothermal fluid circulation at the regional and mine scale, and to indicate the types of processes that led to development of ore mineralisation. A preferred model is presented based on the observed geology, alteration patterns, ore textures, mineral chemistry, whole rock chemistry and isotopes. A detailed thermodynamic model was not completed as part of this study as there is a lack of material with suitable fluid inclusions that might shed more light on the properties of primary mineralising fluids. Thermodynamic modelling has been undertaken in the past by various workers (e.g. Green et al., 1981; Green, 1983; Huston, 1988; Huston and Large, 1989) and evidence collected during this study does not warrant major changes to those models.
Chapter 2: Regional Geology

2.1. Introduction

The Rosebery massive sulphide deposit is hosted within the Central Volcanic Complex, an element of the Mount Read Volcanics (Campana and King, 1963) that was emplaced along the eastern margin of the Dundas Trough during the late Middle Cambrian (Corbett, 1992; Perkins and Walshe, 1993).

The Dundas Trough (fig. 2.1) comprises a succession of Late Proterozoic to Cambrian volcanic and sedimentary rocks underlain and laterally bounded by deformed Precambrian rocks of the Tyennan region to the east and the Rocky Cape region to the northwest (Corbett, 1992). The earliest units deposited in the Dundas Trough are represented by the Success Creek Group, which crops out along its western margin. This group comprises a ~1km thick sequence of sandstone and carbonate rocks which unconformably overlie Precambrian basement, and are interpreted to have been deposited in a shallow water shelf environment (Corbett, 1992). The Success Creek Group is conformably overlain by the Crimson Creek Formation, a ~4km thick sequence of volcaniclastic turbidites and mudstones and associated basaltic lavas (Brown et al., 1989).

Fault-bounded mafic-ultramafic complexes, comprising ultramafic cumulates, low-Ti basalt and boninitic andesite, occur along the western margin of the Dundas Trough. These were tectonically incorporated with the Crimson Creek Formation prior to unconformable middle Cambrian sedimentation, as evidenced by the presence of detrital magnetite and serpentinite pebbles in basal units of the Western volcano-sedimentary sequences (Rubenach, 1974; Brown, 1986; Berry and Crawford, 1988; Brown et al., 1989; Corbett, 1992).


The Owen Conglomerate conformably to unconformably overlies the Mount Read Volcanics,
Figure 2.1 Bedrock geological map of the Mount Read Volcanics (modified after Corbett, 2002)
and comprises a thick late Cambrian to early Ordovician sequence dominated by siliciclastic alluvial conglomerate and shallow-marine quartz sandstone (Corbett and Lees, 1987; Noll and Hall, 2003). The sediments were derived from Precambrian metasediments to the east and deposited in fault-controlled grabens along the margin of the Tyennan Block (Corbett and Lees, 1987).

### 2.2. Mount Read Volcanics

The Mount Read Volcanic belt comprises an eastern-central part made up largely of volcanic and intrusive rocks, and a broader western part dominated by volcano-sedimentary sequences. On the basis of lithological association the Mount Read Volcanics have been subdivided into the Sticht Range Beds, the Eastern quartz-phyric sequence, the Central Volcanic Complex, the Western volcano-sedimentary sequences and the Tyndall Group (Corbett, 1992).

The lithostratigraphy of the Mount Read Volcanics differs either side of the north-northeast trending Henty fault (fig.2.2), a regional structure active during the Cambrian (Berry, 1989). South and east of the Henty fault the Central Volcanic Complex, Eastern quartz-phyric sequence, Western volcano-sedimentary sequences and Tyndall Group are all present, whereas the Eastern quartz-phyric is absent north and west of the fault (Corbett, 1992, 2002).

**Figure 2.2 Time-space diagram.**

Figure shows the distribution of major lithostratigraphic units in the Mount Read Volcanics to the north and west, and the south and east of the Henty Fault. CVC = Central Volcanic Complex, HFW = Henty Fault Wedge. (modified after Corbett, 1992; Selley, 1997; Gifkins, 2001; and Corbett, 2002)
2.2.1. The Sticht Range Beds

The Sticht Range Beds comprise a sequence of siliciclastic pebble-cobble conglomerate, sandstone and interbedded siltstone, with clastic material mainly derived from Precambrian basement and to a lesser degree from Cambrian volcanics. The sequence is upwardly fining, sits unconformably on Precambrian basement, and is interpreted to represent a transition from fluvial to deeper marine environments (Baillie, 1989; Corbett, 1992).

2.2.2. The Eastern quartz-phyric sequence

The Eastern quartz-phyric sequence is a complex series of quartz-feldspar-phyric lavas (rhyolites, dacites and minor andesites) and volcaniclastic rocks, with abundant quartz-feldspar±biotite porphyry intrusions and related granitoids. Similar porphyritic intrusions cut Precambrian basement, indicating that the volcanics were erupted through and onto continental basement.

This sequence overlies the Sticht Range Beds to the east and north of Mt Murchison, and in turn interfingers with rocks of the Central Volcanic Complex (CVC). To the northwest of Mt Murchison the Eastern quartz-phyric sequence is overlain by the Farrell Slates of the Western volcano-sedimentary sequences (Corbett, 1992; Pemberton and Corbett, 1992; Corbett, 2002). Several units of bedded sandstone, siltstone and shard-rich mudstone occur within the sequence, some with a Precambrian-derived clastic component (McNeill and Corbett, 1992), suggesting deposition under subaqueous conditions (Corbett, 1992). The Eastern quartz-phyric sequence is intruded by the Cambrian Murchison granite in the vicinity of Mount Murchison (Corbett, 1992; Pemberton and Corbett, 1992).

2.2.3. The Central Volcanic Complex

The Central Volcanic Complex (CVC) is a sequence of predominantly feldspar-phyric volcanic rocks, with abundant rhyolitic-dacitic lavas, pumiceous volcaniclastic rocks and massive dome-like lava bodies (Corbett and Lees, 1987; Corbett, 1992). The presence of hyaloclastites and extensive subaqueous pumiceous mass flows indicates that the sequence was deposited in a predominantly subaqueous environment (Allen and Cas, 1990).

South of the Henty fault the CVC comprises feldspar-phyric, rhyolitic to dacitic lavas, syn-volcanic intrusions and minor pumiceous units (Corbett, 1992; Pemberton and Corbett, 1992). The eastern margin of the southern CVC interfingers with the Eastern quartz-phyric sequence and is overlain by the Tyndall Group, or by the Anthony Road andesite in the vicinity of the Tyndall Range. To the west the sequence is flanked by the Western volcano-sedimentary sequences, in the form of the Yolande River sequence (Corbett, 1992). The Darwin Granite,
which intruded the CVC during the Cambrian, crops out at Mt Darwin (Corbett and Lees, 1987; Corbett, 1992).

To the north and west of the Henty fault the CVC predominantly comprises feldspar-phyric rhyolitic to dacitic lava, syn-volcanic intrusions and syn-eruptive pumiceous volcaniclastic units (Lees, 1987; Corbett et al., 1989a; Allen and Cas, 1990; Allen, 1991). Basaltic rocks of the Sterling Valley Volcanics are present in the lower part of the CVC stratigraphy, the main component of which is a >1km thick package of basaltic to dacitic mass flow units, lavas and sills (Crawford et al., 2000; Gifkins, 2001). The northern CVC also contains quartz-feldspar-phyric intrusions, and dolerite dikes of the Henty Dike Swarm (Corbett et al., 1989a).

The northern part of the CVC interfingers with, and is both conformably and disconformably overlain by, the Dundas and Mt Charter Groups of the Western volcano-sedimentary sequences (Corbett and Lees, 1987; Corbett, 1992; Pemberton and Corbett, 1992). The Rosebery, Hercules, Hellyer and Que River ore bodies are hosted within the upper part of the Central Volcanic Complex (Corbett, 2002).

2.2.4. The Western volcano-sedimentary sequences

The Western volcano-sedimentary sequences are an extensive succession of volcano-sedimentary rocks located along the western and northern margins of the CVC. The sequence contains marine fossils in places, and typically comprises interbedded tuffaceous mass flow deposits, turbiditic sandstone, shard-rich tuffaceous mudstone, micaceous siltstone and graphitic black shale. Andesitic lavas and breccias, minor felsic lavas and large intrusive porphyritic bodies also occur in the sequence (Corbett, 1992). The sediments commonly have a volcanic provenance, but also interfinger with sediments of Precambrian derivation (Corbett, 1992; Pemberton and Corbett, 1992). The sequence is interpreted to have been deposited in a sub-wave base marine setting (Corbett, 1992; McPhie and Allen, 1992).

Corbett (2002) recently subdivided Cambrian volcano-sedimentary sequences of the Dundas Trough into pre-Tyndall Group and Tyndall Group units, in the process reassigning stratigraphically higher parts of the Western volcano-sedimentary sequences as previously defined (Corbett, 1992) to the Tyndall Group and its correlates (fig.2.1). North of the Henty fault the Western volcano-sedimentary sequences comprise the lower parts of the Dundas and Mount Charter Groups, whereas south of the fault the sequences are represented by the Yolande River sequence. To the west the Western volcano-sedimentary sequences overlay rocks of the Crimson Creek Formation, and to the east they interfinger the Central Volcanic Complex (Corbett, 1981, 1992; Pemberton and Corbett, 1992).
2.2.5. The Tyndall Group

The Tyndall Group is located along the eastern margin of the Mount Read Volcanic belt and comprises predominantly crystal rich sandstone, volcanic breccia and volcanic conglomerate intercalated with minor welded rhyolitic ignimbrite, felsic to intermediate lavas and intrusions, and non-volcanic sedimentary rocks including limestone, mudstone and sandstone (White and McPhie, 1996). The lower subdivision of the Tyndall Group, the Comstock Formation, comprises mainly crystal-rich sandstone, volcanic breccia and laminated mudstone with a volcanic provenance. The upper subdivision, the Zig Zag Hill Formation, comprises mainly polymict volcanic conglomerate and sandstone of mixed volcanic and Precambrian basement provenance, with minor welded ignimbrite and crystal-rich sandstone. The Tyndall group is interpreted to have been deposited in a below-storm-wave-base submarine environment, with some elements interpreted as having been deposited in a shallower marine environment based on the presence of fossiliferous limestone and welded ignimbrite (White and McPhie, 1996; 1997).

Significant parts of the Western volcano-sedimentary sequences within the Dundas Trough have recently been recognised as Tyndall Group correlates (White and McPhie, 1996; Corbett, 2002), including the upper part of the White Spur Formation in the Rosebery area (Corbett, 2002). The Tyndall Group and it’s correlates represent a continuous sequence that originally extended across much of the Dundas Trough, with lithological variation attributable to localised basins and sediment sources (Corbett, 2002). The Tyndall Group and its correlates are characterised by predominantly quartz-feldspar-phyric detritus derived from the Mount Read belt. Although the Tyndall Group on the eastern side of the Mount Read belt appears to lack input from mafic-intermediate volcanism, to the west there appears to have been major intermediate-mafic volcanic sources in Henty Fault Wedge (Corbett, 2002).

The Tyndall Group is interpreted as the youngest part of the Mount Read Volcanic sequence, on the basis of regional stratigraphic relationships and biostratigraphic and isotopic U-Pb dating (White and McPhie, 1996; Corbett, 2002). The group is conformably overlain by the Owen Conglomerate in places, and has variable basal contact relationships with other units of the Mt Read Volcanics. Close to the Cambrian granites the group contains rounded granite clasts, indicating a younger age than the granites, however the same granites can be seen to have intruded older lithostratigraphic units in the Mount Read Volcanics (White and McPhie, 1996).

The Cu-Au mineralisation and associated alteration at Mt Lyell transgress rocks of the Central Volcanic Complex, and the overlying exhalative Pb-Zn mineralisation is hosted within the base of the Tyndall Group (Corbett, 2001, 2002). The high-grade Au mineralisation of the Henty and Mount Julia ore bodies, located ~25km to the north, are also located at the base of the Tyndall
2.2.6. **Andesitic to basaltic volcanics**

Separate lenses and masses of calc-alkaline andesitic to basaltic volcanics and associated intrusives occur within or between the other lithostratigraphic associations of the Mount Read Volcanics (Corbett, 1992). The rocks typically comprise interbedded flows and breccias associated with small to large intrusive bodies, and the lavas are typically either plagioclase-pyroxene-phyric or plagioclase-hornblende-phyric (Corbett, 1992). The presence of pillows and an association with marine sediments indicates emplacement in a submarine environment (Corbett, 1992; McPhie and Allen, 1992).

2.2.7. **Tholeiitic mafic rocks**

Several occurrences of tholeiitic rocks within the Mount Read Volcanics have been described by Corbett (1992). The Miners Ridge basalt comprises basaltic lavas and breccias, which may represent either upthrust basement or early tholeiitic volcanism associated with evolution of the Mount Read Volcanic belt. The Henty fault wedge basalts and associated rocks comprise a series of tholeiitic pillow lavas cut by basaltic dikes, suggesting emplacement in a submarine environment proximal to an eruptive centre. Interfingering of the Henty fault wedge tholeiitic lavas with calc-alkaline andesitic volcanic rocks in the western part of the wedge suggests contemporaneous eruption of both units. The Henty dike swarm comprises numerous basaltic to doleritic and gabbroic dikes focussed along the Henty and North Henty Faults, and intruding rocks of the northern Central Volcanic Complex.

2.3. **Geochronology of the Mount Read Volcanics**

A Middle Cambrian age has been attributed to the Mount Read Volcanics on the basis of biostratigraphic evidence (Corbett, 2002). Ages derived from U-Pb isotopes give a mean age of 502.6 ±3.5 Ma (table 2.1), although some dates for the Central Volcanic Complex and Tyndall Group appear marginally younger (Perkins and Walshe, 1993). This suggests that the Mount Read Volcanics were emplaced during a relatively short period of time in the late Middle Cambrian (Perkins and Walshe, 1993).

The isotopic U-Pb dates are regarded as imprecise due to complications arising from possible lead loss (Adams et al., 1985), and K-Ar and Rb-Sr techniques returned ages which appear younger than biostratigraphic ages due to partial thermal resetting during Late Cambrian/Early Ordovician and Devonian metamorphic events (Adams et al., 1985; Perkins and Walshe, 1993; Everard and Villa, 1994).
Chapter 2

The U-Pb isotopic ages and the interfingering relationship of the Central Volcanic Complex with the Eastern quartz-phyric sequences south of the Henty Fault, and with the Western Volcano-sedimentary sequences north of the Henty Fault, suggest that these units are broadly contemporaneous (Corbett and Lees, 1987; Corbett, 1992; Pemberton and Corbett, 1992).

The younger U-Pb isotopic ages returned for the Tyndall Group are consistent with the unconformably overlying relationship of this unit with respect to the main part of the Mount Read Volcanics and the Darwin Granite (Corbett, 1992; Perkins and Walshe, 1993; White and McPhie, 1996).

### 2.4. Granites, deformation and metamorphism

The Mount Read Volcanics were subjected to deformation during the Middle to Late Cambrian Delamerian Orogeny (Berry, 1986; Corbett and Lees, 1987; Crawford and Berry, 1992). A more extensive phase of deformation also occurred during the Early to Middle Devonian Tabberabberan Orogeny, which resulted in much of the evidence for earlier deformation being obscured (Berry, 1986; Williams et al., 1989).

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Age</th>
<th>Dating Method</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Owen Conglomerate</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Newton Creek Sandstone</td>
<td>Late Cambrian</td>
<td>biostratigraphic³</td>
</tr>
<tr>
<td><strong>Tyndall Group</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mount Julia Member, Comstock Fm</td>
<td>494±3.8 Ma</td>
<td>U-Pb in zircon⁶</td>
</tr>
<tr>
<td>volcaniclastic sandstone, Comstock Fm</td>
<td>502.5±3.3 Ma</td>
<td>U-Pb in zircon⁶</td>
</tr>
<tr>
<td>Comstock Limestone</td>
<td>late Middle Cambrian</td>
<td>biostratigraphic²</td>
</tr>
<tr>
<td>Winter Brook lava</td>
<td>500.4±6.9 Ma</td>
<td>U-Pb in zircon⁶</td>
</tr>
<tr>
<td>Link Road ignimbrite</td>
<td>505.3±3.6 Ma</td>
<td>U-Pb in zircon⁶</td>
</tr>
<tr>
<td><strong>Crown Hill Andesite</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>489±9 Ma</td>
<td>⁴⁰Ar/³⁹Ar in hornblende⁶</td>
</tr>
<tr>
<td><strong>Anthony Road Andesite</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>502.2± 3.5 Ma</td>
<td>U-Pb in zircon⁶</td>
</tr>
<tr>
<td></td>
<td>501.5±5.7 Ma</td>
<td>⁴⁰Ar/³⁹Ar in hornblende⁶</td>
</tr>
<tr>
<td><strong>Murchison Granite</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>501±5.7 Ma</td>
<td>⁴⁰Ar/³⁹Ar in hornblende⁶</td>
</tr>
<tr>
<td><strong>Darwin Granite</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>510(+61,-21) Ma</td>
<td>U-Pb in zircon⁶</td>
</tr>
<tr>
<td><strong>Western volcano-sedimentary sequences</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Southwell Subgroup</td>
<td>503.2±3.8 Ma</td>
<td>U-Pb in zircon⁶</td>
</tr>
<tr>
<td>Que River Shale</td>
<td>late Middle Cambrian</td>
<td>biostratigraphic³</td>
</tr>
<tr>
<td>dacite, Que River Footwall</td>
<td>~500Ma±16 Ma</td>
<td>U-Pb in zircon⁶</td>
</tr>
<tr>
<td><strong>Central Volcanic Complex</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Jukes lava?</td>
<td>503.3±6.9 Ma</td>
<td>U-Pb in zircon⁶</td>
</tr>
<tr>
<td>dacitic sill, Mt Black Volcanics</td>
<td>494±4.3 Ma</td>
<td>U-Pb in zircon⁶</td>
</tr>
</tbody>
</table>

Table 2.1 Age data for lithostratigraphic units of the Mount Read Volcanics

2.4.1. Cambrian granites and associated porphyries

Cambrian granite exposures in western Tasmania include the Elliott Bay, Timbertops, Darwin, Murchison and Dove granites (Leaman and Richardson, 1989). Regional aeromagnetic and gravity surveys indicate that the Cambrian granite exposures form part of a semi-continuous series of intrusive bodies extending over 60km along the eastern margin of the CVC (Leaman and Richardson, 1989; Large et al., 1996). The largest, the Murchison and Darwin granites, are exposed along the eastern margin of the Mount Read Volcanics (fig.2.1) where they intrude the Central Volcanic Complex (Corbett, 1992). The Darwin and Murchison granites are strongly altered, high-K, magnetite-series granites emplaced at depths estimated to have been between 3 and 7km (Large et al., 2001c; Wyman, 2001). The Murchison granite varies in composition from diorite to granite (58-74wt% SiO₂) (Polya et al., 1986), whereas the Darwin granite comprises two highly fractionated intrusive phases comprising granite and granodiorite (74-78wt% SiO₂) (Jones, 1993; Wyman, 2001).

An age of 501 ± 5.7 Ma (40Ar/39Ar - hornblende) for the Murchison granite, and the relationship between the Tyndall Group and the Darwin granite, indicates that intrusion was broadly synchronous with emplacement of the Mt Read Volcanics (Perkins and Walshe, 1993; Large et al., 1996).

2.4.2. The Cambrian Delamerian deformation

Evidence for Cambrian deformation within the Mount Read Volcanic sequence is present in the form of local and regional unconformities and growth faults (Corbett and Lees, 1987; Corbett et al., 1989b; Corbett, 1992; Berry and Keele, 1993). Clasts of randomly orientated and cleaved volcanic rock in the Tyndall Group support Late Cambrian deformation, as do clasts of Cambrian granite in the Tyndall group which indicate tectonic uplift and unroofing (Corbett et al., 1989b; Corbett, 1992). Regional scale structures active during the Cambrian include the Henty fault, the Great Lyell Fault and a proto-Rosebery Fault located several kilometres west of the current fault trace (Berry, 1989; Berry and Keele, 1993).

On the basis of work throughout western Tasmania, Berry and Keele (1993) related formation of the Dundas Trough and emplacement of the Mount Read Volcanics to the Delamerian orogeny (510-490 Ma), commencing with east-west extension and associated rapid subsidence, active syn-orogenic deposition and major felsic-dominated volcanism. This was closely followed by, and may have overlapped, a north-south compressional event, which produced east-west trending folds during the Late Middle to early Late Cambrian. During the Late Cambrian the last phase of the Delamerian orogeny reactivated earlier extensional faults as reverse faults (e.g. the Henty Fault), and formed major reverse faults and upright open north-trending folds in western Tasmania.
2.4.3. The Devonian Tabberabberan deformation

The Middle Devonian deformation event is responsible for the dominant structural features throughout the Dundas Trough and has been correlated with the Tabberabberan Orogeny of eastern Australia (Williams et al., 1989).

This deformation event in western Tasmania is characterised by a complexity of fold orientations, which has been attributed in part to the influence of pre-existing structural elements. In many areas the fold geometry appears to have been controlled by Cambrian fold trends, which were tightened during the Devonian (Williams et al., 1989). This resulted in cleavage orientations that are not parallel to the axial plane of the folds with which they are associated (Berry, 1986).

Two stages of regional deformation that have been recognised have been related to east-west compressional events, with indications that tectonothermal activity occurred over a period of ~30Ma (Corbett, 1981; Adams et al., 1985). The first stage resulted in the development of open, upright, NNW- to NNE-trending folds, in part due to the tightening of earlier Cambrian folds, and the development of an associated NNW-striking cleavage (Corbett, 1981; Berry, 1986; Williams et al., 1989). The subsequent phase of deformation and post-granite emplacement resulted in development of NW- and WNW-trending folds and thrusts (Corbett, 1981).

Cambrian structures, such as the Henty, Great Lyell and Rosebery faults were reactivated during the Devonian deformation, with significant associated displacement (Berry, 1989; Berry and Keele, 1993; Berry, 1994).

2.4.4. Metamorphism and Devonian granites

Intrusion of both S-type and I-type granites in western Tasmania occurred during the Late Devonian to Early Carboniferous (375-335 Ma), late syn- to post-kinematic with respect to development of the regional NNW-trending Devonian cleavage and associated folding and faulting (McDougall and Leggo, 1965; Berry, 1986; Corbett and Lees, 1987; Williams et al., 1989; Corbett, 1992). The intrusions occurred at high crustal levels, with minor related deformation and narrow contact aureoles containing biotite ± pyrite ± tourmaline (Corbett et al., 1989a; Williams et al., 1989).

The principal outcropping plutons in the Rosebery region include the Meredith, Granite Tor, Pine Hill and Heemskirk granites. Geophysical surveys indicate a spine of granite beneath the Cambrian volcanic sequence that links the Granite Tor, Pine Hill and Heemskirk granites (fig.2.3). A strong spatial relationship between Devonian base metal mineralisation and the
Chapter 2

granite spine suggests a genetic link, with metals sourced from the granites or through remobilisation of Cambrian mineralisation (Solomon, 1981; Solomon et al., 1987; Leaman and Richardson, 1989). Devonian mineral deposits related to granite emplacement include the Sn-W deposits of Renison Bell and Mt Bischoff, and the Ag-Pb-Zn deposits of the Mt Farrell mineral field.

Regional metamorphism reached lower greenschist facies (Corbett and Lees, 1987).

2.5. **Immobile element geochemistry of the Mount Read Volcanics**

Based on the immobile element geochemistry of volcanic lava and intrusive units within the Mount Read Volcanics, Crawford et al. (1992) defined three calc-alkaline to shoshonitic suites (I to III) and two tholeiitic suites (IV to V) and related them to mapped lithostratigraphic units (fig.2.4). A summary of their work follows.

**Suite I**

This suite is volumetrically the largest, incorporating felsic rocks of the Eastern quartz-phyric sequence, andesitic to rhyolitic lavas of the CVC from north of the Henty Fault and felsic lavas of the CVC from south of the Henty Fault, lavas of the Que-Hellyer footwall sequence, andesites of the Henty Fault Wedge, dacites and rhyolites of the Tyndall Group, sheet-like
quartz-feldspar porphyry intrusives which are sited mainly along the western side of the Mount Read Volcanics, and the Darwin and Murchison sub-volcanic granitoids. The least evolved rocks of this suite comprise moderately porphyritic augite + plagioclase-porphyritic andesites, plagioclase + sparsely augite-porphyritic dacites, minor hornblende + plagioclase-porphyritic dacites, and felsic lavas comprising sparsely plagioclase-porphyritic or quartz + plagioclase-porphyritic rhyolites. Suite I andesites are transitional, medium to high K calc-alkaline rocks, and are the least REE-enriched calc-alkaline lavas in the Mount Read Volcanics ((La/Yb)_N = 5-12 (ave. 8.1)). The andesites were interpreted to have compositional affinities with modern transitional medium to high K calc-alkaline orogenic andesites that occur in mature island arcs, in particular those constructed on pre-existing continental crust and active continental margins.

Suite II
This suite comprises both intrusive and extrusive andesites and dacites, mainly from the upper part of the CVC south of the Henty fault. The andesites and dacites are often hornblende-porphyritic, although some andesites are less commonly augite + plagioclase-porphyritic. They are more P_2O_5 and light REE-enriched than suite I ((La/Yb)_N = 10-26 (ave. 16.7)) and have high K calc-alkaline affinities.

Suite III
This suite is dominated by basalt and andesite, and includes the Que-Hellyer hanging-wall sequence, the Lynch Creek Basalts near Queenstown and intrusive basalts of the Howards
Plains area northwest of Queenstown. They demonstrate a broad compositional range from low TiO₂ (0.4-0.5%), low P₂O₅ basalts with (La/Yb)ₙ = 8-12 and flat heavy REE patterns, through to low TiO₂ (0.4-0.8%) but strongly P₂O₅ and light REE-enriched ((La/Yb)ₙ to 34). The suite III lavas range from primitive crystal-rich olivine + chromite + clinopyroxene-porphyritic basalts and clinopyroxene-rich ankaramitic basalts, both lacking plagioclase phenocrysts, to more evolved basalts and andesites which are clinopyroxene + plagioclase-porphyritic. The least enriched rocks of suite III appear to be of transitional medium to high K calc-alkaline composition with modern analogues in the Andes and modern arc systems, whereas the highly P₂O₅ and REE-enriched lavas, which are regarded as shoshonitic, appear to have no modern analogues.

**Suite IV**

This suite comprises the tholeiitic Henty dike swarm and tholeiitic pillow basalts and dolerites of the Henty Fault Wedge. The rocks have high TiO₂ (1.0 to 1.6 %), very low Nb contents (<3ppm), and are only slightly light-REE-enriched ((La/Yb)ₙ = 1.4-3.4). The Henty Fault Wedge basalts are mainly augite + plagioclase-porphyritic, and dolerites of the Henty dike swarm are also augite + plagioclase-porphyritic to aphyric. Compositionally suite IV was considered to comprise typical rift tholeiites, with low Nb indicating a supra-subduction setting during an early phase of rifting.

**Suite V**

This suite includes the Miners Ridge basalts and comprises strong light REE-depleted ((La/Yb)ₙ < 0.5), low TiO₂ (<0.7%) tholeiitic basalts, and higher TiO₂ (to 2.6%) lavas with only slight light REE depletion ((La/Yb)ₙ = 0.9). These rocks, which outcrop at Miners Ridge, were tentatively correlated with rift tholeiitic sequences associated with the Crimson Creek Formation, and therefore were regarded as part of the pre-Mount Read Volcanics basement.

The three calc-alkaline suites (I-III) occur in approximate stratigraphic order (with suites I and II being coeval in places), indicating a magmatic evolution from transitional medium- to high-K rocks through high-K rocks and ultimately to strongly enriched shoshonitic rocks. The calc-alkaline suites are therefore regarded as analogous to post-collisional volcanics that occur in regions of arc-continent collision. Suite IV is considered to be similar to supra-subduction zone basalts erupted during the early phase of arc splitting and back-arc basin development, thereby indicating that a period of tension and rifting of the Mount Read belt post-dated the CVC and predated the Tyndall Group. Suite V is interpreted as an inlier of pre-existing basement onto which the Yolande River sequence was deposited.

Located within the upper Central Volcanic Complex, the Rosebery, Hercules, Que River and Hellyer ore bodies are primarily associated with units of suite I, with the Que-Hellyer hanging-wall sequence comprising suite III rocks. Located within the Tyndall Group the Henty, Mt Julia
and Mt Lyell Pb-Zn ore bodies are also associated with units of suite I, whereas the Mt Lyell Cu-Au mineralisation and alteration transgresses rocks of both suites I and II.

Recent εNd work by Crawford et al. (2000) returned values which indicate that felsic lavas of the CVC, Eastern quartz-phyric sequence and Tyndall Group were largely derived from crustal melts, the Que-Hellyer and Lynch Creek basaltic lavas were derived from shallow lithospheric melts, and tholeiitic basalts within the lower Tyndall Group and dolerites around Hellyer were derived from an asthenospheric mantle source (Crawford et al., 2000).

2.6. Tectonic setting of western Tasmania

Several tectonic models have been proposed for the development of the late Proterozoic to early Palaeozoic successions of western Tasmania, and most of these have been reviewed in detail by Corbett et al. (1989b) and Powell (1992).

The earliest models interpreted the Dundas Trough as a narrow continental rift, with the Mount Read Volcanics forming an arc along it's eastern margin (Campana and King, 1963; Corbett et al., 1972; Brown, 1986; Varne and Foden, 1988). The second generation of models proposed subduction related to the convergence of the Rocky Cape Region and the Tyennan Region, with either west-dipping (Solomon and Griffiths, 1972), or east-dipping oceanic slabs (Solomon and Griffiths, 1974; Crook, 1980; Green, 1983; 1984; Corbett and Lees, 1987).

The most recent and commonly accepted tectonic model is that of Berry and Crawford (1988) and Crawford and Berry (1992), which has recently been refined (Crawford et al., 2000; Crawford et al., 2003). The updated model proposes early attenuation and rifting of Proterozoic continental crust, with deposition of the Crimson Creek Formation and correlates into small rift basins along a passive continental margin (fig. 2.5). Eastward-directed intra-oceanic subduction commenced to the east of the continental margin, resulting in the formation of a fore-arc terrane. Further subduction resulted in arc-continent collision at 505-510Ma, with resultant boninitic fore-arc crust thrusting westward over thinned continental crust. Subsequent gravitational failure led to collapse of the new crustal assemblage. This in turn triggered crustal extension and formation of the half-grabens into which intercalated felsic and back-arc basin-type tholeiitic units of the Mount Read Volcanics were erupted at ~500Ma. The felsic units at this time appear to have been largely derived from crustal melts. Accelerating extension then led to increasingly mafic magmatism, with late andesites and basalts that included abundant high-K and shoshonitic compositions derived from shallow lithospheric mantle. Further crustal rifting then changed the nature of magmatism toward asthenosphere-derived tholeiitic magmas. In the Late Cambrian a change in crustal kinematics terminated crustal extension and generation of new oceanic crust. This in turn led to ponding of basaltic magmas in the lower crust and subsequent felsic magmatism derived from crustal melts that
Figure 2.5 Hypothetical tectonic development of the Tasmanian section of south-eastern Australia shown as crustal cross-sections between 600 Ma and 480 Ma.

(A) Plume-triggered rifting at ~600 Ma produces an east-facing volcanic margin with thick seaward-dipping reflector packages (SDRS). ~580 Ma: eventual continental crustal rupturing and ocean opening.

(B) ~520-515 Ma: east-dipping subduction commences and produces a boninitic forearc lithospheric section and a primitive intra-oceanic arc.

(C) ~510 Ma: arc-continent collision leads to emplacement of allochthons, first of SDRS-type volcanic passive margin basalts that are then overridden by nappes of forearc-derived boninitic lithospheric sections, leading to collapse of the margin and development of the Dundas Trough foreland basin.

(D) 505 Ma: the new crustal collage had commenced to collapse due to post-collisional extension and Mt Read Volcanics post-collisional lavas were erupted in the half-graben formed between ~505 and 497 Ma.

(E) ~495 Ma: continued extension led to exhumation of the underthrust continental crust of the volcanic passive margin and the production of large amounts of coarse proximal siliciclastic molasse (Owen Conglomerate) that filled grabens formed along the collision zone.

(after Crawford et al., 2003)
produced the Tyndall Group. Further extension in the Late Cambrian led to gradual exhumation of the crystalline basement that comprised the passive margin crust. Rapid erosion of the actively emergent crystalline crust then produced the proximal siliciclastic molasse that constitutes the Owen Conglomerate and correlates.

2.7. Summary

The Mount Read Volcanic belt was rapidly emplaced at the eastern margin of the Dundas Trough during the Middle Cambrian as the result of post-collisional volcanism. The succession is dominated by submarine, calc-alkaline volcanic and sedimentary rocks, which on the basis of lithological associations have been divided into the Sticht Range Beds, the Eastern quartzphyric sequence, the Central Volcanic Complex, the Western volcano-sedimentary sequences, and the Tyndall Group. The volcanic rocks are dominated by rhyolitic to dacitic compositions, with locally abundant andesite and basalt. The succession underwent deformation and granite intrusion during both the Cambrian and the Devonian. The Rosebery deposit is hosted within the upper part of the Central Volcanic Complex, at the western margin of the northern Central Volcanic Complex. The principal Cambrian ore deposits of western Tasmania are hosted in the upper part of the CVC or within the base of the overlying Tyndall Group.
Chapter 3 : Local Geology

3.1. Introduction

This part of the study was aimed at examining the Rosebery host succession and defining a stratigraphic context for the Rosebery ore deposit. A total of ~15,000 metres of diamond core were examined from holes located at the northern end of the Rosebery mine. The core was graphically logged at 1:200 scale in a manner similar to that described by McPhie et al (1993), underground mine exposures were examined, and samples were taken for petrographic and whole-rock analysis. Mine site datasets, including grade control and development drilling, underground mapping and whole rock analysis, were also incorporated where appropriate.

The basic stratigraphic sequence for the Rosebery mine was initially established by Hall and others (Hall et al., 1953; Hall and Solomon, 1962; Hall et al., 1965) who considered the ore to be hosted in sheared tuffaceous sediments that were in turn overlain by pyroclastic volcanic rocks. The first detailed lithological descriptions of the mine sequence were those of Brathwaite (1969; 1974) and Eastoe (1973), who considered the host sediment and overlying black shale to have been deposited in shallow marine basins within a succession otherwise dominated by the products of subaerial volcanism. This interpretation of the mine sequence was integral to attributing ore formation to a syngenetic process for the first time. Green et al. (1981) and Green (1983) interpreted the mine footwall volcanic rocks to have been deposited in a subaerial environment, followed by large-scale subsidence and subsequent deposition of the host and hanging-wall sediments in a marine environment. Allen and Cas (1990) later reinterpreted the footwall volcanic rocks as having been deposited in a submarine environment. Lees (1987) mapped out the surface distribution of the main lithological units for the Rosebery-Hercules region, establishing the lateral extent of the Rosebery host units and associated alteration. In a series of studies Allen (1991; 1992; 1993a; 1994a, b) examined the Rosebery mine sequence and established a detailed interpretation of the volcanic and sedimentary facies that comprise the sequence, and this work has been utilised as a framework for later studies (e.g. McPhie and Allen, 1992; McPhie et al., 1993; Allen, 1994a, b; Gifkins and Allen, 2001; Large et al., 2001a).

The structure and deformation history of the mine sequence has been described in detail by Brathwaite (1972), Green (1983), Aerden (1990; 1991; 1992; 1994) and Berry (1986; 1990a, b; 1991; 1992; 1993a, b; 1994), with all workers attributing the dominant mine cleavage, folding and fault development to Devonian deformation. Although most workers accept that deformation has overprinted existing sulphide ore and associated alteration (Brathwaite, 1972; Green, 1983; Berry, 1986, 1990a, b, 1991, 1992, 1993a, b, 1994), Aerden (1990; 1991; 1992; 1994) has interpreted ore formation to be synchronous with Devonian deformation.
3.2. Geology of the northern Central Volcanic Complex

The northern Central Volcanic Complex, north and west of the Henty Fault, is conformably and disconformably overlain by and in fault contact with the Dundas and Mt Charter Groups, which form a part of the Western volcano-sedimentary sequences (fig.2.1). The lower contact of the Central Volcanic Complex has not been recognised (Gifkins, 2001).

The northern Central Volcanic Complex has in the past been difficult to subdivide due to the massive nature of the volcanics, the paucity of bedding and facing data, and the general lack of marker horizons (Corbett et al., 1989a). Informal lithostratigraphic units within the succession include the Mount Black Volcanics, the Sterling Valley volcanics, the Rosebery-Hercules host sequence (previously Primrose Pyroclastics), and the Hollway andesite (Anderson, 1972; Brathwaite, 1974; Corbett, 1986; Corbett et al., 1989a; McNeill and Corbett, 1989). The sequence also includes quartz-feldspar-phyric rhyolite intrusives, basaltic dykes of the Henty Dyke Swarm, and quartz-diorite dykes (McNeill and Corbett, 1989).

Recent work by Gifkins (2001) and Corbett (2002) has established a more comprehensive stratigraphic framework for the northern Central Volcanic Complex (figures 3.1 & 3.2), which in the Rosebery area has been divided into four main stratigraphic units that include the Sterling

---

**Figure 3.1 Stratigraphic relationships north of the Henty Fault.**
Rosebery occurs at approximately the same stratigraphic level as the Hellyer volcanic-hosted massive sulphide deposit. (modified after Corbett, 2002)
Figure 3.2  Geology of the northern Central Volcanic Complex.
Rosebery is located at the western margin of the Central Volcanic Complex within a package of rocks bounded by the Rosebery and Mount Black faults. The Hercules Pumice Breccia Formation that comprises the Rosebery host and footwall extends as far south as Hercules and is overlain by correlates of the White Spur Formation. (modified after Gifkins, 2001 and Lees et al., 1990)
Valley Volcanics, the Mount Black Formation, the Hercules Pumice Formation and the Kershaw Pumice Formation. The Sterling Valley Volcanics are considered to lie at the base of the Central Volcanic Complex (Gifkins, 2001), although Corbett (2002) has suggested the possibility that they may alternatively be correlated with the overlying Tyndall Group. The Kershaw Pumice Formation and the Hercules Pumice Formation have been correlated on the basis of lithofacies characteristics, and with the Mount Black Formation they combine to comprise the Mount Black Volcanics (Gifkins, 2001). The stratigraphic relationships across the northern Central Volcanic Complex place the Rosebery and Hellyer massive sulphide occurrences at approximately the same stratigraphic level (fig. 3.1).

3.2.1. The Sterling Valley and Mount Black Volcanics

The Sterling Valley Volcanics comprise a >1.5km thick sequence of polymict mafic breccia, mafic sandstone and siltstone, and dacitic to basaltic lavas and sills that have a gradational upper contact with the overlying Mount Black Formation, and have been interpreted to represent the medial to proximal facies of a submarine basaltic volcanic centre (Gifkins, 2001).

The >1.6km thick Mount Black Formation comprises the predominantly rhyolitic to dacitic, massive, flow-banded and autobreciated lava, dome, cryptodome and sill component of the Mount Black Volcanics, that has been interpreted to represent the proximal facies of a predominantly effusive and intrusive submarine volcanic complex (Gifkins, 2001).

The ~800m thick Kershaw Pumice Formation conformably overlies the Mount Black Formation and comprises pumice breccia, pumice-rich sandstone and shard-rich siltstone, with lesser pumice-lithic-rich breccia and sandstone, massive flow-banded and brecciated rhyolitic and dacitic lavas, cryptodomes and sills of rhyolitic and dacitic composition (Gifkins, 2001). The laterally extensive pumice breccias are interpreted to be have been derived from a single volcanic centre (Gifkins, 2001).

The >500m thick Hercules Pumice Formation is truncated at the base by the Rosebery Fault, is disconformably overlain by volcaniclastic sediments of the White Spur Formation (McPhie and Allen, 1992; Gifkins, 2001), and is host to the Rosebery, Hercules and South Hercules ore deposits. The Hercules Pumice Formation broadly comprises a sequence of massive to weakly graded, feldspar-phyric pumice breccia and sandstone, overlain by interbedded crystal-lithic sandstone, siltstone and black mudstone, that is in turn overlain by interbedded feldspar-quartz-phyric, crystal-rich sandstone and breccia and black mudstone (Gifkins and Allen, 2001). The depositional environment of the sequence has been interpreted to be entirely subaqueous and below-wave-base (Corbett and Lees, 1987; Allen and Cas, 1990; Allen, 1994a, b; Gifkins, 2001).
3.2.2. The White Spur Formation

The White Spur Formation as described by Corbett and Lees (1987) is principally located in a 5km wide belt south of Hercules. The lower part of the White Spur Formation comprises abundant units of massive to normally graded, quartz-feldspar crystal- and lithic-rich pumiceous volcanoclastic sandstone and siltstone, with erosional bases to many units. The lithic component includes intraclasts of black mudstone and abundant clasts of quartz-feldspar porphyry, feldspar-phyric porphyry, juvenile pumice clasts, and rare clasts of massive sulphide. The lower part of the White Spur Formation is interpreted as having been emplaced as a series of submarine mass flows.

The upper part of the White Spur Formation comprises grey to black siltstone and slate interbedded with volcanogenic greywacke turbidites and minor felsic tuff, and granule conglomerate. The sequence also contains some marine fossils.

3.2.3. Intrusives

The quartz-feldspar-phyric rhyolitic intrusives that cross cut the Central Volcanic Complex sequence are of limited extent. They typically contain embayed quartz euhedra up to 7mm in diameter, chloritised biotite, and feldspar phenocrysts of predominantly plagioclase composition which often occur as glomerocrysts (McNeill and Corbett, 1989). The intrusive bodies are rhyolitic to rhyodacitic in composition, and vary in size from small irregular dykes to large masses more than a kilometre across (Corbett et al., 1989a).

Mafic feldspar-phyric dykes intrude the felsic volcanic rocks of the Central Volcanic Complex. They are more common proximal to the Henty Fault zone, and appear constrained to the Mount Black and Sterling Valley Volcanics (Gifkins, 2001). The dykes are dark green in colour due to the presence of chlorite and epidote, with small plagioclase phenocrysts and occasional relict pyroxenes. They range from small irregular bodies less than a metre in thickness to bodies hundreds of metres in length and tens of metres in thickness (Corbett et al., 1989a). The intrusive contacts are sharp, irregular suggesting intrusion post-lithification of the felsic volcanic rocks (Gifkins, 2001), however the amygdaloidal nature of some dykes (Lees, 1987; McNeill and Corbett, 1989) suggests that emplacement occurred prior to significant burial of the volcanic succession.
3.3. Geology of the Rosebery mine

3.3.1. Previous work

The Rosebery ore body is located within the Rosebery-Hercules host sequence on the western margin of the northern Central Volcanic Complex. In the Rosebery area the lower boundary of the host sequence is marked by the Rosebery Fault (Corbett and Lees, 1987) and the upper boundary by the Mount Black Fault (Allen, 1991; Gifkins, 2001; Gifkins and Allen, 2001).

Initial attempts to define the stratigraphy of the Rosebery mine sequence were based upon a mix of lithological and alteration characteristics (fig.3.3).

The first detailed accounts of the mine sequence (Hall and Solomon, 1962; Hall et al., 1965) described it as a series of pyroclastic units that were subdivided, from footwall to hanging-wall, into the footwall pyroclastics, the quartz schist, the host rock, the black slate and the massive pyroclastics. The sequence was generally regarded as conformable, although the absence of black slate in places was interpreted as due to a possible disconformity at the base of the massive pyroclastics.

A revised stratigraphy of the mine sequence (Brathwaite, 1969, 1974) divided it into the Primrose Pyroclastics and the overlying Mt Black Volcanics. The Primrose Pyroclastics were further subdivided into a number of local units, including from footwall to hanging-wall the footwall pyroclastics, the footwall schist, the host rock, the black slate and the massive pyroclastics. Burton (1975) interpreted the lower contact of the footwall pyroclastics as a locally faulted and strongly sheared unconformity. Sainty (1986) renamed the massive
Figure 3.4 Stratigraphy of the Rosebery mine sequence.
Generated from logging of exploration and development drilling at the northern end of the mine, mainly sections 960mN to 1440mN. The Mount Black fault separates the top of this section from the overlying rocks of the Mount Black Formation.
pyroclastics as the hanging-wall epiclastics, a term more in keeping with the interpretation that this volcanic material had been reworked as debris flows (Green et al., 1981).

The inclusion of alteration and deformation features in the definition of stratigraphic units has resulted in some ambiguous geological interpretations, such as the true position of the host rock/footwall contact (Lees et al., 1990). Recent studies, some of which have included extensive mapping at both regional and mine scales, have focussed on primary lithological features, facies analysis and complementary immobile element analysis as tools for defining a stratigraphic framework (Allen and Cas, 1990; Allen, 1991, 1992, 1993a; Allen, 1994a, b; Large and Allen, 1997; Large et al., 1998a; Gifkins, 2001; Large et al., 2001a). The stratigraphic framework produced as a result of these studies (fig.3.4) comprises:

- The footwall volcanics and the overlying transitional stratified volcanioclastics (TSV), which comprise the Rosebery-Hercules Footwall Volcanics (=Hercules Pumice Formation). The Rosebery Fault truncates the base of this unit.
- The Rosebery-Hercules Hangingwall Volcaniclastics (=lower White Spur Formation).
- The Mount Black Formation, which is separated from the Rosebery-Hercules Hangingwall Volcaniclastics by the Mount Black Fault.

It is this last stratigraphic framework that is utilised in this study.

### 3.3.2. The footwall volcanics

The Rosebery-Hercules Footwall Volcanics extend over ~14km of strike and a minimum thickness of ~800m, with the base truncated by the Rosebery Fault (Corbett and Lees, 1987; Allen, 1994a, b). The main unit, informally termed the footwall volcanics, comprises a poorly stratified, massive to weakly graded, predominantly rhyolitic, feldspar-phyric pumice deposit with subordinate coherent to hyaloclastic sills (McPhie and Allen, 1992; McPhie et al., 1993; Gifkins and Allen, 2001; Large et al., 2001a; and this study). Individual beds within the footwall volcanics are up to at least 150m in thickness, have normally graded, stratified flow tops, and sparse lithic clasts at the base (Lees, 1987; Allen and Cas, 1990; Allen, 1991). Lithic clasts include feldspar-phyric, spherulitic and amygdaloidal rhyolite and dacite, basalt and crystal-rich...
Figure 3.6 Immobile element geochemistry of the Rosebery mine sequence.
Figures display a selection of data for: (A) the footwall volcanics with lines highlighting two main trends with the higher Ti:Zr ratio value exhibited by the more dacitic pumice breccia unit. The uniform values for most of the footwall pumice breccia package indicates that material was sourced from a single eruptive event. (B) the transitional stratified volcaniclastics exhibit a range of values as expected for units comprising reworked material of mixed provenance. (C) the quartz-feldspar-phyric sill in the vicinity of K and P lenses, and the poorly feldspar-quartz phric and feldspar-phyric rhyolite sills at the southern and of the mine with values akin to the main footwall volcanics package. (D) the hanging wall black slate unit and related thin pumiceous felsic horizons. (E) the hanging wall volcaniclastics. (F) the dacitic pumice breccia of the footwall and the quartz-feldspar-phyric sill above K lens can be distinguished on the basis of Ti:V ratio values. Data from Large & Allen (1997), Klaere (1998), Solly (1999) and this study. Feldspar sill analyses from Green (1990a) and from Rosebery mine geochemical dataset.
Figure 3.7 Footwall volcanics.
A) Typical quartz-chlorite-sericite augen schist after felsic pumice breccia. [R7455-3.9m, core=35mm diam.]. B) Moderately sericite-altered feldspar-crystal sandstone from graded top of pumice mass flow unit. [R6051-110.6m, core=35mm diam.]. C) Sericite-chlorite schist after feldspathic sandstone from top of pumice mass flow unit. [R6203-D1-110.6m, core=35mm diam.]. D) Pumice clasts preserved by carbonate replacement within fine-grained sericite-chlorite schist after fine-grained volcaniclastic top of pumiceous mass flow unit. [R5192-D1-636.9m, core=35mm diam.]. Inset highlights tube pumice texture within the large pumice clast.
sandstone (Lees, 1987; Allen, 1991). The footwall volcanics are capped by an upper member termed the transitional stratified volcanioclastics (TSV), comprising a discontinuous package of massive to diffusely bedded pumice breccia and sandstone, crystal-lithic sandstone, siltstone and black mudstone that is in places intruded by quartz-feldspar-biotite-phyric dacite lava (Gifkins and Allen, 2001).

The feldspar-phenocryst component of the massive breccia, ranging from 5-15% by volume, is predominantly albite or K-feldspar set in a groundmass of quartz, albite, and K-feldspar (Green et al., 1981). Relict non-welded pumice and volcanic shard textures are sometimes preserved within augen of more competent alteration minerals such as carbonate (fig.3.5).

Immobile element ratios within altered volcanic rocks often reflect precursor volcanic rock types (Winchester and Floyd, 1976; MacLean and Barrett, 1993) and have been utilised to distinguish between different volcanic units within the Rosebery sequence (e.g. Gifkins and Allen, 2001; Large et al., 2001a). Immobile element analysis has shown that the bulk of the footwall volcanics are pumice deposits of a uniform rhyolitic composition (fig. 3.6), except for a thin upper zone with a more dacitic signature that can not be visibly distinguished from the lower footwall rocks in drill core or thin section (Large et al., 2001a). Most variation in texture and composition displayed within the pumice breccia package can be attributed to post-depositional alteration and deformation.

Within the Rosebery mine environs the footwall volcanics grade upward from moderately altered, massive, feldspar-phyric pumice breccia into a schist with varying degrees of silica, chlorite, sericite and carbonate alteration and local development of a quartz-augen texture (fig.3.7) (Hall et al., 1953; Brathwaite, 1974; Burton, 1975; Large et al., 2001a). Flattened chlorite- or sericite-altered lenses of 1-5cm in length and set in a more siliceous matrix are common in the footwall volcanics and are broadly aligned to the regional bedding (Green et al., 1981; Allen and Cas, 1990). Early workers interpreted the phyllosilicate lenses and other features as evidence for welding (Corbett, 1981; Green et al., 1981), but they have more recently been interpreted as the product of heterogenous alteration of the pumice breccia prior to compaction, with the more siliceous matrix preserving relatively undeformed, randomly orientated, non-welded round-vesicle and tube-vesicle pumice textures (fig.3.7d) (Allen and Cas, 1990; McPhie et al., 1993).

Although the footwall volcanics in the mine environs are predominantly pumice breccia, lenses of siltstone up to tens of metres in thickness have been reported to the west of Rosebery (Green et al., 1981) and may represent the graded tops of individual mass flows as described elsewhere in the package (Allen, 1992).
Figure 3.8 Geological cross section 1440mN.
Drilling cross section traversing K lens. The quartz-feldspar phryic sill and ore lens outlines are in part defined by detailed ore definition drilling for which drill traces are not displayed. A thin unit of quartz-feldspar crystal-rich turbidites exists beneath the black slate in 87R, with material probably derived from the peperitic porphyry seen at deeper levels of the mine. A splay off the Mt Black fault has offset mine stratigraphy below ~2500mRL. K-shear represents a late brittle fault crosscutting the ore and surrounding host rocks with a shallow dip to the north and a “top-block” offset to the west of ~20m.
Figure 3.9  Geological cross section 1320mN.
This section traverses the southern extent of K lens. The quartz-feldspar phryic sill and ore lens outlines are in part defined by detailed ore definition drilling for which drill traces are not displayed. The position of the inferred Cambrian fault trace corresponds broadly to a rapid thickening of the TSV and disruption in the quartz-feldspar-phryic sill. Sediments comprising material derived from the sill are present in 80R and 49R. The Mount Black fault clearly truncates the hanging wall stratigraphy.
Figure 3.10 Geological cross section 1200mN.
This section traverses the area between K lens and P lens where there is a rapid thickening in volcanioclastic sediments of the Transitional Stratified Volcaniclastics. The quartz-feldspar phryic sill and ore lens outlines are in part defined by detailed ore definition drilling for which drill traces are not displayed. The position of the inferred Cambrian fault trace corresponds broadly to a rapid thickening of the TSV and disruption in the quartz-feldspar-phryic sill.
Figure 3.11 Geological cross section 1080mN.

This section traverses P lens. The quartz-feldsparphyric sill (overlying P9 sub lens) and ore lens outlines are in part defined by detailed ore definition drilling for which drill traces are not displayed. The position of the inferred Cambrian fault trace corresponds broadly to a rapid thickening of the TSV and a change in the lithological packages that make up the TSV. Mineralisation appears confined to units below the lithic-rich package of the TSV. The upper part of the footwall volcanics comprises a thick feldspar-phyric sandstone that is not present in sections further north, possibly as it had been reworked into units that comprise the TSV. Devonian faulting ($F_3$) appears to transect the stratigraphy without significant offset, with the probable exception of the large-scale fault splay from the Mount Black fault observed in the bottom of R6203-D1.
Figure 3.12 Geological cross section 960mN.
This section traverses P lens. The ore lens outlines are in part defined by detailed ore definition drilling for which drill traces are not displayed. The position of the inferred Cambrian fault trace corresponds broadly to a rapid thickening of the TSV and a change in the lithological packages that make up the TSV. Mineralisation appears confined to units below the lithic-rich package of the TSV, although barite mineralisation does extend higher into the stratigraphy. Devonian faulting ($F_3$) appears to transect the stratigraphy without significant offset, with the probable exception of the large-scale fault splay from the Mount Black fault observed in the bottom of R6053 and R6055.
The footwall volcanics across the northern part of the mine (figures 3.8 to 3.12) comprise predominantly schists with variable degrees of sericite, silica, chlorite and carbonate alteration, augen textures and chlorite or sericite pseudofiamme (fig.3.7). Relict feldspar-phenocrysts are typically replaced by sericite, chlorite or carbonate. Where alteration and deformation are less intense, the footwall volcanics can be seen in part to include massive to normally graded feldspar-phyric pumice units with occasional feldspar-phyric rhyolite clasts at the base.

In areas where the overlying TSV package is of increased thickness, in the vicinity of P-lens for example (fig.3.11), the upper part of the footwall volcanics is a blue-grey, sericite altered, massive, feldspar crystal-rich sandstone (fig.3.7b-d) with rare white to grey, siliceous, rarely amygdaloidal, aphyric rhyolite clasts. The immobile element signature of the sandstone is the same as the pumice breccias of the footwall volcanics (fig.3.6), indicating that it represents the fine-grained top of a pumiceous mass flow unit.

The footwall volcanics are interpreted to be the product of large pyroclastic eruptions that were rapidly emplaced as syn-eruptive mass-flows in a subaqueous, below-wave base environment (Allen and Cas, 1990; McPhie and Allen, 1992; Allen, 1994b). The uniform immobile element ratio suggests a single source for the bulk of the footwall volcanics, with the more dacitic breccia at the top possibly due to chemical zonation within the magma chamber (Large et al., 2001a). Alternatively, the more dacitic unit represents a change in mass-flow provenance that marks the start of TSV sedimentation.

3.3.3. The transitional stratified volcanioclastics (TSV)

The TSV unit within the Rosebery mine approximates the “host rock” unit that is in current use at the mine (Hall et al., 1965; Brathwaite, 1969, 1974; Allen, 1994a; Gifkins and Allen, 2001), although the latter term is more broadly applied to all rocks between the schistose footwall volcanics and the hanging-wall black slate. The same package has also in part been termed the “volcanioclastic sandstone unit” (Large et al., 2001a).

The TSV at the Rosebery mine comprises a series of stratified, pumiceous, crystal-lithic bearing volcanioclastic sediments of felsic composition at the boundary between the footwall volcanics and the Hangingwall Volcanioclastics (Brathwaite, 1974; Green et al., 1981; Allen, 1994a; Large et al., 2001a; and this study). Previous work has recognised the transition from feldspar-phyric pumice breccia and sandstone at the base to interbedded crystal-lithic sandstone, siltstone and black mudstone at the top (Gifkins and Allen, 2001). The TSV displays a wide range of immobile element values, as would be expected for reworked volcanic material of a mixed provenance (fig.3.6).

The volcanioclastic sediments of the TSV are variably altered, cleaved and folded, and it is
Figure 3.13 Transitional Stratified Volcaniclastics.
A) Stratified volcaniclastic sediments (37P level, opposite P10 ore drive facing south and immediately above P10 sub-lens). Coarse volcaniclastic sandstone overlain by diffusely laminated siltstone, both overprinted by quartz-dominant alteration. (pencil ≈150mm). B) Stratified volcaniclastic sediments located between massive ore and overlying quartz-feldspar porphyry unit (40K level, K-nub crosscut facing south). The sediments comprise a fine-grained siltstone overlain by graded volcaniclastic sandstone (turbidite) containing raft of siltstone ripped-up from underlying unit during emplacement. (pen ≈150mm)
Figure 3.13 Transitional Stratified Volcaniclastics.
C) Sericite-chlorite schist after feldspar crystal-rich volcaniclastic sandstone. [R5824-300.4m, core=35mm diam.]. D) Diffusely laminated siltstone with quartz-alteration and thin dark quartz veinlets aligned along cleavage. [127R-1425.8m, core=48mm diam.]. E) Moderately altered and foliated volcaniclastic sandstone with 0.5-1mm clastic quartz grains. [127R-1470.4m, core=48mm diam.]. F) Coarse lithic grit comprising aphyric rhyolite clasts [R5827-329.2m, core=35mm diam.]. G) Quartz-sericite schist after volcaniclastic sandstone with aphyric rhyolite clasts. [R5192-D1-414m, core=35mm diam.]
therefore difficult to correlate individual horizons between drill holes. Primary fabrics such as stratification and grain size are generally only preserved where deformation and alteration are less intense (fig.3.13a-b), whereas even in zones of intense alteration and deformation more competent materials such as quartz crystals and lithic clasts are often preserved (fig.3.13e). A basic subdivision of the TSV in the northern part of the mine has been attempted (fig.3.4, figures 3.8 to 3.12), based on textures and clastic components that have survived alteration and deformation.

The feldspar-phyric package

The lowest package within the TSV is a sequence of moderately to intensely altered, diffusely stratified to well-bedded, pumiceous, feldspar-phyric siltstones and sandstones (fig.3.13 a-d). The well-bedded units are often normally graded, containing occasional rip-up clasts of fine-grained felsic sandstone and black shale, and rare clasts of aphyric rhyolitic lava and pumice. In some instances the siltstone tops of normally graded beds are carbonaceous. Bedding thickness varies between 0.5 and 5m, with unit contacts ranging from diffuse to sharp. The feldspar phenocrysts are typically albite except where replaced by sericite or carbonate.

In areas where the TSV is thickened (figures 3.11 & 3.12), the lower part of the feldspar-phyric package comprises thickly bedded, massive to normally graded, pumiceous, feldspar-phyric sandstones that can be difficult to distinguish from the upper part of the footwall volcanics. This more massive part of the package is overlain by the more stratified part of the package as described above.

The predominance of pumiceous feldspar-phyric material in this package suggests that it has for the most part been locally reworked from the footwall volcanics (Large et al., 2001a). The more massive lower part of the package, where the TSV is thicker, may be the result of local slumping of footwall material into topographic lows. The more stratified part of this package is interpreted to be the product of turbiditic mass flow and suspension sedimentation. The presence of a carbonaceous component at the top of some turbidite units, and carbonaceous mudstone rip-up clasts at the base, suggests local ambient anoxic sedimentation.

The feldspar-quartz-phyric package

The feldspar-phyric package is in places overlain by a sequence of moderately to intensely altered, diffusely stratified to well-bedded, massive to normally graded, pumiceous, feldspar-quartz-phyric sandstones and siltstones (fig.3.13e). The sequence is essentially the same as the upper stratified part of the feldspar-phyric package except that a 0.2-2mm quartz-crystal component is present. The basal parts of normally graded sandstone beds contain rare clasts of aphyric rhyolite lava, pumice and rip-up clasts of carbonaceous mudstone and feldspar-
crystal rich sandstone.

The feldspar-quartz-phyric package was emplaced primarily as turbiditic mass flows, with a mixed provenance comprising feldspar-phyric footwall volcanics, carbonaceous mudstone and more distal feldspar-quartz-phyric volcanics. This package is transitional between the feldspar-phyric package and the lithic-rich package.

**The lithic-rich package**

The uppermost lithological package of the TSV is a series of moderately to intensely altered, massive to normally graded, pumiceous, feldspar-quartz-phyric, crystal- and lithic-rich sandstones, siltstones and gravels and minor carbonaceous siltstones (fig.3.13f-g). The principal lithic component comprises siliceous aphyric rhyolite clasts with occasional clasts of feldspar-phyric rhyolite similar to those found in the hanging-wall units, and irregular rip-up clasts of carbonaceous mudstone. The lithic component ranges from occasional matrix supported clasts at the base of normally graded beds to beds of clast-supported gravel. The package also contains minor units of black carbonaceous mudstone, and some beds of normally graded sandstone fine upward into carbonaceous siltstone and mudstone.

The carbonaceous material in some units allows occasional primary sedimentary textures to be distinguished, including small-scale, randomly contorted and folded bedding at the top of the graded units. This soft-sediment deformation may be related to post-emplacement slumping or to deformation of unconsolidated material during emplacement of overlying sediments. Possible flame textures were observed in some of the fine siltstones. The uppermost graded bed, where not faulted, grades conformably upward into the hanging-wall black slate unit.

This package, with a significantly increased lithic component, appears to represent a sequence of turbiditic mass flows with a provenance increasingly similar to that of the hanging-wall units. Ambient sedimentation again appears to be represented by carbonaceous mudstone deposited under anoxic conditions.

3.3.4. The Hangingwall Volcaniclastics

The Rosebery-Hercules Hangingwall Volcaniclastics are a sequence of interbedded pumiceous, feldspar-quartz-phyric, crystal-rich volcaniclastics and lenses of black mudstone (Allen, 1994a; Gifkins and Allen, 2001; Large et al., 2001a). In the Bastyan Dam-Rosebery-Dalmeny area (fig.3.2 & table 3.1) the sequence has been divided into four separate units on the basis of sedimentary grain size profile, lithic clast composition, quartz- and feldspar-crystal ratio in the matrix, and crystal versus pumice content of the matrix (Allen, 1991). The four units
have been further subdivided in the Rosebery mine area, where the principal volcaniclastic units can be traced for the length of the mine (Hunns, 1991; Allen, 1992; Allen, 1993a, b; and this study).

<table>
<thead>
<tr>
<th>Unit</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Unit 4 (top)</td>
<td>A single thick graded bed of very crystal-rich, relatively quartz-rich volcanic sandstone with a feldspar:quartz crystal ratio of ~75:25, with matrix-supported black mudstone clasts at the base.</td>
</tr>
<tr>
<td>Unit 3</td>
<td>One or more thick graded beds, with feldspar:quartz crystal ratio of ~80:20 in lower units and ~95:5 upper units. Base of unit commonly crystal-rich with larger matrix-supported clasts of altered amygdaloidal basalt and feldspar-quartz-phyric rhyolite. Also present but not diagnostic are clasts of mudstone, siltstone, sandstone and feldspar-phyric lava.</td>
</tr>
<tr>
<td>Unit 2</td>
<td>One or more thick graded beds with feldspar:quartz crystal ratio of ~65:35. Base commonly crystal-rich with larger matrix-supported clasts of carbonaceous mudstone, feldspar-quartz-phyric rhyolite, pumice and minor amygdaloidal mafic rock. Graded beds crystal-rich at base and pumice-rich at top.</td>
</tr>
<tr>
<td>Unit 1 (base) “black slate”</td>
<td>Black-grey mudstone with interbedded coarse-grained feldspar-quartz crystal volcanic sandstone</td>
</tr>
</tbody>
</table>

Table 3.1 Stratigraphy of the Hangingwall Volcaniclastics for the Bastyan Dam-Rosebery-Dalmeny area. (after Allen, 1991)

The Hangingwall Volcaniclastics in the Rosebery mine area contain several lenses of carbonaceous mudstone, the most significant of which immediately overlies the TSV and is locally known as the “black slate” (fig.3.14a-b). The black slate has, except where faulted, a conformable contact with the underlying TSV. The unit is finely laminated, overprinted by a slatey cleavage, and has a non-carbonate carbon content of up to 0.67% (Gee, 1970). The black slate unit is generally ~40-60m in thickness, except where it appears to have been eroded during emplacement of the overlying mass flow units, or affected during later Devonian deformation.

The black slate also contains thin bands of pumiceous feldspar-quartz siltstone and sandstone (fig.3.14b) from 0.5 to 30cm in thickness (Green et al., 1981), some of which are normally graded and contain aphyric rhyolite clasts and black mudstone rip-up clasts. Individual sandstones cannot be traced between drill holes, indicating a discontinuous nature. At the southern end of the mine the black slate contains a ~1m thick graded sandstone bed with a significant lithic component derived from Precambrian basement (Green et al., 1981; Corbett et al., 1989a).

At the northern end of the mine the black slate is interbedded with diffusely stratified to well
Figure 3.14 Hangingwall Volcaniclastics.
A) Carbonaceous mudstone (black slate) with calcite-pyrite-pyrrhotite veinlets parallel to cleavage. [132R-D2-1179m]. B) Carbonaceous mudstone with 10cm band of carbonaceous, pumiceous feldspar-quartz-crystal rich sandstone. [121R-1198.1m]. C-E) Polymict mass flow units with clasts of flow-banded rhyolite and bleached amygdaoidal mafic lava [87R-620.2m], black mudstone rip-up clast in quartz-crystal rich matrix [132R-D2-1164.6m], aphyric and amygdaoidal rhyolite clasts [87R-619.8m]. F-G) Quartz-feldspar-crystal rich sandstones from fine-grained top of mass flow units with [121R-1161.3m] and without [123R-12242.8m] aphyric rhyolite clasts. [all core ≈48mm diam.]
bedded, massive to normally graded, fine to coarse-grained, pumiceous feldspar-quartz-crystal sandstone and siltstone. Some of the beds contain rare rhyolite clasts and black mudstone rip-up clasts at the base. These units are similar in nature to the upper part of the TSV, however hydrothermal alteration is less intense.

The black slate is overlain by a sequence of moderately altered, thickly bedded, massive to normally graded, pumiceous, feldspar-quartz crystal- and lithic-rich breccia, sandstone and siltstone, with minor carbonaceous mudstone (Lees, 1987; Allen, 1992; McPhie and Allen, 1992; Allen, 1993a; McPhie et al., 1993; Gifkins and Allen, 2001; and this study). The contact of the volcaniclastics with the underlying black slate is locally erosional, as demonstrated by abundant carbonaceous shale-rip up clasts within the base of overlying mass flows units, and in places the Hangingwall Volcaniclastics directly overlie sediments of the TSV (Adams et al., 1976; Corbett et al., 1989a; and this study).

The principal lithic components of the Hangingwall Volcaniclastic units are clasts of feldspar ± quartz-phryic and aphyric, rhyolitic to dacitic lavas with occasional amygdaloidal and flow-banded textures, rhyolitic pumice, and carbonaceous mudstone intraclasts (fig 3.14c-g). Also present in variable amounts are clasts of amygdaloidal mafic lava, limestone (mainly calcite altered volcaniclastics), chert, rhyolitic sandstone and siltstone, and mafic siltstone and sandstone (Brathwaite, 1974; Green et al., 1981; Lees, 1987; Allen, 1991; McPhie and Allen, 1992; and this study). Although massive sulphide clasts have been observed in the Hangingwall Volcaniclastics at Bastyan Dam (fig.3.2), West Hercules and possibly in the G10 pit at Hercules (Corbett and Lees, 1987; Allen, 1994a), they have not been reported in the vicinity of the Rosebery mine.

The graded volcaniclastic beds generally comprise lithic- and crystal-rich bases that contain horizons of coarser material, and grade normally into fine-grained pumiceous tops (fig.3.14f-g) (Allen, 1992; Allen, 1993a, b). Interbedded with the graded beds are a number of massive to diffusely stratified feldspar-phryic volcanic sandstone and pumice breccia units, with carbonaceous mudstone rip-up and felsic lava clasts at the base and normal graded tops (Allen, 1992; McPhie and Allen, 1992; Allen, 1993a, b). Several laminated carbonaceous mudstone units are also present at the southern end of the mine (Allen, 1991, 1992; Allen, 1993b).

The top of the Hangingwall Volcaniclastics is marked by the Mount Black Fault, which truncates individual beds at the northern end of the mine (fig.3.2, figures 3.8 & 3.9).

The black slate is interpreted to represent a reduction in volcaniclastic input and a return to ambient marine sedimentation in an anoxic environment, with occasional episodes of minor small-scale turbiditic and water-settled volcaniclastic sedimentation. The overlying sequence is
interpreted as a series of predominantly pumiceous volcaniclastic mass flows derived from subaerial to shallow water pyroclastic eruptions and syn-eruptively emplaced into a below-wave-base marine environment, interrupting ambient carbonaceous mudstone sedimentation (Green et al., 1981; Allen, 1991; McPhie and Allen, 1992). The horizons of coarser material in some of the beds represent surges within the mass flows (Allen, 1992). Although the mass flows eroded into underlying strata there is no evidence for reworking of Rosebery sulphide ore or related hydrothermally altered host rocks.

3.3.5. Intrusives

A large irregular sill of moderately to intensely altered, fine-grained, amygdaloidal, weakly feldspar-quartz-phyric rhyolite lava up to ~300m in thickness, with pumice fragments and perlitic devitrification textures has been mapped within the footwall volcanics at the southern end of the mine (fig.3.19) (Allen, 1991; Hunns, 1991). These textures have been interpreted to indicate that emplacement occurred at a shallow depth beneath the seafloor shortly after deposition of the footwall volcanics, and the alteration intensity has been interpreted to suggest that intrusion probably pre-dated hydrothermal alteration (Allen, 1992). This northerly plunging sill outcrops to the south at Koonya (fig.3.2) and extends two-thirds the way to Hercules (Allen, 1991, 1992). Whole rock analysis undertaken as part of this study returned immobile element ratio values comparable to those of the main footwall pumice breccia, possibly indicating a common magmatic source (fig.3.6).

A moderately to intensely altered, moderately quartz-feldspar-biotite-phyric dacite lava (fig.3.15) is present within the footwall volcanics and lower TSV at the northern end of the mine, overlying much of K lens and parts of P lens and AB lens. The distinctive embayed quartz phenocrysts are ~2-4mm in size and the euhedral feldspar phenocrysts ~1-2mm in size (fig.3.16). The intrusion is peperitic, containing abundant clots of volcaniclastic sediment and displaying complex interfingering with enclosing sediments at both upper and lower margins (fig.3.15b-e). Very rare, intensely altered volcaniclastic clasts also occur within the base of the sill and appear more altered than the enclosing lava (fig.3.15f). In drill holes the sill varies laterally (up-dip) from a thick coherent body with peperitic margins to a coarse hyaloclastite with occasional jigsaw-fit texture (fig.3.17), and then to a series of thin, discontinuous, normally graded, quartz-feldspar-crystal-rich beds that, on the basis of the distinctive quartz and feldspar phenocrysts, appear to be derived from this intrusive (fig.3.15g). Hydrothermal alteration of the sill is significantly more intense at the base and the peperitic margins are often overprinted with disseminated sulphide mineralisation. These features indicate that the sill intruded unconsolidated volcaniclastic sediments near the seafloor prior to significant hydrothermal alteration, with the presence of reworked sill material indicating that it breached the seafloor either during intrusion or through later exhumation. The sill-derived sediments occur below the black slate (figures 3.8 & 3.9) indicating that intrusion occurred prior to
Figure 3.15 Quartz-feldspar-phyric peperite within the TSV.
A) Quartz-feldspar-phyric intrusive overprinted by moderate quartz-sericite alteration that has obscured most of the feldspar phenocrysts. [R6529-112.9m, core≈35mm diam.]. B) Interfingering contact between volcaniclastic sediment and quartz-feldspar-phyric intrusive. (35K level, crosscut looking south, pen ≈150mm). C) Interfingering contact between volcaniclastic sediment and quartz-feldspar porphyry. The sediment has taken on a fine-grained appearance that may reflect contact alteration [122R-1496.7m, core≈48mm diam.].
Figure 3.15 Quartz-feldspar-phyric peperite within the TSV.
D) Interfingering contact between volcaniclastic sediment and quartz-feldspar porphyry. [123R-1398.1m, core=48mm diam.].
E) Clots of fine-grained volcaniclastic sediment with diffuse margins within quartz-feldspar porphyry. [60R-930.1m, core=48mm diam.].
F) Basal contact of quartz-feldspar-porphyry (40K level, K nub looking south). The moderately altered intrusive contains clasts of intensely quartz-altered sediment (pen ≈10mm diam.).
G) Quartz-feldspar crystal-rich sandstone portion of graded turbidite unit. The quartz grains may have been reworked from the peperitic intrusive. [121R-1304.6m, core=48mm diam.]
Figure 3.16 Quartz-feldspar porphyry.
A) Weakly altered intrusive with quartz and feldspar phenocrysts within a fine-grained groundmass. The quartz grains exhibit amorphous relict melt inclusion textures which appear as distinctive white flecks in hand specimen. (Lateral field of view = 8.8mm) xp. B) Albite after plagioclase phenocryst with relict primary concentric zoning and partial replacement by sericite (Lateral field of view = 2.2mm) xp. Sample C00-098, 132R-D2-1252.5m.

Figure 3.17 Schematic of peperitic intrusive facies.
A coherent fabric occurs in the core of the sill, particularly where it is thickest. The margins of the sill display a complex mix of porphyry fragments and sediment (hyaloclastite), with occasional jigsaw fit texture. The hyaloclastic material is in places reworked into proximal mass flows and more distal turbidite beds. This sequence was observed on section 1320mN (Fig. 3.9).

Figure 3.18 Schematic of peperitic sill emplacement.
A) Emplacement of pumiceous breccias of the footwall volcanics and overlying stratified volcanlastic sediments of the TSV as mass flows and turbidite flows, with possible syn-sedimentary faulting active at this time. B) Emplacement of quartz-feldspar-phyric sill at or near interface between unconsolidated sediments of the footwall volcanics and TSV, with resultant doming of TSV. Where the sill breaches the sediment-seawater interface autobrecciation occurs with subsequent reworking of the lava as small-scale mass flows and turbidites. C) Return to ambient carbonaceous mudstone sedimentation.
commencement of hanging-wall deposition (fig.3.18). A thin and discontinuous apophysis of
the sill has intruded into the TSV at a higher stratigraphic level in the vicinity of P-lens
(fig.3.19). In the vicinity of W-lens another apophysis cross cuts the footwall volcanics beneath
the main sill, possibly representing a magmatic conduit. The immobile element signature of
this intrusive is similar to that of the more dacitic pumice breccia unit at the top of the footwall,
although the two can be differentiated on the basis of Ti:V ratio values (fig.3.6).

A mineralised and variably altered, feldspar-phyric rhyolite sill, ~20m in thickness and of
unknown lateral extent, is located within the footwall volcanics immediately under J-lens at the
southern end of the mine. The sill is in part coherent but also comprises autobrecciated lava in
a sediment-rich matrix, indicating the locally peperitic nature of the intrusion. The sill is altered
and mineralised to the same degree as the surrounding footwall volcanicslastics, which
indicates that intrusion occurred prior to the mineralising event at Rosebery (Allen, 1992).
Whole rock analysis of this intrusive returned immobile element ratio values comparable to
those of the main footwall pumice breccia, possibly indicating a common magmatic source
(fig.3.6).

A large, weakly altered, rhyolitic, feldspar-phyric sill has also been mapped within the
Hangingwall Volcaniclastics at the southern end of the mine (Allen, 1992), with a lateral extent
of >600m and a thickness of up to ~100m. The sill contains plastically deformed mudstone
fragments and has a complex mixed and interfingering contact relationship with the enclosing
black mudstone unit; indicative of peperitic intrusion into cohesive, semi-consolidated sediment
(Allen, 1992). It has a planar base and appears to have intruded along bedding planes,
thereby inflating the overlying stratigraphy (Allen, 1992). The degree of alteration and position
within the hanging-wall sequence indicates that intrusion occurred after significant alteration
and mineralisation.

The rhyolitic and dacitic sills present at Rosebery were intruded into wet unconsolidated or
semi-consolidated sediments at shallow levels close to the seafloor. Intrusion occurred before,
during and after the hydrothermal alteration and mineralisation related to the Rosebery ore.
Intrusion of lavas near the seafloor is likely to have led to doming of overlying sediments; a
feature noted in the Green Tuff belt and Abitibi belt volcanic sequences (Cas, 1992), and
would therefore have had a significant effect on seafloor topography. The sills would also have
had a significant impact on local hydrothermal fluid circulation below the seafloor.

Several minor occurrences of dolerite dykes with sharp contacts and chilled margins have also
been encountered in the mine sequence. These are generally 1-2m in thickness and lack
significant alteration or deformation, indicating a post-Devonian, most likely Jurassic age.
3.4. **Structure of the Rosebery mine**

3.4.1. **Bedding (S₀) and early compaction cleavage (S₁)**

(S₁ = S₁ of Allen (1991) and Gifkins and Allen (2001))

Bedding (S₀) is variably preserved throughout the mine (fig.3.13a-b), and it dips 40-60°E and trends toward ~350°N. Normal grading is evident in many of the turbidite beds and indicates that most of the sequence is upward facing.

Chlorite pseudofiamme of the footwall volcanics are commonly aligned with an early phyllosilicate foliation (S₁) that is discordant to regional cleavage and has been interpreted as a diagenetic compaction fabric or early bedding-parallel tectonic fabric with a post- or syn-phyllosilicate alteration timing (Allen and Cas, 1990). Where this fabric was observed it appears to have been crenulated by later deformation.

3.4.2. **Syn-depositional faulting**

TSV thickness measurements were obtained from throughout the mine sequence with the aim identifying possible syn-depositional structures that may have influenced sedimentation and ore formation (fig.3.19). Although Devonian faulting is likely to have had an influence on TSV thickness, previous studies suggest that this is not sufficient to explain the observed thickness variation (Berry, 1991). Detailed mapping of the upper levels of AB lens has previously defined some pre-cleavage bedding disruption that could be attributed to small-scale syn-depositional faulting (Berry, 1991; 1992).

Rapid changes in TSV thickness apparent in long section (fig.3.19) also correspond to lateral lithological changes in the TSV units (figures 3.8 to 3.12). The main changes in TSV thickness are interpreted here to be primarily due to basement faulting that commenced after emplacement of the footwall volcanics and was broadly synchronous with the TSV sedimentation. The hanging-wall units do not appear to have been similarly affected, suggesting that significant fault movement had ceased prior to deposition of the hanging-wall sequence.

Possible fault traces have been interpreted on the basis of TSV thickness variation and lateral changes in the TSV (fig.3.19). The main feature evident is a large-scale, northeast trending fault trace that passes from the top of E-lens toward the lower margin of K-lens. The upper part of this inferred fault trace corresponds to a transfer fault previously interpreted in the upper mine levels (Berry, 1992).

Adjacent to the transfer fault trace are a series of what have been interpreted as normal faults.
Chapter 3

Figure 3.19 Cambrian syn-sedimentary fault traces. TSV true thickness contoured at 10m intervals with interpreted syn-depositional fault traces and distribution of pre- to syn-mineralisation peperitic intrusives. TSV thickness corrected for large-scale Devonian faults only. Data derived from mine drill and underground mapping datasets, including data collected by Graves et al. (1998). Includes small-scale structures identified by Berry (1991, 1992) in the vicinity of AB lens.
of predominantly northwest and northeast orientation. Structural complexity appears to increase to the south, possibly due to the presence of the feldspar phryic sill located within the footwall pumice breccias, which may have affected seafloor topography and thickness of the TSV through doming and propagation of faults into the overlying sediments. The complexity may also be the result of displacement along the transfer fault and propagation or disruption of structures up against the more competent intrusive body. The effects of Devonian deformation are also likely to have played a role in increasing the structural complexity.

The orientation of both the southern margin of the K-lens quartz-feldspar-biotite sill and the P-lens apophysis (fig.3.19) suggests that the east-west trending normal faults linking AB-lens and P-lens may have influenced peperitic sill emplacement. The lack of any major offset in these sills suggests that intrusion post-dates significant fault movement. The interpreted major northeast fault trend also corresponds to regions in which the quartz-feldspar-biotite sill overlying K-lens has become attenuated and discontinuous (figures 3.9 & 3.10). This was perhaps due to the influence on sill emplacement by changes in sediment density and disruption of sediment stratification adjacent to the fault.

Fault structures in a competent basement may have been predominantly discrete and brittle in nature but would most likely have propagated into overlying unconsolidated sediments in a more diffuse and plastic fashion with resultant bedding disruption, tilting and folding, and sediment reworking.

3.4.3. Devonian deformation

\[ S_2 = S_2 \text{ of Aerden (1991), } S_1 \text{ of Berry (1994))} \]
\[ S_3 = S_2 \text{ of Berry (1994), } S_3 \text{ and } S_{2r3} \]
\[ \text{of Aerden (1991), } S_2 \text{ of Allen (1991; Gifkins and Allen, 2001))} \]

The most obvious deformation-related features evident in the Rosebery mine are the product of Devonian deformation, supported by an \(^{40}\text{Ar}/^{39}\text{Ar}\) age of ~400-390Ma in cleavage defining muscovite (Perkins, 1996). The significance of Devonian faulting at Rosebery has been a contentious issue, especially with reference to possible structural control during ore formation in a proposed epigenetic ore model (Aerden, 1991), or the disruption and final position of ore lenses in a syngenetic ore model (Berry, 1990a, b; 1991; 1992).

3.4.3.1. Cleavage

An early cleavage \((S_2)\), preserved in low strain domains and pre-dating the dominant cleavage, was described in petrographic studies by both Aerden (1991) and Berry (1994). This cleavage has been interpreted to be a composite cleavage resulting from a north-northwest-trending Devonian cleavage overprinting an earlier north-trending cleavage of Cambrian or Devonian age (Berry, 1994).
The dominant cleavage ($S_3$) present throughout the mine sequence (fig.3.20) dips steeper than bedding ($\sim$60°E) and trends north at $\sim$350° (Brathwaite, 1972; Berry, 1990b). The $S_3$ cleavage is not evident in the massive sulphide ores, but it is preserved in silicate-rich layers within the ore and in the barite lenses (Brathwaite, 1972; Berry, 1990a). A strong down-dip stretching lineation is associated with $S_3$ and related faulting (Brathwaite, 1972; Berry, 1990a; Aerden, 1991). Aerden (1991) related this cleavage development to movement on the Rosebery Fault, and interpreted it to be a combination of a primary cleavage ($S_3$) with crenulation and reactivation of the earlier cleavage ($S_2$). Berry (1994) also related the dominant cleavage to movement on the Rosebery Fault, simply describing it as an extensional crenulation cleavage.

### 3.4.3.2. Faulting & veining

Associated with $S_3$ cleavage development is the main phase of quartz-carbonate veining evident throughout the mine (fig.3.20), with veins both deformed by and cross cutting the main cleavage. The sulphide content of this vein set varies greatly but it is generally higher in the vicinity of the ore lenses, a feature also noted by Berry (1990a, b).

A series of brittle-ductile faults ($F_3$) displaying prominent shear foliation, cataclastic brecciation, silicification and quartz-carbonate ± sulphide veining (fig.3.20a) formed at the time of $S_3$ cleavage development (Brathwaite, 1972; 1990a, b; 1991; Allen, 1992; and this study). Where these structures parallel the main cleavage they can also occur as subtle brittle-ductile to ductile shears with 2 to 20cm of strong cleavage development that is sometimes obscured by silica alteration (Berry, 1990a, b; 1991; Allen, 1992; Westbrook, 1997). The faults trend parallel to the main cleavage and, although they can dip between 40 to 80° to the east (e.g. figures 3.11 & 3.12), they generally dip at a shallower angle than the main cleavage (Brathwaite, 1972; Berry, 1991). These faults can display significant dip-slip reverse movement (Berry, 1990a, b; Allen, 1992; Berry, 1992; Westbrook, 1997), and Brathwaite (1972) noted displacements of $\sim$1.5m. An imbricate array of reverse faults defined in the upper mine levels through geological and lithogeochemical mapping was interpreted to have had an estimated combined displacement of $\sim$250m, with associated stacking of stratigraphy and ore lenses in the upper levels (Berry, 1990a, 1991, 1992). Overlapping immobile element ratio fields for the TSV and the footwall volcanics, in combination with a degree of uncertainty about the geometry of the contact between the TSV and the footwall volcanics prior to Devonian deformation, leaves some uncertainty as to the true magnitude of movement estimated by Berry (1991; 1992).

This phase of faulting ($F_3$) in the study area dips between 30° and 80° to the east. In areas where the TSV package is thinner (figures 3.8 & 3.9) these faults do not appear to be significant and little disruption of stratigraphy is evident at the macro-scale. They occasionally cross cut massive K-lens ore and minor offsets have been observed. The lower margin of the
quartz-feldspar sill overlying K-lens is a strong locus for faulting due to strain partitioning adjacent to the more competent sill. In areas where the TSV is thicker these faults become more prevalent (figures 3.11 & 3.12) but in most cases no large scale offset in the stratigraphy is apparent. The P-lens sulphide and barite lodes, in particular where hosted in more silica-altered rocks, have acted as a locus for this phase of faulting and are often isoclinally folded in the immediate vicinity of these structures.

A zone of large-scale faulting appears to significantly disrupt stratigraphy below 2600mRL.
Chapter 3

(fig.3.8, figures 3.11 & 3.12). The steeply dipping fault zone, possibly a splay off the Mt Black Fault, repeats stratigraphy with apparent reverse movement of between 10 and 100 metres. This zone of faulting appears to extend at least as far north as 1440mN and may extend to 900mS where Allen (1992) noted similar structures that were interpreted as either splays from the Mt Black Fault or part of the Mt Black Fault itself.

Many of the syn-S₃ faults appear to have acted as conduits for metasomatic fluids related to Devonian granite intrusion, with minor cataclastic brecciation, foliation and vein development indicative of some displacement (fig.3.20c). Quartz-chlorite ± carbonate ± sulphide veins that are part of the granite-related metasomatic alteration event have displacements of up to 1m and at lower levels in the mine exhibit an association with magnetite, tourmaline, garnet, pyrrhotite and fluorite. The quartz-chlorite veins in the study area generally strike parallel to the main cleavage, dip both to the east and to the west, and display oblique movement.

The TSV and overlying black slate locally display meso-scale isoclinal folding with axial planes that parallel the S₃ cleavage (Brathwaite, 1972; Berry, 1990a, b; Aerden, 1991; Allen, 1992). This style of folding is a common feature adjacent to syn-S₃ faulting in the vicinity of P-lens and to a lesser degree in the vicinity of K-lens. The folding is most pronounced in phyllosilicate-altered rocks and in the barite and sulphide lodes. The isoclinal fold axes are doubly plunging with pitch variation in cleavage planes of 0 to 50° (Aerden, 1991). The folds have a predominantly west vergence which precludes macro-scale fold structures and indicates that Rosebery lies on the eastern limb of an antiform, with occasional zones of west facing restricted to minor fold hinges (Berry, 1990a, b).

Several workers have inferred macro-scale folding of the sequence in the upper mine levels to explain increased thickness of the TSV and the stacked sulphide and barite lode distribution (fig.3.21). The folding was attributed to Devonian deformation (Brathwaite, 1972; Green et al., 1981; Green, 1983; Lees et al., 1990) or to a large-scale syn-sedimentary slide (Adams et al., 1976). Brathwaite (1972) attempted to define macro-scale folding using the sulphide ore, the lower contact of the black slate and the most intensely altered and foliated part of the footwall volcanics (the footwall schists) as stratigraphic markers. However, the black slate, where present, often has a faulted contact at the lower margin and the footwall schist is an alteration zone rather than a primary depositional feature. The multiple sulphide lodes were inferred to originally comprise a single stratabound lens but no evidence independent of the ore lenses has been presented to support this idea. These features are unreliable stratigraphic markers and the argument for macro-scale folding is therefore untenable. Further attempts were made to define macroscopic folds by utilising sulphide zonation of the ore as a facing indicator (Green et al., 1981; Green, 1983; Lees et al., 1990), but this again relies on the interpretation of the ore as originally existing as a single horizon with a consistent metal zonation. It was also noted by Lees et al. (1990) that fold hinges in the ore were not readily defined and their
absence was interpreted to be the result of removal by shearing or the result of having been located preferentially between ore lenses. The lack of evidence for macro-scale folding and the restriction of west facing to minor fold hinges (Berry, 1990a, b), indicates that folding has not affected the mine stratigraphy at a macro-scale.

Meso-scale foliation boudinage structures were mapped in the southern part of the mine (Aerden, 1991; Berry, 1991) and interpreted to be syn-\(S_3\) cleavage development. The structures are commonly associated with quartz-carbonate veining (Aerden, 1991; Berry, 1991). Aerden (1991) utilised the meso-scale foliation boudinage structures, the cleavage pattern and sulphide lode distribution to interpret the southern ore lenses as macro-scale foliation boudinage structures with both vertical and horizontal orientations. Detailed mapping by Berry (1991) failed to identify any examples of the vertical foliation boudinage structures inferred by Aerden (1991) and found horizontal foliation boudinage structures to be essentially barren of mineralisation.

3.4.4. Post-Devonian deformation

The most recent set of structures observed at Rosebery are brittle faults (fig.3.22) which vary in thickness from \(~0.1\)m to \(~2\)m and are characterised by quartz-carbonate veining and clay gouge. These structures fold the main cleavage and cross cut all earlier faults and vein sets. Most trend N-S with a steep dip to the west, or occasionally trend E-W with a shallow dip to the north. Westbrook (1997) mapped this phase of faulting in the lower part of J-lens, with a predominantly N-S strike, easterly dip and dip-slip reverse to oblique movements of 4 to 5
metres. Oblique movement along these structures at the northern end of the mine appears to be of the order of metres to tens of metres. The east-west trending and north-dipping “K-shear” offsets the stratigraphy and K-lens sulphide ore by ~20 metres (fig.3.8). Steeply west dipping faults observed in the vicinity of P-lens display offsets of ~1-5 metres. These structures post-date Devonian deformation and are interpreted to be high level, low temperature features.

3.4.5. The Rosebery and Mt Black Faults

The Rosebery Fault truncates the lower margin of the footwall volcanics at Rosebery (fig.3.2). On a regional scale this fault is a major east dipping structure extending for at least 30km from Moores Pimple to Silver Falls with an average dip to the east of 40° and a minimum thrust displacement of 1.5km (Corbett and Lees, 1987). In places it is associated with gouge-fill, silicification and strong cleavage development, and has acted as a pathway for Devonian granite-related fluids resulting in quartz-tourmaline veins with accessory fluorite, pyrite, calcite

\[ \text{Figure 3.22 Post-Devonian deformation.} \]
A) Late brittle fault structure (F₄) cross-cutting ore, main cleavage and white syn-main cleavage quartz-carbonate veins (39P level, P10 ore drive facing north). Lineations on fault surface and offset in the sulphide ore horizon indicate a dextral, dip-slip sense of movement. Fault contains quartz-carbonate veining and clay gouge. B) Cataclastic breccia related to late brittle fault (F₄). The fault cross-cuts main cleavage, syn-main cleavage quartz veins and post-cleavage quartz-chlorite veins. [R5975-30m, core=35mm diam.].

3.4.5. The Rosebery and Mt Black Faults

The Rosebery Fault truncates the lower margin of the footwall volcanics at Rosebery (fig.3.2). On a regional scale this fault is a major east dipping structure extending for at least 30km from Moores Pimple to Silver Falls with an average dip to the east of 40° and a minimum thrust displacement of 1.5km (Corbett and Lees, 1987). In places it is associated with gouge-fill, silicification and strong cleavage development, and has acted as a pathway for Devonian granite-related fluids resulting in quartz-tourmaline veins with accessory fluorite, pyrite, calcite
and sulphides (Corbett and Lees, 1987). Significant movement along the Rosebery Fault was synchronous with main cleavage development ($S_3$) and was followed by brittle reactivation with resultant shear foliation and cataclastic brecciation of quartz-carbonate-sulphide veins, and again followed by late brittle reactivation which formed the clay pug (Allen, 1992).

At Rosebery the contact between the Hangingwall Volcaniclastics and Mt Black Formation is usually defined by the brittle-ductile Mt Black Fault (Allen, 1991, 1992). This fault displays a reverse dip-slip sense of movement and truncates units of the Hangingwall Volcaniclastics and TSV as one moves to deeper parts of the mine (figures 3.8 to 3.12); suggesting that significant movement has occurred (Allen, 1992). The fault often displays cataclastic brecciation that is several metres in thickness and is sometimes obscured by intense silica alteration extending tens of metres from the fault. Altered cataclastic breccia is sometimes difficult to distinguish from the coarser Hangingwall Volcaniclastic rocks because of this alteration. Although the Mount Black Formation and the footwall volcanics are considered part of the Mount Black Volcanics (Gifkins, 2001), in some drill holes the contact between the Mount Black Formation and the underlying Hangingwall Volcaniclastics appears conformable, suggesting that the units are in their correct stratigraphic position (Allen, 1991; 1992; and this study). This would indicate that there might be a complex interfingering relationship between the White Spur Formation and the Mt Black Volcanics.

Alteration, veining and cleavage development associated with the Mt Black and Rosebery Faults indicates displacement and fluid movement that can be related to Devonian deformation, granite intrusion and younger faulting (Corbett and Lees, 1987; Allen, 1992; Berry, 1994). No evidence exists to suggest that the Rosebery Fault was a Cambrian structure. A Cambrian basin-bounding structure has instead been interpreted several kilometres further west on the basis of major stratigraphic changes across an inverted normal fault that also bounds the Dundas Trough further north (Berry, 1993a, b).

### 3.5. Summary

This study has established a geological framework for the northern end of the mine. The study has also identified evidence for mine-scale syn-depositional faulting and it’s effects on local sedimentation and sill intrusion. The timing of sill emplacement at the northern end of the mine has also been established, with important implications for the timing of ore formation.

The pumiceous, feldspar-phyric footwall volcanics at Rosebery were rapidly deposited on the seafloor as a series of syn-eruptive mass flows derived from a single eruptive event, possibly on a basement of coherent dacitic and rhyolitic volcanic rocks (Mt Black Formation). The TSV was deposited as the result of turbiditic mass flow and suspension settling of volcaniclastic sediments between periods of ambient sedimentation that deposited carbonaceous mudstone.
The lower feldspar-phyric package of the TSV was the result of local reworking of the feldspar-phyric footwall volcanics. The upper packages of the TSV mark a change in volcaniclastic provenance with the influx of quartz crystals and rhyolitic lithic material derived primarily from more distal volcanic sources.

Peperitic intrusion of rhyolitic to dacitic lava sills occurred below the seafloor before and after the onset of significant hydrothermal alteration. A peperitic quartz-feldspar-phyric sill overlying the northern ore lenses is intensely altered at the base and is overprinted with disseminated sulphide mineralisation along it's peperitic margins. Aside from very rare, intensely altered lithic fragments, it contains almost exclusively unmineralised sediment clots that are no more altered than the surrounding lava. These textures indicate intrusion probably occurred prior to the onset of significant hydrothermal alteration. In this context, the location of massive sulphide ore beneath this sill is one of the key indicators established during this study demonstrating that ore mineralisation occurred beneath the seafloor. Reworking of material derived from the quartz-feldspar-biotite-phyric sill into thin turbidite beds indicates that intrusion occurred prior to commencement of hanging-wall sedimentation.

Syn-depositional faulting of the footwall volcanics and TSV occurred as a result of doming related to peperitic sill intrusion and the development of growth faults related to basement faulting. Intrusion of the quartz-feldspar-biotite sill at the northern end of the mine appears to have occurred subsequent to most of the displacement along these syn-depositional faults, and significant fault movement had ceased by the time of hanging-wall sedimentation. Intrusion of the quartz-feldspar phryic sill appears to have occurred prior to significant syngenetic mineralisation and associated hydrothermal alteration.

A reduction in volcaniclastic influx marked the return to ambient carbonaceous mudstone sedimentation, with minor volcaniclastic turbiditic and suspension settling sedimentation. The return to conditions of significant volcaniclastic input resulted in deposition of a series of mass flow units derived from a more distal source that included volcaniclastic, sedimentary and metamorphosed basement materials.

Regional studies indicate possible deformation during the late Cambrian but no direct evidence has been preserved at Rosebery. The sequence was subsequently subjected to deformation during the Devonian with resultant metamorphism, pervasive cleavage development, vein development, brittle-ductile faulting and folding, which can be related to both regional deformation and significant reverse movement along both the Rosebery Fault and the Mount Black Fault. Intrusion of Devonian granites generated further vein development, metasomatism and minor oblique faulting. The sequence later underwent minor brittle-ductile faulting and Jurassic dolerite intrusion.
Chapter 4 : Alteration

4.1. Introduction

The Rosebery host sequence has been affected by several phases of alteration that can be attributed to the processes of regional diagenesis, hydrothermal fluid circulation, deformation, metamorphism and granite-related metasomatism. The mineralogy, distribution and paragenesis of the various alteration assemblages were examined with the aim of gaining some insight into the timing of sulphide ore formation; the physicochemical conditions under which ore formation took place; the possible location of hydrothermal fluid conduits; and the processes that occurred after ore formation.

As part of this study alteration details were mapped during the logging of ~15,000m of diamond core across the northern part of the Rosebery mine and where appropriate samples were taken for thin section, whole-rock analysis and electron microprobe analysis. This work has been integrated with data from previous detailed studies.

4.2. Previous work

Numerous previous studies of the Rosebery mine sequence have examined aspects of the style, textures, timing and spatial distribution of alteration (Brathwaite, 1969; Eastoe, 1973; Brathwaite, 1974; Dixon, 1980; Green et al., 1981; Green, 1983; Naschwitz, 1985; Eastoe et al., 1987; Lees, 1987; Solomon et al., 1987; Huston, 1988; Allen and Cas, 1990; Lees et al., 1990; Allen, 1991; Khin Zaw, 1991; Naschwitz and van Moort, 1991; Allen, 1994a, b; Orth and Hill, 1994; Allen, 1997; Khin Zaw et al., 1997; Klaere, 1997; Large and Allen, 1997; Westbrook, 1997; Allen et al., 1998; Large et al., 1998a; Chester, 1999; Solly, 1999; Gifkins et al., 2000; Gifkins and Allen, 2001; Herrmann et al., 2001; Large et al., 2001a), the mineralogy and chemistry of alteration minerals (Brathwaite, 1969, 1974; McLeod, 1978; Dixon, 1980; Green et al., 1981; Green, 1983; McLeod and Stanton, 1984; Plimer and Lees, 1988; Khin Zaw, 1991; Khin Zaw et al., 1997; Large and Allen, 1997; Large et al., 1997; Allen et al., 1998; Large et al., 1998a; Herrmann et al., 2001; Large et al., 2001a), and the changes in whole rock chemistry and resultant geochemical haloes around the mine (Naschwitz, 1985; Naschwitz and van Moort, 1991; Large et al., 1998a, c; Herrmann et al., 2001; Large et al., 2001a, b). In general, these studies concluded that there is a strong spatial relationship between sulphide mineralisation and a range of alteration assemblages that predate the main cleavage development. The alteration assemblages can be related to significant changes in whole rock chemistry relative to the primary host rock compositions, and some alteration minerals display compositional changes moving toward ore. The early alteration assemblages have subsequently been affected by Devonian deformation, and a late-stage alteration phase
related to granitic intrusion overprinted the ore and surrounding host rocks. Observations from these studies are integrated with newly collected data in this chapter.

4.3. Diagenetic alteration

Definitions of diagenesis have generally been formulated in the context of sediments and sedimentary rocks in depositional basins. Ehlers and Blatt (1982) suggested that “diagenesis incorporates all physical, chemical and biological changes that a sediment is subjected to after the grains are deposited but before they are metamorphosed”, and as such the pressure and temperature limits are not readily defined. In the context of volcanic-derived materials, Cas and Wright (1987) defined diagenesis as encompassing “the mineralogical and textural changes associated with lithification and the early stages of burial of any sediment or rock system”. Diagenesis may produce significant textural and mineralogical changes in response to increasing pressure, temperature and fluid flux during burial (Cas and Wright, 1987) and at a regional scale, the resultant alteration mineral assemblages might be expected to display a vertical zonation. In comparison, hydrothermal alteration has been defined by Henley and Ellis (1983) as “a general term embracing the mineralogical, textural and chemical response of rocks to a changing thermal and chemical environment in the presence of hot water, steam or gas”, and as such the boundary between diagenetic and hydrothermal alteration is also gradational. The main influences on hydrothermal processes aside from rock composition, are the permeability of the rock pile, temperature, and the composition of fluids moving through the rock pile (Cas and Wright, 1987).

Diagenetic alteration of the volcaniclastic sediments that constitute the Rosebery mine sequence would have commenced at the time of deposition and continued until the onset of metamorphism. The temperature at which diagenetic processes occurred would have been related to the regional geothermal gradient within the Cambrian basin as well as local scale variations in the geothermal gradient caused by magmatic intrusion, faulting and hydrothermal circulation.

4.3.1. Relict smectitic, zeolitic and feldspathic alteration assemblages

Through detailed examination of the Rosebery-Hercules host sequence, Allen (1997) recognised alteration assemblages that can be attributed to regional diagenesis on the basis of mineralogy, alteration texture, and overprinting relationships with deformation fabrics and other alteration types. The diagenetic assemblages are summarised in table 4.1 and in the following discussion.

Pumice and glass shards in the volcaniclastic sediments often display a thin film of sericite ± chlorite that has been interpreted as the result of early smectite ± opaline silica alteration of
<table>
<thead>
<tr>
<th>Regional and ore-related alteration</th>
<th>Time</th>
</tr>
</thead>
<tbody>
<tr>
<td>Deposition of pumice breccia</td>
<td></td>
</tr>
<tr>
<td>(1) Smectite (montmorillonite) ± opal (glass surfaces)</td>
<td></td>
</tr>
<tr>
<td>(2)* Spheroidal Mn-carbonate (concentric layers, concretionary)</td>
<td></td>
</tr>
<tr>
<td>(3) Mordenite-clinozoisite (replaces glass and infills pore space)</td>
<td></td>
</tr>
<tr>
<td>Sphene-leucoxene-haematite</td>
<td></td>
</tr>
<tr>
<td>(4) ?Analcite (replaces mordenite-clinozoisite)</td>
<td></td>
</tr>
<tr>
<td>K-feldspar-quartz (overgrowth on plagioclase phenocrysts)</td>
<td></td>
</tr>
<tr>
<td>(5) Burial compaction, smectite-sericite fiamme</td>
<td></td>
</tr>
<tr>
<td>(6)* Spotty Mn-carbonate (not concretionary)</td>
<td></td>
</tr>
<tr>
<td>Sub-spheroidal Mn-carbonate veins</td>
<td></td>
</tr>
<tr>
<td>(7)* Sericite-carbonate-leucoxene</td>
<td></td>
</tr>
<tr>
<td>(8)* Quartz ± sericite-py-sp ± ga-cpy</td>
<td></td>
</tr>
<tr>
<td>(9)* Chlorite ± quartz ± py-sp-ga-cpy</td>
<td></td>
</tr>
<tr>
<td>(10) Calcite impregnation ± sp (in TSV/hangingwall)</td>
<td></td>
</tr>
<tr>
<td>(11) Albite (replaces K-feldspar ± analcite; deep subsurface)</td>
<td></td>
</tr>
<tr>
<td>(12) Styloitic S, sericite pseudofiamme</td>
<td></td>
</tr>
<tr>
<td>(13) Metamorphic albite-chlorite-epidote-biotite + retrograde chlorite</td>
<td></td>
</tr>
<tr>
<td>Metamorphic biotite + retrograde chlorite</td>
<td></td>
</tr>
<tr>
<td>(14) Reactivated S₁ ± sericite or chlorite</td>
<td></td>
</tr>
<tr>
<td>(15) Syn-S, sericite-calcite (cleavage related)</td>
<td></td>
</tr>
<tr>
<td>(16) Syn-S, quartz-calcite ± sericite ± sp-py-ga veins</td>
<td></td>
</tr>
<tr>
<td>(17) Synfate-S, Fe-Mn-carbonate ± aspy veins, bleached haloes (rare)</td>
<td></td>
</tr>
<tr>
<td>(18) Synfate-S, network chlorite veins (fault related)</td>
<td></td>
</tr>
<tr>
<td>(19) Late/post-S, quartz carbonate veins, bleached selvagens</td>
<td></td>
</tr>
<tr>
<td>(20) Regional post-S, biotite-tourmaline, chlorite-epidote, carbonate-fluorite</td>
<td></td>
</tr>
<tr>
<td>-tourmaline-arsenopyrite veins (with sericite-dark carbonate vein selvages)</td>
<td></td>
</tr>
<tr>
<td>-at Rosebery includes post-S₁/S₂ magnetite-biotite &amp; pyrrhotite-pyrite zones with later tourmaline-quartz zone</td>
<td></td>
</tr>
</tbody>
</table>

Table 4.1 Alteration types and their relative timing, Rosebery-Hercules area (modified after Large et al., 1998a).
Chapter 4

glass surfaces (assemblage 1 – table 4.1) that commenced at the time of sedimentation (Allen, 1997). The same early alteration is also evident in the amygdales of rhyolite and basalt clasts of the hanging-wall volcanoclastic units (fig.4.1a).

A phase of feldspathic alteration preceded development of the $S_1$ compaction fabric, locally replacing most of the rock and comprising diffuse rounded aggregates of relatively coarse-grained albite ± orthoclase, 5-20mm in diameter, and centred on original plagioclase phenocrysts with fine-grained intergrowth of feldspar and quartz between the albite aggregates (Allen, 1997). The feldspar is predominantly albite, but the least deformed and hydrothermally altered rocks contain relict cores of orthoclase within albite-altered domains (Allen and Cas, 1990). The orthoclase occurs within the plagioclase phenocrysts, in fractures in the phenocrysts, and as rims up to 1mm thickness that enclose the phenocrysts (Allen, 1997). The orthoclase rims have in turn been replaced by albite leading to the interpretation of two phases of feldspathic alteration; the initial phase comprising orthoclase or adularia that nucleated on igneous plagioclase phenocrysts and grew outward into the surrounding groundmass (assemblage 4 – table 4.1), and a second phase of extensive albitisation (assemblage 11 – table 4.1) (Allen, 1997). This feldspar replacement has preserved pumice and shard textures, and has locally preserved relict fan-shaped fibrous textures that have been interpreted as possible zeolite pseudomorphs (assemblages 3 and 4 – table 4.1) (Allen, 1997).

During this study albitised plagioclase phenocrysts in the mine area were commonly found within the peperitic sill overlying K-lens, within weakly altered sediments and within some massive sulphide ore. The feldspar phenocrysts occasionally display primary concentric zonation textures interpreted here as being indicative of replacement by solid-state diffusion rather than dissolution and precipitation (fig.4.1b). This is confirmed by the presence of dissolution voids in the albite pseudomorphs resulting from volume changes during albitisation of the original plagioclase (Morad et al., 1990).

Through comparison with studies of diagenetic mineral formation in volcanic rocks under different physicochemical conditions, including the Neogene marine basins of the Japan volcanic arc, Allen (1997) inferred temperature ranges for the development of diagenetic mineral assemblages in the Rosebery-Hercules area (table 4.2). A distinct vertical or lateral zonation of the alteration assemblages was not recognised in the field. Allen (1997) therefore

<table>
<thead>
<tr>
<th>principle mineral</th>
<th>temperature range</th>
</tr>
</thead>
<tbody>
<tr>
<td>smectite (montmorillonite)</td>
<td>from low temperatures up to ~140°C, at higher temperatures the smectite would have converted to illite and sericite (Simmons and Brown, 1996)</td>
</tr>
<tr>
<td>zeolite (mordenite-clinoepidote)</td>
<td>commenced at 30-40°C and continued up to ~130°C (Henneberger and Browne, 1988; Utada, 1991; Oghara, 1996)</td>
</tr>
<tr>
<td>K-feldspar alteration</td>
<td>occurred mainly below 150°C, but may have extended to 190°C (Munhá et al., 1980; Henneberger and Browne, 1988)</td>
</tr>
<tr>
<td>albite alteration</td>
<td>occurred above 150°C, probably extending to at least 250°C (Munhá et al., 1980)</td>
</tr>
</tbody>
</table>

**Table 4.2 Temperatures of diagenetic mineral development**

Refer table 4.1 for mineral assemblages and paragenesis. (compiled from Allen, 1997).
Figure 4.1 Micro-scale alteration textures
A) Quartz-carbonate filled amygaules in basalt clast from a hanging wall volcaniclastic mass flow. The amygdales are lined with a thin film of sericite-chlorite-quartz after clays formed during diagenesis. [xp, fov=4.4mm, C139]. B) Albite pseudomorph of plagioclase phenocryst in mineralised TSV displaying evidence of original concentric zonation (compare with fig. 3.16). The porosity may be the result of volume changes due to the replacement of plagioclase by albite. [xp, fov=4.4mm, C397]. C) Unzoned carbonate spheroid in sericite altered medium-grained sandstone within the TSV. Although obscured in this image the foliation (dashed line) penetrates the carbonate spheroid. [xp, fov=8.8mm, C477]. D) Concentrally zoned carbonate spheroid in sericite-quartz altered fine-grained top of footwall pumice breccia. [xp, fov=4.4mm, C495]. E) Red-brown sphalerite in pressure shadows of rotated opaque pyrite. Coarse randomly orientated blades of sericite and chlorite are also present in the pressure shadows, with finer grained sericite-chlorite defining cleavage. Altered footwall pumice breccia below K lens ore. [xp, fov=8.8mm, R003]. F) Sericite pseudomorph of feldspar phenocryst in sericite-altered TSV. R6298-78.8m. [xp, fov=8.8mm, C423].
concluded that the albite-dominant alteration was the product of a very high geothermal gradient and the rapid progression of high-grade diagenetic alteration up through the volcaniclastic sequence, thereby preserving vitriclastic rock textures and overprinting any lower grade diagenetic assemblages in the process. The temperature range for the albitisation of plagioclase phenocrysts may have been as low as 75-100°C (Morad et al., 1990). Textural evidence documented by Allen (1997), Large et al. (1998a) and as part of this study suggests that diagenetic alteration commenced prior to ore-related alteration, continued in parallel with the mineralising system, and extended into the hanging-wall after cessation of the mineralising system. Albitisation of plagioclase phenocrysts may have occurred prior to or synchronous with the formation of massive sulphide mineralisation. The preservation of albite phenocrysts within the sulphide ore places important constraints on physicochemical conditions of ore formation that will be discussed at a later point.

4.3.2. Chlorite and sericite pseudofiamme

The altered footwall pumice breccias often contain dark, 1-5cm long, fiamme-like chlorite and sericite lenses (fig.4.2a) that are typically hosted within a more siliceous matrix and are aligned roughly parallel to regional bedding (Allen and Cas, 1990). Although originally interpreted as fiamme and indicative of subaerial eruption (Green et al., 1981; Green, 1983), they have subsequently been interpreted to be the product of heterogenous hydrothermal or diagenetic alteration of non-welded pumice breccia that predated, or was synchronous with, sediment compaction (Allen and Cas, 1990). At the northern end of the mine chlorite pseudofiamme are restricted to the footwall pumice breccias and the upper limit of their distribution correlates broadly to the lower limits of Mn- and Fe-rich carbonate alteration (figures 4.3 to 4.7). This may indicate that formation of pseudofiamme was facilitated by the hydrothermal alteration, or by a complex interaction of processes in the area of transition between hydrothermal and diagenetic regimes. Dark-coloured pseudofiamme analysed with an electron microprobe during this study comprised predominantly chlorite, although one specimen examined appeared to consist of sericite with an abundant dusting of very fine-grained haematite.

4.4. Mineralisation-related hydrothermal alteration

The Rosebery ore lenses are located at or near the top of a blanket-like zone of hydrothermal alteration within the footwall volcanics and the TSV (figures 4.3 to 4.7). A large area of sericite ± quartz alteration related to the southern ore lenses outcrops at the southern end of the mine, extending up to 2km along strike and at least 700m in width (fig.3.2) (Brathwaite, 1974; Naschwitz, 1985; Lees, 1987; Lees et al., 1990; Naschwitz and van Moort, 1991). A second area of alteration related to the northern ore lenses has also been identified underground at the northern end of the mine (Naschwitz, 1985; Naschwitz and van Moort, 1991). The hydrothermal alteration is evident as a transition from relatively unaltered rocks containing
volcaniclastic textures and fresh feldspar phenocrysts, through to schists in which volcaniclastic textures are rare and feldspars have either been destroyed or replaced (Green et al., 1981; Naschwitz, 1985; Naschwitz and van Moort, 1991). The hydrothermal alteration zone has been further subdivided on the basis of mineralogical assemblages that appear spatially and texturally related to ore mineralisation (Brathwaite, 1969, 1974; Green et al., 1981; Green, 1983; Eastoe et al., 1987; Large et al., 2001a). The hydrothermal alteration associated with Rosebery ore includes the sericite-dominant, quartz ± sericite dominant and chlorite-dominant assemblages described below, all of which have been overprinted by subsequent phases of deformation-related alteration and fabrics.

4.4.1. Sericite dominant alteration

The broad sericite-rich alteration halo located beneath and lateral to the ore lenses becomes more intense toward ore. Strong to intense sericite ± quartz ± carbonate ± chlorite alteration encompasses the quartz and chlorite dominant alteration assemblages proximal to the sulphide and barite lodes. The sericite dominant assemblage is typically more intensely foliated than the competent carbonate and quartz dominant assemblages and has therefore preserved fewer primary rock textures (fig.4.2b-c).

In the vicinity of K lens the sericite-dominant alteration is laterally extensive (figures 4.3 to 4.5) and, although it can extend tens of metres into the footwall pumice breccias, it does not typically extend more than a few metres above the main ore position, except where the overlying peperitic quartz-feldspar sill is attenuated or absent. This indicates that sericite-dominant alteration post-dated sill emplacement. In the vicinity of P lens the sericite-dominant alteration extends well up into sediments of the TSV (figures 4.6 & 4.7). Intensely sericite-altered rock is also common within semi-massive and massive sulphide ore.

4.4.2. Quartz ± sericite dominant alteration

Sulphide ore is commonly associated with strata-parallel zones of quartz-rich alteration that typically underlie the ore. Quartz ± sericite altered rocks of the footwall volcanics vary from massive bleached quartz rocks (fig.4.2d), previously termed “quartz schist” (Hall et al., 1965), to schists that commonly display quartz-rich augen within strongly foliated sericite-rich domains (fig.4.2e). Relict feldspar phenocryst textures are common in the massive rocks and within quartz augen, and are typically replaced by carbonate, chlorite, and sericite or occasionally by sulphides where proximal to ore.

In the vicinity of K lens a stratiform zone of intense quartz ± sericite alteration typically underlies the massive sulphide ore. The same alteration assemblage extends several metres into the base of the quartz-feldspar porphyry overlying K lens ore, typically obscuring the
Figure 4.2 Alteration in the Rosebery mine sequence: mineralisation-related.
A) Altered footwall pumice breccia with dark chlorite-rich pseudofamme, pale quartz-rich augen and white carbonate after feldspar phenocrysts. [R6760-3.1m, core≈35mm diam.]

B) Moderate sericite-quartz altered feldspathic sandstone (TSV), with coarse clastic horizon flooded with interstitial calcite and cut by syn- to post-cleavage quartz-carbonate vein. [R6051-247m, core≈35mm diam.]

C) Sericite-quartz-chlorite schist with carbonate augen after footwall pumice breccia. Cut by dark post-cleavage chlorite-filled fracture on left. [87R-817.4m, core≈48mm diam.]

D) Intensely quartz-altered footwall pumice breccia ("quartz schist"), with pyrite ± galena-sphalerite after feldspar phenocrysts. [R6426-79.8m, core≈35mm diam.]
Figure 4.2 Alteration in the Rosebery mine sequence: mineralisation-related.

E) Pale quartz-sericite augen schist with increasing chlorite content toward footwall after TSV sediments (P10 cross-cut, looking south). All cut by late syn- to post-cleavage quartz-carbonate veins. F) Variably altered quartz-feldspar-phryic peperite sill overlying K-lens; L→R weakly foliated, intensely quartz-sericite altered and cross-cut by late syn- to post-cleavage quartz-carbonate veins; moderately sericite-quartz altered with “sandy” fine-grained hyaloclastic texture; moderately sericite-chlorite-quartz altered with white calcite after feldspar phenocrysts and crosscut by post-cleavage quartz-chlorite vein. [127R-1470.4m, 123R-1403.7m, 121R-D1-1285m, core=48mm diam.]
Figure 4.2 Alteration in the Rosebery mine sequence: mineralisation-related.

G) Intensely quartz-pyrite altered TSV siltstone/shale located above P lens ore, with minor syn- to post-cleavage carbonate veinlets. [R5192-D1-771m, core≈35mm diam.].

H) Intensely chlorite-sericite altered TSV sediments (P10 crosscut looking south) underlying intensely quartz-carbonate altered TSV. All cross-cut by late syn- to post-cleavage quartz-carbonate-sulphide veins. (book ≈ 190mm height).

I) Intensely chlorite-sericite altered footwall pumice breccia with minor carbonate, disseminated pyrite-chalcopyrite and late syn- to post-cleavage quartz-carbonate veinlets. (core ≈ 35mm diameter). [R7455-70.5m, core≈35mm diam.]
Figure 4.3 Alteration cross section 1440mN.

This section traverses K lens and shows the alteration halo associated with the massive sulphide ore. The chlorite alteration underlying K lens ore is more laterally extensive than can be displayed at this scale. Intense mineralisation-related quartz+/-sericite alteration is located beneath the massive ore and extends into the overlying volcanics and the base of the quartz-feldspar-phyric sill. An overprinting zone of Devonian quartz+/-sericite alteration is also associated with the Mt Black fault.
Chapter 4

Figure 4.4 Alteration cross section 1320mN.

This section traverses the southern extent of K lens and demonstrates that the alteration directly associated with massive sulphide ore is laterally extensive. Intense quartz+/sercite alteration is located beneath and lateral to the massive ore, and extends into the overlying volcanics and the base of the quartz-feldspar-phyric sill. The alteration appears to extend higher into the stratigraphy in the absence of the quartz-feldspar-phyric sill. An overprinting zone of intense chlorite alteration is attributed to Devonian granite-related metasomatic fluids.
Figure 4.5 Alteration cross section 1200mN.
This section traverses the area between K lens and P lens where there is a rapid thickening in volcanioclastic sediments of the TSV. The mineralisation-related alteration is not as quartz-rich as on adjacent sections, but does appear to extend through the attenuated quartz-feldspar porphyry sill and into overlying volcanioclastic sediments that constitute the footwall to PK lens. An overprinting halo of intense quartz+/- sericite alteration is associated with the Devonian Mt Black fault.
This section traverses P lens. Intense quartz +/- sericite and chlorite alteration associated with the sulphide ore appears confined to units below the lithic-rich package of the TSV, particularly beneath the thick package of siltstone and mudstone (c.f. 960mN). Intense quartz-dominant alteration appears to extend into the footwall beneath the P sub lenses, and corresponds with the position of an interpreted north-east trending Cambrian fault (fig. 3.19). Overprinting zones of intense chlorite and quartz +/- sericite alteration are associated with zones of Devonian faulting and shearing that acted as conduits for granite-related metasomatic fluids.
Figure 4.7 Alteration cross section 960mN.
This section traverses P lens. Intense quartz +/- sercite and chlorite alteration associated with the sulphide ore appears confined to units below the lithic-rich package of the TSV, particularly beneath the thick package of siltstone and mudstone (c.f. 1080mN). Intense quartz-dominant alteration appears to extend into the footwall beneath the P sub lenses, and corresponds with the position of an interpreted north-east trending Cambrian fault (fig. 3.19). Overprinting zones of intense chlorite and quartz +/- sercite alteration are associated with zones of Devonian faulting and shearing that acted as conduits for granite-related metasomatic fluids.
peperitic contact and phenocrysts within the sill (fig. 4.2f). Patchy alteration and mineralisation is also evident at the upper margin of the sill, but alteration is weak within the more massive core of the sill and concentrically zoned feldspar textures are preserved. Quartz ± sericite altered rock is also common within semi-massive and massive sulphide ore. The asymmetric alteration of the sill strongly suggests that the sill pre-dated much of the hydrothermal alteration associated with primary mineralisation, and that the sill acted as an impermeable aquaclude above the volcaniclastic sediments in which the ore formed.

In the vicinity of P lens (figures 4.6 & 4.7) the quartz ± sericite alteration does not have a simple stratiform distribution but in detail occurs as a combination of several strata-parallel zones that both underlie and overlie the ore in addition to a transgressive zone of alteration that crosscuts stratigraphy proximal to the ore lenses. The position of this transgressive alteration zone coincides with the approximate position of an interpreted northeast trending syn-sedimentary fault that can be seen on the host sediment thickness long section (fig. 3.19). The transgressive alteration zone extends to the lower margin of a shale-dominant package in the upper part of the TSV (fig. 4.6), the base of which is also strongly quartz ± sericite altered and exhibits a cherty appearance.

The more intense quartz-dominant alteration has generally obscured most sedimentary and volcaniclastic textures except for quartz phenocrysts and rhyolitic clasts in the upper part of the TSV. The preservation of relict pumice textures in some zones of quartz alteration however, indicates that this alteration occurred prior to significant sedimentary diagenetic compaction and subsequent cleavage development (Allen, 1994a).

4.4.3. Chlorite alteration

Dark green to black chlorite alteration often occurs in the immediate footwall to sulphide ore and typically forms either thin concordant bands of chlorite schist or a component within irregular zones of quartz-chlorite-sericite-altered schist (Green et al., 1981; Berry et al., 1998; and this study). The chlorite alteration is both more common and more intense beneath the chalcopryite-pyrite-rich parts of the sulphide ore, especially in the Cu-rich southern part of the mine (Eastoe et al., 1987; Large et al., 2001a). At the relatively Cu-poor northern end of the mine, zones of intense chlorite alteration are less extensive and chlorite is more commonly present as a component of the quartz- and sericite-dominant mineral assemblages.

Underlying K lens and P lens ore the chlorite-rich alteration is typically present as black chlorite ± sericite-carbonate schist (fig. 4.2h-i), often with associated chalcopyrite, pyrite and dark-brown Fe-rich sphalerite as disseminations, veins or semi-massive to massive sulphide bands. Beneath K lens the chlorite schist is usually present as thin (<1 metre) discontinuous bands that extend laterally beyond the sulphide ore in the footwall to the ore position (Large et
al., 2001a). Chlorite alteration in the vicinity of P lens is also patchy, but a continuous 5 to 10 metre thick zone of intense black chlorite alteration is evident beneath the P6 lode proximal to the transgressive quartz-rich alteration zone (fig.4.7).

4.4.4. Carbonate alteration

Several phases of early carbonate alteration appear spatially related to sulphide mineralisation (types 1-9 of table 4.3), and are generally restricted to within 20m of the stratigraphic level of ore and within 100m lateral distance from ore (Dixon, 1980; Lees et al., 1990; Large and Allen, 1997; Allen et al., 1998; and this study). A carbonate paragenesis has been established by Allen et al. (1998) on the basis of overprinting relationships between the different types of alteration, alteration texture morphologies and deformation-related fabrics (tables 4.1 and 4.3). The mineralisation-related, predominantly Mn-rich carbonates occur as disseminations, grains, spheroids and nodules from a few mm to >20cm in diameter (fig.4.8), and occasionally occur with sufficient density to form massive carbonate lenses (Brathwaite, 1969, 1974; Dixon, 1980; Orth and Hill, 1994; Berry et al., 1998; and this study).

<table>
<thead>
<tr>
<th>Texture</th>
<th>Variations</th>
<th>Internal Structure</th>
<th>Composition</th>
</tr>
</thead>
<tbody>
<tr>
<td>1-6. Nodular or sploty</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1. Large nodules (1-200cm diameter)</td>
<td>dispersed, intergrown</td>
<td>-massive granular, coarse concentric layering</td>
<td>not analysed</td>
</tr>
<tr>
<td>2. Spherial (0.02-1cm spheroids of anhedral grains)</td>
<td>dispersed, distinct spheroids, close-packed, intergrown</td>
<td>-distinct fine concentric layering, faint concentric layering, no visible concentric layering</td>
<td>Ca₂Mg(Fe,Mn)(CO₃)₄, Ca(Mn&gt;Fe,Mg)(CO₃)₂</td>
</tr>
<tr>
<td>3. Spherial-rhombic (0.05-1cm spheroids comprising or rimmed by radiating rhombs)</td>
<td>dispersed, distinct spheroids, close-packed, intergrown</td>
<td>-no concentric layering, coarse concentric layering</td>
<td>not analysed</td>
</tr>
<tr>
<td>4. Rhomnic (0.02-1cm rhombs)</td>
<td>dispersed, random intergrowth</td>
<td>-concentric layering</td>
<td>not analysed</td>
</tr>
<tr>
<td>5. Lozenge (0.02-1cm lozenge-shaped grains)</td>
<td>dispersed, close-packed intergrowth</td>
<td>-concentric layering</td>
<td>(Mn&gt;&gt;Fe)CO₃</td>
</tr>
<tr>
<td>6. Feldspar pseudomorph (after 0.05-0.4cm phenocrysts)</td>
<td>porphyrinic distribution</td>
<td>-patchy, irregular to massive</td>
<td>Ca(Mn&gt;Fe&gt;Mg)(CO₃)₂, (Ca&gt;&gt;Mn)CO₃</td>
</tr>
<tr>
<td>7. Platey (rare) (0.5-3cm tabular plates or lattes ?after anhydrite (gypsum))</td>
<td>dispersed, interconnected</td>
<td>-massive, no distinct spheroidal texture or layering</td>
<td>not analysed</td>
</tr>
<tr>
<td>8. Blebbly (710cm irregular patches)</td>
<td>dispersed, interconnected</td>
<td>-massive</td>
<td>(Mn&gt;Fe&gt;Mg)CO₃, Ca(Mn&gt;Fe,Mg)(CO₃)₂</td>
</tr>
<tr>
<td>9. Massive (irregular compact granular masses 5-200cm)</td>
<td>anhedral grains, close-packed rhomb, close-packed spheroids with carbonate-filled interstices</td>
<td>-concentric layered grains, no layering in grains</td>
<td>Ca(Mn&gt;Fe&gt;Mg)(CO₃)₂</td>
</tr>
<tr>
<td>10. Impregnation (filling or replacement of matrix within non-carbonate rock)</td>
<td>irregular patches, pervasive</td>
<td>-anhedral, non-layered grains</td>
<td>CaCO₃, (Ca&gt;&gt;Mn)CO₃</td>
</tr>
<tr>
<td>11. Veins (carbonate ± quartz veins and their alteration haloes)</td>
<td>early pre-S2 cleavage, pre- to syn-S2 cleavage, syn-S2 cleavage, post-S2 cleavage</td>
<td>-sub-spheroidal, massive to banded, massive</td>
<td>Ca₂Mg(Fe,Mn)(CO₃)₄, Ca(Mn&gt;Fe,Mg)(CO₃)₂, CaCO₃</td>
</tr>
<tr>
<td>12. Limestone (fine-grained, compact, calcitic carbonate)</td>
<td>layer, bed, clasts in mass flow beds</td>
<td>-massive to foliated, fossiliferous (trilobites)</td>
<td>CaCO₃</td>
</tr>
</tbody>
</table>

Figure 4.3 Micro-scale alteration textures
Textural types 1-10 relate to assemblages 2, 6, 7 & 10 of table 4.1. (After Allen et al., 1998)

The spheroidal and nodular carbonates (types 1-5 of table 4.3) occur only within sediments of the footwall volcanics and the TSV, in places preserving pseudomorphs of uncompacted
Figure 4.8  Alteration in the Rosebery mine sequence: mineralisation-related.

A) Intensely sericite-quartz altered TSV sandstone (P37 level, footwall to P10) with 1-2mm pale carbonate spots after clastic feldspar. (pencil ≈150mm). B) Intensely chlorite-sericite altered TSV sandstone (36P level, 1050mN crosscut) with lenses of pale carbonate that contain dark grey clastic quartz and feldspar grains. The same grains appear within the chlorite-sericite alteration which indicates carbonate was formed through replacement. Carbonate predates the main cleavage (S3), (lower scale in cm)
Figure 4.8 Alteration in the Rosebery mine sequence: mineralisation-related.
C) Moderately sericite-quartz altered sandstone with white carbonate blebs after feldspar. [R6804-83.7m, core≈35mm diam.] D) Moderately sericite-altered TSV sediment within white Mn/Fe-carbonate atoll textures (after lithic clasts?). The carbonate atolls are hosted within a more quartz-rich alteration that appears to post-date the carbonate. All assemblages are overprinted by main cleavage (S3). [R6889-77.6m, core≈35mm diam.] E) Sericite-chlorite altered footwall volcanics with Mn-Fe carbonate augen representing a possible carbonate alteration front. [R6426-88.9m, core≈35mm diam.] F) Massive Mn-Fe carbonate within TSV sediments overlying P6 lens. The carbonate is brecciated and cut by minor dark chlorite-sericite and pyrite veins. [R6901-88.2m, core≈35mm diam.]
Figure 4.8 Alteration in the Rosebery mine sequence: mineralisation-related.
G) Massive cream/pink Mn-Fe carbonate within TSV sediments, with minor wispy chlorite-sericite and cut by late syn– to post-cleavage quartz-carbonate vein. [R6203-D1-568.8m, core≈18mm diam.] H) Mn-Fe carbonate nodules and rings within quartz-feldspar phyric peperite sill overlying P-lens (level 37P - P9 drive truck turn looking north). Nodules are elongate parallel to local bedding and the main cleavage (S3) which they predate. (pencil ≈ 190mm). I) White pre– to syn–cleavage calcite impregnation of fine-grained sericite-quartz-altered siltstone (TSV). Folding of the bedding/lamination appears related to main cleavage development. Minor thin calcite veinlets mark small-scale offsets. [127R-1333.5m, core≈48mm diam.]
pumiceous clasts and glass shards (Allen, 1994a; Orth and Hill, 1994; Allen et al., 1998) and sometimes hosting framoidal and colloform pyrite (Dixon, 1980; and this study). Carbonate spheroids that lack internal concentric layering are more commonly associated with pumice breccia and coarse shard-rich sandstone (fig.4.1c), whereas distinctly layered carbonate spheroids are more commonly associated with fine-grained units (fig.4.1d) (Allen et al., 1998). Nodular and spheroidal carbonate of types 1-5, platy carbonate of type 7 and massive carbonate of type 9 typically occur within a sericite-rich or occasionally a chlorite-rich matrix, and usually grade away from ore into sericite alteration containing scattered carbonate blebs and/or carbonate-sericite pseudomorphs after feldspar phenocrysts (fig.4.8a) (Allen et al., 1998). Nodular bodies of carbonate up to 15cm in diameter also occur within the barite ore body (H lens) (Brathwaite, 1969, 1974).

Disseminated to massive carbonate of types 1-9 (table 4.3) occur within a few tens of metres of the stratigraphic position of K lens ore and extend for several hundred metres laterally away from the ore (fig. 4.3). The most intense Mn-rich carbonate alteration in the vicinity of K-lens occurs within the more dacitic pumice breccia unit at the top of the footwall, in which much of the K lens ore is hosted (Large et al., 2001a). Massive carbonate (fig.4.8f-g) occurs within K lens sulphide ore but is more common at the margins, particularly near the up-dip margin in the direction of AB lens. The halo of spheroidal and nodular carbonate alteration is more extensive around the stacked lodes of P lens, extending for up to 200m of stratigraphic thickness and in places strongly altering the peperitic quartz-feldspar phyric sill overlying P9 (fig.4.8h). Around K lens and P lens the lower stratigraphic limit of spheroidal and nodular Mn-rich carbonate appears to broadly correspond to the upper limit of chlorite pseudofiamme, and the upper limit of spheroidal and nodular Mn-rich carbonate alteration corresponds to the lower limit of calcite impregnation of volcanic sediments and calcite replacement of feldspar phenocrysts.

White calcite matrix impregnation of type 10 (fig.4.8i), involving volumes of rock up to 20 metres in thickness and hundreds of metres in lateral extent, occurs predominantly within the TSV and basal parts of the Hanging-wall Volcaniclastics and locally within the footwall volcanics and the porphyry sill overlying K lens (Dixon, 1980; Allen et al., 1998; and this study). Textural gradation into surrounding unaltered sediments indicates that this carbonate has replaced clastic material and infilled porosity, with only weak to moderate alteration of contained lithic clasts and feldspar crystals (Allen et al., 1998). Calcite impregnation occurs above both K lens and P lens, but is more pervasive in sediments located between the two ore lenses (fig.4.5).

The presence of spheroidal and nodular to massive carbonate alteration within intrusive sills and individual mass flow and turbidite units, the preservation of uncompacted pumice textures in carbonate-altered domains, the textural continuity between some carbonate-altered domains
and surrounding rocks, and the altered feldspar cores of some carbonate spheroids supports
the premise that carbonate precipitation was initiated by nucleation on pumice and feldspar
crystals within porous and permeable sediments at a shallow depth below the seafloor (Allen,
1992; Orth and Hill, 1994) rather than through deposition on the seafloor (Eastoe, 1973; Dixon,
1980). Mn-rich carbonate with colloform textures has been recorded in rock material contained
within massive sulphide ore (Dixon, 1980; Green et al., 1981), indicating that at least some
carbonate deposition also occurred in open space, either within porous unconsolidated
sediments or above the seafloor.

Massive carbonate lenses located adjacent to the southern ore lenses have been attributed to
both syngenetic mineralisation-related processes (Brathwaite, 1969, 1974) and to later
deformation-related processes (Lees et al., 1990). “Pisolitic” and “oolitic” spheroidal and
nodular textures within the massive carbonate lenses and an interfingering relationship with
the sulphide ore (Brathwaite, 1969, 1974; Burton, 1975) are consistent with a syngenetic
development, whereas the deformation-related and granite metasomatism-related carbonate
veins tend to be massive and smaller in scale. The interfingering of some massive carbonate
lenses with sulphide ore and the presence of disseminated pyrite and sphalerite-galena veins
within the carbonate (fig.4.8f) (Brathwaite, 1969, 1974; Orth and Hill, 1994; and this study)
indicate that that mineralising fluids may have partially replaced carbonate rocks during
sulphide ore formation (Orth and Hill, 1994).

4.5. Devonian alteration

The Rosebery mine sequence has also been affected by alteration that can be attributed to the
effects of metamorphism, tectonic deformation and metasomatic fluids related to late syn- to
post-deformation granite intrusion on the basis of overprinting and crosscutting relationships
with earlier alteration assemblages, sulphide and barite mineralisation, and various
deformation related features.

4.5.1. Metamorphic and deformation-related alteration

Syn- to post-deformation quartz-carbonate veins occur throughout the mine sequence
containing varying amounts of quartz, carbonate, sericite, chlorite, barite and sulphides
(fig.4.9). Gangue mineral content varies with respect to the surrounding rocks, for example
there is generally an increased chlorite content in veins that traverse chlorite-altered rocks
(Allen, 1997). Sulphide content of the veins also typically increases proximal to the ore lenses
(Brathwaite, 1969, 1974; Berry, 1990a, b; and this study), although in comparison with
massive ore, the chalcopyrite content of the quartz-carbonate veins is often elevated with
respect to the other sulphides. Carbonate within the veins is generally calcite, but veins
located proximal to ore often display compositions closer to those of the mineralisation-related
Figure 4.9 Alteration in the Rosebery mine sequence: deformation-related.
A) Variably altered quartz-feldspar phryic peperite cut by syn- to post-cleavage quartz-carbonate (calcite) veins. [R6529-108.3m, 109.9m & 112.9m core=35mm diam.] B) Sericite-chlorite-quartz-carbonate altered footwall volcanics proximal to K lens ore containing disseminated brown sphalerite-pyrite-galena, and crosscut by quartz-carbonate vein containing galena-chalcopyrite-pyrite-sphalerite. [R6587-78.6m, core=35mm diam.] C) Late-syn to post-cleavage quartz-chlorite-sphalerite-pyrite vein in quartz-sericite-chlorite schist (TSV). [R6587-26m, core=35mm diam.]
Fe- and Mn-rich carbonates. These features may be due to “sweating” of components into the veins from surrounding rocks (Allen, 1997), or there may be a degree of physical incorporation of wall rock and remobilisation of components within the vein.

Brittle-ductile $F_3$ faults throughout the mine sequence are often associated with significant amounts (>20%) of quartz - carbonate ± sulphide veining, and intense shear foliation and cataclastic brecciation may also be developed within or adjacent to the faults. Quartz, sericite and carbonate alteration (predominantly calcite) is typically constrained to material within the fault zones and in places obscures cataclastic textures. Around the Mount Black Fault the same alteration assemblage can extend for several tens of metres into surrounding wall rocks due to an increase in the amount of small-scale faulting and fracturing adjacent to the fault.

The Cambrian alteration assemblages have been overprinted by textures related to later deformation (fig.4.8e), with spaced cleavage developed in quartz-rich rocks and strong cleavage developed in the more phyllosilicate-rich assemblages (Berry, 1990a, b). There is also evidence for the growth of sericite, quartz, carbonate, chlorite, barite and sulphides along cleavage planes and in the pressure shadows of more resistant materials such as pyrite (fig.4.1c,e). Textures within the sericite component of pre-tectonic alteration at Rosebery have been compared to those developed during the deformation of phyllosilicate-rich rocks at Mt Lyell, with relict pre-deformation micas occurring as randomly orientated aggregates inclined with respect to domains of more elongate foliation-defining micas formed during deformation through a process of dissolution and micro-fracturing of early micas and commensurate growth of new micas (Cox and Etheridge, 1989; Perkins, 1996). Sericite samples from J lens and H lens returned $^{40}$Ar/$^{39}$Ar ages that correspond to the Devonian Tabberabberan Orogeny in Tasmania, thereby supporting the idea that pre-Devonian micas were compositionally re-equilibrated or recrystallised during subsequent deformation (Perkins, 1996). Similar processes of recrystallisation or dissolution and re-precipitation have affected other mineral components such as chlorite, quartz and carbonate.

Regional metamorphism in the Rosebery area reached lower greenschist facies with resultant albite-chlorite-epidote bearing assemblages in volcanic rocks (table 4.1) not affected by earlier mineralisation-related hydrothermal alteration (Corbett and Lees, 1987; Eastoe et al., 1987; Allen, 1997). Stratigraphic reconstruction suggests 6 to 8km of overburden and a commensurate pressure of between 1.6 and 2.1kb at the time of peak metamorphism (Green et al., 1981; Green, 1983). Initial pressure estimates based on sphalerite geobarometry returned values of 3.1 to 5.3kb, but these were considered high when compared to geological estimates and were thought to be due to re-equilibration of sphalerite composition below 248°C (Green et al., 1981; Green, 1983). Temperature estimates for the metamorphic event based on arsenopyrite-pyrite geothermometry returned values of 370°C to 450°C, although the upper limit is considered to be too high when the albite-sericite - chlorite - epidote - rare biotite
assemblage observed in the volcanic rocks is taken into consideration (Green et al., 1981; Green, 1983). At a macro-scale the alteration related to deformation and metamorphism appears to have developed within a closed system, in that no significant externally-derived chemical components were introduced at that time (Allen, 1997).

4.5.2. Granite-related metasomatic alteration

At the southern end of the mine massive sulphide lenses and host sediments have been overprinted by a transgressive alteration zone (fig.4.10) comprising iron oxides, sulphides and silicate minerals (Solomon et al., 1987; Khin Zaw et al., 1997). Early workers attributed this phase of alteration within the massive sulphide ore to Devonian metamorphism (Brathwaite, 1969, 1974; Green et al., 1981; Green, 1983), or in part to Cambrian seafloor exhalative processes in the case of some tourmaline-bearing rocks (Plimer and Lees, 1988). More recent work on F lens indicates that the metasomatic fluids which generated these assemblages were related to Devonian post-orogenic granite intrusion (Solomon et al., 1987; Khin Zaw, 1991; Khin Zaw and Large, 1996; Khin Zaw et al., 1997; Khin Zaw et al., 1999).

In the vicinity of F-lens the Fe-S-O-bearing alteration assemblages have been divided by Khin Zaw (1991) and Khin Zaw et al. (1997) into the magnetite ± biotite ± chalcopyrite zone, the pyrrhotite ± pyrite zone and the tourmaline ± quartz ± magnetite zone. A summary of this work completed at the southern end of the mine follows.

At the southern end of the mine the magnetite ± biotite ± chalcopyrite zone is mainly confined to areas below 17-level (~2850mRL) where magnetite forms massive bodies with pyrite, biotite and locally haematite. Biotite in this zone is massive, mottled and green to dark green in colour, and haematite ± quartz veins up to 0.5m across occur within biotite-rich alteration. This zone also contains chlorite, sericite, K-feldspar and recrystallised chalcopyrite, and at depth within the mine locally contains garnet ± biotite and garnet ± helvite ± tourmaline assemblages.

Zones of pyrrhotite ± pyrite alteration are present from 14-level (~2990mRL) down to 17-level where they give way to magnetite-rich assemblages. The pyrrhotite-pyrite zone also contains tourmaline, magnetite and biotite. The pyrrhotite-pyrite assemblage transgresses folded primary massive sulphide ore where it is extensively replaced by pyrrhotite and pyrite (fig.4 in Brathwaite, 1972). The effects of this alteration on massive sulphide ore will be discussed in the mineralisation chapter.

The tourmaline ± quartz ± magnetite zone envelopes and overprints the massive pyrrhotite-pyrite zone and comprises irregular and patchy quartz ± tourmaline veins that cut the host rock, the sulphide lenses and other gangue minerals related to the granitic metasomatism. The tourmaline often forms networks of thin, banded veinlets that clearly post-date cleavage in the
host rocks (fig.4.11a). Patches of pyrrhotite, pyrite and magnetite ± haematite with minor chlorite, fluorite and carbonates were also noted within this zone. On the basis of cross cutting relationships the tourmaline - quartz ± magnetite zone appears to post-date both the magnetite - biotite ± chalcopyrite and pyrrhotite - pyrite alteration zones.

The same metasomatic assemblages extend to the deeper mine levels around J lens and locally include tremolite intergrown with chlorite, and grains of cassiterite dispersed through tourmaline-magnetite-pyrrhotite-altered rocks and pyrrhotite-rhodochrosite veins (Westbrook, 1997; Berry et al., 1998).

Within the northern part of the Rosebery mine the granite-related metasomatic assemblages are a minor component above ~2200mRL, but become more significant below ~2200mRL where they are observed in deep drilling of the down-dip extensions of K lens, P lens and W lens. The drill coverage is unfortunately not of sufficient density to allow for spatial distribution of the individual assemblages to be established, however at a mine-scale there is an overall northerly plunge to the upper limit of significant granite-related metasomatic alteration that is consistent with the northerly dip interpreted for the underlying granite (fig.2.3) (Leaman and Richardson, 1989).

Figure 4.10 Geological cross section through F lens showing distribution of granite-related replacement zones. (modified after Khin Zaw et al., 1997).
Figure 4.11 Alteration in the Rosebery mine sequence: Devonian metasomatism.

A) Quartz-tourmaline-carbonate-pyrite vein crosscutting foliation in quartz-sericite altered footwall volcanics. [R7252-7.4m, core=35mm diam.] B) Chlorite-pyrite-quartz vein with broad chlorite alteration halo extending into surrounding sericite-altered TSV sandstones. (Level 31P - crosscut between P6 and P7 sub lenses). The vein and broad halo post-date cleavage (S₃) and thin white syn– to post-cleavage quartz-carbonate veins. (pencil = 19mm). C) Metasomatically chlorite-altered TSV with disseminated pyrite-pyrrhotite. [R5382-D1-652.5m, core=35mm diam.]
D) Metasomatically altered footwall volcanics and ore comprising dark sphalerite with grey magnetite (after haematite), disseminated pyrite and chlorite. The sample is cut by a green chlorite vein. [127R-1505.9m, core=24mm diam.] E) Chlorite-altered TSV with slightly oxidised wispy veinlets and lenses of chalcopyrite-pyrite-pyrrhotite and disseminated pyrite euhedra. [R5382-D1-657.2m, core=35mm diam.] F) Syn- to post-cleavage barite-chlorite vein with coarse brown sphalerite, chalcopyrite, galena and minor Fe-Mn carbonate. [R5382-D1-689.8m, core=35mm diam.] G) Chlorite-sericite altered mylonite (TSV) with pale brown sphalerite-galena. [R5382-D1-691.7m, core=35mm diam.]
Figure 4.11 Alteration in the Rosebery mine sequence: Devonian metasomatism.

H) Sericitic mylonite (TSV) with brown sphalerite-galena. [R5382-D1-691.7m, core≈35mm diam.] I) Sericite-quartz altered TSV cut by late-syn to post? cleavage alteration front comprising dark brown sphalerite-pyrite with halo of green chlorite, Fe-Mn carbonate and disseminated pyrite euhedra. [R6203-D1-569.9m, core≈35mm diam.] J) Post-cleavage chlorite vein cross-cutting syn– to post-cleavage quartz-carbonate vein in quartz-chlorite-sericite altered footwall volcanics. [R5824-153.8m, core≈35mm diam.] K) Quartz-chlorite-sericite altered augen schist (footwall volcanics). [R5824-153.9m, core≈35mm diam.]
Throughout the northern part of the mine a series of quartz-chlorite ± carbonate ± pyrite ± pyrrhotite veins cross cut the sulphide ore, the main cleavage and syn-deformational quartz-carbonate veins, displaying oblique displacements of up to ~1 metre (figures 4.11b & 3.20c). The quartz-chlorite veins vary from 0.2cm to 50cm in thickness and typically display a dark green chlorite-rich alteration selvage that is often an order of magnitude greater in lateral extent than the corresponding vein. The chlorite alteration associated with a relatively innocuous set of widely spaced, narrow quartz-chlorite veinlets can in places extend for tens of metres. At lower levels in the mine the same vein set contains magnetite, tourmaline and fluorite, suggesting a genetic relationship with granite-derived metasomatic fluids. The same quartz-chlorite vein set is also associated with brittle-ductile faults such as the Mount Black Fault. Of a similar timing are massive pink rhodochrosite veins up to ~0.5m in thickness.

At lower levels in the mine, down-dip of P9 lens (fig.4.6) and P7 lens (fig.4.7), the host sequence has been significantly affected by a zone of chlorite-rich alteration and intense foliation development that has transgressed and obscured stratigraphic boundaries and destroyed primary rock textures such as bedding and grain size variation. This zone of alteration and foliation development is up to ~50m in thickness and extends for at least several hundred metres down-dip. The alteration is characterised by abundant bright to dark green chlorite (fig.4.11c,e) with lesser sericite, and boudinaged and fractured remnants of more structurally competent rocks affected by earlier quartz, carbonate, sulphide and barite alteration. The chlorite-rich alteration zone contains occasional bands of disseminated to massive magnetite up to ~0.4m thickness, and is host to a broadly dispersed sulphide component that includes disseminations and anastomosing “stockwork” veins of chalcopyrite, euhedral pyrite and pyrrhotite, wispy lenses and laminae of pale brown to purple sphalerite, and rare disseminated galena and euhedral arsenopyrite. The sulphide content of this alteration zone is highest at the expected primary ore position, with occasional bands of semi-massive to massive sphalerite, galena, pyrite and barite that display varying degrees of folding, brecciation and foliation development. White bands and veins of barite parallel and crosscut foliation (fig.4.11f), and in places contain quartz, sulphides, chlorite, carbonate, rare haematite and tourmaline. Carbonate is present as pink rhodochrosite veining with a syn- to post-foliation timing and post-cleavage quartz-tourmaline ± chlorite ± sulphide veins are also evident. The chlorite-rich alteration zone appears to extend down-dip from F3 brittle-ductile faults and probably developed by permeation of Devonian metasomatic fluids up along these structures. A mylonitic fabric has developed in parts of the alteration zone (fig.4.11g-h) and appears synchronous with metasomatic chlorite development. The same style of metasomatic chlorite-rich alteration and intense foliation development is also present across the lower part of J lens where there has been tourmaline growth in the pressure shadows of rotated pyrite euhedra (Westbrook, 1997), and proximal to the Mount Black Fault (fig.4.5) and related structures. Broad zones of chlorite alteration are also present in the footwall volcanics (figures 4.6 and 4.7) where they appear to overprint quartz-sericite augen and chlorite pseudofiamme
The spatial distribution, timing and mineralogy of the late chlorite-rich alteration assemblages present at the northern end of the mine suggests that they constitute a more distal component of the granite-related metasomatic alteration assemblages described previously by Khin Zaw (1991) and Khin Zaw et al. (1997) from the southern end of the mine. There are also indications of minor displacement along the quartz-chlorite veins and chlorite-altered mylonite zones synchronous with metasomatic alteration. This suggests that the metasomatic assemblages are the result of granite-related fluids that have permeated up along pre-existing F3 fault structures, the phyllosilicate-rich footwall schists and small-scale crosscutting fractures whilst the faults were active.

A lithostatic pressure value of $3.0 \pm 0.5$kb at the time of granite metasomatism has been calculated using sphalerite geobarometry (Khin Zaw and Large, 1996), although this figure is at variance with a later reported value of $2.5 \pm 0.5$kb for the same study (Khin Zaw et al., 1999). Micro-thermometry and laser Raman spectroscopic studies carried out on fluid inclusions within quartz, fluorite and helvite indicate that the biotite-magnetite and pyrrhotite-pyrite assemblages formed due to the action of moderate to high temperature ($\geq 330^\circ$C), more saline (10-25wt% NaCl equivalent), CO2-bearing fluids, whereas the tourmaline-quartz veining formed from lower temperature ($\leq 300^\circ$C) and less saline ($\leq 10$wt% NaCl equivalent) fluids (Khin Zaw et al., 1999). The high temperature, high salinity and CO2-bearing characteristics of the early metasomatic fluids support the hypothesis that they were derived from Devonian granites (Khin Zaw et al., 1999).

4.6. Mineral chemistry

4.6.1. Carbonate

As discussed previously, carbonate occurs with a range of textures and has formed throughout much of the paragenetic sequence. Previous workers have noted the presence of a predominantly manganese-rich suite of carbonates spatially and temporally related to mineralisation at Rosebery (fig.4.12), which in order of abundance comprises rhodochrosite (MnCO3), ferroan rhodochrosite [(Mn,Fe)CO3], kutnahorite [CaMn(CO3)2], manganese siderite [Fe,(Mn)CO3], dolomite [CaMg(CO3)2], and calcite (CaCO3) (Brathwaite, 1969, 1974; Dixon, 1980; Green et al., 1981; Green, 1983; Large et al., 1998a; Large et al., 2001a). A strong spatial relationship exists between the manganese content of carbonate and the position of massive ore, with spheroidal and nodular carbonate (types 1-6 in table 4.3) within the ore and the footwall dacitic pumice breccia unit containing $>20$ mole% MnCO3, as does carbonate within the main altered footwall pumice unit albeit with a slightly higher Fe-content (Allen et al., 1998; Large et al., 2001a). Concentrically zoned carbonate spheroids (fig.4.1d) display a range
Chapter 4

of compositions from rhodochrosite to manganoan ankerite indicative of fluid chemistry fluctuation during formation, whereas the less obviously zoned spheroids (fig.4.1c) tend toward a more internally consistent composition that includes rhodochrosite, kutnahorite and manganoan ankerite (Allen et al., 1998). In contrast the carbonate impregnation and feldspar phenocryst alteration in the hanging-wall (fig.4.8i) typically comprises calcite with <4mole% MnCO₃ regardless of stratigraphic position, except in the base of the porphyry sill overlying K lens where the calcite contains up to 8.1mole% MnCO₃ (Allen et al., 1998; Large et al., 2001a). Calcite with the lowest Fe- and Mn-contents occurs above massive ore in the same position as the more paragonitic white micas (Large et al., 2001a). Calcite is almost the only carbonate developed at the ore position lateral to significant sulphide mineralisation (Large et al., 2001a).

Syn-cleavage veins (fig.4.9) typically contain calcite with <4mole% MnCO₃ and <2mole% FeCO₃ except where they transect zones of ore-related Mn- and Fe-bearing carbonate alteration (fig.4.8g), in which case they display compositions similar to the surrounding Mn- and Fe-bearing carbonate alteration that they postdate (Allen et al., 1998). Late syn- to post-cleavage veins contain calcite, but may also contain Fe- and Mn-rich carbonate even distal to the ore-related Fe- and Mn-rich carbonate alteration halo (Allen et al., 1998). Some late syn- to post-cleavage veins contain weakly manganiferous ankerite, have bleached alteration haloes, and occur in the hanging-wall and footwall proximal and distal to ore (Allen et al., 1998; Large et al., 1998a).

Fe-rich compositions >12mole% FeCO₃ are restricted to syn-cleavage and post-cleavage alteration parageneses (Allen et al., 1998), and post-cleavage veins linked to granite-related metasomatic fluids contain either Fe>Mn manganoan siderite or Mn>Fe ferroan rhodochrosite (Allen et al., 1998; Large et al., 1998a).

Figure 4.12 Carbonate compositions. Rosebery carbonate compositions tend toward Mn-rich proximal to ore and Ca-rich distal to ore, although post-mineralisation phases in syn– to post-deformation vein sets also tend toward more Ca-rich compositions. (microprobe data from Dixon, 1980; Green et al., 1981; Green, 1983; Naschwitz, 1985; Allen, 1997; and Large et al., 1998)
4.6.2. Chlorite

Chlorite occurs as fine foliation-defining blades, as coarse overgrowths on rotated pyrite, as replacement of feldspar phenocrysts and pumice fragments, as coarse randomly orientated crystals within and adjacent to various quartz ± carbonate veins, and as coarse intergrowths with magnetite ± sphalerite-pyrite.

It is evident within the data compiled from this and previous studies at Rosebery (McLeod, 1978; Green et al., 1981; Green, 1983; Naschwitz, 1985; Khin Zaw, 1991; Large et al., 1998a) and Hellyer (Jack, 1989; Fulton, 1999; Gemmell and Fulton, 2001), that chlorite compositions at Rosebery fall predominantly within the ripidolite compositional field and generally display more Fe-rich compositions than chlorites from the more mafic Hellyer mine host rocks (fig.4.13). This indicates that the predominant control on Fe- and Mg-content of chlorites at Rosebery is related to the rhyolitic to dacitic bulk rock composition. Chlorites in rocks from the Noranda mine sequence are similarly found to be more Fe-rich in felsic rocks than in mafic rocks (Hannington et al., 2003b). Previous workers have described a higher Fe-content in chlorite hosted by sphalerite-rich ore and lower Fe-content in chalcopyrite-rich ore (McLeod, 1978), a positive correlation between the Fe-content of ore and that of chlorite (Green et al., 1981; Green, 1983), or an increased Mg-content in chlorite within the host unit proximal to ore (Large et al., 2001a). However, interrogation of the entire chlorite microprobe dataset for Rosebery undertaken as part of this study failed to establish any consistent trends in Fe- or Mg-content related to ore composition or proximity, the various paragenetic phases of chlorite, stratigraphic units or the intensity of alteration.

Volcanic hosted massive sulphide deposits are commonly associated with either Mg-enrichment of chlorites (e.g. Urabe et al., 1983; McLeod and Stanton, 1984; Gemmell and Fulton, 2001; Paullick et al., 2001; Hannington et al., 2003a) or Fe-enrichment of chlorites (e.g. Urabe et al., 1983; Hannington et al., 2003b) in comparison with regional metamorphic chlorites found in the surrounding sequences. No consistent Fe- or Mg-enrichment trend is evident within the Rosebery chlorites.

Chlorite within semi-massive and massive sulphide ore appears to exhibit a lower Si-content in comparison with the rest of the mine sequence (fig.4.13), and chlorites in a sample of ore affected by Devonian metasomatism display compositions that extend up into the brunsvigite field. Chlorite in the Rosebery environment is Mn-enriched in the footwall volcanics and units of the TSV within a few tens of metres of sulphide ore (fig.4.14), however this trend is not evident in the hanging-wall units. Mn-enrichment of chlorite has similarly been noted as a possible indicator of sulphide ore proximity in the Noranda Mine Sequence (Hannington et al., 2003b).
Figure 4.13 Chlorite compositions.
A) Rosebery chlorite compositions fall predominantly into the ripidolite compositional field. Chlorites in "primary" ore specimens generally display lower Si-contents, although chlorites from a sample of metasomatised ore are more Si-enriched. B) Chlorite compositions in predominantly felsic rocks at Rosebery display more Fe-rich compositions than chlorites in predominantly andesitic and basaltic rocks of the Cambrian Hellyer mine sequence. (Rosebery data - McLeod, 1978; Green et al., 1981; Green, 1983; Naschowitz, 1985; Khin Zaw, 1991; Large et al., 1998; and this study. Hellyer data - Jack, 1989; Fulton, 1999; and Gemmell and Fulton, 2001)
Figure 4.14 Chlorite compositions.
A) Altered footwall pumice breccias within a few tens of metres of sulphide ore display significantly elevated Mn-contents relative to less altered and more distal equivalents. B) The same trend is evident in proximal and distal units of the TSV. C) Chlorites from the relatively unaltered Hangingwall Volcaniclastics display a more restricted range of Mn-values, and do not overlap the ore sample chlorites in the same way that the proximal footwall and TSV chlorites do. (Rosebery data - McLeod, 1978; Green et al., 1981; Green, 1983; Naschitz, 1985; Khin Zaw, 1991; Large et al., 1998; and this study)
Chlorites that appeared to be in textural equilibrium with surrounding metasomatic minerals within samples from F lens have been utilised to calculate temperature ranges of ~267-329°C and 270-352°C for the Devonian metasomatic event (Khin Zaw, 1991; Khin Zaw et al., 1997).

4.6.3. Sericite

Fine-grained white micas (sericite) are a common mineral phase in both the altered and “unaltered” rocks of the Rosebery area, occurring as fine foliation-defining flakes, overgrowths on rotated pyrite grains, aggregates in fractures and in the pressure shadow of augen, wispy aggregates after pumice, fine aggregates after feldspar and as coarse flakes in sphalerite-rich ore (fig.4.1). The sericites are predominantly phengitic with elevated Si- and Fe+Mg-contents (Green et al., 1981; Green, 1983; Khin Zaw, 1991; Khin Zaw et al., 1997).

Values for the Fe/(Fe+Mg) ratio in white micas parallel those of chlorites in some volcanic hosted massive sulphide systems and can be an indicator of proximity to sulphide mineralisation (e.g. Hannington et al., 2003a, b). Previous workers have described Fe-enrichment of sericites in the vicinity of metasomatised F lens ore (Khin Zaw, 1991; Khin Zaw et al., 1997) and an Fe-content in sericites below K lens that decreases toward ore commensurate with an increase in Mg (Large et al., 1997). However, interrogation of the entire Rosebery sericite microprobe dataset undertaken during this study again failed to establish any consistent relationship between Fe- and Mg-content that can be related to stratigraphy, the paragenetic phase of sericite, proximity to ore or intensity of alteration. The Fe/(Fe+Mg) ratio in sericites broadly correlates to that of chlorites from the same samples (fig.4.15), and it is therefore consistent that the lack of any trend in the chlorites is paralleled in the sericites. In contrast to the Waterloo deposit in Queensland, where the Fe+Mg-content of white micas decreases toward mineralisation with a commensurate increase in Fe/Mg ratio (Monecke et al., 2001), there does not appear to be a consistent change in the Fe/Mg ratio or total Fe+Mg-content of sericite proximal to mineralisation at Rosebery. The Si-content and total Fe+Mg-content is, however, found to be lower in sericites from primary semi-massive and massive ores when compared with sericites from ores that have been subjected to Devonian metasomatism (fig.4.15).

Na enrichment is present in some sericites as either a solid solution substitution of Na for K, or as very fine paragonite or albite intergrowths (Large et al., 1997). In the vicinity of K lens Na-enriched sericites without apparent albite intergrowths are hosted within sedimentary units of the TSV that are hanging-wall to ore, whereas in TSV sediments located several hundred metres lateral to ore the sericites are not significantly enriched in Na (Large et al., 2001a). Herrmann et al. (2001) utilised SWIR (short wavelength infrared) and PIMA (portable infrared mineral analyser) spectra as an indicator of white mica composition within a series of drill holes at northern end of mine (K lens and AB lens). White micas in TSV sediments were found
Figure 4.15 Sericite compositions.
A) Fe/(Fe+Mg) ratios in sericites broadly parallel those of chlorites within the same rock specimens. Mica compositional data averaged for each sample. B) Sericite compositions in “primary” ore tend more towards muscovite whereas the metasomatised ore contains more phengitic sericites. (Rosebery data - McLeod, 1978; Green et al., 1981; Green, 1983; Naschwitz, 1985; Khin Zaw, 1991; Large et al., 1998; and this study)
to be least phengitic (<0.2 Fe + Mg) and most sodic (0.1-0.3 Na/Na+K) above ore, but again tend to background compositions away from ore.

Ba-enrichment of sericite has previously been noted in the footwall schists (Gee, 1970) and in host sediments proximal to ore (Large et al., 1997), however the original microprobe analyses for Ba must be treated with caution due to significant Ti signal overlaps (Large et al., 1997). Unlike Hellyer, where the Ba content of white micas in the hanging-wall increases toward ore (Gemmell and Fulton, 2001), the Hangingwall Volcaniclastics at Rosebery display ranges of Ba and Na values typical of “unaltered” volcanics distal to mineralisation (Herrmann et al., 2001). Sericites in less altered and laterally distal footwall volcanics and TSV units display similar “background” compositional ranges. Sericites in the footwall proximal to ore may possibly be enriched in Ba, whereas sericites within TSV sediments located above significant ore development are enriched in Na.

4.7. Alteration halo geochemistry

Commensurate with the zonation of mineral assemblages around the Rosebery mineralisation is a zonation in related elements. Geochemical studies by Smith (1975), Green et al. (1981), Green (1983), Naschwitz and van Moort (1991) and Large et al. (2001a) have demonstrated that Mg, Mn, Rb, K, Ba, and S are enriched in the footwall alteration zone proximal to mineralisation whereas Na, Ca, and Sr are depleted (fig.4.16). In the vicinity of K-lens elevated Mn-values are principally confined to zones of Mn-rich carbonate alteration proximal to sulphide mineralisation (Large et al., 2001a). Mn-rich chlorites also occur proximal to ore (fig.4.14). The depletion of Na and enrichment of K in the footwall alteration zone can be linked to the destruction of albite and replacement by sericite, chlorite and quartz (Smith, 1975; Large et al., 2001a). Elevated K values in the vicinity of K lens are also in part due to the more dacitic composition of the uppermost footwall pumice breccia and the overlying quartz-feldspar phyric sill (Large et al., 2001a). Variation in the enrichment and depletion of these elements can also in places be attributed to syn-tectonic vein and fault development (Large et al., 2001a), although this appears to occur at a smaller and more localised scale. Similar patterns of elemental enrichment and depletion in the footwall to mineralisation are also evident in other volcanic hosted massive sulphide systems (Large, 1992; Ohmoto, 1996), and have led to the generation of alteration indices as indicators of alteration (fig.4.16). Statistical analysis of whole rock and PIXE data has been utilised to define enrichment and depletion trends in the Hangingwall Volcaniclastics suggestive of some hydrothermal alteration during the Cambrian (Naschwitz, 1985; Naschwitz and van Moort, 1991; Aung Pwa et al., 1992), although no evidence of significant hydrothermal alteration is evident in hand specimen or thin section.
Figure 4.16 Downhole geochemical plots - 120R.
Variation in whole rock alteration chemistry through a profile in the K lens area (~1320mN). A) MnO, B) Na₂O, C) K₂O, D) Ishikawa alteration index : 100(MgO+K₂O)/(MgO+K₂O+CaO+Na₂O), E) Chlorite, carbonate, pyrite index: CCPI = 100(FeO+MgO)/(FeO+MgO+K₂O+Na₂O), F) S/Na₂O ratio, G) Ba/Sr ratio. (modified after Large et al., 2001a)
4.8. Discussion

The major control on fluid circulation in modern seafloor hydrothermal systems is related to changes in permeability created by faulting, mineral dissolution and precipitation associated with diagenesis and hydrothermal alteration of the host sequence, magmatic intrusion and the inherent permeability of the original host sequence (Alt, 1995a; Hannington et al., 1995; Humphris, 1995; Scott, 1997).

At the northern end of the mine the Cambrian alteration pattern suggests that a significant amount of hydrothermal fluid was focussed along mine-scale syn-sedimentary faults, with intense quartz alteration locally mapping out the fluid pathways (fig.4.17). Changes in sediment permeability, in the form of intrusive sills (K-lens) and thick siltstones (P-lens), then promoted lateral fluid flow at the top of the footwall pumice breccia and within coarser sediments of the TSV. This is indicated in part by the predominantly stratiform nature of the proximal alteration zones, and by the intense basal alteration of a peperitic sill overlying K lens, and parts of P lens and AB lens. A similar pattern of alteration exists at the base of the thick shale package overlying P lens. Several ancient ore-forming hydrothermal systems display similar stratiform alteration patterns within permeable clastic rocks located beneath less permeable units that have acted as aquacludes during hydrothermal fluid circulation, including Kidd Creek and Seneca in British Columbia (McKinley et al., 1996; Koopman et al., 1999) and Matsuki in northeast Japan (Kuroda, 1983).

The lack of a consistent Fe- or Mg-enrichment trend within chlorites and sericites at Rosebery may in part be due to the effects of diffuse lateral fluid flow and broad heterogenous alteration, in contrast with the more discrete fluid conduit at Hellyer that focussed fluid flow and enhanced the chlorite compositional gradient within surrounding host rocks.

4.9. Summary

The Rosebery-Hercules host sequence was subjected to low temperature diagenetic alteration that commenced immediately after deposition of the volcanic sediments, producing clay and possibly zeolite assemblages. The low temperature alteration assemblages were rapidly overprinted by higher temperature feldspathic alteration generated by an elevated geothermal gradient that was possibly related to shallow-level magmatic intrusion. Analogy may be drawn with the hydrothermal-diagenetic alteration assemblages surrounding the Kuroko deposits of Japan (Iijima, 1974; Pisutha-Arnond and Ohmoto, 1983; Ohmoto, 1996).

Hydrothermal circulation through the unconsolidated volcaniclastic sediments commenced prior to significant sediment compaction and lithification, and continued in parallel with the regional diagenetic alteration, forming a zonation of quartz, sericite, chlorite and carbonate-rich
Cambrian: A) In the case of K lens most of the mineralisation-related hydrothermal alteration post-dates the quartz-feldspar-phyric sill overlying massive sulphide ore. The sulphide ore is immediately underlain by thin (typically <5m) discontinuous chlorite-dominant alteration. A broader stratiform quartz-sericite dominant alteration halo surrounds the ore, extending into the footwall and into the base of the overlying sill. The quartz-sericite-dominant alteration extends laterally for several hundred metres in association with disseminated sulphide mineralisation. Fe/Mn-carbonate alteration is also spatially associated with ore, occurring as disseminations, blebs, nodules and occasionally as zones of massive carbonate. This carbonate alteration is most intense at the ore position and for several metres below the ore position. In contrast the stratigraphic overlying ore contains Ca-carbonate as disseminations, veinlets and zones of replacement to tens of metres in thickness. A broad envelope of sericite-dominant alteration extends for several hundred metres into the footwall but only a few metres into the TSV. Chlorite-sericite pseudofiamme are present within the footwall augen schists but are absent within the quartz-sericite and Fe/Mn-carbonate alteration envelopes.

Devonian: C) Deformation resulted in folding and faulting of the mine sequence. Quartz-carbonate veining formed through local chemical remobilisation of existing minerals, with the extent of Fe/Mn-carbonate and Ca-carbonate veining reflecting Cambrian carbonate mineral distribution. Larger faults contain more intense quartz-carbonate veining and may exhibit haloes of quartz +/- sericite +/- carbonate alteration, particularly where cataclastic brecciation is evident.

D) Metasomatic fluids related to Devonian granite intrusion moved upward along exiting brittle-ductile faults and along zones of shearing such as within the footwall schists. The alteration in the upper mine levels comprises predominantly chlorite with sericite and quartz. Tourmaline, magnetite, pyrrhotite, fluorite, helvite and garnet occur at lower levels within the mine.
alteration assemblages around the sulphide ore (fig.4.17a-b). The alteration assemblages are similar to those associated with other volcanic hosted massive sulphide hydrothermal systems (Lydon, 1984, 1988; Large, 1992; Ohmoto, 1996).

This study has identified narrow transgressive zones of intense quartz alteration beneath the northern ore lenses that correspond spatially with interpreted Cambrian faults, and these faults are interpreted here to have been conduits for hydrothermal upflow. More competent components of the alteration have in places preserved uncompacted pumice fragments, indicating that primary mineralisation-related hydrothermal alteration occurred prior to significant sediment compaction. The base of the quartz-feldspar-phyric peperitic sill overlying K lens, parts of P lens, and parts of AB lens is intensely altered. Intense alteration is typically located within volcaniclastic units underlying the sill, however where the sill is attenuated or absent the intense alteration extends further up into the overlying TSV succession. Sedimentary material incorporated within the sill during emplacement is typically altered to the same degree as the enclosing sill. These features, identified during the current study, strongly indicate that the peperitic sill was emplaced prior to significant hydrothermal alteration and that it acted as an impermeable barrier to upwelling hydrothermal fluids. It was also discovered during this study that the base of a thick shale package overlying P lens is similarly intensely altered, suggesting that it too acted as a barrier to upwelling fluids. The peperitic sill and fine-grained stratified sediments overlying the footwall volcanics would have promoted lateral flow of an upwelling hydrothermal fluid. Strong to intense alteration attributed to the main phase of hydrothermal activity does not extend into units of the Hangingwall Volcaniclastics, suggesting that the system had waned by the time ambient carbonaceous mudstone sedimentation had resumed.

The Rosebery-Hercules host sequence, including the Rosebery area, was subjected to greenschist metamorphism during the Devonian Tabberabberan Orogeny. Existing hydrothermal alteration assemblages within the mine were not significantly affected in terms of mineralogy, however the less competent phyllosilicate-rich assemblages were significantly deformed with resultant remobilisation, recrystallisation, re-equilibration and realignment of minerals such as sericite. The more competent quartz- and carbonate-rich assemblages were less affected by deformation and metamorphism, sometimes preserving primary rock textures such as uncompacted tube pumice. Quartz ± carbonate vein sets developed during deformation, often in association with brittle-ductile faulting (fig.4.17c). Metasomatic fluids associated with late syn- to post-deformation granite intrusion subsequently permeated along brittle-ductile faults and through the footwall schists (fig.4.17d), resulting in the introduction of new elemental components and generating new mineral assemblages which include magnetite, haematite, biotite, chlorite, pyrrhotite, tourmaline, garnet, helvite and cassiterite.
Chapter 5: Mineralisation

5.1. Introduction

The total historical production and current resources for Rosebery stand at ~32.7Mt @ 14.5% Zn, 4.4% Pb, 0.58% Cu, 145ppm Ag and 2.2ppm Au (P. Edwards pers. comm., 2002), with the ore occurring within a series of sulphide and barite-rich lenses that are distributed over ~3000m of strike and extend to at least 1500m depth below surface (fig.5.1). The lodes may originally have been more laterally extensive but any up-dip extension of the ore has been removed by erosion.

This part of the study was aimed at defining the characteristics of ore mineralisation at Rosebery and to this end information was collected from underground exposures and from the ~15,000m of diamond core logged across the northern part of the mine around K lens and P lens. Where appropriate samples were taken for petrographic analysis and observations from previous studies were also incorporated into this work. Most of the work focussed on mineralisation in and around K lens and P lens.

Although the ore lens morphology, mineralogy and mineralogical zonation have been well documented in previous studies, there has been little work completed on primary ore textures and a paragenetic sequence for mineralisation. A significant part of the work completed during this study has included a detailed examination of ore textures in thin section and hand specimen, with an emphasis on relict textures that might be related to primary ore genesis. The characterisation of primitive sulphide textures, completed as part of this study, is critical in understanding primary ore genesis and the effects of subsequent deformation and alteration. Also examined for the first time are the controls on the spatial distribution of the ore relative to interpreted mine-scale growth faults.

5.2. Previous work

Previous studies of the Rosebery ores have examined aspects of the primary mineralogy, morphology and extent of the ore lenses (Williams, 1960; Brathwaite, 1969, 1974; Green et al., 1981; Green, 1983; Huston, 1988; Huston and Large, 1988; Aerden, 1991; Klaere, 1997; Solly, 1999; Vallerine, 2000) and the effects of Devonian metasomatism on the ore (Vokes, 1983; Solomon et al., 1987; Khin Zaw, 1991; Khin Zaw and Large, 1996; Khin Zaw et al., 1997; Westbrook, 1997; Khin Zaw et al., 1999).

Early workers including Stillwell (1934) and Williams (1960) undertook short descriptive studies on the ore mineralogy within the sulphide lodes at Rosebery.
Figure 5.1 Ore lens distribution and Cambrian fault traces.
TSV true thickness contoured at 10 m intervals with interpreted syn-depositional fault traces and ore lens outlines. Data (841 points) derived from mine drill and underground mapping datasets, including data collected by Graves et al. (1998). Includes small-scale structures identified by Berry (1991, 1992) in the vicinity of AB lens. Ore lens outlines from mine site production plan 2002/2003.)
Brathwaite (1969; 1974) examined the mineralogy and morphology of sulphides within the ore and barite lodes in the upper mine levels, concluding that aside from a few minor occurrences of framboidal and “colloform” structures in pyrite the bulk of the ore displayed textures indicative of metamorphic recrystallisation. The ore lens morphology, mineral zonation, and banding within the ore were utilised to formulate an exhalative mode of ore formation. The pyrrhotite-pyrite replacement of sphalerite-galena ore was attributed to a metamorphic origin.

Green et al. (1981) and Green (1983) also examined the ore mineralogy, morphology and mineralogical zonation. Ore lens morphology and layering within the ore were utilised as evidence to support an exhalative depositional sequence from early chalcopyrite-pyrite ore to later sphalerite-galena ore, with a final phase of barite mineralisation.

Huston (1988) and Huston and Large (1988) examined the ore mineralogy and mineral zonation at the northern end of the mine with the aim of establishing a geological context for precious metal occurrences within the ore system. Mineralogical zonation within the ore was interpreted to be the result of zone refinement within exhalative sulphide mineralisation that was originally deposited on the seafloor.

Aerden (1991) utilised the ore mineral distribution within the southern sulphide lodes to support his interpretation of a Devonian epigenetic origin for the ore body.

Studies of individual ore lenses by Klaere (1997), Westbrook (1997), Solly (1999) and Vallerine (2000) were primarily descriptive in nature and included work on ore lens morphology, mineral zonation and ore mineralogy. Aside from a few occurrences of relict framboidal and colloform pyrite the ore was interpreted to exhibit predominantly metamorphic textures.

Work completed by several workers including Vokes (1983) and Solomon et al. (1987) established that the pyrite-pyrrhotite replacement of sulphide ore at the southern end of the mine was the result of chemical remobilisation by metasomatic fluids derived from Devonian granite. In a series of studies Khin Zaw (1991), Khin Zaw and Large (1996), Khin Zaw et al. (1997) and Khin Zaw et al. (1999) further defined the physical and compositional properties of the metasomatic fluids and their effects on primary mineralisation. A study of the intensely metasomatised J lens ore by Westbrook (1997) was primarily descriptive in nature and confirmed the metasomatic alteration pattern described by previous workers.

5.3. **Ore lens morphology and distribution**

The Rosebery ore body is made up of a series of ore lenses that are podiform to sheet-like in morphology and occur either as a single lens or as a series of stacked sub-lenses (figures 3.8
to 3.12). The lenses that comprise single, generally sheet-like bodies of ore are typically located within TSV sediments less than 50m in total thickness (e.g. K lens, AB lens, F lens, J lens) (fig.5.1), whereas multiple stacked sub-lens systems are located within TSV sediments greater than 50m in total thickness (e.g. C lens, D lens, E lens, G lens, H lens, P lens). The individual lenses vary in size from less than 100m up to 500m in lateral extent and they are typically less than 20m in thickness. In long section two trends are apparent in the spatial distribution of the ore lenses at a mine scale (fig.5.1). The southern lenses plunge from the central part of the mine southeast toward RS lens and T lens, and the northern ore lenses plunge northeast from the central part of the mine toward K lens. The area between these two trends contains no significant mineralisation except for the small, poddy and sub-economic V, M and RS lenses. The ore lenses are located along the edge of what is interpreted here to be to be a fault-bound basin, the margins of which are now defined by a rapid increase in TSV sediment thickness. The two principal trends in ore lens distribution also correspond to the two areas of intense footwall alteration delineated in earlier studies (Naschwitz, 1985; Naschwitz and van Moort, 1991).

The delineation of ore lenses is by definition based on economic parameters and they therefore mostly comprise bodies of semi-massive to massive sulphide, although in some cases the concentration of disseminated or stringer style mineralisation within the host rock and barite bodies is sufficient to warrant mining. The semi-massive and massive sulphide ore lenses typically have sharp upper and lower contacts (Hall et al., 1953; Brathwaite, 1969, 1974; Burton, 1975; and this study), although the lower contacts are often underlain by host rock containing disseminated or stringer sulphide mineralisation. At their lateral margins the massive sulphide ore typically grades through disseminated sulphide mineralisation into massive carbonate or altered host rock (Brathwaite, 1969, 1974; Burton, 1975; Berry et al., 1998), with the massive ore lenses sometimes displaying an interfingering or “feathering out” relationship with these units (Hall et al., 1953; Berry et al., 1998).

Most of the ore lenses are hosted within the top of the footwall pumice breccia and the overlying basal units of the TSV, with the remainder occurring at stratigraphically higher levels within the TSV. At the northern end of the mine K lens is hosted primarily within the upper units of the footwall pumice breccia (fig.3.8), whereas the P sub-lenses are located stratigraphically higher within the TSV (figures 3.11 to 3.12). The immobile element signature of K lens ore is the same as the footwall pumiceous mass flow units within which it is hosted (fig.5.2), which strongly indicates sulphide replacement of the host sediments. At the southern end of the mine, G lens and H lens occupy higher stratigraphic levels within the TSV than the underlying D lens and E lens (fig.5.3). The lateral margins of some ore lenses appear to traverse stratigraphy, with the up-dip margins sometimes extending into higher levels within the TSV. At deeper levels in the mine, AB lens underlies the quartz-feldspar-phyric peperite sill and is hosted within the upper part of the footwall pumice breccia, whereas at shallower levels AB
Figure 5.2 Immobile element ratios—K lens ore samples
A) Ore samples taken from a drill traverse at ~2400mRL plot along the main trends for the footwall pumice breccia and the more dacitic upper pumice breccia units defined by Large et al. (2001a). This indicates that K lens sulphide ore is hosted within and probably replaces the pumice breccia units. Useful immobile element data is returned even from massive sulphide ore because of the preservation of Ti and Zr bearing minerals such as those illustrated below. (includes unpublished data of Large et al., 2001a).
B) Relict grain after augite? now defined in part by rutile grains in quartz matrix and overprinted by galena, sphalerite and euhedral pyrite grains. [rl, fov=550µ, C306].
C) Relict grain after augite?) now defined in part by rutile grains in sphalerite-pyrite matrix. [rl, fov=550µ, C285].
Chapter 5

Lens extends up into the TSV package (Allen, 1992; Reid, 1993).

North of ~1000mN AB lens is overlain by a continuous sheet-like body of massive barite that is for the most part separated from the underlying massive sulphide by altered sediments (Huston and Large, 1988). Similarly the baritic H lens is located in the hanging-wall of D and E sulphide lenses, separated from the sulphide lodes by an interval of altered and weakly mineralised sediments (fig.5.3) (Green et al., 1981; Green, 1983). G lens sulphide ore in turn overlies the baritic H lens and they are again separated by altered and weakly mineralised sediments (Green et al., 1981; Green, 1983). The barite lenses in places grade laterally into and interfinger with massive sulphide ore (Brathwaite, 1969, 1974).

5.3.1. K lens and P lens

Located at the northern end of the mine, K lens and P lens represent two possible end-members in terms of lens morphology and stratigraphic setting (figures 3.8 & 3.12). K lens comprises a single tabular body of semi-massive to massive sulphide that occasionally bifurcates and contains lenses of less mineralised host rock. It is hosted within the upper part of the footwall pumice breccia and in places extends into sediments of the TSV. The upper ore margin is typically sharp, whereas the lower margin is gradational into disseminated sulphides within altered footwall rocks. The lens is almost entirely overlain by a quartz-feldspar-phyric peperite sill (fig.3.15) that, except where disrupted by later deformation, is always separated from the ore by a thin layer of altered and weakly mineralised sediment. Where the lower peperitic sill margin is somewhat embayed, the TSV sediments between the ore and the sill are thicker and display an increased level of disseminated mineralisation and alteration overprint. In a few locations, massive sulphide ore extends into and crosscuts the sill but, as these sulphide occurrences also crosscut syn-cleavage quartz-carbonate veining, this is most likely the result of sulphide remobilisation during deformation.

Although minor barite occurs throughout K lens there is a significant increase in barite toward the upper and lateral margins of the sulphide ore, particularly in the direction of AB lens, and there are minor barite lenses located above the quartz-feldspar-phyric sill. Massive carbonate also becomes more prevalent toward and outside the lateral margins of the ore. Intense hydrothermal alteration associated with K lens ore is confined to the footwall side of the peperitic sill.

In contrast, P lens is a somewhat anastomosing series of stacked disseminated to massive and often baritic sulphide sub-lenses hosted within the top of the footwall pumice breccia and at various stratigraphic levels within sediments of the TSV. The exact number of sub-lenses is difficult to ascertain due to the anastomosing and sometimes discontinuous nature of the mineralisation, with initial drilling defining 11 possible sub-lenses (Carnie, 1997) of which 5 are
currently being mined. Although P lens ore in places is as massive as that of K lens, it
generally contains more lenses, bands and fragments of altered host rock. Where the sub-
lenses are hosted within or overlain by stratified sediments, the coarser units appear to have
been preferentially mineralised. Unlike K lens, sulphide mineralisation occurs both above and
below the quartz-feldspar peperitic sill where it is present (fig.3.11), and strong mineralisation-
related alteration extends well into units of the TSV (fig.4.6). In comparison with K lens, the
occurrence of barite in association with P lens is much more extensive. Discontinuous lenses
of massive banded barite and barite altered sandstone are located within and immediately
overlying the P6, P7 and P9 sub-lenses (fig.3.12), whereas the P10 sub-lens is hosted within
massive barite and barite-rich sandstone. The P13 sulphide sub-lens overlies the baritic P10
sub-lens, from which weakly mineralised host rocks separate it. The pattern of sulphide sub-
lens and barite distribution within the stacked P lens system is somewhat analogous to the
complex stacked distribution of sulphide and barite lenses at the southern end of the mine,
which includes lenses D, E, G and H (fig.5.3).

5.4. Ore mineral zonation

The sulphide ore lenses predominantly comprise pyrite, sphalerite, galena and chalcopyrite
with minor tetrahedrite, tennantite and arsenopyrite, whereas the massive barite bodies
contain sphalerite and galena with minor chalcopyrite, pyrite, tennantite and tetrahedrite
(Green et al., 1981). Significant amounts of pyrrhotite, magnetite and haematite are also
present where the lodes have been overprinted by later metasomatism (Khin Zaw et al., 1997;
Khin Zaw et al., 1999).

Vertical mineralogical zonation through the ore lenses has been documented throughout the
mine. The sequence at the southern end of the mine typically comprises basal disseminated to
stringer pyrite-chalcopyrite-rich ore with discontinuous lenses of massive pyrite-chalcopyrite in
chlorite-altered rocks, which is overlain by sphalerite-galena-pyrite-rich disseminated to
massive sulphide ore that is often enriched in pyrite and chalcopyrite toward the base
(Grathwaite, 1969, 1974; Green et al., 1981; Green, 1983; Allen, 1992). The underlying pyrite-
chalcopyrite mineralisation is more restricted in lateral extent than the overlying sphalerite-
galena-rich ore (Brathwaite, 1969, 1974). Massive barite or barite-rich sandstones often overlie
sulphide ore lenses, typically with intervening intervals of weakly mineralised sediments
(Green, 1983; Huston, 1988). Significantly the vertical mineralogical zonation in places is
reversed. Within G lens, for example, pyrite-chalcopyrite enrichment occurs in the upper part
of the sulphide lode (figures 5.3 to 5.4) (Adams et al., 1976; Aerden, 1991). The barite
mineralisation that comprises H lens is present in both the hanging-wall and footwall to G lens
(fig.5.3).

Vertical mineralogical zonation of the ore is also apparent within the northern ore lenses. The
Chapter 5

Figure 5.3 Cross Section of Rosebery ore zone at ~150mN
This cross section illustrates the stacked ore lenses at the southern end of the mine and the down-dip trend to pyrite-chalcopyrite-rich ore. (modified after Green et al., 1981).

Figure 5.4 Copper and zinc distribution - 15 level.
Cu and Zn distribution on 15-level derived from mine grade control datasets, which at these concentrations are contained almost entirely within chalcopyrite and sphalerite respectively. Cu concentrations are elevated in the stratigraphically lower lenses, and in the position where the lenses merge and the TSV rapidly thickens. (modified after Aerden, 1991).
chalcopyrite-pyrite enrichment toward the base is however not as intense, consistent or laterally extensive as beneath the southern ore lenses. AB lens comprises sphalerite-galena ore with a pyrite-chalcopyrite-enriched base, and overlying massive barite mineralisation separated from the sulphide ore by an intervening interval of weakly mineralised sediment (Huston, 1988), and P6 sub-lens exhibits the same mineralogical zonation (fig.3.12). K lens and most of the P sub-lenses exhibit weak, patchy pyrite-chalcopyrite enrichment toward the base of sphalerite-galena ore, however the same style of mineralisation is also present at higher levels within the sphalerite-galena ore. Massive barite associated with K lens most commonly occurs at the top of the sphalerite-galena ore without intervening sediment. Barite is also present at the base of the sphalerite-galena ore in some instances. The P7/P9 and P10/P11 sub-lenses tend to contain significant amounts of disseminated barite, with lenses of massive barite present along both upper and lower margins of the sulphide ore.

Some previous workers (Brathwaite, 1974; Green et al., 1981; Green, 1983; Lees et al., 1990) utilised mineral zonation within the ore as a facing indicator, to aid in the interpretation of macro-scale folding of the southern ore lenses (fig.3.21). The use of mineralisation as a marker horizon, and the internal mineral zonation as a facing indicator, is based on the premise that the southern ore lenses originally formed as a single exhalative sheet-like body on the seafloor. The other assumption made was that the mineral zonation within the ore was consistent, and that the inverted mineral zonation within G lens was the result of macro-scale folding. The presence of ore lenses within different levels of the stratigraphy (Allen, 1992; and this study) suggests that the ore did not originally form as a single sheet-like body. The inverted mineral zonation of G lens might therefore be alternatively explained through the formation of multiple exhalative ore lenses through time, or as the product of sub-seafloor replacement. As discussed in a previous chapter, the lack of evidence for macro-scale folding and the restriction of west facing to minor fold hinges (Berry, 1990a, b), indicates that folding has not affected the mine stratigraphy at a macro-scale.

Lateral mineral zonation in the ore lenses is not as well documented. Green et al. (1981) described broad trends of decreasing chalcopyrite content toward the south in the southern ore lenses and decreasing chalcopyrite content toward the north in the northern ore lenses. In more detailed work, Aerden (1991) utilised grade control data from the southern ore lenses to demonstrate that chalcopyrite is concentrated in the area where G lens and E lens adjoin F lens, and along the down-dip margins of the same ore lenses (figures 5.3 & 5.4). The down-dip margins of these lenses are also enriched in pyrite (Sainty, 1986). Chalcopyrite-pyrite enrichment is also evident along the southern margins of F lens and J lens, and along the down-dip margin of J lens (Sainty, 1986). Although this zonation in F lens and J lens ore may be a primary feature, it is possible that the zonation may also in part reflect the removal of sphalerite, galena and tetrahedrite from the sulphide ore by granite-related metasomatic fluids (Khin Zaw et al., 1999).
5.5. Ore textures and fabrics

5.5.1. Disseminated mineralisation

Intensely quartz-altered footwall rocks beneath K lens and P lens typically contain disseminated sulphide mineralisation. Within quartz-altered footwall rocks, sphalerite occurs as dark brown wispy lenses up to ~2cm in length and aligned broadly parallel to stratigraphy (fig.5.5a), a morphology that is similar to that of the chlorite-pseudofiamme found elsewhere in the altered footwall pumice breccias. Pyrite, sphalerite and occasionally galena form disseminated grains and clots that have a similar dimension and pattern of distribution to that of feldspars and their phyllosilicate-altered equivalents in the footwall pumice breccias. Several occurrences of stringer-like and stockwork sphalerite-rich sulphide veining also cross cut intensely quartz-altered footwall pumice breccias in the footwall to K lens ore (fig.5.5b).

Within stratified volcaniclastic sediments in the vicinity of P lens disseminated sphalerite-rich mineralisation occurs preferentially within coarser-grained volcaniclastic sandstone units (fig.5.5c), whereas the finer-grained units are preferentially quartz-altered. Barite lenses and bands within the TSV sometimes exhibit clastic textures comparable to the surrounding sediments (fig.5.5d), indicating that barite has at some time replaced the host sediment matrix.

K lens ore is typically more massive and does not contain as much altered host rock material as P lens, but in places the massive ore in K lens is transitional to a disseminated and semi-massive style of mineralisation hosted by intensely sericite ± quartz ± carbonate altered rocks. This style of mineralisation appears diffuse and patchy in hand specimen (fig.5.5e) and occurs above, below and within massive sulphide ore over thicknesses of several metres. This style of mineralisation contains atoll, spongiform and colloform textures preserved in pyrite, indicating a primary origin.

5.5.2. Massive sulphide ore

In many parts of the mine the most prominent feature of the massive ore is a compositional banding (fig.5.6) that is in places aligned parallel to bedding and elsewhere varies from parallel to normal to cleavage (Brathwaite, 1972, 1974; Aerden, 1991; and this study). Within the Cu-rich southern ore lenses the banding is prominent, and comprises alternating layers of pyrite-rich, sphalerite-rich, chalcopyrite-rich and silicate-rich material from 0.1 to 10cm in thickness and a lateral extent of up to 20m (Brathwaite, 1974; Green, 1983). This style of banding is absent in much of the sphalerite-galena-rich ore of F lens (Green, 1983) and K lens, and most of P lens with the exception of the P13 sub-lens. Instead, the banding in these Cu-poor lenses is more diffuse in nature, and similar to banding within the massive barite bodies (fig.5.6). The banding in barite and sulphide lodes may in places be complexly folded at a meso-scale, and
Figure 5.5 Disseminated mineralisation.
A) Quartz-altered footwall pumice breccia with wispy lenses of dark red-brown sphalerite oriented broadly parallel to stratigraphy. [R6051-55.6m, core≈35mm diam.].
B) Quartz altered footwall pumice breccia with wispy brown sphalerite lenses aligned broadly parallel to stratigraphy and crosscut by zoned sphalerite veins of same generation. [R6849-2.0m, core≈35mm diam.].
C) Well-stratified quartz-sericite-altered grey siltstones and sandstones above P lens with pale brown sphalerite-rich mineralisation almost exclusively hosted in the coarser sediment bands. [30P crosscut, 5m below P7 sub-lens].
Figure 5.5 Disseminated mineralisation.
D) White barite replacing sericite-carbonate altered volcaniclastic sandstone unit with quartz and lithic grains retained in the barite domains. [R6805-54.1m, core≈35mm diam.].
E) Disseminated red-brown sphalerite-galena mineralisation in sericite-quartz altered pumice breccia unit. [R6587-82.8m, core≈17mm diam.].

Figure 5.6 Banded barite and sulphide ore.
A) Intensely folded, massive banded barite and carbonate-sericite-pyrite schist. [39P crosscut, hanging wall to P10 sub-lens].
B) Banded sphalerite-pyrite (dark) and pyrite-chalcopyrite ore (light). The ore banding is folded and in places offset by late brittle faults (F4).
an axial planar cleavage is often preserved in the silicate and barite layers (Berry, 1990a). An axial planar cleavage is not typically preserved in the folded sulphide layers but Solomon and Groves (2000) have observed some examples of axial planar cleavage associated with meso-scale folding of sulphide ore. Brathwaite (1972) and Solomon et al. (1987) have also published an example of small-scale symmetrical folding in banded sulphide ore that exhibits a weak axial planar cleavage.

Another form of banding is evident in the relatively undeformed parts of K lens and P lens where sphalerite-rich massive sulphide contains fine-grained, sometimes colloform pyrite-quartz bands or veins that range from 1cm to ~30cm in thickness (fig.5.7a-b). The pyrite-quartz veins are aligned both parallel to, and at an angle to, adjacent host rock bedding and cleavage. A few examples contain fragments of sphalerite-rich ore with diffuse margins, which indicates that some of these veins post-date the sphalerite-rich ore. The sphalerite-rich ore and the pyrite-quartz bands both contain primitive framboid, atoll and colloform pyrite morphologies, indicating that this form of banding is a feature of primary ore formation. The sphalerite-rich ore also contains bands of coarser-grained pyrite-quartz ± chalcopyrite (fig.5.7c) that appear broadly synchronous with primary ore formation, and may represent higher temperature equivalents of the colloform pyrite-quartz veins. These pyrite-chalcopyrite bands may also be a precursor to the prominent compositional style of banding present within the southern Cu-rich ore lenses.

Pyrite-rich and sphalerite-rich banding produced during deformation is also evident in shear zones, particularly in the zones of dilation such as within the pressure shadow of boudinaged lithic material (fig.5.7d). The pyrite in these cases is crystalline and often coarser-grained, and sphalerite in the selvages appears darker in colour due to Fe enrichment.

In places the sulphide ore exhibits a fine millimetre scale compositional banding, typically comprising laminae of pyrite and dark red-brown sphalerite (fig.5.7e), in association with rotated and boudinaged fragments of silicate and pyrite (Berry, 1990b; Solomon and Groves, 2000). This banded ore in places exhibits a mylonitic fabric and shear banding with an orientation consistent with reverse movement (Berry, 1990b). Formation of this style of fine banding has been attributed to tectonic deformation by some workers (Berry, 1990b; Solomon and Groves, 2000). This style of banding is most likely due to the combination of primary pyrite-rich banding and veining (e.g. fig.5.7a-c) that is attenuated during deformation with the formation of new pyrite-rich domains as the result of chemical sulphide remobilisation (e.g. fig.5.7d).

5.5.3. Lithic and baritic fragments

Lithic fragments are a common feature of the semi-massive to massive sulphide ore of K lens
and P lens. A few rare examples are clearly epiclastic fragments of amygdaloidal rhyolite lava (fig.5.8a), whereas some appear to represent mineralised fragments of quartz-altered host rock (fig.5.8b).

The majority of lithic material contained within semi-massive to massive sulphide ore comprises platy to well-rounded, occasionally mineralised, intensely quartz and/or sericite altered rock fragments with the degree of boudinage, cleavage and folding development largely dependant on phyllosilicate content (fig.5.8c-f)). The altered host rock fragments do not anywhere exhibit relict textures indicative of coarse-grained volcaniclastic sediments, but are typically fine-grained in nature. The orientation of cleavage within the sericite-altered fragments varies significantly (fig.5.8d), and in underground exposures the fragments are typically distributed in planes that parallel the main cleavage and local ore banding (fig.5.8f). The distribution and morphology of lithic fragments within sulphide ore, the lithological similarity of the lithic fragments to surrounding host rocks, the random orientation of cleavage and fold hinges in different lithic fragments, and the presence of similar lithic clasts within shears in sulphide ore can together be interpreted as durchbewegung fabrics (Vokes, 1969). These fabrics were generated during deformation by disruption, rotation and “milling” of lithic material derived from relict altered host rock horizons within the massive ore and from the surrounding wall rocks.

Another significant feature of K lens and P lens massive sulphide ore is the presence of white, angular to well-rounded, recrystallised barite fragments (fig.5.8g). Similar barite fragments have been reported from AB lens, and in one example pass laterally into discontinuous streaks parallel to the ore banding (Green et al., 1981). The barite fragments are up to ~5cm in size and have diffuse to sharp margins. The barite fragments may also represent durchbewegung textures as the result of ore deformation, but the original morphology is equivocal. Green et al. (1981) interpreted interlayered barite, carbonate and sulphide laminae elsewhere in the mine as being a primary feature (e.g. fig.5.6), and wispy barite veins observed in the footwall of F lens were interpreted to post-date sulphide ore. The barite fragments may represent the remnants of massive barite replaced by sulphide, or they may represent the physically reworked remnants of barite laminae and veins that were originally present within the sulphide ore. Both processes probably played a role in the formation of these barite fragments.

5.5.4. Vein-style mineralisation

As discussed previously, massive pyrite-quartz veins in places cut the massive sulphide and barite ore (fig.5.7a-c). These veins contain primitive textures preserved in pyrite and have been variably affected by later deformation. Deformed chalcopyrite-pyrite stockwork-like veins are an occasional feature of the chlorite-rich alteration beneath the massive sulphide ore (Sainty, 1986; Allen, 1992), although they are rare beneath K lens and P lens ore. Occasional
Figure 5.7 Banding in sulphide ore. A) Colloform pyrite-quartz veins cutting quartz-altered rock and sphalerite-rich ore. [R7251-36.1m, core=35mm diam.]. B) Colloform pyrite-quartz band in sphalerite-rich ore. The same vein set contains rare fragments of sphalerite ore with altered rims. This indicates that the pyrite veins post-date the sphalerite-galena ore. [R6476-117.4m, core=35mm diam.]. C) Pyrite bands or vein in sphalerite-rich ore. Vein to left contains a chalcopyrite-quartz core. [R7251-47.6m, core=35mm diam.]. D) Syn-deformation pyrite vein developed in lee of boudinaged host rock (see fig. 5.18a) and exhibiting Fe-rich sphalerite selvage. (33K crosscut). [fov=14cm]. E) Banded pyrite-sphalerite ore (surface open cut). [fov=8.5cm].
Figure 5.8 Lithic material in sulphide ore.
A) Amygdaloidal rhyolite clast in massive sphalerite ore with zoned sphalerite infill. [R5964-88.5m, core≈17mm diam.]. B) Clast of laminated quartz-sulphide altered host in massive sphalerite-rich ore. [R6426-107.6m, core≈35mm diam.]. C) Randomly oriented quartz-altered host rock with spaced cleavage in massive pyrite-rich ore. [122R-1579.5m, core≈48mm diam.]. D) Randomly oriented sericite-altered host rock with folded cleavage in massive sphalerite-rich ore. [R5192-D1-636.9m, core≈35mm diam.].
Figure 5.8 Lithic material in sulphide ore.
E) Boudin of quartz-altered host rock in banded sphalerite-pyrite ore within sheared sulphide ore [33K crosscut - see also fig. 5.19a, fov=16cm].
F) Randomly oriented sericite-altered host rock strung out along banding in massive sphalerite-rich ore. [31P ore drive, fov=20cm].
G) White sugary barite fragments in massive sphalerite-rich ore. [R6426-112.5m, core=35mm diam.].
Figure 5.9  Pre-deformation veins.
A) Pyrite-quartz vein in quartz-altered volcanic siltstone. (R5824-372.7m). [core ≈ 35mm diam.]
B) pyrite-quartz-albite vein in quartz-altered volcanic siltstone. Dark grey quartz pseudomorphs after anhydrite? in vein selvage. (R5824-350.6m). [core ≈ 35mm diam.]

Figure 5.10  Bladed mineral textures.
A) Quartz after barite blades sited within slightly more quartz-rich domain in sericite-altered host rock. [R7251-41m, core ≈ 35mm diam.]
B) Bladed textures after barite/anhydrite? preserved in brown (oxidised) carbonate and sugary white barite. [R5384-242m, core ≈ 35mm diam.]
examples of weakly deformed sphalerite-rich veins are present in the intensely quartz-altered footwall (fig.5.5b) and appear to have been formed at the same time as the wispy sphalerite lenses. The base of the intensely quartz-altered shale package overlying P lens is cut by a set of highly deformed pyrite-quartz and quartz-barite-pyrite veins (fig.5.9), which may contain albite grains. In a few cases quartz psuedomorphs, possibly after anhydrite, occur in the selvages of the veins (fig.5.9b). All of these vein types are everywhere affected by the main (S₃) cleavage. Some contain primitive pyrite morphologies, and some are found only within or proximal to bodies of massive sulphide and barite. They all display a strong spatial association with specific alteration assemblages associated with primary ore formation, all of which suggests that they formed as part of the primary ore forming process.

Syn- to post-deformation quartz-carbonate-sulphide and sulphide-rich veins and faults are overall a minor component of the ore (fig.4.9b-c), and typically carry sphalerite, pyrite, chalcopyrite, galena, minor tetrahedrite and other sulphosalts. These veins are found throughout the host sequence and they exhibit an increase in sulphide content proximal to ore. Late- to post-deformation vein sets related to Devonian metasomatism carry only a minor sulphide component, including chalcopyrite, pyrite, pyrrhotite, sphalerite, minor galena tetrahedrite and other sulphosalts (fig.4.11a,f).

5.5.5. Other primary textures

Several examples of coarse bladed textures after barite or anhydrite have been preserved as quartz, carbonate and barite pseudomorphs within the ore and surrounding host rock (fig.5.10). Although rare, these textures indicate that sulphate mineralisation occurred within the sediments.

5.6. Sulphide mineralogy and textures

Almost all of the sulphide and barite mineralisation at Rosebery exhibits metamorphic textures at the microscopic scale (fig.5.11). Sphalerite, galena, chalcopyrite, tetrahedrite and barite typically display granoblastic textures consistent with post-deformational static annealing (Brathwaite, 1974; Green et al., 1981). Minerals within the massive ore more resistant to recrystallisation, such as pyrite and arsenopyrite, are typically present as porphyroblastic anhedral to euhedral crystals and aggregates within a matrix of annealed sulphide or barite (Brathwaite, 1974; Green et al., 1981). Relict primary textures in samples from previous studies appear to have been rare occurrences or were not recognised. In the current study approximately half of the mineralised specimens examined from K lens and P lens contained relict primary textures, most preserved within pyrite. These textures, documented in detail for the first time, are key indicators of the timing and mode of primary ore genesis.
5.6.1. Pyrite

Pyrite occurs as fine disseminations throughout the host sequence, as disseminations, bands, veins and lenses within and proximal to massive sulphide ore and massive barite, as a component in syn- to post-deformation quartz-carbonate veins and faults, and as a component in alteration and veining associated with Devonian metasomatism.

Pyrite withstands deformation more readily than many other sulphides due to its micro-hardness and resistance to ductile deformation, which leads it to behave as a brittle substance under typical crustal conditions, although it can exhibit a tendency to recrystallise at greenschist grades of metamorphism and above (Craig and Vokes, 1993). Pyrite in the Rosebery ore is the mineral most likely to have preserved primary textures that can be related to ore formation and subsequent changes. Previous workers have described occasional relict frambooidal and colloform morphologies preserved in pyrite within the sulphide ore of K lens, AB lens and J lens (Brathwaite, 1974; Westbrook, 1997; Fander, 1998; Solly, 1999), although most of these examples appear to have been heavily recrystallised. This study focussed on samples from K lens, P lens and the surrounding host rocks, as they appear to have been less affected by deformation and metasomatic alteration than most other parts of the mine.
Pyrite framoids, framboidal overgrowths and fine-grained euhedra occur within weakly sericite-quartz altered fine-grained volcaniclastic sediments throughout much of the TSV, and in microcrystalline quartz and Mn-rich carbonate proximal to ore (fig.5.12). Very fine-grained pyrite in weakly altered host sediments has also overgrown and intergrown with possible primary rock textures (fig.5.12b-d) and it appears to have formed in the same environment as the pyrite framoids. The framoids range from ~2.5 to 42.5 µm in diameter (mode ~10 µm) with no apparent variation relative to the presence or absence of ore-related sulphides (fig.5.13). Comparison of the relatively large mean diameter and broad size range of framoids in the relatively unaltered units of the TSV with framboid populations in recent sedimentary environments (fig.5.13) indicates a likely formation within sediments at low temperatures (<200°C) during early diagenesis, rather than by precipitation in the overlying water column (Wilkin et al., 1996; Wilkin and Barnes, 1997). Relict pyrite framoids are occasionally found within the sphalerite and galena component of sulphide ore, and in places crystalline pyrite contains circular relict framboidal pyrite cores. The relict framoids preserved in a sulphide matrix often have irregular morphologies indicating a degree of dissolution, and nowhere are framoids to be seen post-dating other sulphides. However, framoids are present within some pyrite-quartz veins that post-date massive sphalerite-galena ore (fig.5.7a-b). The framoids are in part hosted within cryptocrystalline chert and in part within coarser-grained quartz that also contains pyrite with colloform, atoll and crystalline morphologies (fig.5.12e-f). Framboids are also occasionally evident within sediment clots incorporated into the peperitic intrusion overlying K lens.

Pyrite within the ore horizon also exhibits a range of spongiform, atoll, radial and skeletal morphologies. These textures are best preserved in quartz, sphalerite-galena-rich and pyrite-rich ore, becoming more crystalline and less well preserved with increasing chalcopyrite content. Quartz associated with these textures in sulphide ore is more coarsely crystalline than that which elsewhere hosts the pyrite framoids (fig.5.12e-f). Spongiform pyrite is a common feature of the ore specimens examined during this study (fig.5.14a-c), with relict examples occurring in more than half of the samples examined. The spongiform pyrite is most commonly intergrown with galena and minor sphalerite, chalcopyrite, tetrahedrite and quartz. In places the spongiform pyrite occurs as bands up to ~1cm in thickness within massive sulphide ore. Only in a rare examples does spongiform pyrite overprint more crystalline pyrite (fig.5.14c), suggesting that pyrite with primitive textures generally predates the more crystalline pyrite throughout the ore.

Pyrite atolls are present both within the massive ore and within related pyrite-quartz veins (fig.5.7a-b). The atolls exhibit a range of morphologies (fig.5.14d-I), with many containing a pyrite framoid at the core. A sample of colloform pyrite vein from K lens appears to display a range of textures from which a paragenetic sequence may be drawn. The textures include simple pyrite framoids in microcrystalline quartz (c.f. fig.5.12a), spherical accumulations of
quartz with radially and concentrically arranged pyrite that typically contain pyrite frambooids at the centre (fig.5.14d-e), atoll and colloform morphologies located at the margins of spherical silica accumulations (fig.5.14f), and variations of the above due to further sulphide overgrowth and infill (fig.5.14g-k). In places, more coarsely crystalline equivalents also occur (fig.5.14l). The indication is that pyrite frambooids were nucleation sites for spherical mineral accumulations, possibly a hydrated silica gel, onto which more crystalline pyrite and other sulphides could then precipitate from hydrothermal or diagenetic fluids. An analogous paragenesis has been proposed for a range of similar pyrite atoll morphologies found within syngenetic base metal deposits in the Tasman fold belt of New South Wales (England and Ostwald, 1993).

Colloform pyrite textures are typically intergrown with fine-grained galena, sphalerite and minor chalcopyrite and tetrahedrite (fig.5.15a-d). The relict colloform textures are typically more recrystallised in samples with less quartz and higher chalcopyrite content (fig.5.15e-f). Skeletal pyrite morphologies occur within a sulphide matrix and can be attributed to several modes of formation. In places pyrite appears to have overgrown incipient minerals such as sphalerite and anhydrite that have later recrystallised or been replaced (fig.5.15g-k). Rare examples may be after partial replacement of pyrite or after conversion of pyrrhotite to pyrite (fig.5.15l).

Within ore specimens, primitive fine-grained pyrite textures are sometimes intimately intergrown with coarser, more euahedral pyrite and arsenopyrite (e.g. fig.5.14a) but are more commonly recrystallised and overprinted, overgrown and replaced by coarser pyrite and arsenopyrite (e.g. fig.5.15f). Where they exist as isolated grains, the coarser, more crystalline pyrite is anhedral to euahedral in morphology and exhibits caries texture where in contact with sphalerite (fig.5.16a-b). Lack of obvious Fe enrichment in sphalerite surrounding pyrite exhibiting caries texture suggests that dissolution of the pyrite occurred prior to annealing of the sphalerite. Rounded, anhedral grains typically contain inclusions of galena, sphalerite and chalcopyrite and appear to have formed through recrystallisation of spongiform pyrite (fig.5.16a). The more euahedral grains comprise clean crystalline pyrite but often contain anhedral cores of variably infilled and recrystallised relict framboidal, atoll and spongiform pyrite (fig.5.16c-d). The recrystallised spongiform cores still retain fine inclusions predominantly comprising galena, sphalerite and quartz, with grains of electrum occasionally observed at the core margins. These spongiform pyrite cores, and the distribution of inclusions therein, are not consistent with formation through the growth of pyrite and incorporation of adjacent mineral components during metamorphic deformation (Craig and Vokes, 1993). A few examples of porous pyrite cores exhibit textures consistent with the conversion of marcasite to pyrite (fig.5.14e), and these do not contain the sulphide inclusions that elsewhere suggest derivation from relict spongiform pyrite. The crystalline pyrite euhedra and rims typically appear clean and free of inclusions (fig.5.16c), but in places compositional zonation is visible (5.16g). Crystalline pyrite euhedra and rims on earlier pyrite appear to have formed at various
Figure 5.12  Pyrite textures –framboids and overgrowths.
A) Pyrite framboids with traces of sphalerite and galena in quartz matrix. [rl, fov=550µ, C261].
B) Fine-grained pyrite overgrowth with minor sphalerite on relict glass shard?. [rl, fov=1.1mm, C658].
C) Fine pyrite overgrowth on relict augite? grains now defined by rutile. [rl, fov=1.1mm, C558].
D) Irregular fine-grained pyrite overgrowth on relict pumice fragment? [rl, fov=1.1mm, C184].
E) Framboidal pyrite in microcrystalline quartz (LHS) overprinted by spongiform pyrite-galena and
coarser-grained euhedral pyrite in relatively coarse-grained quartz matrix (RHS). [rl, fov=2.2mm,
C184].
F) As previous image. tl-xn.
Chapter 5

Figure 5.13 Framboid size distribution.
A) Pyrite framboid size distribution in mineralised samples and in weakly to moderately altered TSV sediments. The similarity in size distribution and overprinting of framboids by ore-related sulphides indicates a likely formation prior to ore-related mineralisation for most framboids.
B) Plots of the mean versus the standard deviation and mean versus the skewness of framboid size distributions. Data includes modern and ancient examples of framboids within sediments underlying sulphidic water columns (euxinic), and within sediments underlying oxic or dysoxic water columns. The Rosebery TSV framboid data plots within the oxic-dysoxic field. (modified after Wilkin et al., 1996).
Figure 5.14 Pyrite textures—spongiform and atolls.
A) Spongiform pyrite-galena intergrown with euhedral to anhedral pyrite in sphalerite-quartz. [rf, fov=2.2mm, C799].
B) Spongiform pyrite in galena-quartz-sphalerite. [rl, fov=2.2mm, C797].
C) Spongiform pyrite and euhedral pyrite grain. The spongiform pyrite appears to have intergrown with or overgrown the euhedral pyrite grain. [rl, fov=1.1mm, C338].
D) Concentrically and radially distributed fine-grained pyrite in quartz with framoidal core. The radial texture may indicate pyrite nucleation within a gel medium or the concentric accumulation of quartz and fine-grained pyrite layers [rl,fov=130µ, C465].
E) Coalesced spherical domains of quartz with fine disseminated pyrite and coarser-grained pyrite rims. [rl,fov=340µ, C465].
F) Coalesced spherical domains of quartz with fine disseminated pyrite and coarser-grained pyrite-sphalerite rims, some containing framoidal pyrite cores. Coarser-grained rims appear to have developed at the same time as the more crystalline pyrite and sphalerite in the top left of the image. [rl,fov=550µ, C465].
Figure 5.14 Pyrite textures—spongiform and atolls.

G) Coalesced pyrite atolls and spongiform to euhedral pyrite. Some of which have frambooidal cores and are filled with sphalerite. [rl, fov=550µ, C465].

H) Banded pyrite-sphalerite atoll with frambooidal pyrite core. [rl, fov=220µ, C465].

I) Concentric and radial textures in pyrite atoll with frambooidal pyrite core and galena fill. [rl, fov=280µ, C465].

J) Pyrite atolls, some of which are overprinted with euhedral pyrite. [rl, fov=340µ, C465].

K) Coalesced mass of pyrite frambooids with spongiform pyrite rim and overprinted by more euhedral pyrite (top). [rl, fov=550µ, C465].

L) Recrystallised pyrite possibly after atoll texture with relict frambooidal pyrite cores. [rl, fov=280µ, C465].
Figure 5.15 Pyrite textures—colloform and skeletal.
A) Colloform pyrite-galena in sphalerite-quartz matrix. [ril, fov=1.1mm, C362].
B) Colloform pyrite-galena in sphalerite-quartz matrix. [ril, fov=1.1mm, C362].
C) Colloform and euhedral crystalline pyrite in sphalerite-quartz matrix. [ril, fov=1.1mm, C838].
D) Spongiform pyrite-galena spheroid in quartz matrix. [ril, fov=440µ, C362].
E) Relict colloform pyrite-galena in sphalerite matrix. [ril, fov=1.1mm, C799].
F) Relict radial, spongiform and crystalline euhedral pyrite in sphalerite-chalcopyrite-galena. [ril, fov=550µ, BV227].
Figure 5.15 Pyrite textures—colloform and skeletal.
G) Skeletal pyrite in galena-sphalerite with chalcopyrite fill. Structure after early pyrite overgrowth and replacement along sphalerite grain boundaries. Sphalerite subsequently replaced by chalcopyrite. (c.f. fig.14 Hannington et al. (1999)). [rl, fov=550µ, C285]
H) Similar structures and origin to previous image. [rl, fov=550µ, BV223]
I) Skeletal pyrite with galena, quartz, chalcopyrite, sphalerite. Structure after overgrowth and growth along grain boundaries of sphalerite or galena? [rl, fov=550µ, BV223]
K) Skeletal pyrite in sphalerite. Structure after overgrowth or intergrowth with radial mineral structure—anhdyrite? [rl, fov=1.1mm, C422]
L) Skeletal pyrite with galena in sphalerite. Structure possibly after conversion of pyrrhotite to pyrite (c.f. fig.6a Murowchick (1992)). [rl, fov=550µ, BV078].
Figure 5.16 Pyrite textures—other.

A) Typical anhedral pyrite grains with very fine-grained inclusions and displaying caries texture in sphalerite-galena. The anhedral grains represent relict spongiform pyrite masses. [rl, fov=2.2mm, C345].

B) Typical euhedral pyrite with caries texture in galena. Note curved cleavage pit traces in galena. [rl, fov=1.4mm, BV127].

C) Euhedral pyrite in quartz-sercite matrix. Note anhedral spongiform pyrite cores with very fine-grained inclusions and rimmed by "clean" pyrite. [rl, fov=1.1mm, BV219].

D) Euhedral to anhedral pyrite grains in sphalerite-chalcopyrite. Sphalerite appears to replace chalcopyrite and euhedral pyrite grain contains spherical framboidal pyrite core. [rl, fov=1.1mm, C284].

E) Anhedral pyrite grain with porous texture in core. The porous texture in the pyrite core is possibly after inversion of marcasite to pyrite (c.f. fig 5b Murowchick (1992)) [rl, fov=1.4mm, BV020].

F) Enlargement of porous pyrite core in previous image.
Figure 5.16 Pyrite textures—other.

G) Concentrically zoned anhedral pyrite in sphalerite with fractures infilled with sphalerite-galena. [rl, fov=1.8mm, C717].

H) Typical “clean” euhedral pyrite in quartz-altered footwall to ore. [rl, fov=2.2mm, C823].

I) Augen of “clean” pyrite and sphalerite in sericite-quartz schist. The sphalerite development was focused in the pressure shadow of the pyrite grains through either physical or chemical remobilisation processes. [rl, fov=2.2mm, C320].

J) Augen of anhedral spongiform pyrite mass and sphalerite in sericite-quartz schist. [rl, fov=2.2mm, BV210].

K) Fractured sphalerite-pyrite-chalcopyrite ore located proximal to post-Devonian brittle fault. There is little evidence of sulphide remobilisation related to the fractures. [rl, fov=2.2mm, C794].

L) “Clean” euhedral pyrite grains in banded galena-sphalerite. The pyrite grains in sphalerite do not exhibit caries texture and in hand specimen the ore appears to have been deformed, indicating that these pyrite grains developed during deformation. [rl, fov=2.2mm, C365].
times during and after initial sulphide precipitation. The earliest coarse crystalline pyrite is intimately intergrown with fine-grained spongiform and atoll pyrite textures and is therefore broadly synchronous with these early pyrite textures. Overprinting and rimming of early sulphides by later pyrite is likely to have occurred during further hydrothermal, diagenetic and metamorphic processes, but the lack of textures preserved in surrounding sulphides restricts a more precise paragenesis. The crystalline pyrite euhedra in the quartz-altered footwall (fig.5.16h) display similar morphologies to the crystalline pyrite overgrowths and euhedral grains in the massive ore, although they typically lack inclusions or relict spongiform cores. The more crystalline morphologies suggest higher temperatures of formation in comparison with the relict frambooidal and spongiform pyrite morphologies.

Pyrite augen developed during deformation of phyllosilicate-rich rocks, and in many cases sulphide, silicate and carbonate minerals have precipitated in the pressure shadow of the more competent pyrite (fig.5.16i-j). In some cases pyrite displays fracturing that appears related to the main cleavage development and predates annealing of surrounding sulphides (e.g. fig.5.16e-g), however, fracturing is more prevalent in rocks affected by Devonian metasomatism and post-Devonian brittle faulting (fig.5.16k).

Pyrite within a sulphide-rich matrix will recrystallise and form euhedral grains during deformation (Craig and Vokes, 1993). Clean, concentrically zoned, anhedral to euhedral pyrite grains of this type in a sphalerite and galena matrix occur within ore in the pressure shadows of large boudinaged blocks of altered host rock and in the banded ore present in the same shear zones. The same style of pyrite also occurs in sulphide ore emplaced into the sill overlying K lens during Devonian deformation, and occasionally within the massive sulphide ore (fig.5.16l). Vokes (1983) noted similar pyrite characteristics within J lens ore.

5.6.2. Arsenopyrite

Arsenopyrite occurs as a minor component within the sulphide ore and in altered host rocks proximal to ore, but does not occur in association with massive barite (Brathwaite, 1969; 1974; and this study). It is occasionally present in syn- to post-cleavage quartz-carbonate-sulphide and sulphide veins and within the alteration and veining related to Devonian metasomatism.

The earliest arsenopyrite occurs as fine euhedra and rare needles intergrown with and replacing spongiform pyrite (fig.5.17a). Arsenopyrite euhedra are also intergrown with both spongiform and more coarsely crystalline pyrite (fig.5.17b). As for pyrite, the arsenopyrite in sphalerite-rich ore often displays caries texture; although where crystal margins are in contact with interstitial quartz they are well preserved (fig.5.17b-c). Very fine arsenopyrite euhedra are present as inclusions within tennantite and tetrahedrite, and occasionally in galena after it has replaced the tetrahedrite.
As with pyrite, the arsenopyrite is seen in places to have developed silicate mineral growth in pressure shadows during deformation of the ore, and appears to have been preserved during later metasomatic overprinting (fig.5.17d).

5.6.3. Sphalerite, galena and chalcopyrite

Sphalerite, galena and chalcopyrite occur as individual grains, coalesced masses, bands and veinlets in disseminated to massive ore and are a significant component in syn- to post-deformational quartz-carbonate veins and fault zones, and in veins and alteration associated with granite-related metasomatism. These sulphides everywhere display textures indicative of metamorphic annealing (Brathwaite, 1974; Green et al., 1981) (fig.5.17e-g), except in the proximity of post-Devonian brittle faults where brittle fracturing is evident (fig.5.16k). Kinking of sphalerite twin lamellae (Brathwaite, 1969, 1974) and curvature of cleavage pit trails in galena (fig.5.16b) (Klaere, 1997; Vallerine, 2000; and this study) also indicate that some deformation occurred syn- to post-annealing. As all but the latest deformation fabrics in these sulphides are obscured, the annealing process must have occurred in the last stages of, or subsequent to, Devonian granite-related metasomatism. This is also supported by the fact that compositionally zoned sphalerite grains only occur in association with metasomatised ore (Khin Zaw and Large, 1996), whereas all other sphalerite grains have a uniform internal composition that was achieved during recrystallisation and annealing.

Sphalerite has a marked colour variation in hand specimen, typically varying between a pale honey-yellow and a dark tan-brown, ranging to a dark red-brown and dark purple-brown in ore affected by Devonian metasomatism. Most sphalerite exhibits the fine “chalcopyrite disease” texture of Barton (1978) with chalcopyrite located along the margins of twin lamellae and at the margins of sphalerite crystal, becoming more prevalent in sphalerite affected by Devonian metasomatism. Although sphalerite may originally have exhibited chalcopyrite disease textures as the result of coprecipitation (Kojima, 1990), replacement (Barton, 1978; Barton and Bethke, 1987; Bortnikov et al., 1991) or exsolution processes (Barton and Bethke, 1987), the current siting of chalcopyrite along twin and grain boundaries (fig.5.17h) is indicative of recrystallisation during metamorphic annealing (Marshall and Gilligan, 1987).

In the primary ore, very fine-grained sphalerite and galena occur interstitial to spongiform pyrite and as inclusions within the core of more crystalline pyrite, whereas chalcopyrite is more often associated with more coarsely crystalline pyrite. In places, sphalerite, galena and chalcopyrite replace other sulphides (figures 5.15g-j, 5.16d, 5.17c), a process that would have occurred after initial sulphide precipitation through overprinting hydrothermal, metamorphic and metasomatic processes. Rutile in massive sphalerite (fig.5.2c) displays spatial patterns that are probably the result of sulphide replacement of primary volcanic hornblende grains (after augite?), and some skeletal pyrite is the result of sulphide replacement of minerals such
Figure 5.17 Arsenopyrite, sphalerite, galena, tetrahedrite, electrum.
A) Fine-grained arsenopyrite intergrown with very fine-grained and framboidal pyrite in quartz-sericite schist. [rl, fov=2.2mm, C261].
B) Intergrowth of euhedral arsenopyrite and pyrite in sphalerite band crosscutting relict spongiform pyrite and galena. [rl, fov=550µ, C364].
C) Arsenopyrite euhedra and anhedral to euhedral pyrite within and partially replaced by galena. Note tetrahedrite masses without inclusions. [rl, fov=550µ, BV220].
D) Fine-grained arsenopyrite euhedra and anhedral pyrite in sphalerite-pyrrhotite. Arsenopyrite and pyrite both embayed. [rl, fov=2.2mm, C371].
E) Galena-sphalerite-chalcopyrite displaying typically annealed textures. Note coarse grained electrum grain at galena grain boundaries. [rl, fov=2.2mm, C381].
F) Fine quartz veinlet traversing relict spongiform pyrite-galena, euhedral "clean" pyrite and sphalerite, with later infill and replacement by chalcopyrite. [rl, fov=550µ, BV226].
Figure 5.17 Arsenopyrite, sphalerite, galena, tetrahedrite, electrum.

G) Relict colloform and spongiform pyrite-galena in sphalerite disrupted and partially replaced by later chalcopyrite, all of which are crosscut by later quartz veinlet. [rl, fov=2.2mm, C364].

H) Chalcopyrite disease in sphalerite with chalcopyrite grains distributed along twin and grain boundaries, and along fine veinlet. [rl, fov=550µ, C794].

I) Tetrahedrite with galena-sphalerite-pyrite and containing fine-grained inclusions of arsenopyrite and chalcopyrite. [rl, fov=550µ, C329].

J) Tetrahedrite in galena with sphalerite-pyrite and containing fine-grained inclusions of arsenopyrite and chalcopyrite distributed along internal structural boundaries. [rl, fov=550µ, C337].

K) Relict tetrahedrite mass with arsenopyrite in chalcopyrite-quartz with sphalerite-pyrite-galena. The arsenopyrite was originally present as inclusions in tetrahedrite prior to replacement by chalcopyrite-quartz. [rl, fov=550µ, C364].

L) Tetrahedrite, chalcopyrite and arsenopyrite in galena with sphalerite and pyrite. The galena appears to have partially replaced tetrahedrite with inclusions of chalcopyrite and arsenopyrite. [rl, fov=1.1mm, C801].
as anhydrite.

Within the syn-deformational quartz-carbonate veins and faults the sulphides are again annealed, and typically exhibit a coarser-grained nature. Sphalerite and galena are also the principal sulphide components present in the pressure shadows of rotated pyrite grains and silicate augen (fig.5.16i-j).

5.6.4. Tennantite, tetrahedrite and the other sulfo-salts.

Tennantite and tetrahedrite are a minor component in both the sulphide and barite lodes, but are typically absent in the footwall mineralisation. They occur in close spatial association with galena as small anhedral grains, often with abundant inclusions of arsenopyrite and chalcopyrite (fig.5.17i-l). Tennantite and tetrahedrite occur as components in syn- to post-deformation quartz-carbonate veins.

Other sulfo-salts form a trace component within the Rosebery ores and include pyrargyrite, bornite, bourbonite, boulangerite, jordanite, meneghinite, bismuthinite, kobellite, cosalite, miargyrite, sulvanite and cubanite. Although observed on very rare occasions within massive ore or as inclusions within pyrite, these minerals most commonly occur as components of late syn- to post-cleavage veinlets and quartz-carbonate-sulphide veins (Stillwell, 1934; Williams, 1960; Brathwaite, 1969; Green et al., 1981; Green, 1983; Huston and Large, 1988; Khin Zaw et al., 1997; Klaere, 1997; Vallerine, 2000; and this study).

5.6.5. Gold and electrum

Around 85% of Au in the mill feed occurs as electrum (Huston et al., 1992) and it is an economically significant trace component of the sulphide and barite ore. Electrum occurs in association with pyrite, sphalerite, galena, arsenopyrite, chalcopyrite and barite in the primary sulphide ores and with a similar range of minerals in the barite bodies, typically along grain boundaries, fractures and veinlets or as inclusions (table 5.1). Electrum most commonly occurs with pyrite in sphalerite-galena-rich sulphide ore (Huston and Large, 1988; and this study), although Vallerine (2000) noted that coarse grains of electrum in sphalerite and galena were prominent within the high grade Au zones. These two associations were also reflected in an examination of the mill zinc residue (Henley and Steveson, 1978). In this study very fine-grained electrum (<10µ) was occasionally observed in spongiform pyrite, and most commonly within the more crystalline pyrite, often in association with sulphide inclusions within or at the edges of irregular and circular cores that are interpreted to be relict spongiform and frambooidal pyrite. The fine-grained electrum in pyrite was often more readily observed after the pyrite had developed a light tarnish. Coarse electrum, up to ~200µ in size, was also observed within galena and sphalerite (figures 5.16l & 5.17e).
Electrum is present within the quartz-carbonate-sulphide and sulphide vein set that formed syn- to post-deformation (Table 5.1). Au also occurs in the pyrrhotite-pyrite ore formed during the granite-related metasomatism of F lens, and is present as electrum marginal to pyrite grains, in microfractures within euhedral pyrite grains, as inclusions within pyrrhotite, and in chalcopyrite veins cutting pyrrhotite (Khin Zaw, 1991).

Fine-grained electrum associated with spongiform and colloform pyrite and with the cores of more crystalline pyrite was emplaced at the time of primary ore formation, either as fine-grained inclusions due to coprecipitation or in solid solution within the pyrite crystal matrix. Au was almost certainly associated with other sulphide species at the time of ore formation. Subsequent deformation, metasomatism and annealing is interpreted to have resulted in recrystallisation and remobilisation of Au, resulting in a coarse-grained electrum along fractures and grain boundaries (Khin Zaw, 1991; Huston et al., 1992). Electrum was also remobilised into syn- to post-deformation quartz-carbonate-sulphide and sulphide veins at this time.

<table>
<thead>
<tr>
<th>Au occurrence</th>
<th>mineral association</th>
</tr>
</thead>
<tbody>
<tr>
<td>sphalerite-galena-pyrite ore</td>
<td>chalcopyrite-arsenopyrite-tetrahedrite aggregate (Brathwaite, 1969; Huston et al., 1992; Westbrook, 1997; Vallerine, 2000); micro-veins of galena or chalcopyrite in pyrite (Huston and Large, 1988); fine-grained pyrite – recrystallised relict spongiform? (Vallerine, 2000, and this study); in sphalerite or galena filling fractures within pyrite or at pyrite margins (Vallerine, 2000, and this study); galena and sphalerite as inclusions and at grain boundaries (Klaere, 1997; Vallerine, 2000, and this study); chalcopyrite and galena inclusions in fractured arsennopyrite (Vallerine, 2000)</td>
</tr>
<tr>
<td>pyrite-chalcopyrite ore</td>
<td>association not described (Brathwaite, 1974; Green, 1983)</td>
</tr>
<tr>
<td>massive barite mineralisation</td>
<td>chalcopyrite (Huston and Large, 1988); discrete grain between barite grains (Huston and Large, 1988; Vallerine, 2000); galena (Huston and Large, 1988); galena, tetrahedrite, sphalerite, chalcopyrite, pyrite, barite (Huston et al., 1992)</td>
</tr>
<tr>
<td>altered host beneath massive barite</td>
<td>chalcopyrite (Huston and Large, 1988)</td>
</tr>
<tr>
<td>distal pyrite mineralisation</td>
<td>micro-veins of galena or chalcopyrite in pyrite (Huston and Large, 1988); with galena-tetrahedrite interstitial to lathe-like pyrite at edge of colloform pyrite mass (Huston and Large, 1988)</td>
</tr>
<tr>
<td>in cross-cutting chalcopyrite vein</td>
<td>in cracks and as inclusions in recrystallised arsennopyrite (Huston and Large, 1988)</td>
</tr>
<tr>
<td>syn- to post-cleavage sulphide-quartz-carbonate veinlets in unmineralised host rock</td>
<td>chalcopyrite (Huston and Large, 1988); sphalerite-galena-chalcopyrite-quartz (Huston et al., 1992; Vallerine, 2000); coarse-grained tetrahedrite ± chalcopyrite ± sphalerite ± galena (Stillwell, 1934; Williams, 1960; Brathwaite, 1969; Green, 1983; Westbrook, 1997; Vallerine, 2000)</td>
</tr>
<tr>
<td>bismuthinite vein crosscutting recrystallised massive pyrite-chalcopyrite ore</td>
<td>bismuthinite-maldonite-bismuth-carbonate (Huston and Large, 1988)</td>
</tr>
<tr>
<td>remobilised massive sulphide in K lens sill</td>
<td>arsenopyrite-sphalerite in tetrahedrite (Solly, 1999)</td>
</tr>
<tr>
<td>pyrrhotite-chalcopyrite-pyrite-sphalerite vein</td>
<td>pyrite-sphalerite (Westbrook, 1997)</td>
</tr>
<tr>
<td>pyrite-pyrrhotite zone</td>
<td>pyrite, pyrrhotite, chalcopyrite (Khin Zaw, 1991)</td>
</tr>
</tbody>
</table>

Table 5.1 Electrum and gold mineral associations.
5.6.6. Gangue mineralogy

The majority of the gangue material in the sulphide ore comprises sericite, quartz, carbonate, barite, chlorite and minor rutile as intergrowths and inclusions in the sulphides, and within the lithic bands and fragments. In ore affected by Devonian metasomatism, the ore may also contain tourmaline, magnetite, haematite, garnet, fluorite and helvite.

Albite is locally a significant component within the sulphide and barite ore (Brathwaite, 1969, 1974; Green et al., 1981; Vokes, 1983; and this study), and sometimes within the pre-deformational pyrite-quartz veins in quartz-altered shale above P lens. K-lens does not contain significant albite, but instead contains anhedral aggregates of quartz interstitial to the predominant sphalerite grains. P lens contains albite as a common component within the semi-massive and massive barite and sulphide mineralisation, but it is relatively rare in the P6 sub-lens that also contains interstitial quartz as in K lens. The albite grains are present in massive sphalerite and pyrite (fig.5.18a-j), and in massive barite (fig.5.18k-l), but are absent in the presence of chalcopyrite. In massive sulphide ore the margins of albite grains not in contact with sulphide or barite exhibit varying degrees of sericite, quartz and carbonate alteration (fig.5.18g-h), and in places quartz overgrowths occur within the pressure shadows of albite grains (fig.5.18j). Albite is not a significant component of highly sheared and metasomatised ores.

The albite occurs as sub-angular to sub-rounded grains, up to 3mm in size. The smaller albite grains are clean with no apparent internal textures except for twinning that extends across the grain (fig.5.18d-e), however most of the albite grains greater than 0.5mm in size exhibit some evidence of relict igneous concentric zoning (fig.5.18g-k). The larger albite grains exhibit void and inclusion patterns similar to those generated by volume reduction during albitisation of original igneous plagioclase grains (Morad et al., 1990). In a few instances the larger albite grains appear to have been overprinted by sulphides, and they occasionally contain fractures that have been infilled by sulphide.

The larger albite grains display sizes, spatial distributions and relict internal textures that indicate they are psuedomorphs of original volcanic plagioclase grains that were preserved during sulphide and barite mineralisation. Later fluids caused destructive alteration of grains not protected by massive sulphide or barite. Fine-grained albite that does not display internal concentric zonation and void textures is most likely the result of fragmentation and physical reworking of igneous feldspar crystals prior to final volcaniclastic deposition. Rare examples of clean albite rims on altered cores (fig.5.18l) may be the result of differential sericite-quartz alteration within feldspar grains, or possibly the result of very localised albite precipitation after initial alteration of feldspar grains.
Figure 5.18 Albite occurrences in sulphides and barite.

A) Grey albite grains in pyritic band underlain by sericite schist and overlain by thin discontinuous band of silicified shale and massive sphalerite-galena ore at base of P10 sub-lens. [fov=9cm, P10 ore drive, R065].

B) Grey feldspar grains and minor silicified lithic fragments in massive sphalerite-galena ore. [R5192-D1-631.7m, core=35mm diam.].

C) Albite grains in sphalerite-galena-pyrite groundmass. [tl-xn, fov=8.8mm, C547].

D) As for (C). [rl].

E) Albite grain in sphalerite-galena-quartz groundmass. Note presence of quartz and sphalerite inclusions, and weak rounded texture in middle of large grain that in places disrupts twinning. [tl-xn, fov=2.2mm, C547].

F) As for (E). [rl].
Figure 5.18 Albite occurrences in sulphides and barite.

G) Albite grain in sphalerite-pyrite-galena and sericite-quartz schist. Sericite replacement of the albite has occurred where the sulphides do not enclose the albite. Note concentric texture to albite grain that may be relict of primary feldspar texture (c.f. fig. 4.1b). [tl-xn, fov=2.2mm, C575].

H) As previous image [rl].

I) Albite grain crosscut by later sericite-pyrite alteration. [tl-xn, fov=4.4mm, C353].

J) Albite grain with relict primary concentric texture (after plagioclase) cut by sulphide vein. The grain is hosted within sphalerite-galena-pyrite and is overgrown by quartz that probably developed in the grain pressure shadow during deformation. The porous texture exhibited in the albite would be due to volume reduction during albitisation of primary plagioclase. [tl-xn, fov=4.4mm, C397].

K) Albite exhibiting relict circular internal fabric and hosted within barite. [tl-xn, fov=4.4mm, C400].

L) Rare example of an albite grain in barite-sericite-carbonate exhibiting sericite-altered core. The grain may be the result of partial sericite alteration of a primary feldspar grain prior to overgrowth by later albite. The small adjacent fragmental grains of albite do no display this texture. [tl-xn, fov=2.2mm, C705].
5.7. Deformation-related ore remobilisation

The Rosebery ores have undoubtedly been subjected to deformation with commensurate evidence for folding faulting and shearing of the ore evident throughout the mine (figures 5.6i-k & 5.19). Although post-deformation annealing has obscured much of the textural evidence in thin section, there is still abundant evidence for deformation and remobilisation within the ore.

Marshall and Gilligan (1987) defined metamorphic remobilisation as the micro-, meso- and macro-scale translocation, resulting in modified concentration and distribution, of pre-existing massive and semi-massive mineralisation. They argued that remobilisation of sulphides can occur by solid-state physical transfer (mechanical), liquid state transport including solution melting and wet diffusion (chemical) or by a combination of the above (mixed). Gilligan and Marshall (1987) defined a set of criteria by which remobilised sulphide ores might be recognised, and these have been applied to the Rosebery ore body.

Macro-scale evidence of remobilisation at Rosebery is difficult to establish due to intermittent underground exposure and drill hole spacing. Meso-scale evidence for remobilisation in hand specimen and outcrop includes; the presence of elongation lineations within the ore and host rocks defined by the presence of sulphide mineral pressure shadows around high-strength minerals and lithic fragments (fig.5.19e); foliations within sulphide bodies that are defined by the preferred distribution of sulphide species (figures 5.7e & 5.8e); entrainment of host fragments (fig.5.8f); folding of sulphide layering (fig.5.6); boudinage development in higher-strength layers with partitions infilled with remobilised sulphides (fig.5.19f); durchbewegung textures (fig.5.8c-d,f); piercement structures (fig.5.19b,d); and boudinage of more competent sulphide layers (fig.5.19g).

Micro-textural evidence for remobilisation includes the effects of deformation at a grain scale including; fracture of pyrite (fig.5.19h); poikiloblastic (fig.5.16c) and porphyroblastic (fig.5.5l) pyrite textures; preferential exsolution of chalcopyrite from sphalerite along marginal zones of deformed grains (fig.5.17h); bent cleavage pits in galena (fig.5.16b); domainal sulphide fabric in mixed ores; pressure shadow mineralisation (fig.5.16i-j); and the sulphide infilling and healing of fractures in high-competence phases within mixed ores (5.16g). The almost ubiquitous nature of caries texture in pyrite within sphalerite-rich ore, including porphyroblastic pyrite grains, indicates that dissolution of the pyrite occurred during deformation and prior to annealing of the sulphide matrix.

Further evidence of sulphide remobilisation at the meso-scale is the syn-tectonic quartz-carbonate vein set that lacks a significant sulphide component distal to ore but contain significant amounts of sulphide proximal to ore, indicating that the sulphides were most likely sourced from the primary ore. The same vein set is in places disrupted, deformed and
Figure 5.19 Ore remobilisation.
A) Shearing within massive banded sphalerite-pyrite ore hosted within altered footwall pumice breccia. The ore is banded (see fig. 5.7d) and contains boudinaged blocks of pyrite and quartz-altered host rock (see fig. 5.8e). [33K crosscut].
B) Folding of stratified TSV units located between main ore zone (LHS) and porphyry sill (RHS). Syn-tectonic quartz-carbonate veins are deformed (A), crosscut cleavage and folded bedding (B), and are boudinaged within massive sulphide ore (C), suggesting syn-tectonic ore remobilisation. Banding of the ore within the fold axes lacks symmetry and fold closures, suggesting remobilisation and emplacement of sulphides during deformation. Massive ore within the porphyry post-dates syn-tectonic quartz-carbonate veining (see fig.5.19c) and is in places banded, indicating that this is a piercement structure generated as the result of ore remobilisation. [40K, K nub drive].
C) Sulphide veins in quartz-feldspar porphyry sill crosscutting syn-tectonic quartz-carbonate veins [40K, K nub drive].
D) Syn-tectonic quartz-carbonate veins crosscut by later sphalerite-galena piercement veins. [R6475-124.3m, core≈35mm diam.]
Figure 5.19 Ore remobilisation.
E) Pyrite mass hosted in semi-massive sphalerite-rich ore with sugary white barite developed in pressure shadow indicative of rotation during deformation. [R6011-383.1m, core≈35mm diam.]
F) Boudin of chlorite-sericite altered host rock in sphalerite-rich ore. Coarse chalcopyrite-pyrite-galena developed in dilatational zone between host rock boudins during deformation. [R6847-8.0m, core≈35mm diam.]
G) Boudinage and necking of primary pyrite band in banded sphalerite-pyrite ore. [33K crosscut].
H) Fractured pyrite grains in annealed sphalerite. [rl, fov=2.2mm, C359].
boudinaged within massive sulphide ore (fig.5.19b). Both Au and Ag were chemically remobilised during deformation, resulting in the formation of relatively coarse-grained electrum in moderately deformed ore and in syn-deformational quartz-carbonate-sulphide veins (Huston and Large, 1988; Huston et al., 1992). The presence of occasional syn-deformational tetrahedrite and pyrargyrite veins to several centimetres in thickness, both within and proximal to primary sulphide and barite mineralisation, further indicates that remobilisation of Ag occurred during deformation. The enrichment of some tetrahedrite-tennantite grains in Ag, as indicated by the commensurate exsolution of chalcopyrite and arsenopyrite (fig.5.17l), suggests that a significant amount of the Ag may have been remobilised out of galena. Syn-deformational tetrahedrite veins may also contain significant amounts of electrum (Vallerine, 2000), further indicating that Au remobilisation occurred during deformation. The restriction of mineralised syn-deformational veins to within a few tens of metres of primary mineralisation, suggests that Au and Ag remobilisation and enrichment within the veins was a localised process. As such, although syn-deformational veins may be enriched in Au and Ag, any change in overall grade of massive ore is not significant.

Although not all the criteria listed by Gilligan and Marshall (1987) are present, there is sufficient evidence to demonstrate syn-tectonic remobilisation of the Rosebery ore. Most of the evidence for remobilisation described can be attributed to physical and chemical processes that occurred during metamorphism and deformation. Further sulphide remobilisation occurred during the Devonian metasomatism event after most of the tectonic deformation had occurred, but there is little evidence of sulphide remobilisation during the development of late brittle faults.

5.8. Devonian metasomatism

The effects of granite-related metasomatism on the sulphide and barite ores have been examined in detail by previous workers (Vokes, 1983; Solomon et al., 1987; Khin Zaw, 1991; Khin Zaw and Large, 1996; Khin Zaw et al., 1997; Khin Zaw et al., 1999), and the resultant alteration assemblages generated during metasomatism of rocks at the southern end of the mine are summarised in chapter 4. The metasomatic alteration mineral assemblages in the vicinity of F lens and J lens have been divided into the initial magnetite-biotite ± chalcopyrite and the pyrrhotite – pyrite zones, and the overprinting tourmaline - quartz ± magnetite zone (Khin Zaw, 1991; Khin Zaw et al., 1997). Alteration of the primary sulphide ore to these alteration assemblages, after folding and deformation of those lenses, is succinctly illustrated in a sample of folded sphalerite-galena-pyrite ore cross-cut by pyrrhotite-replacement alteration, that was initially collected by Brathwaite (1969) and subsequently photographed by Solomon et al. (1987) and Khin Zaw et al. (1997).

At the southern end of the mine granite-related metasomatic fluids caused significant
replacement of primary sphalerite-galena-pyrite ore by a pyrrhotite-pyrite ± chalcopyrite assemblage, resulting in removal of sphalerite, galena and tetrahedrite from the ore, and retention of pyrite, chalcopyrite and Au (Khin Zaw et al., 1999). Adjacent to the pyrrhotite-pyrite replacement front sphalerite is typically enriched in Fe (Vokes, 1983; Solomon et al., 1987; Khin Zaw et al., 1999). Au remobilised from the magnetite-biotite ± chalcopyrite alteration zone was precipitated in the pyrite-pyrrhotite zone, resulting in enriched Au grades relative to the precursor sulphide ore (Khin Zaw et al., 1999).

As the down-dip limits of the P lens sulphide lenses are not well defined it is difficult to ascertain the extent of metasomatic alteration on primary ore in this position. Metasomatic alteration on the down-dip margins of P lens has, however, resulted in formation of patchy pyrrhotite-pyrite and magnetite-biotite ± chalcopyrite alteration in the ore position. The zone of chlorite-rich alteration and intense foliation development discussed in chapter 4 has disrupted the ore below P7 and P9 sub-lenses, and contains a broadly dispersed sulphide component that includes disseminations and anastomosing “stockwork” veins of chalcopyrite, euhedral pyrite and pyrrhotite, wispy lenses of pale brown to purple sphalerite, and rare disseminated galena and euhedral arsenopyrite (fig.4.11c-h). The sulphide content of this alteration zone is highest at the expected primary ore position, with occasional bands of relict primary ore in the form of semi-massive to massive sphalerite, galena, pyrite and barite that display varying degrees of folding, brecciation and foliation. The lack of massive bodies of pyrrhotite-pyrite would suggest that metasomatism in this area overprinted primary mineralisation that was predominantly disseminated in nature. An example of a chlorite-sphalerite alteration front within sericite-altered host rocks adjacent to this area may be the result of sphalerite precipitation after zinc was remobilised from nearby primary ore (fig.4.11i).

5.9. Discussion and summary

This study has shown that the Rosebery ore lenses are spatially related to the margins of a fault-bounded basin, the limits of which are now defined by a rapid increase in TSV thickness. The ore lenses are podiform to sheet-like in morphology and occur either as a single lens or as a series of stacked sub-lenses. The more laterally extensive sheet-like ore lenses tend to occur where the TSV sediments are less than 50m in total thickness, whereas those that comprise multiple stacked sub-lenses occur within stratified TSV sediments that are greater than 50m in thickness. Throughout the mine the ore lenses are hosted within the upper part of the footwall pumice breccia unit and at various levels within the TSV sediments. In places individual ore lenses appear to transgress stratigraphy.

Many of the sulphide lenses are overlain by massive barite or barite-rich sandstones that may be separated from the ore by weakly mineralised host rocks. Most sulphide lenses display a vertical mineral zonation, with an upper sphalerite-galena-pyrite-rich disseminated to massive
sulphide ore that is often enriched in pyrite and chalcopyrite toward the base. This may be underlain by a disseminated to stringer pyrite-chalcopyrite-rich ore that contains discontinuous lenses of massive pyrite-chalcopyrite in chlorite-altered rocks. Within some lenses the zonation appears inverted, and the vertical zonation is less pronounced within the relatively Cu-poor northern ore lenses. In the southern sulphide lenses the sphalerite-rich ore also grades laterally into more chalcopyrite-pyrite-rich ore, and this change coincides spatially with rapid changes in TSV sediment thickness that may indicate proximity to a Cambrian growth fault. The sulphide ore typically grades laterally into, or interferes with massive barite, massive carbonate or weakly mineralised host rock at the margins.

The sulphide paragenesis as presented in this study (table 5.2), differs markedly from that presented by Aerden (1992; 1993), who proposed the paragenetic sequence pyrite → quartz-chlorite → carbonate → late-stage pyrite → sphalerite + galena + chalcopyrite. Aerden (1992; 1993) failed to recognise colloform, spongiform and skeletal pyrite morphologies within the Rosebery ore, textures that clearly demonstrate early co-precipitation of pyrite, sphalerite, galena, chalcopyrite, electrum and sulphate minerals.

Diagenetic pyrite framboids and fine-grained overgrowths on clastic material developed within sediments shortly after deposition (table 5.2), and probably continued to develop at the low-temperature margins of the hydrothermal system that formed the primary ore mineralisation.

Examination of primary sulphide and barite mineralisation (assemblages 2 & 3 in table 5.2) during this study has identified primitive textures preserved within pyrite that are indicative of a broadly syn-sedimentary ore formation, rather than related to later deformation and metamorphism. These textures include colloform, spongiform, framboidal, skeletal and atoll morphologies, many of which are the result of pyrite precipitation in conjunction with other sulphides and sulphates. The textures are considered to be analogous to those observed within primary VHMS ores at Kidd Creek (Hannington et al., 1999a), the Kuroko district (Mukaiyama et al., 1974; Yamaoka and Asakura, 1974; Yui and Ishitoya, 1983), the Hellyer deposit in western Tasmania (McArthur, 1996), and various deposits within the Tasman Fold Belt of New South Wales (England and Ostwald, 1993). Volcaniclastic albite grain preservation appears to have occurred with this initial phase of ore mineralisation.

As identified during this study, the preferential mineralisation of coarser-grained sediments within the stratigraphy proximal to P lens (fig.5.5c) is consistent with sulphide replacement of sediments as mineralising fluid moved along more permeable horizons within the TSV, or possibly with physical reworking of exhalative sulphide material during turbidite emplacement. The lack of reworked sulphide or mineralised host rock clasts anywhere within the TSV succession suggests that physical reworking of sulphide material did not occur. The immobile element signature of K lens ore is the same as the footwall pumiceous mass flow units within
Table 5.2 Rosebery mineralisation paragenesis.
The paragenetic sequence commences with diagenetic / low temperature hydrothermal framboidal pyrite that forms within unconsolidated volcaniclastic sediments of the TSV. This is in turn followed by primary ore mineralisation, the broad paragenesis of which is based on thin section and hand specimen observation. A more detailed paragenetic sequence for ore mineralisation is not possible due to ore recrystallisation. The primary ore is subsequently overprinted by relatively minor deformation-related mineralisation comprising material chemically remobilised from the primary ore. The granite-related metasomatic mineralisation is locally pervasive and includes a significant introduced iron component. Sulphide annealing occurred during the final stages of metasomatism or sometime thereafter.

<table>
<thead>
<tr>
<th>Assemblages</th>
<th>Time</th>
<th>Cambrian pre-tectonic</th>
<th>Devonian syn-tectonic (S/S)</th>
<th>granite-related metasomatism</th>
</tr>
</thead>
<tbody>
<tr>
<td>(1) pyrite</td>
<td><em>primary ore mineralisation</em></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- trace sphalerite-galena</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- (pyrite as framboïd, fine-grained euhedral and spongeform overgrowths)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- formed within sediments prior to/distal to ore</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- minor disseminations and thin veinlets</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(2) <em>barite-pyrite-galena-sphalerite</em></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- minor tetrahedrite, no arsenopyrite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- trace electrum, gold and chalcopyrite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- (pyrite with colloform, spongeform, atoll, skeletal and more euhedral morphologies)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- disseminated to massive mineralisation</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- mainly lateral and upper margins of sphalerite-rich ore, barite clots/bands within ore, pyritic baritic veins in hanging wall shales, minor late cross-cutting veins in ore</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- overprints framboidal pyrite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(3) <em>sphalerite-galena-pyrite + barite</em></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- minor tetrahedrite-arsenopyrite-chalcopyrite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- trace electrum and native gold</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- (pyrite with colloform, spongeform, atoll, skeletal and more euhedral morphologies)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- disseminated to massive mineralisation</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- overprints framboidal pyrite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(4) <em>sphalerite-galena-pyrite-chalcopyrite</em></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- minor tetrahedrite-arsenopyrite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- trace electrum and native gold</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- (pyrite as euhedral crystals, aggregates and overgrowths on primitive pyrite morphologies)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- veins and disseminated to massive mineralisation</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(5) <em>pyrite-chalcopyrite-sphalerite</em></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- minor arsenopyrite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- trace tetrahedrite, electrum, native gold</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- (pyrite as predominately more euhedral/crystalline morphologies)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- veins and disseminated to massive mineralisation</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- mainly lower margins of sulphide lodes and proximal to inferred Cambrian faults</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(6) pyrite-sphalerite-chalcopyrite-galena</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- minor sulphosalts, barite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- trace electrum, native gold, native bismuth</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- within syn-tectonic quartz-carbonate and sulphide veins, as mineral growth in low-pressure domains and disseminations in faults/shears</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(7) pyrite-pyrrhotite &amp; magnetite-biotite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- minor chalcopyrite-sphalerite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- trace sulphosalts, wolframite, native bismuth, with gold and electrum in pyrite-pyrrhotite zone</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- massive replacement of primary ore</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(5) tourmaline-quartz ± sulphides</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- minor sphalerite, galena, chalcopyrite, pyrrhotite</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- as post-S/S, veins and within major faults</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
which it is hosted, strongly indicating sulphide replacement of the host sediments. The fine-grained nature of remnant lithic material within both K lens and P lens is also consistent with near total sulphide replacement along coarse sediment horizons within the volcaniclastic succession. The more diffuse style of primary banding within the ore (fig.5.6b) may be the product of sediment infill and replacement by sulphides during the lateral movement of metalliferous fluid through unconsolidated volcanoclastics. Analogy may be drawn with the Cu-rich sulphide replacement of coarse-grained horizons within turbidite units beneath the modern Bent Hill deposit, where veining is not developed and primary sedimentary features have been preserved (Duckworth and the Leg 169 Shipboard Scientific Party, 1997). The occurrence of stacked ore lenses where TSV thickness exceeds 50m may be related to the presence of multiple horizons of high permeability that were amenable to acting as conduits for hydrothermal fluids.

A subsequent phase of ore mineralisation (assemblages 4 & 5 in table 5.2) produced more crystalline pyrite that locally post-dates the primitive pyrite morphologies, and probably represents increased hydrothermal fluid temperatures. This phase of mineralisation was most likely associated with fluids responsible for volcanoclastic albite destruction, and may also have been involved in the dissolution of existing barite to produce the barite fragments present within sulphide ore. The most intense pyrite-chalcopyrite mineralisation occurs at the base of sulphide lenses and within lenses proximal to Cambrian fault traces, probably indicating high temperature hydrothermal fluid upflow along growth faults and lateral flow of hydrothermal fluids along the base of existing sulphide mineralisation. Intense compositional banding of the sulphide ore is also more prominent within the Cu-rich southern ore lenses. Formation of the intense compositional banding may be due to the same high temperature hydrothermal fluids that produced the footwall and massive pyrite-chalcopyrite mineralisation; with pyrite-chalcopyrite bands forming as hydrothermal fluids flowed along more permeable horizons within the ore. Albite is not a significant gangue component of the Cu-rich southern ore lenses.

A late pyritic and baritic pre-deformation vein set identified during this study locally crosscuts primary ore, and probably represents a final phase of mineralisation associated with assemblages 2 and 3 (table 5.2) that formed as the hydrothermal system waned.

Disseminated mineralisation overprinting the peperitic margins of the sill in the hanging-wall to K lens, and the lack of sulphide or barite clasts within the sill, is consistent with the sill having been intruded into unconsolidated sediments prior to significant sulphide mineralisation. The intensely altered sill apophysis overlying the lower components of P lens is most likely of a similar timing relative to P lens mineralisation. The apparently transgressive nature of mineralisation relative to the stratigraphic units described previously is consistent with a replacement mode of ore formation.
The ore mineral zonation and banding has been interpreted as the product of high and low temperature phases of exhalative sulphide deposition (e.g. Brathwaite, 1969, 1974; Green et al., 1981; Green, 1983; Solomon et al., 1990), or alternatively as a product of directional replacement along host-rock cleavage during heterogenous cleavage-parallel extension and cleavage development (Aerden, 1991). Aerden (1991) also utilised the transgressive nature of the ore lenses to support a deformation-related syn-tectonic replacement mode of ore formation. The validity of these interpretations and related genetic models will be discussed in detail in the final chapter.

detailed evidence of deformation and sulphide remobilisation related to Devonian deformation was identified during this study, and is present in thin section and hand specimen. At the meso-scale bands of more competent material such as pyrite are boudinaged, dismembered and rotated, with growth of sulphide and gangue minerals in associated low-pressure domains. The ore is tightly folded in places with cleavage typically preserved within lithic and baritic bands, and rarely within the sulphide bands. The ore itself exhibits durchbewegung and occasional mylonitic fabrics in hand specimen. In thin section the more competent minerals such as pyrite and arsenopyrite have been deformed, fractured, rotated and brecciated, with growth of sulphide and gangue minerals in low-pressure domains. In contrast, the less competent minerals such as sphalerite, galena, chalcopyrite and tetrahedrite exhibit textures indicative of post-deformation annealing. Aerden (1991; 1992) rejected the interpretation that the sulphides had been annealed, instead utilising the lack of tectonic fabric within sulphides to interpret that ore genesis was synchronous with deformation. The primitive textures preserved throughout the primary ore and documented here are not consistent with this interpretation, but instead point to a Cambrian ore formation at or near the seafloor.

Syn-tectonic quartz-carbonate veins typically contain a remobilised sulphide component that decreases with distance from the ore (table 5.2), and when present within the ore they may be variably folded, brecciated and boudinaged. Syn-tectonic sulphide veins and piercement structures are sometimes developed within structurally competent host rocks within and proximal to massive ore, and in places sulphides were remobilised into the axis of folds. Precious metals were preferentially remobilised out of primary ore and enriched within the syn-tectonic vein set.

Devonian granite-related metasomatic fluids have remobilised sphalerite, galena and tetrahedrite out of the ore and replaced them with pyrite-pyrrhotite and magnetite-biotite ± chalcopyrite assemblages. The pyrite-pyrrhotite zone has been enriched in Au, due to the remobilisation of Au out of the magnetite-biotite ± chalcopyrite alteration zone. The presence of both recrystallised sulphide textures and compositional zoning of some sphalerites within metasomatised ore suggests that sulphide annealing occurred during or after the final stages of metasomatism.
Chapter 6 : Metal Distribution within Ore

6.1. Introduction

The aim of this part of the study was to examine the major and trace metal distribution within the ore with a view to gaining an insight into the physicochemical processes that controlled ore formation. These characteristics might then also be utilised to define primary features within the mineralising system, such as the pathways along which hydrothermal fluids passed. The spatial zonation of metals within ore examined during this study can be related to interpreted growth faults that are interpreted to have acted as hydrothermal conduits during ore formation.

6.2. Previous work

Several studies, most relying on the mine grade control dataset, have examined the distribution of major and minor metals throughout the Rosebery ores. Examination of the major metal distribution in cross section has been undertaken on F lens (Brathwaite, 1969, 1974), E lens, G lens and H lens (Green et al., 1981; Green, 1983), AB lens (Huston, 1988; Huston and Large, 1988, 1989), J lens (Khin Zaw, 1991; Westbrook, 1997), P lens (Klaere, 1997), V lens (Chester, 1999) and K lens (Solly, 1999; Vallerine, 2000). These studies confirm the general trend of a chalcopyrite-pyrite-rich zone stratigraphically underlying Zn-Pb-Ag-Au-rich ore, except for G lens where the pattern appears inverted (Adams et al., 1976). Although the southern ore lenses often include basal zones of massive chalcopyrite-pyrite ore (Brathwaite, 1969, 1974; Green et al., 1981; Green, 1983) the northern ore lenses more typically display disseminated chalcopyrite and pyrite at the base of sphalerite-galena-pyrite ore (Klaere, 1997; Solly, 1999; Vallerine, 2000; and this study).

Green et al. (1981) and Green (1983) also examined the lateral variation in Pb, Zn, Cu, Ag, Au and Fe values in grade control data across an “unfolded” longitudinal projection of the southern ore lenses. Although broad scale zonation of most elements was noted, the veracity of the interpretation is difficult to ascertain due to the schematic presentation of the data and lack of evidence to support the interpreted macro-scale folding of the ore. Aerden (1991; 1992) examined the southern ore lenses in plan projection and highlighted the Cu and Fe enrichment of E lens, F lens and G lens in the area that they adjoin (fig.5.4).

Smith (1975), Huston (1988) and Smith and Huston (1992) examined the trace element content of the ore and surrounding host rocks, identifying a Ba-Ag-Sb association in the upper barite mineralisation, a Pb-Zn-Ag-Au-Cd-As-Sb-Hg-Tl association within the sphalerite-galena-pyrite ore, and a Cu-Bi-As-Fe association within the underlying chalcopyrite-pyrite ore. Limited trace element analyses of K lens and P lens ores by Vallerine (2000) and Klaere (1997) are
consistent with these elemental associations.

Huston and Large (1987) examined ores from various mines and prospects, including Rosebery, and utilised zinc ratio (100Zn/(Zn+Pb)) values to differentiate between Cambrian syngenetic ores and Devonian granite-related epigenetic ores in western Tasmania.

6.3. Methodology

Average values of Pb (wt%), Zn (wt%), Au (ppm), Ag (ppm) and Fe (wt%) were compiled for ore grade intercepts across K lens and P lens, with raw values derived from the Rosebery mine grade control dataset. Approximate spatial coordinates and true intercept thicknesses were scaled off mine grade control sections and plans, and the occurrence of barite within or adjacent to the ore intercept was recorded from drill logs compiled by mine site personnel. Data was also collected from poorly mineralised and barren drill holes to aid in defining the lateral limits of ore occurrences. Where possible the ore intercept data was corrected to remove the effects of Devonian faulting with interpreted offsets of more than ~10m (e.g. “K shear” in fig.3.8). The P10, P11 and P13 sub-lenses were not examined in detail due to a limited amount of drill coverage.

Raw K lens assay data was gridded utilising an isotropic point kriging method. The dataset was then contoured on the basis of the data point counts, using the 0-25%, 25-50%, 50-75%, 75-90% and 90-100% subdivisions. To allow direct comparison of metal distribution within K lens and P lens, precisely the same assay value limits were applied to the P lens data. The domains of elevated metal values appear reasonable in the light of geological features such as syn-depositional fault structures, changes in TSV thickness, stratigraphic position and the spatial distribution of barite.

6.3.1. Results

The spatial domains of elevated values within the grade control data are primarily the product of relative elemental enrichment, with mineralisation intensity making a somewhat lesser contribution. Through examination of the elevated elemental domains across K lens (fig.6.1) two spatial associations are evident. Domains of elevated Fe values correlate with domains of elevated Cu values (fig.6.1h). In contrast, domains of elevated Ag values correlate well with domains of elevated Au, and with the presence of barite within or immediately adjacent to the ore intercept (fig.6.1i). The Zn ratio (100Zn/(Zn+Pb)) is elevated proximal to the northeast trending fault trace and exhibits a marked gradient across the upper part of K lens (fig.6.1g). The Ag-Au-barite and Fe-Cu associations within K lens display a moderately antithetic relationship, and elevated Zn ratio domains appear to have a moderate spatial association with elevated Fe-Cu domains. The domains of elevated Pb and Zn values variably overlap the Fe-
Figure 6.1 Major element distribution in K lens ore.
Contour plans of average ore intercept grades as derived from the mine grade control dataset. The view is from hanging wall side of the ore and includes inferred fault traces derived from figure 5.1. (A) Ag, (B) Au, (C) Cu ...
Figure 6.1 Major element distribution in K lens ore.
Contour plans of average ore intercept grades as derived from the mine grade control dataset. The view is from hanging wall side of the ore and includes inferred fault traces derived from figure 5.1. (D) Fe, (E) Pb, (F) Zn ...
Figure 6.1 Major element distribution in K lens ore.

(G) Zn ratio \((100\times \text{Zn}/(\text{Zn} + \text{Pb}))\). Within the ore, zinc ratio values are elevated proximal to northeast trending fault. (H) Coincident Fe and Cu contours define a high temperature trend orthogonal to the northeast trending fault. (I) Au and Ag both exhibit a positive spatial relationship with barite occurrences, and a moderately antithetic relationship with the elevated Fe-Cu domains of the preceding plan.
Figure 6.2 Major element distribution within P lens ore: P6 and P7-P9 sub lenses. Contour plans of average ore intercept grades as derived from the mine grade control dataset. The view is from hanging wall side of the ore and includes inferred fault traces derived from figure 5.1, cross sections (figs. 4.5 to 4.7), and trends within the grade control data. (A) Ag, (B) Au, (C) Cu ...
Figure 6.2 Major element distribution within P lens ore: P6 and P7-P9 sub lenses. Contour plans of average ore intercept grades as derived from the mine grade control dataset. The view is from hanging wall side of the ore and includes inferred fault traces derived from figure 5.1, cross sections (figs. 4.5 to 4.7), and trends within the grade control data. (D) Fe, (E) Pb, (F) Zn...
Figure 6.2 Major element distribution within P lens ore: P6 and P7-P9 sub lenses. 

(G) Zn ratio (100xZn/(Zn+Pb)). Domains of elevated zinc ratio values correspond spatially to domains of elevated Fe-Cu values. Coincident Fe and Cu contours define a high temperature zone that appears antithetic to the area of elevated Ag values. (I) Au exhibits a moderate spatial relationship with barite occurrences in the P7-P9 sub-lenses, and an antithetic relationship with the elevated Cu domain.
Cu and Ag-Au domains.

Although absolute metal values are lower within P lens ore due to moderate ore dilution caused by a higher lithic component, similar associations are evident in the stratigraphically lower P sub-lenses (fig.6.2). An elevated Fe-Cu domain dominates most of the P6 sub-lens, with some indication of an elevated Ag-Pb domain along the up-dip margin. Elevated Fe domains within the stratigraphically higher P7-P9 sub-lenses are spatially associated with domains of elevated Cu and Zn-ratio values. The Fe-Cu-Zn ratio domains are strongly antithetic to domains of elevated Ag values, and moderately antithetic to domains of elevated Au-Pb-Zn values and barite. The strongly baritic P10 sub-lens, located stratigraphically above the P7 and P9 sub-lenses, lacks any significant domain of Fe enrichment but is instead predominantly Ag-enriched.

6.4. Discussion

6.4.1. Metal associations across the ore lenses

The principal ore mineral components of the sulphide ore at Rosebery are sphalerite, pyrite, galena, chalcopyrite, minor tetrahedrite-tennantite and arsenopyrite, with trace levels of native Au, electrum and other sulphosalts. Pb, Zn, Ag, Au, Cd, Sb, As, Hg, Tl and Mo are strongly associated with the sphalerite-galena-rich ore and baritic mineralisation, whereas the chalcopyrite-rich ore is relatively enriched in Cu, Bi, In and Fe (Smith and Huston, 1992; Vallerine, 2000; and this study). These associations are considered analogous to those assemblages observed within Kidd Creek ores, with the low temperature suite of elements within Zn-rich ore which includes Zn, Cd, Ag, Pb, Sn, Sb, As, Hg, ± Tl, ± W, and the high temperature suite found within chalcopyrite-rich ores and stringer zones which includes Cu, Co, Bi, Se, In, Ni, ± As, ± Sn, ± Ag (Hannington et al., 1999a). A similar low temperature elemental suite (Pb, Zn, As, Sb, Cd, Ag, Au, Hg, Tl, Mo, Ge) in the Zn-Pb-rich ore and a high temperature (Cu, Fe, Bi) suite in the Cu-Fe-rich ore has been identified within the Kuroko deposits (Shimazaki, 1974).

The antithetic relationship between the Fe-Cu-Zn ratio association and the Ag ± Au-Pb-Zn-barite association (figures 6.1 & 6.2) was most likely the product of a lateral physicochemical gradient active at the time of ore formation, probably dominated by changes in temperature. The Cu-Fe-Zn ratio value association is interpreted to be the product of elevated temperature, whereas the Ag ± Au-Pb-Zn-barite association is interpreted to be the product of lower temperature.

Metal trends within the P sub-lenses suggest movement of hydrothermal fluids up along Cambrian syn-depositional faults that crosscut stratigraphy, with subsequent lateral fluid
movement toward the north and northeast (fig.6.3). The metal domain patterns exhibited within the P7-P9 sub lens suggest a fault intersection along which vertical fluid flow may have been focussed.

Metal trends within the up-dip half of K lens suggest that elevated fluid temperatures occurred proximal to a large northeast trending syn-depositional fault (fig.6.3) and decreased toward the northwest. Linear patterns exhibited by the high-temperature domains may reflect pathways along which the majority of lateral fluid movement occurred during sub-seafloor replacement, or they could reflect local topographic lows at the time of early high-temperature hydrothermal venting and seafloor sulphide deposition. Drill coverage of the footwall beneath the high temperature domain was not sufficiently detailed to definitively locate fluid upflow zones of the type mapped beneath P lens, however the northeast trending fault is a prime candidate for the location of one or more fluid upflow zones beneath K lens.

**Figure 6.3 Interpreted temperature gradients across northern ore lenses.**
Interpreted temperature gradients across (A) P6 and P7-P9 sub-lenses and (B) K lens. (H = high temperature domain, L = low temperature domain). The arrows may also represent the direction of hydrothermal fluid flow during primary ore formation.
6.4.2. Application of the zinc ratio

Thermodynamic modelling of the chemistry associated with Pb and Zn in ore forming fluids by Huston and Large (1987) indicates that the Zn ratio for a solution saturated in both Pb and Zn is dependent only on the temperature and the activity of chloride, and is independent of pH, \(a_{\text{H}_2\text{S}}\) and \(f_{\text{O}_2}\). The modelling by Huston and Large (1987) also indicates that the Zn ratio will typically increase with decreasing temperature and increasing salinity, except for a trend reversal that is evident between 200°C and 250°C (fig.6.4) for hydrothermal solutions containing at least 1.0m NaCl\(_{eq}\).

The Zn ratio values in K lens decrease from the footwall up into the Zn-Pb ore of K lens (fig 6.5). The Zn ratio values in both K lens and P lens are highest in areas of elevated Cu and Fe values, and decrease toward peripheral domains of elevated Ag values (figures.6.1 and 6.2). This suggests that the Zn ratio in these settings at the time of ore formation was highest in areas of elevated temperature and lowest in areas of reduced temperature. Several alternative processes may be invoked to explain this trend. The observed variation of Zn ratio values across K lens may be the product of a saturated hydrothermal fluid precipitating across a temperature gradient of between 200°C and 250°C, coincident with the trend reversal evident in the modelling by Huston and Large (1987). Alternatively it could have been produced by sphalerite precipitation proximal to the source from hydrothermal fluids saturated in Zn and undersaturated in Pb, coupled with more distal galena precipitation where the fluids had became saturated in Pb due to decreased temperature and/or salinity. A third alternative is

Figure 6.4 Zn ratio value (100\(\text{Zn}/(\text{Zn}+\text{Pb})\)) - variation with temperature and salinity. Modelled variation of zinc ratio value with changes in temperature and salinity for a hydrothermal solution saturated in Pb and Zn. (after Huston and Large, 1987).
Zn ratio values are elevated in the footwall beneath K lens ore in the area from which group A samples have returned pyrite values with elevated Co:Ni and Bi:Pb ratios and sphalerite with elevated Fe contents. Contours based on Zinc ratio values calculated for individual samples from mine grade control dataset.

that following initial sphalerite-galena precipitation the hydrothermal fluids passing laterally through the mineralised zone became warmer, with resultant dissolution of galena and remobilisation to a more distal position where it then re-precipitated. Pb may have been remobilised as an aqueous sulphur complex along with Ag, As, Sb, Hg and Tl (Hannington et al., 1999b), which could account for the association of these elements with Au in K lens ore. Although no diagnostic evidence is available to indicate which process was predominant, sphalerite-galena-rich ore probably formed in the temperature range of 200°C to 250°C (Huston and Large, 1989), which is in the range of the trend reversal evident in the geochemical modelling (fig.6.4).

The spread of data above the minimum cut-off line in the Pb vs Zn plots (fig.6.6) can be attributed to variation in Pb saturation within the hydrothermal fluid, while the lower cut-off limit represents a minimum resulting from saturation in both Pb and Zn for a particular set of temperature and salinity conditions (Huston and Large, 1987). Huston and Large (1987) reported changes in standard deviation values for the zinc ratio data upward through AB lens. A high standard deviation (14.7) was evident within altered footwall volcanics, which they attributed to the influence of higher temperatures of formation. A low standard deviation within barite gangue (9.2) overlying AB lens was attributed to a narrower range of physicochemical conditions at the time of ore formation. Similarly, sulphide ore in K lens in the absence of barite displays a higher standard deviation than sulphide intercepts that contain barite (table 6.1). The same trend to lower standard deviation values in ore intercepts containing barite extends to the P6 (6.7 c.f. 7.7), P7 (7.2 c.f. 8.1), P9 (8.1 c.f. 11.8), P10 (10.9 c.f. 11.9), P11 (5.9 c.f. 7.4) and P13 sub-lenses (3.2 c.f. 9.8). Utilising the assumptions about fluid composition and thermochemical modelling of Huston and Large (1987), the range of Zn ratio cut-off values exhibited by the northern lenses (table 6.1) would indicate temperatures in the range of ~190°C to 260°C, which is consistent with the temperature range suggested by Huston and Large.
Table 6.1 Lower Zinc ratio cut off values for northern ore lenses.
(AB lens data from Huston and Large, 1987).

<table>
<thead>
<tr>
<th>Lens Type</th>
<th>Zn ratio mean</th>
<th>Zn ratio std. deviation</th>
<th>Zn ratio lower cut-off</th>
<th>n</th>
</tr>
</thead>
<tbody>
<tr>
<td>AB lens – pyrite dominant massive ore</td>
<td>71.4</td>
<td>13.1</td>
<td>60</td>
<td>85</td>
</tr>
<tr>
<td>AB lens – sphalerite-galena massive ore</td>
<td>74.5</td>
<td>9.6</td>
<td>60</td>
<td>379</td>
</tr>
<tr>
<td>AB lens – barite ore</td>
<td>68.9</td>
<td>9.2</td>
<td>65.5</td>
<td>94</td>
</tr>
<tr>
<td>K lens – sulphide ore</td>
<td>74.5</td>
<td>10.3</td>
<td>59</td>
<td>381</td>
</tr>
<tr>
<td>K lens – sulphide ore with barite</td>
<td>70.0</td>
<td>7.4</td>
<td>59</td>
<td>111</td>
</tr>
<tr>
<td>P6 lens</td>
<td>73.4</td>
<td>7.5</td>
<td>63</td>
<td>100</td>
</tr>
<tr>
<td>P7 lens/P9 lens</td>
<td>77.2</td>
<td>9.2</td>
<td>63</td>
<td>220</td>
</tr>
<tr>
<td>P10 lens</td>
<td>73.4</td>
<td>12.3</td>
<td>63</td>
<td>137</td>
</tr>
<tr>
<td>P11 lens</td>
<td>78.5</td>
<td>6.8</td>
<td>67</td>
<td>42</td>
</tr>
<tr>
<td>P13 lens</td>
<td>74.7</td>
<td>9.4</td>
<td>63</td>
<td>33</td>
</tr>
</tbody>
</table>

Figure 6.6 Zn ratio “cut-off” values.
Zn vs Pb plots of ore intercepts calculated from mine grade control dataset. The lines represent the lower Zn ratio “cut-off” value as discussed in the text. (A) K lens, (B) P6 sub lens, and (C) P7-P9 sub lens.
(1989) for formation of the sphalerite-galena ore. There is no consistent variation in the Zn ratio values related to the stratigraphic position of the various P sub lenses. The P lens Zn ratio cut-off values are marginally higher than the cut-off values for K lens and AB lens, possibly indicative of overall lower temperature and salinity conditions. The preservation of detrital albite in much of P lens and the paucity of it in K lens is consistent with this interpretation.

Huston and Large (1987) utilised the mean and standard deviation of zinc ratio values to distinguish between sulphide deposits of differing genetic origin in western Tasmania. Although there may be problems with the statistical approach undertaken (Anderson, 1997), the zinc ratio values of Cambrian syngenetic ores do demonstrate a more restricted range and a lower standard deviation than for the Devonian epigenetic ores.

### 6.4.3. The southern ore lenses

Lateral element variation across K lens and P lens is indicative of lateral fluid movement and related elemental remobilisation. It is in this light that the elemental distribution across the southern ore lenses may be re-examined. The stratigraphic and structural setting for the southern lenses can be considered analogous to that of P lens, with sulphide ore present at various stratigraphic levels of the footwall and TSV units. E lens, F lens and G lens are located proximal to interpreted syn-depositional fault structures thought to be responsible for rapid changes in TSV thickness (fig.5.1). The lower margins of E lens and F lens, and the upper margin of G lens are enriched in chalcopyrite and pyrite, as are the lateral margins where these lenses coalesce (figures.5.3 to 5.4). The distribution of chalcopyrite-pyrite enrichment is consistent with lateral fluid movement of high temperature fluids through and adjacent to sphalerite-galena ore with replacement of the primary ore by chalcopyrite-pyrite, possibly in the fashion proposed by Huston and Large (1989). The enrichment of chalcopyrite-pyrite in the vicinity of a possible syngenetic fault suggests that this may have been a locus for upflow of hydrothermal fluids.

### 6.5. Summary

The elemental associations and their distribution within Rosebery ore are analogous to those in other volcanic hosted massive sulphide systems such as Kidd Creek and the Kuroko district. The low temperature elemental suite found within the sphalerite-galena-pyrite ore comprises Pb-Zn-Ag-Au-Cd-As-Sb-Hg-Tl and the high temperature suite within chalcopyrite-pyrite ore comprises Cu-Bi-Fe-In.

Identified within the sphalerite-galena-pyrite ore of K lens and P lens during this study, is a Fe-Cu-Zn ratio value association interpreted to be the result of elevated fluid temperatures, with a
Ag ± Au-Pb-Zn-barite association that is interpreted to be related to lower temperatures of formation. Within these ore lenses the high temperature association occurs proximal to interpreted syngenetic faults that acted as conduits for fluid upflow, whereas the low temperature association is located distal to these structures.

The elemental zonation across K lens and P lens is consistent with lateral movement of hydrothermal fluids through and beneath the ore. Zn ratio value variation from footwall to ore, and laterally across the sphalerite-galena ore of K lens and P lens, may be the product of sulphide precipitation from a Zn-Pb saturated hydrothermal fluid in the range of ~250°C to ~200°C.

The stacked southern ore lens system is considered here to represent a high temperature analogue of the stacked P sub lens system. Elemental zonation exhibited within the southern ore lenses is considered to be more intense than that observed within the northern ore lenses. It is consistent with high temperature fluid upflow along syn-depositional fault traces, in conjunction with lateral hydrothermal fluid flow around and through the sphalerite-galena ore. This resulted in overprinting and replacement of the sphalerite-galena ore by chalcopyrite-pyrite proximal to the zones of fluid upflow, possibly in the fashion put forward by Huston and Large (1989).
Chapter 7 : Sulphide Trace Element Chemistry

7.1. Introduction

The aim of this part of the study was to examine the major and trace elements within the sulphide mineral component of the ore with a view to gaining insight into the kind of physicochemical processes that may have played a role in producing the observed texture, composition and distribution of different sulphides during ore formation.

7.2. Previous work

Several studies have been undertaken on the trace element content of sulphides within the Rosebery ore. Loftus-Hills and Solomon (1967) and Loftus-Hills (1968) analysed Rosebery pyrite concentrates as part of a regional study, and were able to differentiate between pyrites of hydrothermal origin within volcanic rocks and sedimentary or diagenetic pyrites within shales. Groves and Loftus-Hills (1968) examined the Cd content of sphalerites in a similar study, and successfully differentiated between ores of Cambrian syngenetic and Devonian granite-related origin.

The composition of individual sulphosalts within the Rosebery ore was initially examined by Stillwell (1934) and Williams (1960). Smith (1975), Henley and Steveson (1978), Huston (1988), Huston and Large (1988) and Vallerine (2000) undertook more detailed examination of the tennantite-tetrahedrite series, identifying it as the major silver-bearing mineral within Rosebery ore. Huston (1988) was able to differentiate between primary and remobilised tetrahedrite-tennantite on the basis of Ag and Sb content. The chemistry of Au and electrum in the Rosebery ores has been examined by Huston (1988), Huston and Large (1988), Huston et al. (1992) and Vallerine (2000), with Au interpreted to have originally been transported in the footwall and Cu-rich ores as AuCl₂ and in the Zn-rich massive ore as Au(HS)₂, with a significant part of the Au remobilised into fractures and along grain boundaries during subsequent metamorphism (Huston and Large, 1988, 1989).

Henley and Steveson (1978) examined the precious metal content of several sulphide species, including pyrite. Green et al. (1981) and Green (1983) utilised the Co and Ni content of pyrites in the footwall at Rosebery in an attempt to identify the position of a possible fluid conduit beneath the southern ore lenses. The Se content of pyrite was examined by Huston et al. (1995b), who concluded that the low Se content of Rosebery pyrites might be attributed to the lack of magmatic fluid input. Huston et al. (1993) and Huston et al. (1995b) further examined the trace element content of pyrite, chalcopyrite and sphalerite in the Rosebery ore, dividing the elemental occurrences into three types: (1) elements that occur principally as inclusions,
elements that occur as non-stoichiometric substitutions within the crystal lattice, and (3) elements that occur as stoichiometric substitutions within the crystal lattice. Huston et al. (1996) examined the silver content of the principal ore minerals in the sulphide ore and barite mineralisation.

Green et al. (1981) and Green (1983) analysed sphalerite across the southern ore lenses, noting increased Fe-content within ore lenses enriched in chalcopyrite and pyrite and low Fe-sphalerite within the baritic lode. Khin Zaw (1991) and Khin Zaw & Large (1996) examined the Fe, Mn and Cd content of sphalerite from Rosebery ores, distinguishing between Cambrian syngenetic and Devonian granite-related ores on the basis of the elemental ratios and calculating estimates of pressure conditions under which Devonian replacement processes took place.

7.3. Analytical methods

Mineralised rock specimens were collected from two diamond drill hole traverses across the K lens ore body (fig.7.1). The traverses were designed to extend laterally out of the massive sulphide ore into weakly mineralised host rocks outside of the ore lens, although drill coverage restricted the extent of one traverse. The selected drill holes were spaced at ~20m intervals across the lens, and where possible they were selected so as to avoid any ore obviously overprinted by Devonian metasomatism or late stage brittle faulting. Samples from each drill hole were collected at ~2m intervals (true thickness) vertically through the mineralised horizon. The samples were then mounted into 2.5cm circular resin blocks and polished. Polished mounts were also prepared from syn-tectonic quartz-carbonate-sulphide veins and metasomatised ore found within the Rosebery mine, and from examples of Devonian granite-related, vein-hosted sulphide mineralisation from mines and prospects elsewhere in the Rosebery district.

Figure 7.1 K lens geochemical traverses. K lens ore outline showing the two drill hole profiles sampled for geochemical analysis. Ore outline from 2002/2003 mine production plan.
Samples were examined under reflected light and significant sulphide textures were recorded. Individual sulphide grains selected for analysis were digitally photographed and their position electronically recorded. The samples were carbon coated and individual sulphide grains were analysed on a Cameca SX50 electron microprobe located in the Central Science Laboratory at the University of Tasmania. On each polished mount 1-4 grains of each sulphide species were analysed, with 1-3 adjacent analyses carried out on each grain. Several samples were also subjected to electron microprobe analysis after the laser ablation study so as to confirm the Tl content of selected pyrite grains. A specimen containing coarsely-crystalline lorandite (Tl\textsubscript{3}As\textsubscript{2}S\textsubscript{6}) from the Lanmuchang Tl mine in China (Guozhu, 1996) was prepared as a polished probe mount and utilised to calibrate the electron microprobe for Tl analysis. Microprobe analysis for Se was undertaken on all sulphide grains, however the results were discounted as the Se results have been significantly elevated by signal interference related to the presence of As. Microprobe analyses for As within sphalerite were similarly discounted due to significant signal interference related to the presence of Fe.

After microprobe analysis the sulphide grains were subjected to laser ablation - inductively coupled - plasma mass spectrometry (LA-ICP-MS) analysis. The laser spot target on each sulphide grain was selected so as to avoid obvious inclusion material. The spots on each sulphide grain were ablated utilising either a Merchantek LUV266X laser ablation system or a Merchantek UP-213 laser ablation system, each with a frequency quadrupled Nd:YAG laser and producing a 266nm or 213nm wavelength infrared laser beam respectively. The laser repetition rate was set within a range of 5-10 Hz, the percentile power was set within a range of 45-60% (mJ/pulse) and the beam diameter in a range of 10-50\(\mu\)m. The power setting was adjusted so as to avoid condensate material accumulating around the ablation pit and to prevent the spalling of mineral fragments during ablation. Ablation was undertaken in a helium atmosphere and the ablated material was subsequently mixed with argon gas before introduction to an HP4500 induced plasma mass spectrometer (ICP-MS) operated in time resolved mode with one peak per mass and dwell times of 50msec per mass. A glass standard synthesised from certified high sulphide rock standards (RTS-4 and CZN-1 (Canmet, 1994)) as part of a similar study by Clark (2001) was utilised as the standard (STDGL-1) against which the analyses in this study were calibrated. The glass standard was analysed twice at the beginning and end of each analytical run, with 2-20 sulphide spots analysed during each run. Data was collected for a suite of elements that included Ti, V, Mn, Co, Ni, Cu, Zn, As, Mo, Ag, Sn, Sb, Au, Tl, Pb and Bi. The equipment was not set up to record major element data such as Cu in chalcopyrite and Zn in sphalerite.

Raw time-resolved data spectra derived from the HP4500 were reviewed and analyses were rejected where the laser had obviously ablated into the surrounding matrix or where large sulphide inclusions were encountered below the surface of the target grain. The raw data was reduced utilising LAMTRACE software (Jackson, 1996b), and the analytical detection limit was
set at 3x the calculated limit of detection for each element in each analysis. Laser data reduction requires the input of a known elemental value for each sulphide species. Ideal stoichiometric Fe values were utilised for this purpose as the initial microprobe analyses demonstrated that the Fe contents of pyrite, arsenopyrite and chalcopyrite were within a few percent of ideal stoichiometric values. The Fe content of sphalerite was found to be much more variable between samples, so a single value could not be utilised. The Fe content of sphalerite was, however, found to be reasonably consistent over a single probe mount, so the average Fe value for each probe mount was utilised to reduce the raw laser data. A similar strategy has been utilised in other studies (e.g. Blevin and Jackson, 1998).

7.4. Comments on LA-ICP-MS analysis

The accuracy of LA-ICP-MS analysis can be affected by several factors, including variation in ablation characteristics of different elements and target matrices (Morrison et al., 1995; Perkins et al., 1997; Chen, 1999), background versus signal counting statistics (Norman et al., 1996) and the precision of internal standard elemental concentrations (Norman et al., 1996). Even with these inherent errors the technique still permits analytical accuracies to within an order of magnitude (Norman et al., 1996; Norman et al., 1998). This level of analytical uncertainty is typically much lower than the variation in analyses observed during this study that can be directly attributed to heterogeneity in the elemental distribution within each sulphide grain (fig.7.2). Average values for elements in common to both the LA-ICP-MS and electron microprobe techniques appear to be of the same order of magnitude which, taking into account the sample volume difference for the two analytical techniques and the marked heterogeneity of elemental distribution, would indicate that the LA-ICP-MS technique can be regarded as quantitative.

Figure 7.2 Electron microprobe backscatter images of pyrite.
This pyrite mass from K lens ore (sample C828) displays typical elemental heterogeneity. (A) Fe - Kα peak (light grey = pyrite). (B) As - Lα peak. (C) Tl - Lα peak. A peak overlap for Tl and Pb means this image is displaying counts for both elements. (field of view ~580µ, white = high count) Images courtesy of D. Steele.

The maximum total amount of all trace elements in any single LA-ICP-MS analysis retained for this study was less than 3wt%, with ~90% of analyses containing trace element totals below 1wt%. Several features observable within the time-resolved LA-ICP-MS data profiles indicated
the occurrence of significant elemental associations within the ablated sulphide grains (fig.7.3). Discrete elemental peaks observed during pyrite analysis demonstrated a strong association between Au and Ag, between Pb-Sb-Ag ± Bi ± Tl and between Ti-V ± Mn. The Cu ± Pb ± Ag ± Sb ± Bi association was also observed within sphalerite grains. Time-resolved LA-ICP-MS data for some elements displayed more gradational changes with time, without a consistent trend toward higher or lower values with time that might be attributed to processes of differentiation during ablation and delivery to the ICP-MS. Within pyrite the elements that displayed this kind of behaviour included As, Sb, Co, Ni and Tl, whereas trace elements within sphalerite, chalcopyrite and tetrahedrite did not exhibit this kind of pattern.

Figure 7.3 Time-resolved LA-ICP-MS data profile.
This profile exhibits a response created by ablation of an electrum inclusion within pyrite.

For the purpose of evaluating all subsets within the analytical data by graphical and statistical means it was at times necessary to assign a value to analyses that were below their calculated detection limit. The LA-ICP-MS data is not simple to deal with, as there is a unique detection limit calculated for every element of each analysis, and this leads to a range of detection limits for each element within the data set. It therefore follows that the range of detection limit values overlaps with the range of “measured” values that are above their uniquely calculated detection limits (fig.7.4). To permit graphical plotting of the data, any analysis that was below its calculated detection limit has been assigned a value of 0.5x the calculated detection limit (0.5xdl). A line defining the 95th percentile of the 0.5xdl range has been plotted on the x-y graphs to indicate that beneath this line there is a mixture of “measured” and 0.5xdl data, and therefore interpretation of data below these lines should be undertaken with caution. The trace element datasets approximate lognormal distributions with the range of each element typically
extending over several orders of magnitude, and the trace element data has therefore been displayed as log-log plots. Where elemental ratio values are plotted only those analyses above detection limits have been utilised.

7.5. Results

On the basis of relict pyrite morphology and stratigraphic position, specimens of K lens mineralisation were divided into three main groups that included:

- **Group A** - mineralised footwall to ore samples containing only euhedral to subhedral pyrite without primitive morphologies (e.g. fig.5.16h).
- **Group B** - ore samples containing pyrite with relict “primitive” pyrite morphologies (e.g. fig.5.16j).
- **Group C** - ore samples containing only euhedral to anhedral pyrite without primitive morphologies (e.g. fig.5.17k).

Several samples of weakly mineralised host, syn-deformation quartz-carbonate-sulphide veins and metasomatised ore from Rosebery, and Devonian granite-related mineralisation were also analysed.
7.5.1. Pyrite

Electron microprobe analysis of the pyrites returned values below detection limit for all minor and trace elements (Table 7.1) with the exception of As (ave. 0.14wt%). Overall the group B pyrites are relatively enriched in As in comparison with group A and C pyrites, a result also reflected in the LA-ICP-MS data.

LA-ICP-MS analysis of pyrites returned a significant number of values above detection limit for most trace elements. Group A pyrites are relatively enriched in Co and Bi (table 7.1), whereas group B samples contain pyrites with elevated levels of Mn, Ni, As, Ag, Au, Pb, Sb and Tl. Group C pyrites typically exhibit ranges for these elements that overlap or are transitional between the A and B pyrite groups. Group C pyrites are however slightly Sn-enriched in comparison with group A and B pyrites.

<table>
<thead>
<tr>
<th>PYRITE</th>
<th>Group A - footwall</th>
<th>Group B - primitive</th>
<th>Group C - non-primitive</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>average*</td>
<td>maximum</td>
<td>average*</td>
</tr>
<tr>
<td>microprobe (n=96)</td>
<td>(n=147)</td>
<td>(n=130)</td>
<td></td>
</tr>
<tr>
<td>Mn (wt%)</td>
<td>bdl</td>
<td>0.030</td>
<td>bdl</td>
</tr>
<tr>
<td>Co (wt%)</td>
<td>bdl</td>
<td>0.154</td>
<td>bdl</td>
</tr>
<tr>
<td>Ni (wt%)</td>
<td>bdl</td>
<td>0.019</td>
<td>bdl</td>
</tr>
<tr>
<td>Cu (wt%)</td>
<td>bdl</td>
<td>0.862</td>
<td>bdl</td>
</tr>
<tr>
<td>As (wt%)</td>
<td>0.096</td>
<td>0.532</td>
<td>0.202</td>
</tr>
<tr>
<td>Ag (wt%)</td>
<td>bdl</td>
<td>0.025</td>
<td>bdl</td>
</tr>
<tr>
<td>Au (wt%)</td>
<td>bdl</td>
<td>0.039</td>
<td>bdl</td>
</tr>
<tr>
<td>LA-ICPMS (n=76)</td>
<td>(n=164)</td>
<td>(n=126)</td>
<td></td>
</tr>
<tr>
<td>Ti (ppm)</td>
<td>45.1</td>
<td>1091</td>
<td>28.1</td>
</tr>
<tr>
<td>V (ppm)</td>
<td>0.16</td>
<td>1.35</td>
<td>0.48</td>
</tr>
<tr>
<td>Mn (ppm)</td>
<td>69.6</td>
<td>1388</td>
<td>288</td>
</tr>
<tr>
<td>Co (ppm)</td>
<td>58.3</td>
<td>710</td>
<td>14.0</td>
</tr>
<tr>
<td>Ni (ppm)</td>
<td>3.75</td>
<td>36.7</td>
<td>27.4</td>
</tr>
<tr>
<td>Cu (ppm)</td>
<td>277</td>
<td>7815</td>
<td>308</td>
</tr>
<tr>
<td>Zn (ppm)</td>
<td>162</td>
<td>10425</td>
<td>1550</td>
</tr>
<tr>
<td>As (ppm)</td>
<td>624</td>
<td>6430</td>
<td>1799</td>
</tr>
<tr>
<td>Mo (ppm)</td>
<td>15.8</td>
<td>283</td>
<td>13.1</td>
</tr>
<tr>
<td>Ag (ppm)</td>
<td>3.09</td>
<td>39.7</td>
<td>30.1</td>
</tr>
<tr>
<td>Sn (ppm)</td>
<td>0.44</td>
<td>3.59</td>
<td>0.59</td>
</tr>
<tr>
<td>Sb (ppm)</td>
<td>7.23</td>
<td>108</td>
<td>114</td>
</tr>
<tr>
<td>Au (ppm)</td>
<td>1.12</td>
<td>27.3</td>
<td>6.90</td>
</tr>
<tr>
<td>Ti (ppm)</td>
<td>0.18</td>
<td>2.17</td>
<td>90.9</td>
</tr>
<tr>
<td>Pb (ppm)</td>
<td>198</td>
<td>3969</td>
<td>1080</td>
</tr>
<tr>
<td>Bi (ppm)</td>
<td>23.9</td>
<td>850</td>
<td>1.19</td>
</tr>
<tr>
<td>Co/Ni</td>
<td>15.5</td>
<td>-</td>
<td>0.51</td>
</tr>
<tr>
<td>Bi/Pb</td>
<td>0.1207</td>
<td>-</td>
<td>0.0011</td>
</tr>
</tbody>
</table>

*Analyses below detection limit were assigned 0.5x detection limit for purposes of calculating average values.

Table 7.1 Electron microprobe and LA-ICP-MS results - pyrite.

Refer to text for characteristics of groups A, B and C.

Some elements in the pyrite dataset, for example Sb, Ag, Pb and Cu, display evidence of moderate to strong positive correlation in at least one of the sample groups (fig.7.5a-f) whereas other elements such as As and Cu display no obvious correlation with each other (fig.7.5i). Several combinations of elemental associations, with the simplest being Pb-Sb-Ag-Cu, Au-As-Sb ± Tl and Au-Sb-Ag-Cu, can explain the observed pattern of element correlation (table 7.2). The Pb-Sb-Ag-Cu association evident in the group A, B and C pyrites (fig.7.5a-c)
Chapter 7

extends to include Bi in the group A pyrites (fig.7.5d). The Au-As-Sb ± Tl and Au-Sb-Ag-Cu associations are primarily evident within group B and group C pyrites (fig.7.5e-l). Although not exhibiting an obvious linear correlation, high levels of Tl are commonly associated with elevated Au, Sb and As within group B pyrites (fig.7.5g-i), an association confirmed by subsequent electron microprobe analysis where a strong Tl-Sb-Hg-As association was noted (table 7.3).

Table 7.2 Elemental correlation trends - pyrite.

extends to include Bi in the group A pyrites (fig.7.5d). The Au-As-Sb ± Tl and Au-Sb-Ag-Cu associations are primarily evident within group B and group C pyrites (fig.7.5e-l). Although not exhibiting an obvious linear correlation, high levels of Tl are commonly associated with elevated Au, Sb and As within group B pyrites (fig.7.5g-i), an association confirmed by subsequent electron microprobe analysis where a strong Tl-Sb-Hg-As association was noted (table 7.3).

Table 7.2 Elemental correlation trends - pyrite.

<table>
<thead>
<tr>
<th>Sample</th>
<th>grain</th>
<th>pyrite morphology</th>
<th>W%(As)</th>
<th>W%(Ag)</th>
<th>W%(Sb)</th>
<th>W%(Hg)</th>
<th>W%(Tl)</th>
</tr>
</thead>
<tbody>
<tr>
<td>C01-828 py1</td>
<td>embayed subhedral pyrite</td>
<td>0.517</td>
<td>0.042</td>
<td>0.012</td>
<td>&lt;0.022</td>
<td>&lt;0.024</td>
<td></td>
</tr>
<tr>
<td>C01-828 py1</td>
<td>embayed subhedral pyrite</td>
<td>0.177</td>
<td>&lt;0.012</td>
<td>&lt;0.010</td>
<td>&lt;0.022</td>
<td>&lt;0.024</td>
<td></td>
</tr>
<tr>
<td>C01-828 py1</td>
<td>embayed subhedral pyrite</td>
<td>0.116</td>
<td>&lt;0.012</td>
<td>&lt;0.010</td>
<td>&lt;0.022</td>
<td>&lt;0.024</td>
<td></td>
</tr>
<tr>
<td>C01-828 py2</td>
<td>anhedral relict spongiform mass</td>
<td>0.408</td>
<td>0.020</td>
<td>&lt;0.010</td>
<td>&lt;0.022</td>
<td>0.089</td>
<td></td>
</tr>
<tr>
<td>C01-828 py2</td>
<td>anhedral relict spongiform mass</td>
<td>0.344</td>
<td>0.017</td>
<td>0.026</td>
<td>0.029</td>
<td>0.156</td>
<td></td>
</tr>
<tr>
<td>C01-828 py2</td>
<td>anhedral relict spongiform mass</td>
<td>0.369</td>
<td>&lt;0.012</td>
<td>0.038</td>
<td>0.031</td>
<td>0.163</td>
<td></td>
</tr>
<tr>
<td>C01-828 py3</td>
<td>anhedral relict spongiform mass</td>
<td>0.343</td>
<td>&lt;0.012</td>
<td>0.033</td>
<td>&lt;0.022</td>
<td>0.170</td>
<td></td>
</tr>
<tr>
<td>C01-828 py3</td>
<td>anhedral relict spongiform mass</td>
<td>0.387</td>
<td>0.024</td>
<td>0.064</td>
<td>0.046</td>
<td>0.188</td>
<td></td>
</tr>
<tr>
<td>C01-828 py4</td>
<td>anhedral relict spongiform mass</td>
<td>0.438</td>
<td>0.030</td>
<td>0.080</td>
<td>&lt;0.022</td>
<td>0.242</td>
<td></td>
</tr>
<tr>
<td>C01-828 py4</td>
<td>anhedral relict spongiform mass</td>
<td>0.270</td>
<td>&lt;0.012</td>
<td>&lt;0.010</td>
<td>&lt;0.022</td>
<td>0.032</td>
<td></td>
</tr>
<tr>
<td>C01-828 py4</td>
<td>anhedral relict spongiform mass</td>
<td>0.297</td>
<td>0.016</td>
<td>0.067</td>
<td>0.036</td>
<td>0.203</td>
<td></td>
</tr>
<tr>
<td>C01-828 py4</td>
<td>anhedral relict spongiform mass</td>
<td>0.573</td>
<td>0.034</td>
<td>0.056</td>
<td>&lt;0.022</td>
<td>0.261</td>
<td></td>
</tr>
<tr>
<td>C01-828 py5</td>
<td>anhedral relict spongiform mass</td>
<td>0.103</td>
<td>0.024</td>
<td>&lt;0.010</td>
<td>&lt;0.022</td>
<td>&lt;0.024</td>
<td></td>
</tr>
<tr>
<td>C01-828 py6</td>
<td>anhedral relict spongiform mass</td>
<td>0.109</td>
<td>&lt;0.012</td>
<td>&lt;0.010</td>
<td>&lt;0.022</td>
<td>&lt;0.024</td>
<td></td>
</tr>
<tr>
<td>C01-828 py7</td>
<td>embayed subhedral pyrite</td>
<td>0.102</td>
<td>0.015</td>
<td>&lt;0.010</td>
<td>&lt;0.022</td>
<td>&lt;0.024</td>
<td></td>
</tr>
</tbody>
</table>

Table 7.3 Electron microprobe Tl analyses - pyrite.

The occurrence of trace elements within pyrite is in general agreement with the types proposed by Huston et al. (1993) and Huston et al. (1995b) that includes: (1) elements that occur principally as inclusions, (2) elements that occur as non-stoichiometric substitutions in the lattice, and (3) elements that occur as stoichiometric substitutions. If the elemental associations evident in the analytical data are taken into consideration in conjunction with trends within time-resolved LA-ICP-MS data profiles, petrographic observations and previous studies, the principal occurrence of elements within pyrite can be established. Pyrite appears to contain inclusions of galena (Pb, Sb, Ag, Bi), various sulphosalts (Cu, As, Sb, Ag, Bi, Sb,
Figure 7.5 LA-ICP-MS analysis scatter plots - pyrite.
(A) Pb vs Ag. (B) Pb vs Sb. (C) Pb vs Cu. (D) Pb vs Bi. (E) Au vs As. (F) Au vs Sb.
Figure 7.5 cont... LA-ICP-MS analysis scatter plots - pyrite
(G) Tl vs Au. (H) Tl vs As. (I) Tl vs Sb. (J) Au vs Ag. (K) Au vs Cu. (L) As vs Cu.
Pb, Tl, Hg), electrum (Au, Ag, Hg), arsenopyrite (As), sphalerite (Zn, Mn, Cd) and chalcopyrite (Cu). Elemental substitution of Fe by Co, Ni, As, Sb, Mn, Sn, Mo, (possible low Au, Ag, Cu, Tl) has occurred within the crystal lattice.

7.5.2. Sphalerite

Electron microprobe analysis of sphalerites returned values below detection limit for all minor and trace elements with the exception of Fe, Mn and Cd. Sphalerites exhibit negligible variation in Fe, Mn and Cd content within individual grains, with most samples exhibiting less than 0.5wt% variation in Fe content between sphalerite grains within a single sample. Group A sphalerites typically exhibit elevated Fe values, whereas group B and C sphalerites exhibit elevated Zn values and marginally elevated Cd values (table 7.4 and fig.7.6). Group B sphalerites exhibit a greater range of Cd values than either group A or group C sphalerites. Within all sphalerites an increased Fe or Mn content corresponds to a decrease in both Zn and Cd values.

During LA-ICP-MS analysis of sphalerites only Ag, Mn, Sn, Sb, Cu, Pb, and Bi returned a significant number of values above their respective detection limits (table 7.4). Group A sphalerites from the footwall are confirmed to be relatively enriched in Fe, Mn, Cu, and As, whereas the group B and C sphalerites from the ore are enriched in Zn, Ag, Sb and Pb. Group A and C sphalerites are both enriched in Bi in comparison with group B sphalerites.

<table>
<thead>
<tr>
<th>SPHALERITE</th>
<th>Group A - footwall</th>
<th>Group B - primitive</th>
<th>Group C - non-primitive</th>
</tr>
</thead>
<tbody>
<tr>
<td>microprobe</td>
<td>average*</td>
<td>maximum</td>
<td>average*</td>
</tr>
<tr>
<td>Mn (wt%)</td>
<td>0.165</td>
<td>0.312</td>
<td>0.133</td>
</tr>
<tr>
<td>Fe (wt%)</td>
<td>5.41</td>
<td>8.36</td>
<td>3.19</td>
</tr>
<tr>
<td>Ni (wt%)</td>
<td>bdl</td>
<td>bdl</td>
<td>bdl</td>
</tr>
<tr>
<td>Cu (wt%)</td>
<td>0.334</td>
<td>3.13</td>
<td>0.054</td>
</tr>
<tr>
<td>Zn (wt%)</td>
<td>60.7</td>
<td>66.8</td>
<td>63.6</td>
</tr>
<tr>
<td>As (wt%)</td>
<td>bdl</td>
<td>bdl</td>
<td>bdl</td>
</tr>
<tr>
<td>Ag (wt%)</td>
<td>0.060</td>
<td>0.214</td>
<td>0.179</td>
</tr>
<tr>
<td>Cd (wt%)</td>
<td>0.173</td>
<td>0.214</td>
<td>0.179</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>LA-ICP-MS</th>
<th>(n=76)</th>
<th>(n=164)</th>
<th>(n=126)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mn (ppm)</td>
<td>2043</td>
<td>2994</td>
<td>1248</td>
</tr>
<tr>
<td>Cu (ppm)</td>
<td>1751</td>
<td>12125</td>
<td>586</td>
</tr>
<tr>
<td>As (ppm)</td>
<td>20.5</td>
<td>3.83</td>
<td>10.2</td>
</tr>
<tr>
<td>Ag (ppm)</td>
<td>6.4</td>
<td>19.3</td>
<td>15.3</td>
</tr>
<tr>
<td>Sn (ppm)</td>
<td>1.36</td>
<td>2.07</td>
<td>1.33</td>
</tr>
<tr>
<td>Sb (ppm)</td>
<td>3.44</td>
<td>15.4</td>
<td>13.2</td>
</tr>
<tr>
<td>Pb (ppm)</td>
<td>36.5</td>
<td>367</td>
<td>327</td>
</tr>
<tr>
<td>Bi (ppm)</td>
<td>3.47</td>
<td>13.2</td>
<td>0.42</td>
</tr>
</tbody>
</table>

*Analyses below detection limit were assigned 0.5x detection limit for purposes of calculating average values.

Table 7.4 Electron microprobe and LA-ICP-MS analyses - sphalerite.

Refer to text for characteristics of groups A, B and C.

Positive correlations occur between Pb, Ag and Sb within group A, B and C sphalerites, extending to include Bi in the group A sphalerites (fig.7.7a-d). A weak inverse correlation occurs between the Pb-Ag-Sb content and the Fe and Mn contents in sphalerite, with group A
Figure 7.6 Microprobe analysis scatter plots - sphalerite.
(A) Cd vs Fe. (B) Cd vs Mn. (C) Zn vs Fe.

Table 7.5 Elemental correlation trends - sphalerite.

<table>
<thead>
<tr>
<th></th>
<th>Mn</th>
<th>Cd</th>
<th>Fe</th>
<th>Zn</th>
<th>As</th>
<th>Cu</th>
<th>Pb</th>
<th>Ag</th>
<th>Sb</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cd</td>
<td>N</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Fe</td>
<td>N</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Zn</td>
<td>N</td>
<td>N</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>As</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
</tr>
<tr>
<td>Cu</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
</tr>
<tr>
<td>Pb</td>
<td>-</td>
<td>N</td>
<td>-</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
</tr>
<tr>
<td>Ag</td>
<td>-</td>
<td>N</td>
<td>-</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
</tr>
<tr>
<td>Sb</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
</tr>
<tr>
<td>Bi</td>
<td>-A</td>
<td>-A</td>
<td>-A</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
<td>N</td>
</tr>
</tbody>
</table>

+/- = positive/negative correlation, N = no correlation, A = group A, B = group B

Table 7.5 Elemental correlation trends - sphalerite.
Figure 7.7 LA-ICP-MS analysis scatter plots - sphalerite.
(A) Pb vs Ag. (B) Pb vs Bi. (C) Pb vs Sb. (D) Sb vs Ag. (E) Pb vs Fe. (F) Pb vs Mn. (Fe and Mn analyses undertaken with microprobe.)
sphalerites exhibiting elevated Fe and Mn and reduced levels of Pb-Ag-Sb (fig.7.7e-f).

The lack of compositional variation of sphalerite over distances of several centimetres is consistent with homogenisation during annealing and recrystallisation, and as a result of recrystallisation most inclusions within sphalerite are located along grain and crystallographic boundaries. If the elemental associations evident in the analytical data are taken into consideration along with trends within time-resolved LA-ICP-MS data profiles, petrographic observations and previous studies, the principal occurrence of elements within sphalerite can be established. Sphalerite appears to contain inclusions of galena (Pb, Sb, Ag, Bi), various sulphosalts (Sb, Ag, Bi, Pb, Cu), electrum (Au, Ag, Hg) and chalcopyrite (Cu). Cd, Fe, Mn, As, Co and Sn have substituted for Zn within the crystal lattice.

7.5.3. Tetrahedrite and tennantite

Electron microprobe analysis of tennantite-tetrahedrite in Rosebery ore (n=79) returned values above detection limit for Ag, Cu, As, Sb, Fe, Te and Zn (table 7.6). The analysed grains display a reciprocal substitution for the element pairs Cu-Ag, As-Sb and to a lesser degree Fe-Zn, along with a non-linear inverse relationship between Ag and As typical of the tennantite-tetrahedrite series (fig.7.8) (Sack and Loucks, 1985; Johnson et al., 1986; Seal et al., 1990).

The fine size of tennantite-tetrahedrite grains in the ore samples limited the number of LA-ICP-MS analyses to 9, with Mn values to 182ppm (ave. 124ppm), Sn values to 214ppm (ave. 28ppm), and Bi values to 114ppm (ave. 52ppm).

7.5.4. Galena

Electron microprobe analysis of galena from group B and C samples returned a significant number of values above detection limit for Ag, Zn and Sb (table 7.7). Group B galena is on average more enriched in Sb, whereas group C is more enriched in Zn and Ag. Zn is most likely present as inclusions of sphalerite, whereas Ag and Sb substitute for Pb in the galena lattice. Galena was not analysed by LA-ICP-MS during this study.
Figure 7.8 Microprobe analysis scatter plots - tennantite-tetrahedrite.
(A) Ag vs Cu. (B) Sb vs As. (C) Zn vs Fe. (D) Ag vs As.

Table 7.7 Electron microprobe analyses—galena.
Refer to text for characteristics of groups A, B and C.
7.5.5. Arsenopyrite, chalcopyrite & pyrrhotite

Electron microprobe analysis of chalcopyrite returned values above detection limit for As, Ag and Au (Table 7.8). Microprobe analysis of arsenopyrite (n=25) did not return values above detection limit for minor and trace elements. Analysis of pyrrhotite (n=9) from several samples affected by Devonian metasomatism returned only As values above the detection limit, with average values of 1000ppm (max. 1107ppm).

The paucity of LA-ICP-MS data for arsenopyrite and chalcopyrite (Table 7.9), primarily due to small target relative to laser beam size, means that differences between sample groups could not be evaluated. The analyses reveal that chalcopyrite contains significant levels of Ag and Sn, and that arsenopyrite contains significant levels of Co, Ni, Sb, Au and Tl, in comparison with pyrite and sphalerite.

<table>
<thead>
<tr>
<th>CHALCOPYRITE</th>
<th>Group A - footwall</th>
<th>Group B - primitive</th>
<th>Group C - non-primitive</th>
</tr>
</thead>
<tbody>
<tr>
<td>average*</td>
<td>maximum</td>
<td>average*</td>
<td>maximum</td>
</tr>
<tr>
<td>Mn (wt%)</td>
<td>bdl</td>
<td>0.210</td>
<td>bdl</td>
</tr>
<tr>
<td>As (wt%)</td>
<td>0.098</td>
<td>0.606</td>
<td>0.067</td>
</tr>
<tr>
<td>Ag (wt%)</td>
<td>0.020</td>
<td>0.048</td>
<td>0.020</td>
</tr>
<tr>
<td>Au (wt%)</td>
<td>bdl</td>
<td>0.046</td>
<td>bdl</td>
</tr>
</tbody>
</table>

*Analyses below detection limit were assigned 0.5x detection limit for purposes of calculating average values.

Table 7.8 Electron microprobe analyses—chalcopyrite.
Refer to text for characteristics of groups A, B and C.

The paucity of LA-ICP-MS data for arsenopyrite and chalcopyrite (Table 7.9), primarily due to small target relative to laser beam size, means that differences between sample groups could not be evaluated. The analyses reveal that chalcopyrite contains significant levels of Ag and Sn, and that arsenopyrite contains significant levels of Co, Ni, Sb, Au and Tl, in comparison with pyrite and sphalerite.

<table>
<thead>
<tr>
<th>CHALCOPYRITE</th>
<th>ARSENOPYRITE</th>
</tr>
</thead>
<tbody>
<tr>
<td>average*</td>
<td>average*</td>
</tr>
<tr>
<td>maximum</td>
<td>maximum</td>
</tr>
<tr>
<td>Ti (ppm)</td>
<td>bdl</td>
</tr>
<tr>
<td>V (ppm)</td>
<td>bdl</td>
</tr>
<tr>
<td>Mn (ppm)</td>
<td>bdl</td>
</tr>
<tr>
<td>Co (ppm)</td>
<td>bdl</td>
</tr>
<tr>
<td>Ni (ppm)</td>
<td>bdl</td>
</tr>
<tr>
<td>Cu (ppm)</td>
<td>bdl</td>
</tr>
<tr>
<td>Zn (ppm)</td>
<td>1125</td>
</tr>
<tr>
<td>As (ppm)</td>
<td>55.8</td>
</tr>
<tr>
<td>Mo (ppm)</td>
<td>bdl</td>
</tr>
<tr>
<td>Ag (ppm)</td>
<td>213</td>
</tr>
<tr>
<td>Sn (ppm)</td>
<td>178</td>
</tr>
<tr>
<td>Sb (ppm)</td>
<td>21.5</td>
</tr>
<tr>
<td>Au (ppm)</td>
<td>bdl</td>
</tr>
<tr>
<td>Tl (ppm)</td>
<td>bdl</td>
</tr>
<tr>
<td>Pb (ppm)</td>
<td>298</td>
</tr>
<tr>
<td>Bi (ppm)</td>
<td>4.85</td>
</tr>
</tbody>
</table>

*Analyses below detection limit were assigned 0.5x detection limit for purposes of calculating average values.

Table 7.9 LA-ICP-MS analyses - chalcopyrite and arsenopyrite.

7.5.6. Discussion

Variation in absolute elemental contents and elemental ratios within individual sulphide species may reflect physicochemical controls at the time of ore formation or changes brought on by later metamorphism and metasomatism. As discussed in an earlier chapter, the
indications are that within Rosebery ore more competent pyrite and arsenopyrite have been little affected by metamorphism, whereas sphalerite, galena and tennantite-tetrahedrite appear to have been variably remobilised, recrystallised and annealed. The near ubiquitous presence of pyrite in and around the ore therefore makes it the most suitable candidate as an indicator of changes in physicochemical controls at the time of primary sulphide deposition.

7.5.7. Co:Ni in pyrite

Co and Ni substitute for Fe in pyrite as part of a solid solution series between FeS$_2$, NiS$_2$ and CoS$_2$ (Fleischer, 1955) with Co, and to a lesser degree Ni, incorporated into pyrite in preference over Fe (Springer et al., 1964). Although limited experimental work by Bezmen and Tikhomirova (1975) indicates Co:Ni ratio values of pyrite may decrease with increasing temperature of formation, further work by Arutyunyan et al. (1980) indicates that the Co:Ni ratio values of pyrite will generally increase with increasing temperature. Empirical evidence indicates that the Co and Ni content of pyrite, along with the Co:Ni ratio value, will decrease in the transition from Cu-rich to Pb-Zn-rich ores (Walshe and Solomon, 1981; Hannington et al., 1999a; Serranti et al., 2002), thereby demonstrating a relationship to changes in physicochemical factors such as temperature. It has been argued that metamorphism is capable of causing increases (Fleischer, 1955) or decreases (Itoh, 1976) in the Co content of pyrite, with the distribution of Co becoming more homogenous with increasing metamorphic grade (Itoh, 1976). As there is little evidence of metamorphic recrystallisation within the Rosebery pyrites it is likely that the Co, Ni and Co:Ni ratio values present at the time of ore formation have been preserved.

Trace element analysis of pyrites from various ores at Mt Lyell reveals significant differences in the Co content and Co:Ni ratio value between Cu-rich and Zn-Pb-rich ores (Loftus-Hills and Solomon, 1967; Loftus-Hills, 1968; Raymond, 1996). These differences have been attributed by Raymond (1996) to the variation in Co content of the hydrothermal fluid as the result of changes in temperature or salinity. A recent genetic model has presented the Mt Lyell ore bodies as a hybrid magmatic-seawater system with disseminated Cu-rich ore within a large feeder zone overlain by exhalative Zn-Pb ore (Corbett, 2001), and in this context there is a systematic decrease in Co content and Co:Ni ratio values upward through the ore-forming system (fig.7.9). A similar trend is evident in the Co:Ni ratio values within the Rosebery system (fig.7.9). Green et al. (1981) and Green (1983) analysed a series of pyrite concentrates (n=44) from the footwall immediately beneath the southern ore lenses at Rosebery by wet chemical techniques. The pyrites returned relatively high Co (ave. 270ppm), Ni (ave.43.4ppm) and Co: Ni ratio values (ave.6.22/max.251). Samples of ore from the southern ore lenses at Rosebery examined as part of a study by Loftus-Hills and Solomon (1967) and Loftus-Hills (1968) (n=16) returned relatively low Co (ave. 6.2), Ni (ave. 10.9) and Co:Ni ratio values (ave. 0.56). Pyrites in the footwall to K lens (group A) examined as part of this study (fig.7.10) display Co:Ni ratio
Figure 7.9 Co vs Ni diagram for pyrite from Rosebery and Mt Lyell

Figure 7.10 LA-ICP-MS analysis scatter plots - Co and Ni in pyrite.
(A) Pyrite in the footwall to K lens exhibits elevated Co:Ni values relative to the overlying ore. (B) The Co:Ni value displays a strong negative relationship with Ti content in Rosebery pyrite.
values (ave. 15.5) that are higher than the overlying sphalerite-galena-rich ores (ave. 1.63). A further trend is evident within the ore at a micro-scale as the more crystalline group C pyrites typically display higher Co:Ni ratio values (ave. 3.03) than group B pyrites with relict primitive morphologies (ave. 0.51). This suggests that pyrite in the sphalerite-galena ore exhibiting primitive morphologies may have formed at lower temperatures than the more euhedral crystalline pyrites within the ore, and that pyrites in the footwall formed at yet more elevated temperatures.

The zone of elevated Co (fig.7.11) and Co:Ni ratio values (ave. 23.0) for footwall pyrite in the study by Green et al. (1981) and Green (1983) is coincident with a significantly increased chalcopyrite-pyrite content within the overlying ore (mine levels 15 and 17 in fig.3.21), and was therefore interpreted to mark the position of a possible hydrothermal conduit. The up-dip part of this zone underlies the southern ore lenses at a point where both E lens and G lens coalesce with F lens (figures 5.3 & 5.4). This position is also coincident with a rapid change in TSV thickness most likely to have been caused by syn-depositional faulting (fig.5.1). The area beneath the southern ore lenses outside the 300ppm Co envelope returned an average Co:Ni ratio value of ~2.2, probably indicating a lower temperature of formation in comparison with pyrite within the inferred hydrothermal conduit.

![Diagram](image)

**Figure 7.11 Southern ore lenses - ≥300ppm Co in footwall pyrites.**

Diagram shows the zone of elevated Co values within footwall pyrite samples as analysed by Green et al. (1981) and Green (1983). Co:Ni values are also elevated within this zone (ave. 22.0). The zone corresponds to a significant increase in Cu-Fe mineralisation within the overlying Zn-Pb ore (refer mine levels 15 and 17 in fig. 3.21)

Analysis of K lens footwall pyrites as part of the current study also returned significantly lower Co (ave. 58.4ppm) and Co:Ni ratio values (ave. 15.5). This suggests that pyrites in the footwall beneath the northern ore lenses formed at lower temperatures than pyrites within the inferred
hydrothermal conduit beneath the southern ore lenses. This is further supported by the observation that the southern ore lenses (D, E, F, G) overlying the area of elevated Co and Co:Ni ratio values in the footwall pyrites are significantly more enriched in Cu (ave. 0.7 to 1.3wt%) and Fe (10.5 to 23.3wt%) in comparison with the northern ore lenses (AB, K, P) which have much lower Cu (ave. 0.1-0.4wt%) and Fe (6.5-11.7wt%) contents (Berry et al., 1998).

The Co and Ni content and the Co:Ni ratio values have been utilised empirically in several studies to distinguish between pyrites of differing genetic origin (e.g. Fleischer, 1955; Loftus-Hills and Solomon, 1967; Loftus-Hills, 1968; Bralia et al., 1979; Xuexin, 1984; Bajwah et al., 1987; Brill, 1989). Pyrites of sedimentary or diagenetic origin are typically found to display low Co values and consistent Co:Ni ratio values of less than 1 (fig.7.12), whereas pyrites of submarine hydrothermal origin commonly display elevated Co values and Co:Ni ratio values that are more variable and are typically greater than 1 (Hegemann, 1943; Fleischer, 1955; Bralia et al., 1979).

Loftus-Hills and Solomon (1967) and Loftus-Hills (1968) analysed a series of pyrite concentrates as part of a regional study in western Tasmania aimed at differentiating between pyrite of sedimentary and diagenetic origin within shales and pyrites of hydrothermal origin within volcanic rocks and associated sulphide ores. Sedimentary and diagenetic pyrites in shales were found to be characterised by the combination of a high Ni content and an average Co:Ni ratio of less than 1, Cu-rich Cambrian ores were characterised by a high Co content and average Co:Ni ratios greater than 1, and Pb-Zn-rich Cambrian ores were characterised by low Co and Ni contents and average Co:Ni ratio values of less than 1.

Pyrite within stratigraphically deeper Cu-rich ores from Mt Lyell and within Rosebery footwall mineralisation exhibits average Co:Ni ratio values >1, a characteristic typical of submarine hydrothermal deposits (fig.7.12). However, the Zn-Pb-rich ores from Mt Lyell and Rosebery exhibit low Co and Ni contents and Co:Ni ratio values below 1, a characteristic more typical of sedimentary or diagenetic pyrites (fig.7.12). The majority of submarine hydrothermal deposits plotted in figure 7.12 are located within geological sequences that contain a significant component of mafic rocks, whereas felsic rocks dominate the Tasmanian sequence. It is suggested therefore that the availability of Co and Ni within the stratigraphic sequence controlled the degree to which hydrothermal fluids were enriched in these elements. This may explain the overall lower levels of Co and Ni in the Cambrian Zn-Pb-rich ores in western Tasmania but it does not explain why the Co:Ni ratio values are less than 1. Diagenetic pyrite, in the form of framboïds and fine-grained overgrowths, is a relatively minor component of the Zn-Pb-rich ore at Rosebery and probably contributes little to the low Co:Ni ratio values. Low temperature hydrothermal pyrite can in some cases take on a sedimentary signature if formed within organic-rich sediments (e.g. Zhang et al., 2002). However, although carbonaceous shales are present within the stratigraphic sequence, the Cambrian ores are primarily hosted...
within volcanic-derived rocks lacking evidence of a significant organic carbon component. The switch to a predominance of Ni within pyrite relative to Co could have been the result of preferential partitioning of Co into sulphide species other than pyrite within the Zn-Pb ore, however this is not supported by analytical data from this study. It is more likely that the preferential incorporation of Co into higher temperature pyrites within the footwall mineralisation resulted in the depletion of Co within the hydrothermal fluids relative to Ni. The relative concentration of Ni and Co in the hydrothermal fluids then became the dominant factor controlling how much Co and Ni substituted for Fe in pyrite within the Zn-Pb rich ores that
Chapter 7

In contrast to the Cambrian syngenetic ores in western Tasmania, the Devonian-Carboniferous granite-related ores (fig.7.13) contain pyrite with similar average Co:Ni ratio values to the sedimentary pyrite (Loftus-Hills and Solomon, 1967; Loftus-Hills, 1968). Pyrite in late-syn- to post-deformation quartz-carbonate-sulphide veins within the Rosebery mine, and in fault-related mineralisation of similar timing from the Chamberlain and Salisbury prospects, both returned elevated Co and Ni values with average Co:Ni ratio values greater than 1 (Loftus-Hills and Solomon, 1967; Loftus-Hills, 1968; and this study), which places them within the syngenetic hydrothermal field. In contrast, pyrite within granite-related Pb-Zn ore from the Murchison, South Comet, Zeehan Queen and Zeehan Montana mines and the Heemskirk granite exhibited low Co and Ni values and average Co:Ni ratio values below 1 (Loftus-Hills and Solomon, 1967; Loftus-Hills, 1968; and this study). This would suggest that the Co and Ni content of pyrite might be utilised not only to distinguish between Cambrian syngenetic and Devonian granite-related ores in western Tasmania, but also between vein-hosted sulphides that were derived directly from granite-related fluids or those that were remobilised from syngenetic ores during deformation.

![Figure 7.13 Average Co vs Ni values in pyrite of Devonian origin - Tasmania.](image)

**Figure 7.13 Average Co vs Ni values in pyrite of Devonian origin - Tasmania.**

The syn-deformational quartz-carbonate-sulphide veins at Rosebery plot above a Co:Ni ratio value of 1, as do samples from the fault-related Chamberlain and Salisbury prospects. In contrast granite-related Zn-Pb ores exhibit much lower Co values and plot below Co:Ni ratio value of 1. Data from Loftus-Hills & Solomon (1967), Kitto (1996) and this study.
7.5.8. Bi:Pb in pyrite and sphalerite

The presence of Ag, Bi and Sb in galena can be explained by the coupled substitution of Ag with Bi or Sb in the form $\text{Ag}^{+} + \text{(Sb}^{3+},\text{Bi}^{3+}) \rightleftharpoons 2\text{Pb}^{2+}$ (Foord and Shawe, 1989). A solid solution exists at high temperatures between galena (PbS) and matildite (AgBiS$_2$), but at temperatures below 200°C the solid solution becomes unstable and decomposes (Craig and Vaughan, 1990). Similarly, substitution of significant amounts of Sb into galena will only occur at high temperatures (350-400°C), with a tendency for Sb and Ag to form sulphosalts at lower temperatures (Foord and Shawe, 1989).

The levels of Bi, Sb and Pb within Rosebery pyrites extend well above maximum levels that might be expected from simple elemental substitution within the pyrite matrix (Craig and Vaughan, 1990). The molar levels of Ag, Sb and Bi in comparison with Pb vary significantly within Rosebery pyrites (table 7.10). Although there may be some substitution of these elements for Pb in galena inclusions within pyrite, the molar ratios of these elements with respect to Pb are elevated to such a degree that they can only be explained by the presence of Ag, Sb and Bi-bearing sulphosalts.

The levels of Ag, Bi, Pb and Sb in sphalerites returned from the LA-ICP-MS analyses are below maximum levels that might be attributed to elemental substitution into the sphalerite matrix (Craig and Vaughan, 1990), however the strong correlation between these elements suggests that they too are associated with galena and sulphosalts inclusions within the sphalerite matrix. This interpretation is supported by the coincident trend toward elevated Bi in both pyrites and sphalerites of group A, and the trend toward elevated Sb and Ag in both pyrite and sphalerite of group B and group C samples. Extremely high levels of Ag in some cases correspond to elevated Au values and can therefore be attributed to electrum inclusions within the sphalerite.

Within the Shakanai mine in the Kuroko district of Japan, Bi is concentrated in the high temperature siliceous and yellow ores (Nishiyama, 1974). Bismuth minerals are also described from the stringer zones and interaction zone at the base of massive sulphide lenses in the Iberian Pyrite Belt but are absent in the overlying massive sulphide ores (Marcoux et al., 1996). The presence of Bi-sulphosalts in the group A pyrites and sphalerites therefore suggests higher temperatures of formation in comparison with group B and C pyrites. This is further supported by the strong positive correlation between Bi:Pb and Co:Ni ratio values, and between Bi:Pb ratio values and the Fe-content sphalerites within the same samples (fig.7.14).
Figure 7.14 LA-ICP-MS analysis scatter plots - Bi and Pb in pyrite.
(A) Pyrite in the footwall to K lens exhibits elevated Co:Ni values relative to the overlying ore. (B) to (D) The Bi:Pb values display a strong negative relationship with elements such as Tl, Au and As within pyrite.

Table 7.10 Elemental ratios of Bi, Sb, Ag vs Pb - pyrite and sphalerite.
Refer to text for characteristics of groups A to C.
7.5.9. TI in pyrite

As a trivalent ion TI can substitute into sulphide minerals such as pyrite, galena and sphalerite, and as a monovalent ion it can substitute for alkali metals in minerals such as micas and potassic feldspars (Smith and Carson, 1977; McGoldrick et al., 1979; Sobott et al., 1987).

In previous studies, TI within the Rosebery alteration halo was found to exhibit a very strong correlation with K and Rb in altered rocks that contained significant amounts of potassic micas (Smith, 1975; Smith and Huston, 1992), a feature attributed to monovalent TI substitution for K that occurred during hydrothermal alteration. TI has also been found to occur in Rosebery ore concentrates, with the highest levels present within the pyrite concentrate (79ppm) and within the galena concentrate (30ppm) (Smith, 1975; Smith and Huston, 1992). The current study only returned significant TI values from pyrite, with the highest values evident in pyrite that exhibited relict primitive textures (tables 7.1 & 7.3, fig.7.5). Although a few elevated TI values occur within sphalerite, tennantite-tetrahedrite, arsenopyrite and chalcopyrite, most LA-ICP-MS analyses returned values below detection limits. This indicates that TI measured in most of the sulphide concentrates during the earlier study is likely to have been due to the presence of TI-enriched pyrite.

TI enrichment to levels of thousands of ppm is often associated with colloidal pyrite and marcasite in low temperature hydrothermal deposits (Fleischer, 1955; Smith and Carson, 1977; Sobott et al., 1987), with marcasite often significantly more enriched than pyrite (Fleischer, 1955). Examples of the relationship with colloform pyrite have been recorded within Carlin-type deposits in a strong association with elevated Au, Hg, Sb and As (Sobott et al., 1987; Guozhu, 1996; Zhong and Baogui, 1996), within the Pb-Zn rich parts of several Kuroko deposits (Murao and Itoh, 1992) and in the upper part of the volcanic-hosted massive sulphide ore at Mt Chalmers (Huston et al., 1995b). These occurrences are considered analogous to the presence of elevated TI in pyrites that exhibit relict primitive textures in the Rosebery Zn-Pb ore. TI precipitation as a sulphide component is thought to be the result of cooling and oxidation of hydrothermal fluids (Ewers and Keays, 1977; McGoldrick et al., 1979; Sobott et al., 1987). It has also been observed that precipitation of TI within the Broadlands hydrothermal system is strongly temperature dependent, with appreciable quantities not depositing from mineralising solutions until they have ascended to within 200m of the earth’s surface (Ewers and Keays, 1977). TI has also been found to be associated with galena, pyrite and cinnabar mineralisation forming during low temperature (~85°C) hydrothermal venting in ~10m of water depth (Prol-Ledesma et al., 2002). The enrichment of TI in pyrites at Rosebery in association with As-Sb ± Au-Hg, is interpreted to be indicative of relatively low-temperatures, supported by the observation that pyrites containing elevated TI values exhibit primitive morphologies, low Co:Ni ratio values (fig.7.10) and low Bi:Pb ratio values (fig.7.14).
A limited number of analyses (n=8) were conducted on pyrite within two syn-deformational quartz-carbonate-sulphide veins from Rosebery. Although most analyses returned values below detection limits, a single analysis did return a value of 311ppm Tl. This indicates that Tl was mobile during deformation, and that deformation may have played a role in expanding the extent of the Tl enrichment halo surrounding the Rosebery system as described in previous studies (Smith, 1975; Smith and Huston, 1992; Large et al., 2001a). It is possible that some remobilised Tl may also have substituted for K within sericite during metamorphic recrystallisation of the micas.

7.5.10. Au in pyrite

In a previous study by Huston et al. (1992), pixeprobe analysis of 26 pyrite grains for Au returned only two values above the detection limit of 20ppm (max. 50ppm). The LA-ICP-MS analyses in this study (table 7.1) are undoubtedly influenced by the presence of native Au and electrum inclusions, and even when a “top-cut” of 10ppm is applied to the dataset the average Au content of K lens pyrites still varies from 0.47ppm in group A and 0.38ppm group C, to 1.25ppm in group B. These values are higher than the 0.2ppm calculated for AB lens pyrites by Huston et al. (1992), and are closer to the value of 2ppm Au in pyrite from zinc residues reported by Henley and Steveson (1978).

Au in this study displays variable associations with Ag, As, Sb and Tl within pyrites of group B and C (fig.7.5e-l). The correlation between Au and Ag is reasonably strong in comparison with the other elements, probably due to the relatively common occurrence of electrum grains within pyrite. Au values appear to become significantly elevated above ~100ppm As and ~10ppm Sb, probably in part be due to the substitution of As and Sb for Fe within pyrite. This may have caused distortion of the pyrite lattice and permitted the non-stoichiometric substitution of Au within pyrite.

The pyrites of group B are considered to be analogous with much of the pyrite contained within sulphide ore of the nearby Hellyer VHMS deposit. The Hellyer pyrite typically exhibits primitive textures with little metamorphic recrystallisation (McArthur, 1996) and, due to its presence in pyrite as very fine inclusions, only 14-21% of the Au was amenable to metallurgical recovery (Bottrill et al., 1992).

The presence of Au and electrum along fractures within Rosebery pyrite grains and in syn-tectonic veins supports the argument for remobilisation of Au during deformation. LA-ICP-MS analysis of pyrite (n=8) within syn-deformation quartz-carbonate veins returned a single Au value of 11.5ppm, further indicating mobilisation of Au during deformation. However, as inclusions are also observed at the margins of anhedral pyrite cores within larger pyrite grains and there is little evidence for significant metamorphic recrystallisation of the pyrites, this
would suggest that at least some of the larger Au and electrum grains are relict primary features.

Huston and Large (1989) and Huston (2000) proposed that much of the Au within Zn-Pb ore was transported in hydrothermal solution as thio-complexes and that precipitation was controlled by increased pH due to seawater mixing, or to a larger degree by a reduction in the activity of reduced sulphur brought on by the precipitation of pyrite as illustrated by the following equation:

\[ 4\text{Au(HS)}_2^- + 2\text{H}_2\text{O} + 4\text{H}^+ \rightleftharpoons 4\text{Au}^0 + 8\text{H}_2\text{S}_{(aq)} + \text{O}_2(g) \]

Precipitation of pyrite with primitive morphologies would have provided more surface area in comparison with coarser pyrite euhedra and would therefore have provided more sites for the nucleation and coprecipitation of Au and electrum. If other elements, such as Tl, Hg, As, Sb and Pb, are transported in a similar fashion as aqueous sulphur complexes (Hannington et al., 1999b) then precipitation of pyrite would also facilitate co-precipitation of these elements in association with Au. This may explain the strong association of these elements with pyrite.

Au and electrum grains were not specifically analysed as part of this study. Electrum grains within and adjacent to a variety of sulphide species exhibit a range of fineness (1000Au/(Au+Ag) from 320-970 fine (Smith, 1975; Huston and Large, 1988; Huston et al., 1992; Vallerine, 2000). Smaller grains tend to be more Ag-rich whereas larger grains are more Au-rich, and occasionally have Ag-rich rims (Huston et al., 1992) which is a feature also noted in Kuroko ores (Shimazaki, 1974). Electrum analysed by Vallerine (2000) contained up to 15.1wt % Hg (ave. 1.42wt%), with grains of lowest fineness containing the highest Hg content.

### 7.5.11. Ag in galena, chalcopyrite and sphalerite

Huston et al. (1996) examined the Ag content of the principal ore minerals within the Rosebery sulphide ore and associated barite mineralisation in the vicinity of AB lens, and the values they reported for the Ag content of chalcopyrite (ave. 220ppm), galena (ave. 0.22wt%) and sphalerite (<11ppm) within sphalerite-galena ore are consistent with the results derived during this study. The results of the study by Huston et al. (1996) indicated that the coupled substitution of Ag with (Sb,Bi) had occurred during the formation of galena at Rosebery.

Results reported by Huston et al. (1996) from AB lens indicate that Ag content of galena decreases from chalcopyrite-pyrite stringer ore (ave. 0.96wt%) to the sphalerite-galena ore (0.22wt%) and the overlying barite mineralisation (ave. 0.06wt%). The Ag content of chalcopyrite in the same study returned elevated values in the chalcopyrite-pyrite ore (ave. 170ppm) and overlying sphalerite-galena ore (ave. 200ppm) in comparison with the barite ore (ave. 12wt%). In a study of K lens sulphides Vallerine (2000) also noted that the average Ag content of galena in Cu-rich ore was higher (0.75wt%) than in the massive sphalerite-galena
ore (0.056wt%) or the barite-rich mineralisation (0.04wt%). A similar trend toward a higher Ag content within galena proximal to the feeder zone at Hellyer was reported by McArthur (1996). Huston et al. (1996) related the compositional trends in chalcopyrite and galena within different ore types to temperature, redox conditions, $ΣBi:ΣSb$ values in the hydrothermal fluid and the fractional crystallisation of tetrahedrite. Chalcopyrite was interpreted as the preferred host for Ag in sulphide ores formed under reduced and higher temperature conditions. The coupled substitution of Ag and (Bi,Sb) into galena was proposed to occur under cooler reduced conditions in conjunction with elevated $ΣBi:ΣSb$ values in the hydrothermal fluids. The partitioning of Ag into tetrahedrite was proposed to occur under cooler and more oxidised conditions where the fluids displayed lower $ΣBi:ΣSb$ values. Elevated $ΣBi:ΣSb$ values in hydrothermal fluids associated with chalcopyrite-rich ores favoured the coupled substitution of Ag with Bi, whereas with lower $ΣBi:ΣSb$ values in sphalerite-galena ore coupled substitution with Sb was favoured. More Ag can be accommodated within galena in association with Bi than with Sb, and therefore the Ag content of galena decreases markedly from the chalcopyrite-pyrite stringer ore to the sphalerite-galena ore (Huston et al., 1996). The difference in Ag content of chalcopyrite between the chalcopyrite-pyrite stringer and barite ores was attributed to a change from reduced to oxidised conditions (Huston et al., 1996). The similarity in Ag content of chalcopyrite within chalcopyrite-pyrite and sphalerite-galena ores may indicate that controlling factors other than redox conditions were predominant.

Galena within samples containing crystalline group C pyrites is on average more Ag-enriched than galena within samples containing group B pyrites with primitive morphologies (table 7.7), indicating that group C samples may have been formed under more elevated temperatures and reduced conditions in comparison with the group B samples.

7.5.12. Cd, Mn and Fe in sphalerite

Elemental substitution of Cd, Mn and Fe for Zn in sphalerite is a feature of a range of ore deposit styles (Craig and Vaughan, 1990). The Cd content of sphalerite is dependant on a range of physicochemical factors, including the temperature of formation, pH and Cd content of the hydrothermal fluid, the concentration and nature of complexing ligands, and the concentration of reduced sulphur (Mookherjee, 1962; Kase and Horiuchi, 1996; Schwartz, 2000). In the absence of complexing ions, partitioning of Cd into sphalerite increases with decreasing temperature, however this trend can be reversed by changes in pH, the presence of complexing ligands and high concentrations of reduced sulphur (Schwartz, 2000). Through a series of sulphide deposits in Japan the Cd:Zn value in sphalerite has apparently been preserved during contact metamorphism despite a related increase in the Fe, Mn and Co content (Kase and Horiuchi, 1996). It is probable that the original Cd:Zn value within Rosebery sphalerites has been preserved despite later recrystallisation and annealing.
Several studies have utilised the Cd content of sphalerites to characterise different ore deposit types. Groves and Loftus-Hills (1968) analysed sphalerite from Rosebery ore as part of a regional study aimed at characterising sulphide deposits of Cambrian and Devonian age on the west coast of Tasmania. They were able to differentiate between deposit styles on the basis that sphalerites from Devonian granite-related deposits are significantly more enriched in Cd and Fe than those from Cambrian ores (fig. 7.15). In comparison with an extensive range of deposit data compiled by Schwartz (2000), the Cd values exhibited by Rosebery sphalerites (ave. 1781 ppm) are most similar to syngenetic volcanic hosted deposits (ave. 2360), whereas the values exhibited by samples of Devonian granite-related mineralisation examined during this study (ave. 3630 ppm) are more typical of epigenetic vein-hosted deposits (ave. 4100 ppm). A similar pattern has been described for a series of Japanese massive sulphide deposits, with elevated Cd values associated with granite-related vein-hosted Sn-W deposits and lower Cd values associated with Kuroko ores (Tsukimura et al., 1987).

Within sphalerite from Rosebery, the Fe and Mn have substituted for both Cd and Zn in a proportional fashion that appears to have preserved the original Cd:Zn value. This would indicate that this ratio is more suitable for the empirical characterisation of ore systems than the absolute Cd content. Sphalerites from the Devonian granite-related sulphide ores display

![Figure 7.15 Fe vs Cd in sphalerite from various Tasmanian deposits.](image)

Range of Fe and Cd values displayed within sphalerite from deposits of different genetic origin. Microprobe point data from the current study is plotted for comparison. Modified after Groves & Loftus-Hills (1968).
significantly higher Cd:Zn ratio values in comparison with sphalerites from various Rosebery ore lenses (table 7.11). There appears to be no significant change in Cd:Zn ratio values for Rosebery sphalerites due to metasomatic alteration despite related changes in the Fe and Mn content. Sphalerites within syn-tectonic quartz-carbonate-sulphide veins from K lens and fault-related mineralisation from the Chamberlain prospect also returned values similar to primary ores, indicating that in both cases that the sphalerite may have been remobilised from Cambrian ores rather than the direct product of Devonian granite-related mineralisation.

The Fe content of sphalerite is principally controlled by pressure, the activity of sulphur and temperature, although the effect of pressure is minor in the presence of pyrite (Ohmoto et al., 1983; Scott, 1983; Tsukimura et al., 1987). The Fe content of sphalerite can also be related to host sequence compositions and the relative availability of Fe (Tsukimura et al., 1987). Green et al. (1981) and Green (1983) analysed sphalerite from various ore lenses at Rosebery, noting that the Fe content of sphalerite decreased in the transition from Cu-rich to barite-rich ore, and that the average Fe content of sphalerite decreased stratigraphically upward through D lens. The same upward stratigraphic decrease in the Fe-content of sphalerite has also been described for Kuroko (Urabe, 1974) and Kidd Creek ores (Hannington et al., 1999a). Although the Fe-content of sphalerite is variable within massive K lens ore the footwall contains only dark-coloured Fe-rich sphalerite, whereas sphalerite in massive sphalerite-galena ore is predominantly light tan to honey-brown in colour. Within K lens the Fe content of sphalerite is higher in the presence of more crystalline pyrite morphologies (fig.7.16) and exhibits a positive correlation with the Co:Ni and Bi:Pb ratio values in pyrites within the same samples. This indicates that temperature was likely to have been a significant control on the Fe content of sphalerite within the primary sulphide ore.

<table>
<thead>
<tr>
<th>Rosebery - primary ore</th>
<th>Cd:Zn*</th>
<th>Mn:Zn*</th>
<th>Fe:Zn*</th>
</tr>
</thead>
<tbody>
<tr>
<td>K lens</td>
<td>1.6x10^-3</td>
<td>2.6x10^-3</td>
<td>7.0x10^-2</td>
</tr>
<tr>
<td>AB, D, F, G, K lenses</td>
<td>1.5x10^-3</td>
<td>3.6x10^-3</td>
<td>5.2x10^-2</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Deformation-related veins &amp; faults</th>
<th>Cd:Zn*</th>
<th>Mn:Zn*</th>
<th>Fe:Zn*</th>
</tr>
</thead>
<tbody>
<tr>
<td>quartz-carbonate-sulphide veins</td>
<td>1.6x10^-3</td>
<td>1.6x10^-3</td>
<td>5.6x10^-2</td>
</tr>
<tr>
<td>Chamberlain prospect</td>
<td>1.9x10^-3</td>
<td>5.1x10^-3</td>
<td>7.2x10^-2</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Rosebery – metasomatised ore</th>
<th>Cd:Zn*</th>
<th>Mn:Zn*</th>
<th>Fe:Zn*</th>
</tr>
</thead>
<tbody>
<tr>
<td>K lens</td>
<td>2.2x10^-3</td>
<td>14.2x10^-3</td>
<td>23.1x10^-2</td>
</tr>
<tr>
<td>J lens</td>
<td>1.1x10^-3</td>
<td>6.6x10^-3</td>
<td>11.4x10^-2</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Devonian granite-related mineralisation</th>
<th>Cd:Zn*</th>
<th>Mn:Zn*</th>
<th>Fe:Zn*</th>
</tr>
</thead>
<tbody>
<tr>
<td>Murchison &amp; South Comet mines</td>
<td>4.1x10^-3</td>
<td>0.8x10^-3</td>
<td>13.7x10^-8</td>
</tr>
<tr>
<td>Zeehan mineral field</td>
<td>3.0x10^-3</td>
<td>6.0x10^-3</td>
<td>10.7x10^-2</td>
</tr>
</tbody>
</table>

* values expressed as average molar ratios

Table 7.11 Elemental ratios of Cd, Mn and Fe vs Zn in sphalerite - Rosebery and Devonian ores
Figure 7.16 Fe content of sphalerite vs elemental ratios and contents of pyrite. Scatter plots include all data points from the sample groups: A - footwall, B - ore contains pyrite with relict primitive morphologies, and C - ore contains pyrite without primitive morphologies. Fe content of sphalerite as determined by electron microprobe vs elemental ratios and contents of pyrite from the same sample as determined by LA-ICP-MS.

(A) vs Co:Ni in pyrite. (B) vs Bi:Pb in pyrite. (C) vs Tl in pyrite. (D) vs Au in pyrite.

Figure 7.17 Fe content of sphalerite vs Fe content of tennantite-tetrahedrite series.
Rosebery sphalerites are relatively homogenous in terms of major element compositions at the grain scale and over distances of a few centimetres. Similar sphalerite homogeneity has been described at the grain and hand specimen scale within metamorphosed Besshi-type deposits in Japan, contrasting with sphalerites of typically heterogenous composition found within skarn, vein and undeformed volcanic-hosted mineralisation (Mizuta, 1988). In a study of metamorphosed Ba-Zn-Pb ores from the Foss deposit, Moles (1983) attributed variation of Fe and Mn content within sphalerite to intra-crystalline diffusion, buffering of sphalerite compositions with pyrite and pyrrhotite, partitioning of elements between sphalerite and surrounding mineral phases such as carbonate, and the exsolution of Fe-rich components within the sphalerite grains. Within Rosebery ore the crystallographic continuity and lack of compositional variation within metamorphosed sphalerite grains, the morphology of grain boundaries consistent with annealing, and the paucity of inclusions within sphalerite grains indicates that recrystallisation of sphalerite and intra- to inter-crystalline diffusion were the main mechanisms of compositional homogenisation during metamorphism. The apparent equilibrium between the Fe-content of sphalerite and tetrahedrite-tennantite within the same samples supports this idea (fig.7.17). The lack of evidence indicating the diffusion of Fe into sphalerite from adjacent embayed pyrite grains suggests that, if any Fe exchange had occurred, it did so prior to sphalerite recrystallisation. Although changes in sphalerite composition have been attributed to metamorphism (e.g. Moles, 1983; Kase and Horiuchi, 1996) the changes occur at a micro-scale and metamorphosed deposits often preserve compositional zonation at the ore body scale (e.g. Moles, 1983; Hannington et al., 1999a). Although homogenisation of the Rosebery sphalerites appears to have occurred at the centimetre scale during metamorphism, compositional variation at the ore body scale appears to have been preserved.

The most significant changes in sphalerite composition within Rosebery ore can be attributed to granite-related metasomatism, and include an increased Fe-content (Solomon et al., 1987; Khin Zaw and Large, 1996; and this study), an increased Mn-content (table 7.11), and the development of a gradational compositional zonation (Khin Zaw and Large, 1996). These features are comparable to textures in Japanese plutonic-related skarn and vein-style mineralisation described by (Mizuta, 1988). Fe-rich alteration assemblages at Rosebery indicate that metasomatic fluids introduced Fe during granite-related metasomatism, whereas late rhodochrosite veining supports the argument that Mn was most likely remobilised from pre-existing Mn-rich carbonates surrounding the primary ore. A sample of metasomatised Zn-Pb ore from K lens returned elevated Co values (ave. 29.2ppm) in comparison with sphalerite within primary Zn-Pb ore (ave. bdl & max. 4.01ppm), suggesting that metasomatism may also have increased the Co content of sphalerite either by introduction with the metasomatic fluids or by remobilisation from pyrite. In contrast, sphalerites from Devonian granite-related mineralisation are enriched in Fe but not in Mn. Sphalerites from deformation-related veins have Fe and Mn contents similar to those found in primary Rosebery sphalerites (table 7.11).
7.5.13. Trends in the tennantite-tetrahedrite series

Data compiled from this and earlier studies show that the full range of tennantite-tetrahedrite compositions is present within Rosebery ore (fig.7.18). This is a feature also noted in the Kuroko ores where tennantite-tetrahedrite with a broad compositional range formed under apparently similar physicochemical conditions (Shimazaki, 1974). Work by Smith and Huston (1992) indicates that remobilisation into veins and fractures has resulted in a trend toward the Cu-As-rich tennantite end-member. Pyrargyrite has also been observed in late-syn to post-deformation veins.

The presence of arsenopyrite and chalcopyrite in tetrahedrite-tennantite (fig.5.17i-j) may be evidence for the diffusion of Ag and Sb from surrounding galena into the tetrahedrite-tennantite matrix during diagenesis or metamorphism. Similarly, a positive correlation between the Fe content of tennantite-tetrahedrite and sphalerite in the same samples (fig.7.17) indicates that the tennantite-tetrahedrite has probably undergone post-depositional equilibration with surrounding sulphides. The observation that tennantite and tetrahedrite appear to be restricted to ores with relatively low-Fe sphalerite (<5.2wt% Fe) (fig.7.19) could, however, indicate a common physicochemical control on the Fe content of sphalerite and the stability of tennantite-tetrahedrite that operated during primary ore formation. The absence of tennantite-tetrahedrite...
in the samples with Fe-enriched sphalerites at Rosebery, particularly in the chalcopyrite-pyrite ore, is considered analogous to the Shakanaï ore deposit where tennantite-tetrahedrite is only found in black ore and not in the underlying siliceous or yellow ores (Nishiyama, 1974). Tennantite-tetrahedrite exhibits a marked low temperature solubility (Spycher and Reed, 1989; Hannington et al., 1999a) and is absent within Rosebery mineralisation where sphalerite contains elevated Fe-contents (>5.2wt% Fe). The elevated Fe-content of sphalerite at Rosebery appears to be the result of elevated temperatures of formation, and these conditions most likely prevented tennantite-tetrahedrite from precipitating.

Although minerals of the tennantite-tetrahedrite series can exhibit systematic compositional variation with changes in paragenetic timing and distance from fluid source within some hydrothermal systems (e.g. Hackbarth and Petersen, 1984; Sack and Loucks, 1985; Serranti et al., 2002), there is insufficient data to establish similar trends in the Rosebery system.

7.6. Summary

The combined use of the electron microprobe and LA-ICP-MS techniques has permitted a comprehensive study of trace element characteristics of sulphide within the Rosebery ore. This was not previously achievable, due to the relatively low levels of trace elements within Rosebery sulphides.

Rosebery sulphides typically contain trace element levels below detection limits achievable by conventional electron microprobe techniques. The LA-ICP-MS technique utilised in this study has permitted detection limits up to several orders of magnitude lower than electron microprobe techniques, with a spatial resolution not achievable through bulk sampling chemical analysis. This has for the first time permitted a detailed compositional characterisation of Rosebery sulphide species.

Physicochemical controls at the time of ore formation and during subsequent deformation and metasomatism are responsible for the trace element components of the Rosebery ore. Samples of sphalerite-galena ore that formed at relatively low temperatures displays features that include:

- Relict primitive pyrite morphologies indicative of rapid precipitation.
- Elevated levels of Mn, Ni, As, Ag, Sb, Au, Tl, and Pb in pyrite.
- Low levels of Co and Bi in pyrite with commensurate low Co:Ni and Bi:Pb ratio values.
- Sphalerite containing low levels of Fe and Cu, and elevated levels of Zn, Ag, Sb and Pb.
- Galena containing relatively low Sb and Ag.

The early pyrite with relict primitive morphologies is overgrown and replaced by more euhedral
and crystalline pyrite. Samples lacking pyrite with primitive pyrite morphologies display features indicative of higher temperatures of formation including:

- Euhedral and more coarsely crystalline pyrite morphologies.
- Elevated levels of Bi and Sn in pyrite with commensurate high Bi:Pb ratio values.
- Moderate levels of Ni, Ag, Sb, Tl, Pb, and Co, and moderate Co:Ni ratio values in pyrite.
- Low levels of Mn, As and Au in pyrite.
- An increased Fe content in sphalerite.
- Galena with elevated levels of Sb and Ag.

Samples from the immediate footwall to the sphalerite-galena ore are typically more enriched in chalcopyrite-pyrite. These samples exhibit evidence of higher temperatures than samples from the sphalerite-galena ore including:

- Coarsely crystalline and often very euhedral pyrite morphologies.
- High levels of Co and Bi in pyrite with commensurate elevated Co:Ni and Bi:Pb ratio values.
- Sphalerite with high levels of Fe, Mn, Cu, Sb and Bi.
- Overall low levels of most other trace elements in pyrite and sphalerite.
- Relatively minor occurrences of tennantite-tetrahedrite and galena.

The association of elements such as Au, Ag, Tl, Hg, As, Sb and Pb with the low temperature pyrite may be in part the result of coprecipitation, particularly if these elements were transported within low temperature fluids as aqueous sulphur complexes. The pyrite morphologies and trace element content indicate that after initial formation of the sphalerite-galena ore the system heated up, resulting in a later phase of overprinting high temperature mineralisation. Elevated Co and Co:Ni ratio values within pyrite from the footwall to Rosebery sphalerite-galena ore indicate that significant parts the southern ore lenses were probably associated with higher temperatures of formation than the northern lenses. The zone of elevated Co and Co:Ni ratio values in footwall pyrite at the southern end of the mine coincides with the potential fluid upflow zone related to formation of the southern ore lenses.

The Rosebery ore can be empirically differentiated from Devonian granite-related ores on the basis that:

- Cd:Zn ratio values for Rosebery sphalerite are significantly lower (1.6x10^-3) than the values found within sphalerite from Devonian ores (3.0-4.1x10^-3).
- Average Co:Ni ratio values for Rosebery pyrite higher than the average values found within pyrite from Devonian ores.
- Pyrite in Devonian ores displays markedly higher Ni values.
Chapter 8 : Isotopes

8.1. Introduction

The aim of this part of the study was to examine and characterise isotopic signatures of the mineralisation and alteration associated with the Rosebery ore deposit. This has involved the compilation and review of carbon, oxygen, sulphur, strontium and lead isotopic data from previous studies, and the integration of new sulphur isotope data collected as part of this study from the northern part of the mine.

The isotope values are compared to the Standard Mean Ocean Water standard (SMOW) for δ¹⁸O analyses, the Peedee belemnite standard (PDB) for δ¹³C analyses, and Cañon Diablo Troilite standard (CDT) for δ³⁴S analyses.

8.2. Sulphur isotopes

8.2.1. Previous studies

Sulphur isotope studies have previously been undertaken by Stanton & Rafter (1966), Solomon et al. (1969), Green et al. (1981) and Green (1983), Solomon et al. (1988), Khin Zaw (1991), and Davidson and Kitto (1997). The bulk of analysis has been focussed on the sulphide and barite lodes, although the more recent study by Davidson and Kitto (1997) extended to rocks of the surrounding mine sequence.

Stanton and Rafter (1966) analysed 26 mixed sulphide ore samples from lenses A to F, which returned δ⁳⁴S values of 8.8 to 15.5‰.

Solomon et al. (1969) analysed mineral separates from the footwall schist, sulphide and barite lodes, the hanging-wall black slate, and syn-tectonic veins from within the hanging-wall succession. δ³⁴S values for sulphides within the sulphide lodes ranged between 9.0 and 13.8‰ (excluding galena), whereas barite returned values between 38.5 and 40.1‰. Sulphides from the barite lode exhibited values between 17.4 and 19.2‰, with values for the barite ranging between 34.6 and 41.2‰. Pyrite from the black slate exhibited a wide range of values from 2.3 to 25.3‰. Temperature estimates based on δ³⁴S values in coexisting sulphides yielded no consistent values, however a temperature of 170°C was calculated from coexisting barite-pyrite within a sample from H lens. It was concluded that the δ³⁴S values in barite were indicative that sulphur was derived from either seawater or connate waters.

Green et al. (1981) and Green (1983) analysed samples from three sections across the
sulphide and barite lodes at the southern end of the mine, including material from sulphide ore, footwall schist, barite lode and oxide-bearing mineralisation overlying both the sulphide and barite lodes. In line with an inferred exhalative sulphide depositional sequence the $\delta^{34}$S values within sulphide ore were interpreted to increase stratigraphically upward through an "unfolded" interpretation of the mine sequence, with slightly lower $\delta^{34}$S values evident in the chalcopyrite-pyrite ore (7.8 to 9.8‰) in comparison with sphalerite-galena ore (values to 17.2‰). The $\delta^{34}$S values returned for sulphides within the footwall schist were found to be similar to those of the massive ore immediately overlying. Sulphides associated with the barite lodes returned higher $\delta^{34}$S values (14.5 to 19.8‰) than those associated with the sulphide lodes (7.8 to 17.2‰). $\delta^{34}$S values for barite in the sulphide lodes (39.6 to 40.1‰) were comparable to values returned for the same mineral in the barite lodes (34.6 to 41.2‰). $\delta^{34}$S values from pyrite in pyrrhotite-replaced ore (16.1‰) exhibited similar values to pyrite in nearby sphalerite-galena ore, and it was interpreted that Devonian metasomatism had not significantly influenced the Cambrian $\delta^{34}$S signature. Sulphide samples from oxide-bearing assemblages overlying the barite and sulphide lodes displayed elevated $\delta^{34}$S values (11.5 to 18.9‰ above sulphide, 20.7 to 27.6‰ above barite). Temperature estimates based on $\delta^{34}$S data for barite-sulphide minerals pairs originally analysed by Solomon et al. (1969). The resultant calculated temperature range of 255° to 298°C was interpreted to reflect the temperature of barite lode formation. The source of sulphur was attributed to a significant component of partially reduced seawater sulphate combined with a magmatic sulphur input derived either directly from a magmatic fluid or through leaching of volcanic rocks.

In the context of a regional study related to the sulphur isotope signatures of mineralisation in western Tasmania, Solomon et al. (1988) also argued that the sulphur at Rosebery was derived from reduced seawater sulphate, with a magmatic sulphur component derived from either leaching of volcanic rock or from direct magmatic input. The proposed model is discussed in detail later in this chapter.

Khin Zaw (1991) analysed sulphide ore from J lens and concluded that Devonian metasomatic pyrrhotite-pyrite replacement of the sphalerite-galena ore had resulted in homogenisation of the primary $\delta^{34}$S values at the hand specimen scale without significantly changing the overall $\delta^{34}$S signature of the ore.

Davidson and Kitto (1997) examined the $\delta^{34}$S signature of pyrite at the northern end of the mine as part of a study aimed at identifying possible Cambrian faults within the Mount Read Volcanics. A total of 143 analyses were undertaken on pyrite taken from several drill holes in a traverse across AB lens and north along strike from the ore lenses. A zone of elevated $\delta^{34}$S values in pyrite within the footwall north of AB lens (~5 to 46‰, ave.23.5‰) was interpreted as a region of hydrothermal down-flow related to Cambrian hydrothermal fluid circulation during
formation of the primary sulphide ores.

8.2.2. Current study

8.2.2.1. Aim

The aim of the current study was to obtain sulphur isotope data in the vicinity of K lens and P lens and, in conjunction with the historical data, to examine the lateral variation in $\delta^{34}$S values at a mine scale. A series of samples from several P sub-lenses were analysed to test whether there is a change in isotopic values upward through the stratigraphy as observed within the southern ore lenses by Green et al. (1981) and Green (1983). Samples taken from K lens were in part aimed at determining isotopic differences between Cu-Fe- and Au-Ag-barite-enriched ores. Samples of various pre-deformational sulphide veins were examined with the aim of comparing their isotopic signatures with those of the main lodes.

8.2.2.2. Analytical methods

In this study pyrites within most samples were analysed by laser ablation microprobe extraction due to their fine grain size. Rock samples were prepared as doubly polished thin sections of ~100µ thickness, before removal from their glass mounts and cleaning with solvent. Individual pyrite grains were then ablated with a Quantronix TEM 117 Nd-YAG laser located within the Central Science Laboratory at the University of Tasmania. Analysis was undertaken utilising the method outlined by Huston et al. (1995a). Typically several grains within each sample were ablated to provide sufficient material for analysis.

Conventional methods were utilised for isotopic analysis of all the sphalerites and the remainder of the pyrites. The low Fe-content of sphalerite grains had resulted in significant transmission and dispersion of the laser beam, thereby preventing laser analysis. Instead ~12mg of sample was collected from polished thin section material subsequent to removal from the glass mount and cleaning. Samples were examined under magnification in both transmitted and reflected light to ensure minimal impurities. Samples were analysed in the Central Science Laboratory at the University of Tasmania using the method outlined by Robinson and Kusabe (1975).

Samples of barite were obtained with the use of a micro hand drill, with approximately ~60 mg of material collected for analysis. The samples were analysed at the Central Science Laboratory at the University of Tasmania utilising the method outlined by Coleman and Moore (1978).
8.2.2.3. Results

The analytical results from this study are summarised in table 8.1. Sulphides in the immediate footwall to K lens returned $\delta^{34}S$ values similar to the values within the overlying sulphide ore, an observation also made by Green et al. (1981), Green (1983), and Davidson and Kitto (1997) elsewhere in the mine. There appears to be no relationship between the $\delta^{34}S$ value and alteration mineralogy within the footwall samples. A small number of samples from the TSV and peperitic sill immediately overlying sulphide ore also returned $\delta^{34}S$ values comparable to the ore values. Fe-Zn-Cu-enriched K lens ore with dark brown Fe-rich sphalerite returned an average $\delta^{34}S$ value of 13.2‰ (n=11), whereas Au-Ag-Pb-barite-enriched K lens ore with pale honey-brown sphalerite returned a higher average value of 15.4‰ (n=4). There was no clear trend in $\delta^{34}S$ values vertically through the K lens ore.

<table>
<thead>
<tr>
<th>n= type range (‰)</th>
<th>ave. (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td>K lens - footwall</td>
<td>9 py, sp 11.7 to 17.0</td>
</tr>
<tr>
<td>K lens – sulphide ore</td>
<td>15 py, sp 9.7 to 17.6</td>
</tr>
<tr>
<td>K lens – overlying TSV / porphyry</td>
<td>3 py, sp 12.1 to 14.9</td>
</tr>
<tr>
<td>pyritic pre-deformation veins cross-cutting K lens ore</td>
<td>2 py 12.8 to 13.3</td>
</tr>
<tr>
<td>P lens – sulphide ore</td>
<td>27 py, sp 8.7 to 16.7</td>
</tr>
<tr>
<td>P lens – barite lodes</td>
<td>3 py 10.8 to 23.5</td>
</tr>
<tr>
<td>pyritic pre-deformation veins cross-cutting P lens ore</td>
<td>7 py, sp 11.7 to 17.4</td>
</tr>
<tr>
<td>pyritic pre-deformation veins in shale above P lens ore</td>
<td>2 py 16.6 to 17.7</td>
</tr>
<tr>
<td>K lens barite lodes/clots in ore</td>
<td>8 ba 33.8 to 47.3</td>
</tr>
<tr>
<td>P lens barite lodes</td>
<td>10 ba 35.6 to 44.2</td>
</tr>
</tbody>
</table>

Table 8.1 Summary of sulphur isotope results from current study—K lens and P lens.

P lens sulphide ore returned $\delta^{34}S$ values that in comparison with K lens were more homogenous and exhibited a lower average value (table 8.1). There was no obvious trend in $\delta^{34}S$ values upward through the stratigraphy, either within or between the P sub-lenses. Sulphides from the semi-massive and massive barite lodes located in the vicinity of P lens returned higher $\delta^{34}S$ values than the P lens and K lens sulphide ores.

Pyritic pre-deformation veins and bands cross-cutting the sulphide ore of K lens and P lens (e.g. fig.5.7) display values similar to those of the massive and semi-massive ore. In contrast, pre-deformation pyritic veins within the thick shale unit overlying P lens (fig.5.9) returned elevated $\delta^{34}S$ values in pyrite, similar to those in the barite lodes.
8.2.3. Discussion and interpretation

8.2.3.1. Rosebery δ³⁴S signatures

When integrated with data from earlier studies several features become evident in the Rosebery sulphur isotope dataset. Some of the δ³⁴S signatures within the mine sequence at Rosebery can be related to Cambrian seawater, the δ³⁴S signature of which has been inferred from Cambrian evaporites to be ~30‰ (Claypool et al., 1980) after a minor equilibration fractionation correction of ~1.65‰ (Thode and Monster, 1965).

The δ³⁴S signature of pyrite within the hanging-wall black slate extends over a broad range below ~30‰ (fig.8.1). This is consistent with the biogenic reduction of coeval seawater (Ohmoto and Rye, 1979), an interpretation initially made by Solomon et al. (1969) on a much smaller dataset. The material analysed included disseminated and nodular pyrite, massive lenses and thin deformation-related veinlets interpreted to be the result of local remobilisation (Davidson and Kitto, 1997).

Sulphides analysed within the TSV included finely disseminated, euhedral to framboidal pyrite, pre-deformation veins and disseminations of pyrite and sphalerite related to ore. The range of δ³⁴S values within the TSV is greater than that of the primary ore and less than that of the black slate. The δ³⁴S distribution most likely reflects a signature related to the biogenic reduction of seawater combined with the δ³⁴S signature of ore-related fluids.

The δ³⁴S signature of the ore sulphides is markedly different to those within the black slate and surrounding TSV units, displaying a much narrower range of values and a somewhat bimodal distribution of values (fig.8.1). The ~12‰ mode mainly comprises sulphide δ³⁴S analyses from AB, D, E, G and P lenses, whereas the ~16‰ mode mainly comprises analyses from F, H and K lenses. The δ³⁴S values for barite extend above the value expected for unmodified Cambrian seawater. The difference between the δ³⁴S values for barite and coexistent sphalerite or pyrite in the same rock specimens is in the range of 19.8 to 28.8‰ and averages 22.3‰ (n=7). The same difference in δ³⁴S values is also reflected in histogram plots of the ore-related sulphide and barite datasets (table 8.2 & fig.8.1), with a difference of ~24‰ between the barite modal δ³⁴S value and the upper sulphide modal δ³⁴S value.

The footwall δ³⁴S signature is more complex, with modal δ³⁴S values at ~13‰ and ~17‰ that represent an ore-related signature. The δ³⁴S values below ~0‰ are from within post-deformation Devonian veinlets and may represent a Devonian magmatic signature (Davidson and Kitto, 1997). The scatter of δ³⁴S values above ~20‰ is part of a broader range of values (~5 to 46‰) from disseminated pyrite within a drill hole ~1200m north of AB lens. This population has been interpreted by Davidson and Kitto (1997) to be the result of seawater
Figure 8.1 Rosebery δ²⁸S data for principal stratigraphic units and ore mineralisation.

A) Hangingwall volcaniclastics - black slate unit and contained volcaniclastic turbidite units only.
B) Hangingwall volcaniclastics - mass flow units only.
C) TSV units - samples located both proximal and distal to ore.
D) Footwall volcanics - samples located both proximal and distal to ore.
E) Sulphide and barite mineralisation.
F) Detail of sulphide analyses by ore lens - note bimodal distribution.
G) Detail of barite analyses by ore lens.

Refer to text for discussion and tables 8.2 - 8.3 for further detail. Sulphide analyses include pyrite, sphalerite and pyrrhotite only. (data from Solomon et al., 1969; Green, 1983; Khin Zaw, 1991; Davidson and Kitto, 1997; and this study).


Chapter 8

down-flow during shallow hydrothermal circulation at the margins of the Rosebery ore system with a commensurate partial reduction of seawater sulphate, possibly combined with an overprinting ore-related $\delta^{34}\text{S}$ signature. The broad spread of values is consistent with biogenic reduction of sulphate, but it is not clear whether the original sulphate signature was derived from locally circulating seawater ($\delta^{34}\text{S} \approx 30\%$) or from modified seawater related to the ore forming hydrothermal fluid ($\delta^{34}\text{S} \approx 40\%$).

$\delta^{34}\text{S}$ values for sulphides in pyrrhotite-pyrite-rich metasomatised ore (range 16.1 to 17.2\%, ave. 16.6\%) are similar to $\delta^{34}\text{S}$ values within nearby sphalerite-galena ore (range 15.4\% to 18.6\%, ave. 16.9\%). This suggests that metasomatic replacement of the ore did not have a significant effect on the $\delta^{34}\text{S}$ values but may have simply homogenised the isotopic signature (Green et al., 1981; Khin Zaw, 1991).

### 8.2.3.2. Rosebery $\delta^{34}\text{S}$ - spatial trends

The major ore lenses at Rosebery can be characterised on the basis of average $\delta^{34}\text{S}$ values and metal contents. Sulphide and barite ore lenses with average sulphide $\delta^{34}\text{S}$ values above $\sim$12\% exhibit relatively low average Fe-Cu values and high average Au values (table 8.3 & fig.8.2), whereas ore lenses with average sulphide $\delta^{34}\text{S}$ values below $\sim$12\% are relatively enriched in Fe ($\pm$Cu) and poor in Au. In long section it is clear that a region encompassing ore lenses with low sulphide $\delta^{34}\text{S}$ values at the southern end of the mine passes laterally into regions containing ore lenses with higher $\delta^{34}\text{S}$ values and lower Fe values (fig.8.3). The zone of pyrite mineralisation with elevated $\delta^{34}\text{S}$ values within the footwall north of AB lens ($\sim$5 to 46\%, ave.23.5\% - fig.8.3) is the locus of local fluid down-flow as proposed by Davidson and Kitto (1997).

The outward trend from ore lenses with low sulphide $\delta^{34}\text{S}$ values toward peripheral ore lenses with elevated sulphide $\delta^{34}\text{S}$ values (fig.8.3) may be the result of a reduced local seawater sulphur component incorporated into the sulphide minerals during primary ore formation. This could have occurred as a reduced hydrothermal fluid mixed with increasing volumes of locally

<table>
<thead>
<tr>
<th>analyses</th>
<th>type</th>
<th>range (‰)</th>
<th>modes (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td>hanging wall – black slate</td>
<td>51</td>
<td>py</td>
<td>-6.5 to 30.2</td>
</tr>
<tr>
<td>hanging wall - volcanics</td>
<td>2</td>
<td>py</td>
<td>1.1 to 11.2</td>
</tr>
<tr>
<td>TSV</td>
<td>31</td>
<td>py, sp</td>
<td>-5.0 to 17.7</td>
</tr>
<tr>
<td>footwall volcanics</td>
<td>114</td>
<td>py, sp, (po)</td>
<td>-3.5 to 49.0</td>
</tr>
<tr>
<td>ore - sulphide and barite lodes</td>
<td>119</td>
<td>py, sp (po)</td>
<td>7.1 to 23.5</td>
</tr>
<tr>
<td>barite occurrences</td>
<td>36</td>
<td>ba</td>
<td>33.1 to 47.3</td>
</tr>
</tbody>
</table>

Table 8.2 Summary of sulphur isotope data from all units. (data from Solomon et al., 1969; Green, 1983; Khin Zaw, 1991; Davidson and Kitto, 1997; and this study.)
Figure 8.2 Average sulphide $\delta^{34}S$ vs metal content of major ore lenses.
A) Fe content vs $\delta^{34}S$. Data delineates two trends, one toward the D and E sulphide lenses with high-Fe contents and low $\delta^{34}S$ values, and the other toward the baritic H lens with a low-Fe and high $\delta^{34}S$ value.
B) Cu content vs $\delta^{34}S$ data.
C) Au content vs $\delta^{34}S$ data.
(metal data from Berry et al., 1998; refer table 8.3 for $\delta^{34}S$ data sources)

Figure 8.3 Average sulphide $\delta^{34}S$ vs metal content of major ore lenses.
Long section displaying spatial distribution of ore lenses characterised on the basis of Fe/$\delta^{34}S$ fields (see fig. 8.1a). Field (I): high Fe/low $\delta^{34}S$. Field (II): moderate Fe/low $\delta^{34}S$ values. Field (III): low Fe/high $\delta^{34}S$ values. Field (IV): zone of local down-flow in TSV/footwall defined by Davidson & Kitto (1997) from pyrite $\delta^{34}S$ values in drill hole 109R (range 5-46‰, ave. 23.5‰).
(metal data from Berry et al., 1998; refer table 8.3 for $\delta^{34}S$ data sources)
circulating oxidised seawater away from the major locus of hydrothermal fluid upflow located proximal to D and E lens, commensurate with the partial reduction of locally derived seawater sulphate. The lateral variation in δ\(^{34}\)S values may alternatively be due to the lateral movement of hydrothermal upflow zones with time, in conjunction with an increasing aqueous sulphate component within the hydrothermal fluid. The spatial pattern may also be explained by an early phase of Zn-Pb-rich sulphide and barite mineralisation characterised by elevated δ\(^{34}\)S values, followed by a more spatially restricted phase of Fe-Cu-rich mineralisation characterised by lower δ\(^{34}\)S values produced by a hotter and more reduced hydrothermal fluid. The increase in δ\(^{34}\)S values from Fe-Zn-Cu-enriched K lens ore containing dark brown Fe-rich sphalerite (13.2‰) to the Au-Ag-Pb-barite-enriched K lens ore with pale honey-brown sphalerite (15.4‰) suggests a similar set of processes may also have been active at the ore lens scale.

### Table 8.3 Summary of sulphur isotope results from sulphide and barite lodes.

Sulphide analyses include sphalerite and pyrite only. (mixed sulphide analyses from Stanton and Rafter, 1966. All other data from Solomon et al., 1969; Green, 1983; Khin Zaw, 1991; Davidson and Kitto, 1997; and this study)

<table>
<thead>
<tr>
<th></th>
<th>sulphides</th>
<th>mixed sulphide</th>
<th>barite</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>n=</td>
<td>range (‰)</td>
<td>ave. (‰)</td>
</tr>
<tr>
<td>AB lens</td>
<td>5</td>
<td>9.2 to 10.9</td>
<td>10.4</td>
</tr>
<tr>
<td>C lens</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>D lens</td>
<td>17</td>
<td>7.8 to 18.9</td>
<td>10.5</td>
</tr>
<tr>
<td>E lens</td>
<td>7</td>
<td>9.4 to 13.8</td>
<td>11.5</td>
</tr>
<tr>
<td>F lens</td>
<td>17</td>
<td>13.0 to 17.2</td>
<td>15.3</td>
</tr>
<tr>
<td>G lens</td>
<td>8</td>
<td>9.3 to 12.0</td>
<td>11.2</td>
</tr>
<tr>
<td>H lens (barite lode)</td>
<td>7</td>
<td>14.5 to 19.8</td>
<td>17.3</td>
</tr>
<tr>
<td>J lens</td>
<td>3</td>
<td>7.1 to 10.0</td>
<td>9.0</td>
</tr>
<tr>
<td>K lens</td>
<td>17</td>
<td>9.7 to 17.6</td>
<td>14.0</td>
</tr>
<tr>
<td>P lens</td>
<td>27</td>
<td>8.8 to 16.7</td>
<td>12.3</td>
</tr>
<tr>
<td>P lens (barite lode)</td>
<td>3</td>
<td>10.8 to 23.5</td>
<td>18.5</td>
</tr>
</tbody>
</table>

\[\delta^{34}\text{S}\]

8.2.3.3. Rosebery δ\(^{34}\)S – sulphur isotope geothermometry

δ\(^{34}\)S values derived from coexisting barite-sulphide mineral pairs were used to estimate temperatures of mineral formation utilising the method outlined by Ohmoto and Rye (1979), with sulphate fractionation values from Miyoshi et al. (1984). The use of coexistent mineral pairs assumes that aqueous sulphide and sulphate species within the hydrothermal fluid were in equilibrium prior to rapid barite and sulphide precipitation, which is likely at temperatures above ~250°C (Seal II et al., 2000). δ\(^{34}\)S analyses collected during this study and that of Solomon et al. (1969) returned calculated temperatures in the range of 247-294°C for the H lens and P lens barite lodes (fig.8.4). Barite-sulphide mineral pairs from the AB lens and K lens sulphide lodes returned calculated temperature values in the range of 191-236°C.
Pyrite has typically not been recrystallised during deformation and metamorphism, so it is unlikely that the temperatures derived from the pyrite-barite mineral pairs within the barite lodes reflect metamorphic temperatures. The similarity of pyrite-barite temperature estimates to those derived from sphalerite-barite mineral pairs suggests that isotopic re-equilibration did not occur between sphalerite and barite during metamorphism. Calculated temperatures for the barite lodes appear too high in comparison with temperatures associated with barite occurrences in modern seafloor systems. This was probably due to rapid cooling of the upwelling hydrothermal fluid, followed by precipitation of sulphide and barite minerals that failed to isotopically re-equilibrate at the lower temperatures, thereby preserving disequilibrium $\Delta$SO$_4$-H$_2$S values that are too low for the precipitation conditions (Seal II et al., 2000). The calculated temperatures may, however, reflect the upwelling hydrothermal fluid temperatures at which aqueous sulphide and sulphate species had reached equilibrium, assuming minimal local seawater sulphate input. This would suggest that hydrothermal fluid temperatures were in the range of 250 to 300°C prior to significant mixing with locally circulating seawater. Calculated temperatures for the sulphide lode samples may reflect temperatures under which the barite formed at the margins of AB lens and K lens, however the values are unlikely to reflect the temperature under which most of the sulphide ore formed. The sulphide lode temperature estimates are considered unrealistic and probably reflect isotopic disequilibrium between the sulphide and barite minerals as a result of mineral precipitation during separate phases of mineralisation under different physicochemical conditions.

Mean $\delta^{34}$S values for pyrite, sphalerite and barite from the barite and sulphide lodes were utilised to calculate a "global" temperature estimate for the Rosebery mineralisation. The temperature estimates calculated for pyrite-barite (198-207°C) and sphalerite-barite (202-215°C)
most likely reflect bulk isotopic disequilibrium between the sulphide and barite lodes due to differences in overall physicochemical conditions during mineral precipitation.

8.2.3.4. Regional $\delta^{34}S$ trends

A preliminary attempt to define $\delta^{34}S$ background values in the Mount Read Volcanics through the analysis of least-altered volcanic rocks returned values in the range of $-14.5$ to $24.2\%_o$, which unfortunately indicates that most of the rocks had undergone significant changes in sulphur isotope composition (Davidson and Kitto, 1997). Both hydrothermal and biogenic processes may have been responsible for the range of $\delta^{34}S$ values.

Igneous rocks have historically been attributed $\delta^{34}S$ values in the range of $0 \pm 5\%_o$ (Ohmoto and Rye, 1979; Ohmoto and Goldhaber, 1997), however sulphur isotope variation is evident between different tectonic settings. Mid-ocean ridge basalts exhibit $\delta^{34}S$ values in the narrow range of $0.3 \pm 0.5\%_o$ and reflect depleted mantle compositions (Sakai et al., 1984), whereas arc-related volcanic rocks display a wider range of compositions that reflect $\delta^{34}S$ enrichment due to the incorporation of seawater sulphate through subduction of altered oceanic crust and sediment (Woodhead et al., 1987; Alt, 1994). Volcanic rocks of the Northern Marianas Arc, for example, have an average $\delta^{34}S$ value of $\sim 11\%_o$ and volcanic rocks of the Japanese Island Arc have an average $\delta^{34}S$ value of $\sim 4.4\%_o$ (Ueda and Sakai, 1984; Woodhead et al., 1987). In the context of the tectonic setting for formation of the Mount Read Volcanics, average primary $\delta^{34}S$ values within the predominantly felsic volcanic succession were most likely in the range of $\sim 5$ to $10\%_o$. The $\delta^{34}S$ signature of mafic volcanics in the Mariana backarc basin is in the range of $\sim 0.6 \pm 0.3\%_o$ (Kusakabe et al., 1990), and this may indicate that primitive mafic lavas such as the Hellyer basalt probably exhibited primary values closer to $0\%_o$. A value of $6\%_o$ was similarly assumed by Wyman (2001) for the magmatic $\delta^{34}S$ signature of the Darwin granite on the basis of comparison with more recent analogues.

A regional-scale pre-Devonian reconstruction of the northern Mount Read belt stratigraphy by Berry (1993a) suggests that Rosebery was originally underlain by a $\sim 1.3$km thickness of rhyolitic to dacitic volcanics of the CVC, and a further $\sim 2$km thickness of Crimson Creek Formation that in turn overlay Precambrian rocks. The Crimson Creek Formation along the western side of the Mount Read Volcanic belt comprises mafic volcanioclastics and tholeiitic lavas, with associated fault-emplaced ultramafic-mafic complexes (Corbett et al., 1989b; Corbett, 1992). The $\delta^{34}S$ signature of the tholeiitic lavas and ultramafic-mafic complexes was likely to have originally been close to $0\%_o$, consistent with a mantle origin for the volcanics (Sakai et al., 1984; Crawford and Berry, 1992). Limited isotopic analysis of mafic volcanioclastic turbidites and shales within the Crimson Creek Formation returned $\delta^{34}S$ values from $-22.3\%_o$ to $15.23\%_o$ (ave. $2.2\%_o$) (Heath, 1999), a range that is consistent with the biogenic reduction of seawater within an open marine environment (Ohmoto and Rye, 1979). The overall $\delta^{34}S$
Figure 8.5 €34S data for Cambrian and Devonian deposits in western Tasmania.
Cambrian:- Rosebery, Hercules, South Hercules, Que River, Hellyer, Mt Lyell, Chester, Jukes - Mt Darwin System, Anthony PDT. Devonian:- Farrell Field, Zeehan Field, Lynch Creek.
Sulphide analyses include pyrite, sphalerite, pyrrhotite & chalcopyrite only.
(data from Both, 1966; Solomon et al., 1969; Collins, 1981b; Walshe and Solomon, 1981; Green, 1983; Eastoe et al., 1987; Lees, 1987; Solomon et al., 1988; Jack, 1989; Doyle, 1990; Khin Zaw, 1991; Abbott, 1992; Gemmell and Large, 1992; Khin Zaw and Large, 1992; McGoldrick and Large, 1992; Raymond, 1992; Gadaloff, 1996; Jackson, 1996; Mawdesley, 1996; Davidson and Kitto, 1997; Wyman, 2001; Green and Vicary, 2002; and this study)
signature of the Crimson Creek Formation was likely to have been between 0‰ and ~2‰, a range comparable to that of modern oceanic crust (Sakai et al., 1984; Alt, 1995b).

A limited number of sulphur isotope analyses have been undertaken on Precambrian rocks of the Rocky Cape and Tyennan regions by previous workers, with sandstone and shale of the Rocky Cape Group returning $\delta^{34}$S values in the range of 15.9‰ to 18.1‰ (Yaxley, 1981), Precambrian quartzwacke turbidites west of Zeehan returning $\delta^{34}$S values in the range of 18.5‰ to 19.4‰ (Hajataheri, 1985), and Precambrian metasediments of the Tyennan Block returning values >14‰ (Solomon and Groves, 2000). There is insufficient data to ascertain a representative $\delta^{34}$S value for the Precambrian rocks, however the results to date suggest a value close to ~15‰, possibly reflecting a derivation from Precambrian seawater (~15‰) (Claypool et al., 1980).

Cambrian syngenetic mineral deposits of the Mount Read Volcanics exhibit several distinctive $\delta^{34}$S value distributions (fig.8.5). The syngenetic base metal deposits interpreted to have formed at or near the seafloor such as Hercules, South Hercules, Hellyer, Que River and Mt Lyell (Lees, 1987; Gemmell and Large, 1992; Khin Zaw and Large, 1992; McGoldrick and Large, 1992; Corbett, 2001) display modal sulphide $\delta^{34}$S values in the range of ~8 to 12‰ and modal barite $\delta^{34}$S values in the range of ~40 to 45‰. The Hercules and South Hercules deposits exhibit marginally higher modal values than the Hellyer, Que River and Mt Lyell deposits. This has been attributed to the more mafic volcanic host rock sequences associated with Hellyer and Que River, which would have exhibited higher sulphur contents than their felsic counterparts, thereby increasing the low-$\delta^{34}$S volcanic sulphur component incorporated into the ore-forming hydrothermal fluid (Solomon et al., 1988). In comparison with the other base metal deposits, the Chester high sulphidation pyrite deposit displays relatively low sulphide and barite $\delta^{34}$S values (Davidson and Blake, 2002). The Mt Lyell deposit displays a secondary set of $\delta^{34}$S values that has been attributed to a separate phase of high sulphidation mineralisation (Walshe and Solomon, 1981; Davidson and Blake, 2002; Solomon, 2002b).

Mount Darwin and surrounding mineral prospects display a wide range of sulphide $\delta^{34}$S values and appear to have preserved a Cambrian seawater signature within the barite (fig.8.5). Wyman (2001) identified three overlapping $\delta^{34}$S signatures within mineralisation interpreted to be genetically related to the Cambrian Darwin granite, and attributed them to a progressive change in sulphur source. Early intrusion-related mineralisation produced $\delta^{34}$S values in the range 7 to 9‰ through mixing of magmatic sulphur (~6‰) and rock sulphur equilibrated with seawater (10 to 15‰). A middle phase of mineralisation was associated with increased modified seawater input that produced $\delta^{34}$S values in the range of 10 to 14‰, and the $\delta^{34}$S values increased further to 15 to 19‰ as the modified seawater sulphur input again increased. A final phase of mineralisation was related to the influx of seawater into the granite itself, thereby producing variable $\delta^{34}$S values in sulphide. With further cooling (<300°C) free sulphate
was able to enter the system, which then allowed rapid precipitation of sulphate without isotopic fractionation, thereby preserving Cambrian seawater $\delta^{34}S$ values in the barite (~30‰). A Cambrian seawater signature also appears to have been preserved in association with a barite-galena occurrence in the Anthony Power Development Tunnel near the Murchison granite (Abbott, 1992).

Pb-Zn deposits of Devonian age, including the Zeehan and Farrell mining fields (fig.8.5), exhibit sulphide $\delta^{34}S$ value distributions that overlap with those of the Cambrian deposits. Variability between the Devonian $\delta^{34}S$ distributions most likely reflects a diversity of host rocks from which much of the sulphur was derived, including Cambrian sedimentary (Farrell Field) and mafic volcanic units (Lynch Creek), and Proterozoic and Silurian sedimentary units (Zeehan Field). The sulphur isotope signature of the Farrell Field has been interpreted as the product of granite-derived magmatic sulphur ($\leq$12‰), combined with a significant amount of sulphur remobilised from the surrounding Cambrian volcanic rocks (~18‰) (Solomon et al., 1988; Mawdesley, 1996). $\delta^{34}S$ values within sulphides are generally not suitable for distinguishing between Cambrian and Devonian ores, however $\delta^{34}S$ values within Devonian barite from the Farrell Field appear markedly lower than $\delta^{34}S$ values within barite from most Cambrian deposits (fig.8.5).

In the regional context Rosebery exhibits a sulphur isotope signature that is most similar to the Cambrian syngenetic deposits such as Hercules, Hellyer and Que River deposits.

**8.2.4. Sulphur isotope model - discussion**

The model proposed by Solomon et al. (1988) to explain the sulphur isotope signature of Rosebery mineralisation commences with the downward circulation and heating of seawater ($\delta^{34}S \approx 30‰$) to ~250°C, with commensurate precipitation of anhydrite formed through the reaction of seawater sulphate with Ca (fig.8.6). Anhydrite is likely to have precipitated over the temperature range of ~150 to 200°C (Shanks III et al., 1981). Fractionation associated with precipitation of 90% of seawater sulphate as anhydrite could have lowered the $\delta^{34}S$ of the residual aqueous sulphate by ~2.3‰ (Alt and Chaussidon, 1989). Precipitated anhydrite would subsequently have been dissolved during the influx of cooler seawater due to a retrograde solubility at low temperatures, and the sulphate would have then been recirculated into the local environment (Holland and Malinin, 1979; Shanks III et al., 1981). With further heating the hydrothermal fluid may have leached more Ca from the surrounding rocks, resulting in further depletion of aqueous sulphate through precipitation of anhydrite (Shanks III et al., 1995).

Upon further heating the remnant sulphate in solution would have been reduced through reaction with ferrous iron mineral species such as hornblende, magnetite or chlorite, with isotopic fractionation of sulphur occurring in a fashion that could be viewed as a Rayleigh
Chapter 8

Distillation process (fig. 8.7) (Solomon et al., 1988). At temperatures in the range of 250°C to 350°C the inorganic reduction of seawater sulphate would have been rapid (Shanks III et al., 1981; Ohmoto and Lasaga, 1982). Attainment of equilibrium between sulphate and H₂S as the dominant aqueous sulphide species is also rapid over this temperature range (Ohmoto and Rye, 1979; Shanks III et al., 1981). At temperatures between 250° and 300°C the isotopic fractionation between aqueous sulphate and sulphide would have been between 25‰ and 18‰ (Ohmoto and Rye, 1979; Shanks III et al., 1981). Seal et al. (2000) argued that elevated δ³⁴S values in barite within Cambrian deposits of the Mount Read Volcanics could have been derived through bacterial reduction of seawater sulphate in anoxic bottom waters related to an episode of global oceanic bottom-water anoxia. This mechanism of sulphate reduction is considered unlikely, as the range of values that would have been produced by biogenic reduction of seawater sulphate should exceed the range of values observed in deposits such as Hellyer and Rosebery (Green et al., 1981; Huston, 1999), and there is evidence for the preservation of a ~30‰ seawater signature within some mineralised systems of the Mount Read Volcanics (fig.8.5).

Solomon et al. (1988) proposed that a hot hydrothermal fluid moving through volcanic rocks beneath Rosebery would have leached a significant component of sulphur with a magmatic δ³⁴S signature, and sulphur derived from modified seawater and volcanic rock sources would have been homogenised by mixing within the hydrothermal fluid. The δ³⁴S signature inferred by Solomon et al. (1988) for leached magmatic sulphur was close to 0‰, however the δ³⁴S signature of leached sulphur was more likely to have been higher as hot hydrothermal fluids passed through volcanics of the CVC (~5‰ to 10‰) and Crimson Creek Formation (~0 to...
2‰), and metasediments of the Precambrian basement (~15‰). The relative sulphur input from each of these units would have been dependent on factors such as the sulphur content and volume of rocks through which the hydrothermal fluids passed. Based on the stratigraphy discussed previously, and rock volume calculations for a range of hydrothermal circulation depths and hydrothermal cell aspect ratios, δ³⁴S values for the leached sulphur component is likely to have been in the range of ~5‰ to 10‰. Although a direct magmatic sulphur component related to magmatic fluids cannot be ruled out, the very low Se content of pyrites within the Rosebery ore probably indicates the lack of a significant direct magmatic input (Huston et al., 1995b).

Solomon et al. (1988) proposed that no significant change in the sulphur isotope signature of the hydrothermal fluid occurred during rapid upward movement, and that rapid mixing of hydrothermal fluid with seawater would have prevented any isotopic re-equilibration between

---

**Figure 8.7 Rayleigh isotopic fractionation of Cambrian seawater.**
The diagram illustrates variation in δ³⁴S values of aqueous sulphide and residual sulphate formed by partial reduction of Cambrian seawater (δ³⁴S =30‰) at 250°C. The range of δ³⁴S values exhibited by sulphide (pyrite-sphalerite) and barite associated with Rosebery ore mineralisation is shown for comparison. (modified after Huston, 1999)
the aqueous sulphur species. The rapid mixing of seawater proposed by Solomon et al. (1988) is based on the premise of a focussed fluid flow venting above the seafloor. If the hydrothermal fluid had instead flowed in a diffuse fashion through unconsolidated sediments, it is possible that significant mixing of hydrothermal fluid and unmodified seawater would have occurred prior to fluid venting. The lack of an unmodified seawater $\delta^{34}S$ signature (~30‰) in the Rosebery barite, suggests that the hydrothermal fluid mixed with modified seawater from which a significant proportion of sulphate had been removed through anhydrite precipitation. Locally derived seawater sulphate therefore did not impact significantly on the isotopic signature of the central ore lenses (fig.8.3). The more peripheral ore lenses with $\delta^{34}S$ values above ~12‰ may have incorporated small amounts of locally derived sulphur; either directly from locally circulating seawater or through reduction and remobilisation of anhydrite precipitated marginal to the ore. The pattern of average $\delta^{34}S$ values at the mine scale may also be the result of widespread initial Zn-Pb-rich sulphide and barite mineralisation with elevated $\delta^{34}S$ values, overprinted by a more spatially restricted phase of high-temperature Fe-Cu-rich sulphide mineralisation with low $\delta^{34}S$ values.

Based upon an inferred exhalative sulphide depositional sequence, Solomon et al. (1988) proposed a change from low $\delta^{34}S$ values in an early phase of chalcopyrite-pyrite mineralisation, to higher $\delta^{34}S$ values in a later phase of sphalerite-galena mineralisation due to depletion of volcanic rock sulphur in the high temperature part of the system. The inferred diachronous nature of the ore was initially put forward by Green et al. (1981) and Green (1983), however the lack of macro-scale folding of the northern ore lenses and an inverted sulphide sequence for G lens ore suggests that the mineralisation sequence was not so straightforward. If the chalcopyrite-pyrite and sphalerite-galena mineralisation are instead interpreted as part of a sulphide mound undergoing zone refinement (e.g. Huston, 1988), the relative timing of these ores is reversed, and if the ores are interpreted as part of a sub-seafloor replacement (Allen, 1994a; Doyle and Allen, 2003) the timing of these ore types is open. Regardless of timing, any overall increase in $\delta^{34}S$ values could be due to factors other than depletion of rock sulphur in the high temperature part of the system. An increase in $\delta^{34}S$ values may also have been related to an increased water/rock ratio or an increased sulphate component in the heated seawater due to Ca depletion (Solomon et al., 1988). Rapid heating of seawater to 250°C and reduction of sulphate prior to anhydrite precipitation could also have increased $\delta^{34}S$ values in the hydrothermal fluid (Alt and Chaussidon, 1989). This last process is unlikely to have occurred as part of the normal hydrothermal cycle, but it may have become a significant component as the result of a shallow magmatic intrusion proximal to the site of fluid upflow.

Solomon et al. (1988) suggested that as the hydrothermal system cooled, and after much of the Ca and ferrous components in the volcanic rocks had been consumed, a resultant increase in sulphate within the upwelling hydrothermal fluid would have led to formation of significant
barite mineralisation. The barite lodes have in the past been interpreted to post-date sulphide mineralisation on the basis of an exhalative stratigraphy and occurrences of barite veins cutting sulphide ore (Green et al., 1981; Green, 1983; Solomon et al., 1988). However, barite commonly evident in sulphide ore at the northern end of the mine (fig.5.8g) probably predates significant sulphide mineralisation, suggesting that significant barite mineralisation occurred throughout much of the life of the system and not just as the system waned.

In an upwelling hydrothermal fluid at temperatures of between 100° and 350°C, and with H₂S as the dominant sulphur species, the isotopic fractionation between the fluid and sphalerite or pyrite is ≤1.5‰ (Ohmoto, 1972). The low δ³⁴S values associated with chalcopyrite-pyrite ore (~7.8 to 9.8‰) have been interpreted to represent the product of peak temperatures in the hydrothermal system, with commensurate minimal sulphur derived from modified seawater (Green et al., 1981). This is consistent with a basement rock-derived δ³⁴S signature in the range of 5 to 10‰.

Barite lodes and Zn-Pb-rich sulphide ores with δ³⁴S values above ~12‰, appear to be the product of hydrothermal fluids that carried a significant aqueous sulphate component (δ³⁴S ~40‰) as the result of lower temperatures and higher water/rock ratios within the hydrothermal system; and/or due to depletion of Ca and ferrous mineral components in the volcanic rocks through which the fluids passed. The more Fe-Cu-rich sulphide mineralisation, with δ³⁴S values below ~12‰, appears to be the product of high temperature hydrothermal fluids that lacked a significant aqueous sulphate component. Lateral and vertical variation in δ³⁴S values at the mine scale is also probably due to a combination of the spatial extent of these two phases of mineralisation, in conjunction with varying amounts of a local seawater sulphur component.

8.2.5. Summary

The Rosebery sulphur isotope signature is consistent with a combined input of sulphur from modified seawater and the rocks through which the hydrothermal fluid passed. As the seawater was drawn below the seafloor it was heated, and at temperatures above ~150°C anhydrite was formed through the interaction of aqueous sulphate with Ca derived from the volcanic rocks. The hydrothermal fluid was further heated to temperatures above 250°C, and the isotopic fractionation that occurred during reduction of remnant sulphate can be described as a Rayleigh fractionation process. The aqueous sulphide and sulphate present in the hydrothermal fluid reached rapid equilibrium at the elevated temperatures. The heated hydrothermal fluid also stripped sulphur from the rocks through which it passed, and sulphur from the two sources mixed homogenously within the hydrothermal fluid. The rapidly upwelling hydrothermal fluid did not undergo significant changes. Sulphide mineralisation formed upon mixing of the hydrothermal fluid with cooler seawater. Several factors including cooler
temperatures within the hydrothermal system, increased water/rock ratios and depletion of Ca and ferrous minerals contributed to periods where a significant component of aqueous sulphate was present in the upwelling hydrothermal fluid, leading to formation of barite mineralisation. The system is interpreted to have fluctuated between periods where the upwelling hydrothermal fluid contained negligible aqueous sulphate and periods where aqueous sulphate was a significant sulphur species. The lateral and vertical variation in $\delta^{34}$S values within the ore lenses was due to varying degrees of mixing between the hydrothermal fluid and locally-circulating seawater. Sulphate remobilised during replacement and dissolution of barite mineralisation may also have contributed to the lateral variation in $\delta^{34}$S values.

8.3. Carbon and oxygen isotopes

No new carbon or oxygen isotope data was collected during the current study, as previous studies appear to have been sufficient to characterise the Rosebery system. Oxygen isotope analysis has been undertaken by Solomon et al. (1969) on barite (n=15) and coexisting carbonate (n=3) proximal to AB, E, G and H lenses, with barite returning $\delta^{18}$O values in the range of 9.4 to 12.8‰ and carbonate returning $\delta^{18}$O values in the range of 12.4 to 14.2‰. In further studies, Dixon (1980) analysed 13 samples proximal and lateral to AB, E, F, G and H ore lenses, Khin Zaw (1991) analysed 20 carbonate samples proximal to AB and J lenses, and Large et al. (1998a) analysed a further 24 carbonate samples across the mine stratigraphy in the vicinity of K lens. The samples taken during these studies included both syngenetic ore-related carbonate and Devonian syn-tectonic carbonate as veins and impregnations. The results of these studies are summarised and discussed below.

The preservation of primary colloform textures and uncompacted pumice shards within the syngenetic pre-deformation carbonate probably indicates that the isotopic compositions have not been affected by Devonian metamorphism (Dixon, 1980). The $\delta^{13}$C values for Cambrian hydrothermal carbonate at Rosebery (fig.8.8) display a range of values similar to those found in other volcanic-hosted massive sulphide systems where $\delta^{13}$C values typically range between −5 and 0‰ (Huston, 1999).

The $\delta^{18}$O values within Cambrian hydrothermal carbonate at Rosebery are lower than Cambrian sedimentary carbonates, which typically exhibit $\delta^{18}$O values in the range of 14.5 to 26.5‰ (Veizer and Hoefs, 1976), and the lower $\delta^{18}$O values are most likely caused by a difference in depositional temperatures (Huston, 1999). The range of carbonate $\delta^{18}$O values are consistent with deposition from a hydrothermal fluid with a $\delta^{18}$O value of ~0 to 4‰ at temperatures between 100° and 300°C (Huston, 1999). The Rosebery carbonate $\delta^{13}$C values are consistent with derivation from dissolved bicarbonate in seawater, possibly with the involvement of some oxidised organic carbon (Huston, 1999).
Figure 8.8 Carbon and oxygen isotopes—Rosebery carbonate.
The diagram shows the isotopic signature of Cambrian syngenetic carbonate and Devonian syn-deformation carbonate. (data from Dixon, 1980; Khin Zaw, 1991; and Large et al., 1998a)

Figure 8.9 Carbon and oxygen isotopes - Cambrian vs Devonian systems.
The diagram shows the isotopic signature of Cambrian and Devonian hydrothermal carbonate for several deposits in western Tasmania. (data from Dixon, 1980; Collins, 1981a; Khin Zaw, 1991; Khin Zaw and Large, 1992; Mawdesley, 1996; and Large et al., 1998a)
The South Hercules deposit, located in a similar stratigraphic setting south of Rosebery, has been interpreted to have formed by a sub-seafloor replacement process from a relatively low-temperature (150° to 250°C), near neutral hydrothermal fluid having $\delta^{18}O$ values of 2 to 4‰ and $\delta^{13}C$ values of -1 to -2‰ (Khin Zaw and Large, 1992). The strong linear relationship evident in the $\delta^{13}C$ and $\delta^{18}O$ data has been attributed to the effects of temperature change on equilibrium fractionation between the fluid and precipitating manganiferous carbonate, with lower temperatures leading to increased $\delta^{18}O$ and $\delta^{13}C$ values (Khin Zaw and Large, 1992). Alternatively, Solomon (2002a) has suggested that the same trend might be derived from a hot carbon-rich fluid mixing with cooler entrained seawater. The Rosebery $\delta^{13}C$ and $\delta^{18}O$ data plots in a restricted range, without an apparent linear trend that might be utilised in hydrothermal fluid modelling. If the hydrothermal fluids and processes of carbonate deposition were similar to those at South Hercules, it might be inferred that the Rosebery carbonates formed at the higher end of the temperature range suggested for South Hercules carbonate formation. This is also supported by the higher Cu content of the Rosebery system compared to the South Hercules system.

The syn-tectonic carbonate subjected to analysis at Rosebery comprised predominantly veins and impregnations. Syn-tectonic carbonate proximal to ore returned $\delta^{18}O$ and $\delta^{13}C$ values similar to those seen in the ore-related carbonates (table 8.4), as did syn-tectonic carbonate from the footwall mass flow units. Syn-tectonic carbonate from within the hanging-wall mass flow units exhibited similar $\delta^{13}C$ values associated with marginally elevated $\delta^{18}O$ values. Syn-tectonic carbonate within the black slate and contained turbidite packages returned significantly elevated $\delta^{18}O$ values and reduced $\delta^{13}C$ values. There is no correlation between carbonate composition and the $\delta^{18}O$ or $\delta^{13}C$ values (Dixon, 1980; Large et al., 1998a). The $\delta^{18}O$ and $\delta^{13}C$ values within Cambrian hydrothermal carbonate do not display a linear trend as found in the Hercules South isotopic data (fig.8.9), however the syn-tectonic veins at Rosebery do display a trend toward elevated $\delta^{18}O$ values and reduced $\delta^{13}C$ values (fig.8.8). The $\delta^{18}O$ and $\delta^{13}C$ values in the Devonian carbonates probably represent the mixing of an inherited Cambrian hydrothermal carbonate signature with a signature, related to the oxidation of

<table>
<thead>
<tr>
<th>n</th>
<th>$\delta^{18}O$ range (‰)</th>
<th>$\delta^{18}O$ ave. (‰)</th>
<th>$\delta^{13}C$ range (‰)</th>
<th>$\delta^{13}C$ ave. (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ore position: Cambrian hydrothermal</td>
<td>37</td>
<td>8.7 to 14.3</td>
<td>11.3</td>
<td>-6.8 to -0.5</td>
</tr>
<tr>
<td>ore position: syn tectonic</td>
<td>11</td>
<td>8.9 to 13.9</td>
<td>11.3</td>
<td>-8.8 to -1.6</td>
</tr>
<tr>
<td>footwall: syn-tectonic</td>
<td>6</td>
<td>8.5 to 13.0</td>
<td>11.0</td>
<td>-6.5 to -0.4</td>
</tr>
<tr>
<td>hanging wall: syn-tectonic</td>
<td>8</td>
<td>10.6 to 13.4</td>
<td>11.4</td>
<td>-4.5 to -0.6</td>
</tr>
<tr>
<td>black slate: syn-tectonic</td>
<td>4</td>
<td>12.1 to 13.4</td>
<td>12.7</td>
<td>-5.8 to -8.7</td>
</tr>
</tbody>
</table>

Table 8.4 Summary of carbon and oxygen isotope data in carbonate.
(data from Dixon, 1980; Khin Zaw, 1991; and Large et al., 1998a)
organic carbon in the black slate and adjacent volcaniclastic sediments (Large et al., 1998a).

The $\delta^{13}C$ and $\delta^{18}O$ values for Rosebery carbonate fall within the same general area as other syngenetic deposits from the Mount Read Volcanics (fig. 8.9), and they can be differentiated from the Devonian hydrothermal carbonates that exhibit higher $\delta^{18}O$ and lower $\delta^{13}C$ values.

### 8.3.1. Whole rock oxygen isotopes

In a study of the Rosebery mine sequence, Green (1990a, b) examined the $\delta^{18}O_{wr}$ values for 57 rock samples taken along ~8km of strike extent centred on the mine. The study returned results in the range of 7.6 to 13.5‰, with significant zones of low $\delta^{18}O_{wr}$ values (<10‰) extending for less than 100m laterally away beneath the ore lenses and only a few tens of metres into the footwall beneath the ore horizon. Samples from the Hangingwall Volcaniclastics returned only one value below 10‰.

In a review study undertaken by Herrmann (2002), least altered rocks in the Mount Read Volcanics were found to display $\delta^{18}O_{wr}$ values in the range of 6.4‰ to 15.9‰ (mean = 11.4‰), whereas rocks affected by hydrothermal alteration display values in the slightly broader range of 1.8‰ to 19.0‰ (mean = 10.1‰). A value of 9‰ appears to represent the threshold below which rocks are most likely to have been hydrothermally altered (Herrmann, 2002). The Rosebery $\delta^{18}O_{wr}$ data does not define a broad halo as found beneath other massive sulphide deposits, where significantly lower $\delta^{18}O_{wr}$ values may extend for 100 to 1500m laterally beneath the ore (Huston, 1999).

Metamorphic fluids in equilibrium with the bulk of the volcanic succession outside the Rosebery mine area could have produced fluids with $\delta^{18}O_{wr}$ values of 11.5 to 16.5‰ (Green, 1990b), and equilibration of granite-related metasomatic fluids with the volcanic rocks could also have produced fluids with a similar range of $\delta^{18}O_{wr}$ values (Green, 1990a). Although low $\delta^{18}O_{wr}$ values beneath the Rosebery ore lenses are considered to be primary (Green, 1990a, b), it is possible that much of the $\delta^{18}O_{wr}$ pattern associated with the Rosebery sequence is the result of an early Cambrian hydrothermal signature overprinted by a Devonian hydrothermal signature characterised by elevated $\delta^{18}O_{wr}$ values. This interpretation is consistent with the distribution of deformation and granite related alteration found throughout the mine. Consequently it is unlikely further detailed work on the $\delta^{18}O_{wr}$ patterns at Rosebery will assist in understanding the primary Cambrian hydrothermal flow patterns associated with mineralisation.
8.4. **Strontium isotopes**

Whitford et al. (1992) examined the strontium isotope composition of barite from several Cambrian ore deposits on the west coast of Tasmania and compared them to potential sources. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio values within barite from the Rosebery deposit range between 0.70929 and 0.71231 (n=3), values which Whitford et al. (1992) considered to be primary in nature and therefore unaffected by Devonian deformation and granite-related processes.

In comparison with estimates for the $^{87}\text{Sr}/^{86}\text{Sr}$ values for coeval seawater of ~0.7090 (Burke et al., 1982; Veizer et al., 1999), the surrounding volcanic pile (~0.7065-0.7095) (Whitford et al., 1990) and a possible juvenile magmatic source component, the values exhibited by barite from Rosebery are significantly elevated and must, therefore, have received input from a $^{87}\text{Sr}$-enriched source. Whitford et al. (1992) argued that elevated $^{87}\text{Sr}/^{86}\text{Sr}$ ratio values within barite at Rosebery could be attributed to deep circulation of hydrothermal fluids through $^{87}\text{Sr}$-enriched continental crust onto which the Mount Read Volcanics were originally deposited. Isotopic analysis of Devonian granites was utilised to indicate $^{87}\text{Sr}/^{86}\text{Sr}$ ratio values for the early crust, with $^{87}\text{Sr}/^{86}\text{Sr}$ ratio values in the range of 0.7104 to 0.76.

8.5. **Lead isotopes**

Ostic (1967) undertook Pb isotope analysis of galena samples from A, B and F lenses at Rosebery and found the isotopic signature to be homogenous ($^{206}\text{Pb}/^{204}\text{Pb}$: 18.366 to 18.382, $^{207}\text{Pb}/^{204}\text{Pb}$: 15.740 to 15.756, $^{208}\text{Pb}/^{204}\text{Pb}$: 38.454 to 38.496, n=8). The Rosebery values appeared colinear with elevated isotopic values obtained from the Mt Farrell deposit of Devonian age ($^{206}\text{Pb}/^{204}\text{Pb}$: 18.687 to 18.743, $^{207}\text{Pb}/^{204}\text{Pb}$: 15.780 to 15.797, n=6). The apparent age inferred from the isotopic data was anomalous and did not fit with a Cambrian age for the Rosebery ores (Ostic et al., 1967).

Examination of the Pb isotope signature of ore deposits in western Tasmania undertaken by Gulson and Porritt (1987) and Gulson et al. (1987) included material from the Rosebery deposit. Analyses were undertaken on galena-rich sulphide bands selected from the D, E, F and H ore lenses, with the Pb isotope signature again proving to be homogenous ($^{206}\text{Pb}/^{204}\text{Pb}$: 18.266 to 18.279, $^{207}\text{Pb}/^{204}\text{Pb}$: 15.583 to 15.609, $^{208}\text{Pb}/^{204}\text{Pb}$: 38.041 to 38.083, n=8) and reasonably consistent with the analyses of Ostic et al. (1967). The isotopic signatures of the Rosebery and Hercules ore deposits are coincident and they exhibit similar values to the Cambrian Hellyer and Que River ore deposits (fig.8.10). Devonian deformation- and granite-related vein-style mineralisation displays a more radiogenic signature than the Cambrian deposits with $^{206}\text{Pb}/^{204}\text{Pb}$ ratio values typically more than 1 percent higher.

The Chamberlain and Salisbury prospects are located adjacent to the Rosebery Fault.
Figure 8.10 Pb isotope data - Cambrian vs Devonian.
Devonian mineralisation exhibits a more radiogenic signature than Cambrian mineralisation. (modified after Gulson and Porritt, 1987; and Gulson et al., 1987)

Figure 8.11 Pb isotope data - Cambrian and Devonian.
Plot showing linearity of isotopic data for varying types of mineralisation and some rock data in the Mount Read volcanic belt. Isotopic ratios for ores or high Pb samples are average figures. The line of best fit intersects the lead evolution (growth) curve for other massive sulphide deposits at ~1000m.y. (plot includes data from Cambrian deposits and prospects including Rosebery, Prince Lyell, Que River, Hellyer, and various Voyager prospects, and from Devonian deposits including- Farrell lode, Renison, Tullah, Murchison lode and Queen Hill. (after Gulson and Porritt, 1987)
southwest of Rosebery and comprise Devonian fault-related sulphide mineralisation. Both the Salisbury prospect \( ^{206}\text{Pb}/^{204}\text{Pb}: 18.528, \quad ^{207}\text{Pb}/^{204}\text{Pb}: 15.620, \quad n=3 \) and the Chamberlain prospect exhibit signatures that reflect a Devonian origin \( ^{206}\text{Pb}/^{204}\text{Pb}: 18.420, \quad ^{207}\text{Pb}/^{204}\text{Pb}: 15.600, \quad n=2 \) (Dean and Carr, 1989).

Gulson and Porritt (1987) and Gulson et al. (1987) interpreted colinearity of the Cambrian and Devonian data to indicate a common origin for the Pb (fig.8.11). The intersection of the linear array on the \( ^{207}\text{Pb}/^{204}\text{Pb} \) vs \( ^{206}\text{Pb}/^{204}\text{Pb} \) plot with the massive sulphide Pb-evolution curve, in conjunction with U-Pb ages of inherited zircons and strontium isotope modelling, supports the interpretation that Pb was originally sourced from Precambrian crust with an age of \( \geq 1000\text{Ma} \) and a \( ^{206}\text{Pb}/^{204}\text{Pb} \) ratio value of \( \sim 17.2 \). After the further radioactive decay of \( ^{238}\text{U} \) and a commensurate increase in the \( ^{206}\text{Pb} \) content, the Precambrian crust was recycled during formation of the Mount Read Volcanics. Hydrothermal circulation and intrusion resulted in formation of massive sulphide deposits with \( ^{206}\text{Pb}/^{204}\text{Pb} \) ratio values of \( \sim 18.08 \) to \( 18.36 \). Further radioactive decay of U and Th within the Precambrian sediments and Cambrian volcanics further increased the levels of \( ^{206}\text{Pb} \), and Devonian deformation- and granite-related mineralisation is characterised by \( ^{206}\text{Pb}/^{204}\text{Pb} \) ratio values of \( \sim 18.43 \) to \( 18.61 \).

Further radioactive decay of U and Th within the Precambrian sediments and Cambrian volcanics further increased the levels of \( ^{206}\text{Pb} \), and Devonian deformation- and granite-related mineralisation is characterised by \( ^{206}\text{Pb}/^{204}\text{Pb} \) ratio values of \( \sim 18.43 \) to \( 18.61 \).

The isotopic signature for Rosebery mineralisation fits with a Cambrian age for primary ore formation. As discussed previously, the Devonian fault-related mineralisation that comprise the Salisbury and Chamberlain prospects exhibits elemental signatures within pyrite and sphalerite indicative of a remobilisation from a Cambrian syngenetic precursor. However, the mineralisation displays a typical Devonian Pb isotope signature. This indicates that deformation and granite-related remobilisation of Cambrian ore may have imparted a Devonian Pb isotope signature on the resultant mineralisation. This is at odds with the unsupported assumption by Aerden (1991; 1993), that remobilisation of a precursor sulphide ore and subsequent transport during deformation would not alter the Pb isotope signature of the resultant epigenetic mineralisation. It is likely that if this process had occurred, a more radiogenic signature may have been imparted onto the Rosebery ore than is now evident.

### 8.6. Summary

The Mount Read volcanics were probably enriched in \( ^{34}\text{S} \) due to seawater sulphate incorporated into parental magma during subduction of oceanic crust. Radiogenic Pb was also enriched due to recycling of Precambrian crust. Background \( ^{87}\text{Sr}/^{86}\text{Sr} \) levels suggest that similar enrichment in \( ^{87}\text{Sr} \) due to recycling of Precambrian crust did not occur.

This study found that the Rosebery sulphide and barite mineralisation is consistent with derivation of sulphur from mixing of modified seawater with sulphur contained in volcanic and meta-sedimentary rocks through which the hydrothermal fluid passed. Initial heating of
seawater within the hydrothermal system caused removal of sulphate through precipitation of anhydrite. Further heating caused reduction of remnant sulphate through interaction with ferrous iron minerals and rapid fractionation of sulphur isotopes between aqueous sulphur species. The formation of sulphides with a low $\delta^{34}S$ signature reflects elevated temperatures in the hydrothermal system and a maximum in the relative rock sulphur input. The formation of barite and sulphide with an elevated $\delta^{34}S$ signature reflects a significantly increased sulphate content in the hydrothermal fluid as a result of a decreased temperature, increased water/rock ratio and depletion of Ca and ferrous minerals in the volcanic rocks. The ore-forming hydrothermal fluid probably fluctuated between these two states. It is proposed here that elevated $\delta^{34}S$ values within peripheral sulphide lenses are indicative of the reduction and incorporation of locally derived seawater sulphate, possibly combined with sulphate remobilised from early barite mineralisation. The spatial distribution of $\delta^{34}S$ values across the mine, as outlined in this study, would indicate that the Fe-Cu-rich southern ore lenses are located proximal to the principal zone of hydrothermal upflow at Rosebery. Subsequent Devonian metasomatism has locally homogenised the $\delta^{34}S$ signature during local sulphide remobilisation. $\delta^{34}S$ values within the barite lodes appear to have preserved evidence for hydrothermal fluid temperatures in the range of 250° to 300°C prior to mixing with cooler seawater.

The $\delta^{13}C$ and $\delta^{18}O$ values within syngenetic Cambrian hydrothermal carbonate at Rosebery are consistent with deposition from a hydrothermal fluid at a temperature between 100° and 300°C, with a $\delta^{18}O$ value of ~0 to 4‰, and with carbon derived from dissolved seawater bicarbonate and some oxidised carbon.

Zones of low $\delta^{18}O_{wr}$ values (<10‰) extend for only a few tens of metres into the footwall beneath the ore horizon, with a lateral extent of less than 100m. An elevated $\delta^{18}O_{wr}$ signature related to Devonian metamorphic and metasomatic fluids (11.5-16.5‰) appears to have overprinted and obscured isotopic patterns related to primary ore formation.

Strontium isotope values in barite mineralisation are consistent with the circulation of hydrothermal fluids through Precambrian crust onto which the Mount Read volcanics were erupted.

The Pb isotopes in primary sulphide ore at Rosebery are consistent with a Cambrian ore-forming event involving the leaching of lead from the Cambrian volcanic succession and Precambrian meta-sedimentary basement by ore-forming hydrothermal fluids.

The overall isotopic signatures associated with the mineralised system at Rosebery are comparable to other major syngenetic Cambrian ore deposits located in the Mount Read Volcanics. The Cambrian syngenetic deposits can be differentiated from Devonian epigenetic
deposits on the basis that:

- Hydrothermal carbonates in Cambrian systems exhibit lower $\delta^{18}O$ values and significantly higher $\delta^{13}C$ values than in Devonian systems,
- Cambrian mineralisation displays a lower radiogenic Pb isotope signature than Devonian deposits with $^{206}\text{Pb}/^{204}\text{Pb}$ ratio values typically more than $\sim1\%$ lower.
Chapter 9 : Discussion and Synthesis

9.1. Previous genetic models

Ore formation at Rosebery has been regarded by previous workers as either the result of epigenetic replacement during Devonian deformation and metamorphism (e.g. Aerden, 1991, 1992), or of processes occurring at or near the seafloor at a time broadly synchronous with deposition of the Cambrian succession (e.g. Brathwaite, 1974; Solomon and Walshe, 1979; Green et al., 1981; Huston and Large, 1988; Allen, 1994a) (fig.1.3). Processes related to a Cambrian timing for ore formation may involve massive sulphide formation by hydrothermal fluid venting and mineral precipitation, physical reworking of sulphides on the seafloor, replacement of host sediments or their lithified equivalent, replacement and chemical reworking of earlier ore mineral components, or a combination of these processes.

9.2. Devonian or Cambrian?

The first question to be resolved is whether primary mineralisation is the result of Devonian replacement or the result of Cambrian near-seafloor processes. The key evidence utilised by Stillwell (1934) and Aerden (1991; 1992; 1993) to support a Devonian replacement mode of ore replacement during the D3 deformation event included:

- A paragenetic replacement sequence of pyrite $\rightarrow$ quartz-chlorite $\rightarrow$ carbonate $\rightarrow$ late-stage pyrite $\rightarrow$ sphalerite + galena + chalcopyrite.
- The presence of cleaved host rock within undeformed mineralisation, and the truncation of S3 cleavage by sulphide overgrowth.
- Parallelism between ore shoots, ore banding, and cleavage, and discordance of the ore to bedding.
- The location of mineralisation along interpreted shear planes associated with macro-scale foliation boudinage structures that were inferred from micro- to meso-scale foliation boudinage structures mapped in outcrop, and from cleavage and shear orientations in mine plans and cross sections.
- The Cu-Fe enrichment of sulphide ore along down-dip margins of the southern ore lenses.

The sulphide paragenesis (table 5.2) and alteration paragenesis (table 4.1) as presented in this study differ markedly from that presented by Aerden (1992; 1993). This may be attributed to a failure by Aerden (1992; 1993) to recognise the relevance of colloform, spongiform and skeletal pyrite morphologies within the Rosebery ore, textures that clearly demonstrate early co-precipitation of pyrite, sphalerite, galena, chalcopyrite, electrum and sulphate minerals. Several features indicate that less competent sulphides such as sphalerite have been
recrystallised, including chalcopyrite disease along sphalerite grain margins and typically granoblastic textures even where these sulphides are intimately associated with relict primitive pyrite textures. The apparent truncation of S₃ cleavage by undeformed mineralisation is actually due to the recrystallisation of less competent sulphide minerals and destruction of most deformation-related fabrics contained therein. The preservation of uncompacted pumice textures within alteration zones clearly related to primary mineralisation (table 4.1), indicates that this alteration occurred prior to significant sediment compaction and subsequent Devonian deformation.

Parallelism between ore lenses, ore banding and cleavage is not ubiquitous. It was observed during this study that the ore lenses and banding locally parallel bedding rather than cleavage. Several possible modes of ore banding development have been discussed previously including: formation during sulphide deposition on the seafloor; preservation of primary sedimentary fabrics during sulphide replacement of host sediments as the result of mineralising fluids flowing laterally through permeable horizons within sediments and early sulphide mineralisation; differential sulphide remobilisation and mineral growth during deformation. Banding greater than a few centimetres in thickness is considered to be a primary feature of the ore, in most cases developed during sulphide replacement along permeable sediment and sulphide horizons. The development of fine-scale tectonic banding was related to later deformation. Shearing and attenuation of the ore during deformation would also have rotated any originally bedding-parallel banding closer to the main cleavage orientation.

The discordance of ore to bedding at a macro-scale indicates ore formation through replacement of the host succession, although alteration features indicate that this occurred within unconsolidated sediments during the Cambrian. At the meso- and micro-scale, discordance of ore to bedding can also be explained by chemical and physical remobilisation of primary ore components during deformation. Evidence for the deformation and remobilisation of ore observed during this study included piercement structures, syn-tectonic sulphide and quartz-carbonate-sulphide veins, massive sulphide within the axes of fault-related folding, and deformation-related fabrics within the primary ore itself. Remobilised sulphides are typically located no more than a few tens of metres away from primary ore.

Aerden (1991; 1992; 1993) mapped foliation boudinage structures within the mine at a micro- and meso-scale with both vertical and horizontal orientations. The cleavage orientation and secondary shears on several mine levels and cross sections were then mapped out, and from this macro-scale foliation boudinage structures with both horizontal and vertical orientations were inferred (fig.1.3). These structures were interpreted to control the distribution of mineralisation during ore formation. No other evidence of macro-scale boudinage structures was presented other than the distribution of pyrite-chalcopyrite mineralisation proximal to inferred shear ramps. Berry (1991) mapped meso-scale foliation boudinage structures in
underground exposures at the northern end of the mine that had a horizontal orientation, sub-perpendicular to the stretching direction, but could not locate any with a vertical orientation. The foliation boudinage structures mapped by Berry (1991) were associated with barren quartz-carbonate veining of the same generation as the syn-tectonic quartz-carbonate vein set found throughout the mine sequence, and were therefore interpreted to post-date primary ore formation. This quartz-carbonate vein set crosscuts sulphide ore that contains primitive sulphide textures, however these sulphide textures are not anywhere evident within the veins themselves, indicating a later timing for vein formation. The sulphide content of the vein set diminishes rapidly with distance from massive ore, indicating that the sulphide was locally remobilised from the primary ore. The pyrite-chalcopyrite distribution within the southern ore lenses might also be explained by a high temperature replacement or zone-refinement model associated with ore formation during the Cambrian.

Evidence of ore formation at or near the seafloor has been presented by Brathwaite (1969; 1974), Green et al. (Green et al., 1981) and Green (1983), Huston (1988) Huston and Large (1989) and Large (1990); and this study. Key evidence supporting Cambrian ore formation at or near the seafloor includes:

- The stratiform and stratabound nature of the ore lenses (Brathwaite, 1969, 1974; Allen, 1994a).
- Compositional banding of the ore (Brathwaite, 1969, 1974).
- Massive sulphide clasts occurring within the lowest mass flow unit of the Hangingwall Volcaniclastics on a regional basis, at localities including Bastyan Dam, West Hercules, Hercules and White Spur (Allen, 1994a).
- The asymmetric mineral zonation of ore lenses, typically comprising sphalerite-rich ore with a pyrite-chalcopyrite-rich base, and in turn overlain by barite-rich mineralisation (Brathwaite, 1969, 1974).
- The asymmetric distribution of alteration spatially associated with mineralisation (Brathwaite, 1969; 1974; and this study).
- The interlayering and interfingering of ore with massive carbonate alteration (Brathwaite, 1969, 1974).
- The spatial association of mineralisation with alteration that has preserved uncompacted pumice fragments, indicating that alteration occurred prior to significant compaction and lithification. Outside quartz-carbonate domains the phyllosilicate altered pumice fragments have been overprinted by all of the deformation-related cleavages (Allen, 1994a).
- Relict framoidal pyrite (Brathwaite, 1969, 1974) and other primitive textures within the ore including colloform, atoll, skeletal and spongiform pyrite morphologies (this study).
- Overprinting of the ore and associated alteration by deformation-related structures and fabrics (Brathwaite, 1969; 1974; and this study).
• The $^{34}\text{S}$ values associated with sulphide and barite mineralisation are consistent with sulphur derived from modified Cambrian seawater and the underlying volcanic succession (Green et al., 1981; Green, 1983), and are comparable to other deposits of Cambrian age within the Mount Read Volcanics (Solomon and Groves, 2000; and this study).

• The lead isotope signature of ore mineralisation is comparable to that of other deposits of Cambrian age within the Mount Read Volcanics (Gulson et al., 1987; Gulson and Porritt, 1987).

• The ore lenses demonstrate a spatial relationship with interpreted Cambrian faults (this study).

The stratabound nature of mineralisation is apparent at the regional to mine scale, but at the ore lens scale mineralisation may transgress stratigraphy. This is indicative of a replacement mode of ore genesis, but does not of itself indicate whether the timing was Cambrian or Devonian. As has been discussed, the compositional banding within sulphide and barite lodes is most likely a primary feature of mineralisation, and has been overprinted by subsequent deformation-related fabrics and structures. Clasts of massive sulphide within the basal part of the Hangingwall Volcaniclastics on a regional basis indicate that sulphide mineralisation was on or near the Cambrian seafloor at the same stratigraphic level as Rosebery during mass flow emplacement. This is an interpretation supported by the presence of other Cambrian ore deposits at approximately the same interpreted stratigraphic position, including Hercules, South Hercules, Hellyer and Que River. There is, however, a lack of sulphide clasts within the Hangingwall Volcaniclastics immediately overlying and adjacent to Rosebery.

Although much of the Rosebery ore exhibits a mineral zonation that can be considered analogous to other VHMS systems such as Kuroko and Kidd Creek, in places this sequence is inverted. The typical vertical zonation present within most of the ore lenses at Rosebery probably reflects a stratigraphic orientation at the time of Cambrian ore formation, however it is not diagnostic of a Cambrian origin. As macro-scale folding of the ore is absent, the inverted zonation within G lens could be indicative of a replacement mode of ore formation, or of a fluctuation in physicochemical factors controlling ore deposition on the seafloor.

The Rosebery ore-related alteration assemblages and their zoning may be considered analogous to those within other VHMS systems such as Kuroko and Kidd Creek. The spatial relationship between the primary alteration assemblages and the ore indicates a genetic link, however it is the preservation of uncompacted pumice textures within quartz and carbonate domains that confirms a Cambrian timing of ore formation. The interlayering and interfingering of sulphide ore with massive carbonate further supports the close spatial association of alteration and ore. Colloform textures are present in both the sulphide and carbonate, indicating mineral precipitation into open spaces.
Pyrite framboids within the ore and associated alteration assemblages are not necessarily diagnostic of a Cambrian origin, however the presence of framboids in weakly altered host rocks distal to ore confirms that development of the framboids occurred during sediment deposition and/or diagenesis. This origin is further supported by similar framboid size distributions within both the ore and nearby sedimentary rocks that lack ore-related mineralisation. Other primitive textures within the ore identified during this study, including colloform, atoll, skeletal and spongiform pyrite morphologies, indicate that pyrite precipitated rapidly in association with other sulphide and sulphate minerals. These textures are not consistent with a deformation-related origin.

As discussed previously, the alteration assemblages associated with primary mineralisation have been overprinted by all the deformation events evident within the mine sequence. Massive mineralisation contains abundant evidence to indicate a significant degree of deformation and remobilisation, although late-stage annealing of the less competent sulphide component appears to have destroyed most micro-fabrics. This indicates that ore mineralisation predates the D₃ deformation event, and almost certainly predates all Devonian deformation.

The $\delta^{34}$S signature of Rosebery mineralisation is consistent with the derivation of sulphur from modified Cambrian seawater, combined with a sulphur component leached from the rocks through which the hot hydrothermal fluids passed. The pattern of isotopic zonation at the mine scale is also consistent with mixing of a hydrothermal fluid with increasing amounts of locally circulating seawater away from the primary upflow zone. The range of $\delta^{34}$S values is consistent with other Cambrian massive sulphide deposits located at the same stratigraphic level as Rosebery, including Hellyer and Que River. Aerden (1993) however, argued that the sulphur within Rosebery mineralisation may have been derived through remobilisation of disseminated Cambrian volcanic sulphur during Devonian metamorphism. The sulphide $\delta^{34}$S values associated with Devonian granite-related ores of western Tasmania appear to be controlled to a significant degree by their host rocks, and their $\delta^{34}$S value distributions overlap with those of Cambrian ores. Deposits of Cambrian and Devonian age within the Mount Read Volcanics unfortunately cannot be readily differentiated on the basis of their sulphur isotope signature.

The Pb isotope signature of Rosebery mineralisation is comparable with that of sulphide deposits in western Tasmania of Cambrian age, and is significantly less radiogenic than the Devonian granite-related ores. Aerden (1991) argued that the Rosebery Pb isotope signature may have been derived through leaching of Pb from the Cambrian volcanic succession during metamorphism and deformation. After emplacement of the volcanic succession, radioactive decay of contained U and Th would have continued until the onset of Devonian metamorphism and deformation, thereby increasing the radiogenic signature of contained Pb. Devonian
deformation-related ore derived from disseminated mineralisation within the Cambrian volcanics should therefore have a more radiogenic signature than the Cambrian ore deposits. Rosebery is actually less radiogenic than the Hellyer and Que River deposits, demonstrating that the Pb isotope signature is more consistent with a Cambrian age of primary ore formation.

A Cambrian age of primary ore formation is further indicated by the strong spatial relationship observed between the ore lenses and Cambrian fault structures interpreted during this study (fig. 5.1). It has also been demonstrated during this study that the zonation of metals within ore lenses is also spatially related to the Cambrian faults, in a fashion consistent with the faults having acted as fluid conduits.

Some of the evidence presented is indicative of a replacement mode of ore genesis, but does not indicate the timing of ore formation. The evidence for a Devonian timing of primary ore formation relies almost entirely on microstructural observations that can be discounted if the less competent ore minerals were annealed during deformation. There is sufficient evidence to indicate that this did occur. The weight of evidence is strongly in favour of primary ore formation at or near the Cambrian seafloor, with subsequent chemical and physical remobilisation during Devonian seafloor metamorphism and deformation.

9.3. At or below the seafloor?

The genetic models proposed for formation during the Cambrian place the site of mineralisation either at or beneath the seafloor. Doyle and Allen (2003) proposed several features that might be considered diagnostic or indicative of seafloor ore deposition including the presence of:

- Sedimentary clastic sulphide textures.
- Sulphide chimney textures.
- Exhalite at the ore horizon.
- Fossilised tubeworms and bivalves.
- Facies characteristics indicating a slow accumulation rate for the host rocks, and/or the occurrence of ores between host rock units that were rapidly emplaced.
- An asymmetric alteration pattern of strong footwall alteration and weaker hanging-wall alteration (indicative).

No clasts of sulphide or mineralised rock fragments have been observed within the turbidite and mass flow units that comprise the TSV and Hangingwall Volcaniclastics at Rosebery. Rare clasts of aphyric, occasionally amygdaloidal rhyolite are observed within both the ore and surrounding sedimentary units. These clasts are considered to be a primary clastic component of the mass flow and turbidite units that were preserved during sulphide replacement of the host sediments. The vast majority of lithic material within massive sulphide ore was not
incorporated through sedimentary processes. Most lithic material within the ore comprises
fragments that are intensely altered, have a ubiquitous cleavage development and lack
primary coarse-grained clastic textures. This lithic material is interpreted as having been
derived through the deformation-related disruption and rotation of fine-grained sediment bands
preserved after sulphide replacement of adjacent coarser and more permeable horizons within
the volcaniclastic sediments, and through the physical incorporation of adjacent host rock
material during deformation. Albite grains within sphalerite-rich and barite mineralisation are
also interpreted to be a relict clastic component of the volcaniclastic sediments preserved
during sulphide and barite replacement of surrounding vitriclastic material. This is supported by
the observation that the size, morphology and spatial distribution of albite grains within
massive sulphide ore are essentially the same as the size, morphology and distribution of
albite grains within nearby unmineralised volcaniclastic sediments. If albite grains had been
incorporated into the ore during turbiditic flow of sulphide material on the seafloor, they should
have been concentrated along low-density horizons within sulphide ore due to gravitational
sorting and a large density differential. This has not been observed.

Compositional banding has been interpreted by some workers to have formed through settling
of precipitated sulphides onto the seafloor (e.g. Brathwaite, 1969; Brathwaite, 1974). As
discussed in chapter 5, diffuse banding may instead be related to the preservation of
sedimentary fabric during sulphide replacement of sediments. The more intense compositional
banding, observed mainly within the Cu-rich southern ore lenses, may be attributed to the
lateral flow of hotter mineralising fluids through permeable horizons within sediments and early
low temperature mineralisation. Some compositional banding can also be related to
def ormation and sulphide remobilisation. No other sedimentary textures were observed within
the sulphide ore that might indicate sedimentary reworking of the sulphides, and no textures
were evident within the ore that might be interpreted as chimneys or the reworked fragments of
chimneys. There are no units that could be readily interpreted as an exhalite extending
laterally away from the Rosebery ore. Although there may be some indication of bioturbation
within the TSV (Crawford, 1998), there are no indications of fossils associated with the
mineralisation.

Volcaniclastic sediments of the footwall volcanics and TSV were rapidly emplaced,
predominantly as mass flow and turbidite units. For ore to develop at the seafloor,
vulcaniclastic sedimentation must have been significantly reduced relative to the rate of ore
deposition, as appears to have been the case with the modern Bent Hill massive sulphide
deposit (Zierenberg and Miller, 2000). Alternatively, the turbidites flowed around ore that
comprised part of a topographic high. If volcaniclastic sedimentation had ceased for a period
long enough to allow significant ore development, the development of carbonaceous mudstone
or an exhalite might be expected lateral to ore. If instead volcaniclastic material had flowed
around a topographically elevated sulphide mound, sulphide or mineralised lithic clasts should
have been deposited at the ore margins or entrained into the flow units. No evidence was observed that supports either of these scenarios. In places the ore lenses transgress bedding, something interpreted to be a primary feature. Some ore, for example K lens, occurs within the top of a footwall volcaniclastic mass flow unit. For mineralisation to occur within a unit that is deposited instantaneously in comparison with likely rates of ore formation, a replacement mode of ore formation is necessary. Replacement of sedimentary units by sulphide ore is strongly supported in the case of K lens, where analyses collected as part of this study clearly demonstrate that the semi-massive to massive sulphide ore has the same immobile element signature as the host pumice breccia units (fig. 5.2).

The halo of alteration around Rosebery ore is asymmetric at the mine scale, extending for hundreds of metres into the footwall volcanics and to a maximum of tens of metres toward the base of the Hangingwall Volcaniclastics. The upward extent of alteration is more commonly less than ten metres, except where a system of stacked lenses is present. This asymmetry may be interpreted to indicate widespread hydrothermal fluid flow within the footwall volcanics during formation of ore at the seafloor, followed by limited circulation and weak alteration of the subsequently deposited sediments as the hydrothermal system waned. Alternatively, the asymmetrical alteration pattern is the result of diffuse hydrothermal fluid upflow through unconsolidated sediments of the footwall volcanics and TSV, combined with a significant component of lateral fluid flow beneath relatively impermeable peperitic sills and fine-grained volcaniclastic units at the top of the footwall volcanics and within the TSV.

Doyle and Allen (2003) also proposed several features that might be considered diagnostic or indicative of sub-seafloor ore deposition including the presence of:

- Relics of host rock within the sulphide deposit.
- Facies characteristics indicating very rapid emplacement of the host lithofacies.
- Identification of replacement fronts between the sulphide deposit and host.
- Discordance of the ore with the enclosing lithofacies (indicative).
- The presence of strong hanging-wall alteration, similar in style and intensity to the footwall alteration (indicative).

Semi-massive to massive mineralisation at Rosebery typically contains an intensely altered lithic component comprising bands and rotated fragments that are identical to the surrounding host rock. These are interpreted to be the remnants of volcaniclastic units replaced during mineralisation. The preferential mineralisation of coarser horizons within well-stratified parts of the footwall volcanics and TSV indicates that fine-grained material was more likely to have been preserved within the ore. This may in part explain why no lithic fragments with coarse-grained textures have been preserved within massive ore, although intense alteration may also have been a factor in the destruction of coarse-grained textures within lithic material.
Volcaniclastic sediments hosting the mineralisation at Rosebery predominantly comprise mass flow and turbidite units that were rapidly emplaced on the seafloor. The ore lenses transgress stratigraphy in places, and mineralisation in some cases (e.g. K lens) occurs within the top of a mass flow unit, indicating primary ore formation within the host succession. Several volcaniclastic and early diagenetic features have also been preserved within massive ore, further supporting a replacement mode of ore formation. For example, albite phenocrysts preserved during early low-temperature sulphide mineralisation exhibit the same size, morphology and distribution within the massive ore as those within the surrounding sediments. Quartz grains are similarly preserved within sulphide and barite mineralisation hosted by quartz-crystal-bearing sediments of the TSV. Rutile within massive sulphide ore also retains a small-scale distribution related to the alteration and replacement of volcanic hornblende (after augite?) (fig. 5.2), a texture also observed within unmineralised volcaniclastic sediments. Diagenetic pyrite framboids and overgrowths developed within volcaniclastic sediments during diagenesis and were overprinted by subsequent sulphide and barite mineralisation.

The quartz-feldspar peperite sill overlying K lens, the stratigraphically deeper part of P lens and a significant part of AB lens, was intruded into weakly altered, unconsolidated volcaniclastic sediments of the footwall volcanics and TSV. Emplacement occurred prior to deposition of the basal carbonaceous mudstone at the base of the Hangingwall Volcaniclastics, as evidenced by the presence of hyaloclastite and thin turbidite units derived from the sill that extend to the base of the Black Slate. The sill may have been disrupted to some degree by the influence of Cambrian faults located between P lens and K lens, but it was not offset any significant amount despite large changes in TSV sediment thickness. This indicates that intrusion probably occurred after movement along these faults had effectively ceased, and most likely after the majority of the TSV had been deposited. The sill contains rare intensely altered rock fragments and abundant weakly altered sediment clots, but no sulphide or mineralised lithic material. The peperitic base of the sill is overprinted by intense ore-related alteration and disseminated sulphide mineralisation, whereas the peperitic top of the sill and overlying sediments are relatively unaltered. Where the sill is attenuated or absent, intense hydrothermal alteration extends higher within the TSV. This evidence, collected primarily during the current study, strongly indicates that intrusion occurred prior to the onset of significant hydrothermal alteration and mineralisation. The clear implication is that the northern ore lenses were formed within volcaniclastic sediments beneath the peperitic sill. The base of a thick fine-grained volcaniclastic package overlying P lens is intensely altered in an analogous fashion to the base of the peperitic sill. It also contains a pre-deformation pyritic and baritic vein set interpreted to have developed synchronous with primary ore formation and characterised by $\delta^{34}$S values similar to those in the primary sulphide and barite lodes. It is therefore likely that the fine-grained units were also emplaced prior to ore formation, in which case the stratigraphically higher parts of P lens also developed beneath the Cambrian seafloor.
Alteration and replacement fronts developed within coherent rocks or compacted sediments with relatively low permeability are likely to be well defined. This occurs at Hellyer, where a well-defined sequence of alteration assemblages is tightly focussed about a fault traversing coherent lavas in the immediate footwall to massive sulphide (Gemmell and Large, 1992). In contrast, alteration fronts developed within permeable unconsolidated sediments are likely to be more extensive and gradational in nature as a result of diffuse hydrothermal fluid flow, as is the case beneath sulphide mounds within the sediment-covered Middle Valley system (Turner et al., 1993). An alteration front comprising disseminated sulphide mineralisation and strong to intense quartz ± sericite alteration extends for hundreds of metres laterally away from massive ore lenses at Rosebery, for tens of metres into the footwall, and may extend for several metres into the hanging-wall. Descriptions available from previous studies (Brathwaite, 1969, 1974; Green et al., 1981; Green, 1983) indicate that strong sericite-quartz alteration extends into the hanging-wall above the southern ore lenses, whereas both strong sericite-quartz alteration and intense quartz alteration extend into the hanging-wall above the northern ore lens. Quartz-dominant alteration of the basal contact of the peperitic sill overlying K lens, and of the basal part of the fine-grained volcanioclastics overlying P lens, is of similar intensity to the quartz-dominant alteration within sediments that form the “quartz schist” located stratigraphically below and lateral to massive sulphide ore. The alteration within the base of the peperitic sill is sufficient in places to destroy peperitic textures and phenocrysts within the intrusive. Primary mineralisation and proximal alteration at Rosebery are orientated sub-parallel to stratigraphy, consistent with significant lateral hydrothermal fluid flow within sediments beneath the peperitic sill and fine-grained sediments. The principal replacement front associated with ore formation is therefore located lateral to the ore rather than at the upper or lower margins. At their lateral margins, the massive sulphide ore lenses typically grade through disseminated sulphide mineralisation into massive carbonate or intensely altered host rock. Massive ore typically grades down into disseminated mineralisation within the footwall, whereas the upper ore lens margin typically exhibits a sharp contact with unmineralised host rocks. This distribution of alteration and mineralisation is consistent with lateral fluid flow through permeable horizons within volcanioclastic sediments that are underlain by relatively permeable footwall units and overlain by impermeable hanging-wall units.

There is no evidence within the ore and host succession at Rosebery that is diagnostic of sulphide ore having been present at the seafloor during the Cambrian, although the possibility that some ore lenses formed at the seafloor cannot be dismissed. Evidence collected during this study, including the timing of sill emplacement relative to sulphide ore formation and the similarity of sulphide ore and host sediment immobile element signatures, is strongly indicative of primary ore formation within sediments beneath the Cambrian seafloor.
9.4. Regional-scale hydrothermal circulation

The driving force for regional hydrothermal circulation related to formation of the Rosebery ore deposit is unclear because, due to basal truncation of the Rosebery mine succession by the Rosebery Fault (Corbett and Lees, 1987) and emplacement of a large Devonian granite beneath the Rosebery mine (Leaman and Richardson, 1989), there is uncertainty as to what geological components originally underlay the footwall volcanics. A regional-scale reconstruction of the Rosebery area by Berry (1993a) suggests that ~1.3km of Central Volcanic Complex and ~2km of Crimson Creek Formation overlay Precambrian basement beneath Rosebery during the Cambrian. The Central Volcanic Complex beneath Rosebery most likely comprised ~800m of the footwall volcanics, in turn underlain by more coherent dacitic and rhyolitic facies comprising lavas, domes and sills of the Mount Black Formation (Gifkins, 2001). The Central Volcanic Complex was underlain by the Crimson Creek Formation, a succession of mafic volcaniclastic turbidites and mudstones, and associated basaltic lavas (Brown et al., 1989), most likely in association with slabs of fault-emplaced ultramafic-mafic complex (Berry and Crawford, 1988). Precambrian basement beneath the Crimson Creek Formation most likely comprised sediments and metasediments of the type now exposed in the Rocky Cape and Tyennan regions (Corbett, 1992; Solomon and Groves, 2000).

The magmatism that produced the volcaniclastics and peperitic sills of the mine succession was also most likely responsible for providing the heat that drove regional hydrothermal circulation, probably in the form of large-scale sub-volcanic intrusions located within the succession onto which the footwall volcanics were deposited. The presence of intrusives throughout the Rosebery-Hercules host sequence that both pre-date and post-date mineralisation indicates that magmatism occurred throughout the life of the hydrothermal system. The line of large-scale granitic intrusions located within the Central Volcanic Complex along the eastern margin of the Mount Read Belt may have played a role (fig. 2.1), but the timing of emplacement relative to the lower Tyndall Group suggests that they probably post-dated Rosebery ore formation (White and McPhie, 1996). The depth of magmatic emplacement beneath Rosebery was probably in the range of 3 to 7km, similar to that of the Darwin and Murchison granites (Large et al., 2001c; Wyman, 2001).

In the context of regional hydrothermal circulation related to Rosebery ore formation, parallels may be drawn with processes occurring within modern sediment-covered seafloor systems such as Middle Valley, located along the Juan de Fuca Ridge. Cooling of oceanic crust without sedimentary cover is primarily a convective rather than conductive process, due to the ready influx of cold seawater into the top few hundred metres of the crust (Alt, 1995a). Sediment cover of a few tens of metres thickness in the Middle Valley area significantly restricts access of overlying seawater to oceanic crust, compresses geothermal gradients near the seafloor.
and promotes the advective transport of heat and fluid over large distances (~20km) within the basement (Humphris, 1995; Stein and Fisher, 2001). With increased turbiditic sediment thickness (>160m) direct exchange of seawater between basement and the overlying ocean ceases, although uniform temperatures are maintained at the basement-sediment interface due to continuing circulation within the basement (Humphris, 1995). Seawater recharge of modern hydrothermal systems occurs via the slow flow of seawater through sedimentary cover in combination with rapid fluid down-flow along faults such as those found at axial valley margins (Koski et al., 1988; Alt, 1995a; Stakes and Schiffman, 1999). After emplacement of the footwall volcanics of the Rosebery succession onto the Cambrian seafloor, direct access of seawater to the underlying stratigraphy would have been significantly restricted. Hydrothermal recharge at a regional scale would then have been largely confined to large-scale basin-margin or caldera-related faults, or to areas without significant sediment cover. A thick, laterally extensive volcaniclastic cover provided by the laterally extensive footwall volcanics would have promoted significant lateral hydrothermal fluid flow within the underlying rocks. The depth of hydrothermal circulation is uncertain, however Pb and Sr isotope signatures associated with Rosebery mineralisation indicate that a considerable amount of hydrothermal circulation occurred within Precambrian basement rocks.

In the modern environment, basement highs appear to be responsible for elevated heat flux at the seafloor due to reduced sediment cover, and therefore promote the upflow of hydrothermal fluids along the basement-sediment interface (Alt, 1995a; Goodfellow et al., 1999; Stakes and Schiffman, 1999; Stein and Fisher, 2001). Faults penetrating the basement and overlying sedimentary succession may also accommodate fluid upflow (Hannington et al., 1995; Humphris, 1995; Stakes and Schiffman, 1999). Insufficient data exists to ascertain whether a basement high was present beneath Rosebery, however faulting is evident within the Cambrian succession (Berry and Keele, 1993; and this study). Vertical heterogeneity of permeability within sediment-sill complexes may also impose a strong lateral component on hydrothermal flow at a more local scale (Stakes and Schiffman, 1999).

### 9.5. Local hydrothermal circulation

The distribution of ore lenses in long section suggests that hydrothermal fluid upflow at Rosebery was primarily controlled by Cambrian fault structures (fig. 5.1), probably comprising discrete fault zones within competent basement rocks and compacted sediments that extended upward as broad zones of disruption within overlying unconsolidated sediments. Vertical fluid flow would therefore have become more diffuse in nature as the fluids passed upward into the footwall volcanics and mixed with locally circulating seawater. The ready access of seawater to only the unconsolidated upper portion of some modern sediment-covered seafloor systems (e.g. Humphris, 1995) would indicate that significant local influx and circulation of seawater within the Rosebery host succession was restricted to depths of less
than 200m below the seafloor. The stratiform nature of mineralisation and alteration at Rosebery is consistent with significant lateral fluid flow at the top of the footwall volcanics that was promoted by sediment stratification and peperitic sills. The early development of stratiform diagenetic or hydrothermal alteration zones would have further promoted lateral fluid flow.

At the mine scale, fluid pathways related to hydrothermal fluid upflow are difficult to locate. The only clearly defined fluid conduit identified during this study underlies P lens (figures 4.6 & 4.7). It comprises a transgressive quartz-rich alteration zone that corresponds spatially to an interpreted northeast trending Cambrian normal fault. Elevated Cu and Fe values within the P lens sulphide ore occur proximal to the fault trace (fig.9.1), here interpreted to indicate a zone of elevated temperatures (>280-300°C). An interpreted northeast trending fault appears to have been the locus of hydrothermal fluid upflow linked to formation of K lens ore, however the drill coverage across the lower levels of K lens is not sufficient to confirm the presence of a transgressive alteration zone. Indirect evidence, in the form of elevated Cu and Fe values within K lens sulphide ore proximal to the fault, suggest that this fault was the locus for upwelling hydrothermal fluids (fig.9.2).

P lens comprises less than ~1.5 million tonnes of ore within a series of stacked lenses. The transgressive zone of intense quartz alteration underlying the P lens is ~20m to 30m across, trending laterally beneath the ore for at least 200m and possibly for up to 350m. The Hellyer ore body by comparison, comprises ~17 million tonnes of high grade sulphide ore located stratigraphically above a stringer alteration zone that contains a core of intense quartz alteration ~140m across and extending discontinuously over ~750m of strike (Gemmell and Large, 1992). If the Hellyer system were to be scaled down by a factor of 10, the dimensions of the ore and underlying fluid conduit would therefore be comparable to those observed for P lens at Rosebery. The dimensions of feeder zones associated with the other ore lenses at Rosebery are not likely to be significantly larger than the one located beneath P lens, and might therefore be difficult to locate without close-spaced drilling.

The southern ore lenses, in particular C lens, D lens and E lens, are relatively Fe-Cu enriched in comparison with the northern ore lenses, containing more significant zones of pyrite-chalcopyrite mineralisation both beneath and lateral to massive sphalerite-galena ore (fig.9.3). Pyrite within the footwall of the southern lenses exhibits higher Co, Ni and Co:Ni ratio values in comparison with pyrite from the northern ore lenses, indicative of higher temperatures of formation. The Fe-Cu enriched southern ore lenses also exhibit lower δ³⁴S values than most of the northern ore lenses (figures 8.2 & 8.3), due either to the mixing of upwelling hydrothermal fluids with lesser amounts of locally circulating seawater, or to the overprinting of early low-temperature Zn-Pb-rich mineralisation by a phase of high-temperature Fe-Cu mineralisation characterised by low δ³⁴S values. Together these observations indicate that the southern ore lenses formed at overall higher temperatures than the northern ore lenses. This is consistent
Figure 9.1 Hydrothermal fluid pathways - P lens.
The figure shows the trace of an inferred Cambrian fault and proposed hydrothermal fluid pathway for P lens. Refer to figure fig. 5.1 for inferred fault trace in long section and figure 4.17 for alteration distribution.

Figure 9.2 Hydrothermal fluid pathways - K lens.
The figure shows the trace of an inferred Cambrian fault and proposed hydrothermal fluid pathway for K lens. Refer to figure fig. 5.1 for inferred fault trace in long section and figure 4.17 for alteration distribution.
with the southern ore lenses forming adjacent to what was probably the principal site of hydrothermal upflow at Rosebery, focussed along the intersection of two northeast and southwest trending mine-scale faults (fig.5.1). The zone of hydrothermal upflow is defined, in part, by the elevated Co in footwall pyrite halo located proximal to the southern ore lenses (fig.7.11).

The inability of previous workers to identify feeder zones beneath sulphide lenses may be in part attributed to limited drill hole coverage outside of the massive ore, and to a bias toward a genetic model requiring that hydrothermal fluid upflow should have produced discrete, laterally zoned, transgressive alteration zones with abundant veining of the type found beneath Hellyer (Gemmell and Large, 1992). The paucity of ore-related veining within the Rosebery footwall probably reflects an environment in which the hydrostatic pressure of mineralising fluids never exceeded the lithostatic pressure of surrounding unconsolidated volcaniclastic sediments. With localised decreases in permeability of the volcaniclastic sediments caused by the development of stratiform zones of hydrothermal alteration and mineralisation, the hydrothermal fluids continued to flow laterally through adjacent permeable horizons within the volcaniclastic sequence.

9.6. Mineralisation

Estimates for the size of significant deposits on sediment-free mid-ocean ridges range from 1 to 5 million tonnes whereas, in an area of sediment-covered oceanic crust located along the
Juan de Fuca Ridge, the Middle Valley sulphide deposit may be as large as 50 to 100 million tonnes (Herzig and Hannington, 1995). The development of large sulphide accumulations within areas of sediment-covered oceanic crust, such as at Middle Valley, may be attributed to several factors. Oceanic crust free of sediment cover is prone to rapid cooling due to the ready influx of cold seawater, whereas with increasing sediment cover and reduced seawater influx the resultant heat loss is significantly reduced (Goodfellow and Blaise, 1988; Koski et al., 1988; Humphris, 1995). A reduced rate of heat loss would in turn prolong the life of a hydrothermal system (Goodfellow and Blaise, 1988), thereby allowing more time for sulphides to accumulate. By restricting the influx of overlying seawater, thick sedimentary cover can also lead to a hydrothermal system with lower water:rock ratios, longer fluid residence times, higher temperatures and a greater chemical exchange (Humphris, 1995), leading to more efficient extraction and transport of metals by the hydrothermal fluid.

Sulphide precipitation may take place within sediments as the result of mixing of the hydrothermal fluid with pore water, and/or as the result of reaction with and replacement of host sediments (von Damm et al., 1985; Goodfellow and Blaise, 1988; Herzig and Hannington, 1995). Where hydrothermal fluids vent directly onto the seafloor more than 90% of the total mass flux may be lost to a diffuse hydrothermal plume (Herzig and Hannington, 1995). Diffuse lateral flow of a low-velocity hydrothermal fluid through sediments beneath the seafloor provides more opportunity for mixing of a hydrothermal fluid with locally circulating seawater, more opportunity for the fluids to react with host sediments, and a commensurate increase in the amount of metal deposited from solution. Diagenetic alteration of sediments surrounding some Kuroko deposits may have resulted in an increase in pore water pH to ~8.9 (Iijima, 1974). It is therefore possible that sediment pore waters mixing with the hydrothermal fluid beneath the Cambrian seafloor at Rosebery were also of a significantly higher pH than the overlying seawater. Similarly, the reaction of hydrothermal fluids with feldspar-clay assemblages and organic carbon in the sediments may have led to a significantly increased fluid pH (von Damm et al., 1985; Goodfellow and Blaise, 1988). This may, in part, explain why albite grains were preserved within early low temperature sulphide and barite mineralisation at Rosebery, a feature expected of near neutral rather than more acid conditions.

Sulphide mounds such as the Middle Valley and Bent Hill deposits are interpreted to have precipitated initially as permeable sulphide deposits on the seafloor during venting of metalliferous fluids, with subsequent replacement and recrystallisation (zone refinement) of the deposit core, and precipitation of low temperature minerals at the outer margins of the mound as mineralising fluids permeated through the early sulphide accumulation (Hannington et al., 1995). Formation of the Kuroko deposits was first attributed to the same process by Eldridge et al. (1983), and at a smaller scale a similar process may be responsible for sulphide chimney growth on the modern seafloor (Graham et al., 1988; Hannington et al., 1995). Huston (1988) and Huston and Large (1989) originally interpreted the vertical mineral zonation within the
Rosebery ore lenses to be the result of zone refinement processes associated with vertical high-temperature fluid flow through permeable sulphide ore. However, lateral mineral zonation within the ore lenses is, in places, at least as pronounced as vertical mineral zonation (fig 5.4). The lateral flow of high temperature Cu-Fe-enriched fluids through permeable horizons within sediments and sphalerite-rich mineralisation is interpreted to have produced the lateral metal zonation within Rosebery ore by a process of zone refinement. It may also have been responsible for development of the strong compositional banding within the Cu-rich southern ore lenses. Limited upward permeation of high-temperature fluids into overlying sphalerite-rich ore at Rosebery resulted in basal Cu-Fe enrichment of the sphalerite-rich ore.

Evidence gathered during this study indicates a higher temperature overprint on initial low-temperature mineralisation that is reflected in the morphologies and trace element content of K lens sulphide minerals. Initial low-temperature sphalerite-galena-rich mineralisation is characterised by primitive sulphide textures, including spongiform, atoll, colloform and skeletal pyrite morphologies. The low temperature trace element signature associated with relict primitive sulphide morphologies is defined by: elevated levels of Mn-Ni-As-Ag-Sb-Au-Tl-Pb and low levels of Co-Bi in pyrite; low Co:Ni and Bi:Pb ratio values in pyrite; low levels of Cu-Fe and elevated levels of Zn-Ag-Sb-Pb in sphalerite; and low levels of Sb-Ag in galena. The initial low temperature mineralisation was locally permeated by higher temperature fluids, leading to more euhedral and crystalline phases of pyrite, pyrite overgrowths on the primitive sulphide textures and an increase in chalcopyrite. The higher temperature trace element signature associated with the more crystalline pyrite morphologies within massive ore is defined by: elevated levels of Bi-Sn, moderate levels of Ni-Ag-Sb-Tl-Pb-Co, and low levels of Mn-As-Au in pyrite; elevated Bi:Pb ratio values and moderate Co:Ni ratio values in pyrite; an increased Fe content in sphalerite; and galena with elevated levels of Sb-Ag. Barite clots within sulphide ore are interpreted to be the result of dissolution of early barite bands and veins by higher temperature fluids. An albitised volcanic plagioclase component, initially preserved within low temperature barite and sulphide mineralisation (e.g. P lens), was also subsequently destroyed with the advent of higher temperature and more acid fluids.

The discontinuous stratiform zones of chlorite alteration and associated pyrite-chalcopyrite mineralisation that underlie most sulphide lenses were produced as high temperature hydrothermal fluids flowed laterally within zones of higher permeability immediately beneath the ore lenses. These zones may be analogous to the Deep Copper zone underlying the Bent Hill and ODP sulphide deposits in the Middle Valley area, produced as a result of high temperature fluid flowing laterally through a more permeable sandy turbidite horizon (Goodfellow et al., 1999). In the footwall to massive sphalerite-galena K lens ore this high temperature mineralisation exhibits predominantly crystalline pyrite morphologies, including anhedral to euhedral grains and aggregates. The associated high temperature trace element signature is defined by: high levels of Co-Bi in pyrite; elevated 3 and Bi:Pb ratio values in
pyrite; and high levels of Fe-Mn-Cu-Sb-Bi in sphalerite.

Analogous processes of sub-seafloor zone refinement and replacement have been interpreted for the ancient Kidd Creek deposit (Hannington et al., 1999a) and for recent mineralisation at Site 856, Middle Valley (Krasnov et al., 1994).

9.7. The hydrothermal fluid

The sulphur isotope signature associated with Rosebery mineralisation indicates that the hydrothermal fluid was derived from modified seawater, and the Pb isotope signature within Rosebery ore indicates that the metal component was probably leached from the Cambrian volcano-sedimentary succession and underlying Precambrian basement. Stolz and Large (1992) suggested that the high Au content of Rosebery ore might have been derived through the high-temperature leaching of an Au-enriched package within the strata through which the hydrothermal fluids had passed. Basaltic lavas within the Crimson Creek Formation were suggested as the Au-enriched source rocks, as they were found to have significantly higher background Au contents (to 23ppb) than Precambrian basement rocks (<0.6ppb) or the overlying volcanic and sedimentary rocks of the Mount Read Volcanics (to 5.5ppb and 7.3ppb respectively). Stolz and Large (1992) also conceded that significantly Au-rich source rocks were not necessary for the formation of Au-enriched sulphide ore. If conditions were such that the hydrothermal fluid became saturated at relatively low Au concentrations, they would be unable to extract all available Au from the Au-rich source rock. In this case the presence of an Au-rich source rock would not have a significant impact on the Au content of sulphide ore. In fact, the relatively low Cu and elevated Zn-Pb content of Rosebery ore suggests that felsic rather than mafic rocks were the predominant contributor of metals to the Rosebery system. The average metal content of Rosebery ore is comparable to modern massive sulphide deposits forming within a back-arc ridge setting (table 9.1).

<table>
<thead>
<tr>
<th></th>
<th>Mid-ocean ridges</th>
<th>Back-arc ridges</th>
<th>Rosebery</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Volcanic-hosted</td>
<td>Sediment-hosted</td>
<td>Intraoceanic</td>
</tr>
<tr>
<td>Fe (wt%)</td>
<td>26.4</td>
<td>24</td>
<td>13.0</td>
</tr>
<tr>
<td>Zn (wt%)</td>
<td>8.5</td>
<td>4.7</td>
<td>16.5</td>
</tr>
<tr>
<td>Cu (wt%)</td>
<td>4.8</td>
<td>1.3</td>
<td>4.0</td>
</tr>
<tr>
<td>Pb (wt%)</td>
<td>0.1</td>
<td>1.1</td>
<td>0.4</td>
</tr>
<tr>
<td>As (wt%)</td>
<td>0.0235</td>
<td>.3</td>
<td>0.0845</td>
</tr>
<tr>
<td>Sb (wt%)</td>
<td>0.0046</td>
<td>.06</td>
<td>0.0106</td>
</tr>
<tr>
<td>Ba (wt%)</td>
<td>1.8</td>
<td>7</td>
<td>12.6</td>
</tr>
<tr>
<td>Ag (ppm)</td>
<td>113</td>
<td>142</td>
<td>217</td>
</tr>
<tr>
<td>Au (ppm)</td>
<td>1.2</td>
<td>0.8</td>
<td>4.5</td>
</tr>
<tr>
<td>n</td>
<td>1259</td>
<td>57</td>
<td>573</td>
</tr>
</tbody>
</table>

Table 9.1 Metal contents - Rosebery vs modern hydrothermal sulphides.

Bulk chemical composition of seafloor polymetallic sulphides from modern mid-ocean ridges and back-arc spreading centres in comparison with Rosebery ore. (data from Herzing and Hannington, 1995; Herzig et al., 2002; P. Edwards pers. comm., 2002; and this study - * approximations based on various mine datasets, - values not determined)
As part of a previous study, Huston and Large (1989) carried out thermodynamic modelling of a proposed hydrothermal fluid based on earlier work by Large (1977), Eldridge et al. (1983) and Pisutha-Armond and Ohmoto (1983), with a particular focus on possible modes of Au transport and distribution within the Rosebery system (fig. 9.4). Assumptions made about the fluid modelled by Huston and Large (1989) were based in part on observations made by Brathwaite (1969; 1974), Dixon (1980), Green et al. (1981) and Huston and Large (1988) which included: (1) the presence of chlorite-sericite alteration within the footwall volcanics with minor carbonate, (2) trace amounts of arsenopyrite and pyrrhotite, and only minor amounts of carbonate in the chalcopyrite-pyrite ore, (3) trace amounts of arsenopyrite and significant amounts of carbonate within the sphalerite-galena-pyrite ore, (4) the presence of major carbonate bodies lateral to sphalerite-galena-pyrite ore, and (5) the occurrence of barite and albite without arsenopyrite in massive barite mineralisation. Observations made during the current study indicate that the earlier observations are also valid for the northern ore lenses, although some of the sphalerite-galena-pyrite ores at the northern end of the mine also contain significant amounts of albite. The trace element content of sulphide mineral components further supports the formation of chlorite-pyrite-chalcopyrite assemblages at higher temperatures than the assemblages associated with sphalerite-galena-rich ore. The model of Huston and Large (1989) was based on the premise that vertical zone refinement occurred within a sulphide mound at the seafloor. The same processes are considered to be consistent with predominantly lateral zone refinement that occurred beneath the Cambrian seafloor.

Early sphalerite-galena-pyrite ore and massive barite mineralisation were interpreted by Huston and Large (1989) to have formed at low to moderate temperatures (200-250°C) from near neutral fluids (pH = 5-6) in the pyrite stability field (fig.9.4). This temperature range is consistent with estimates derived during this study. Sphalerite-galena mineralisation was interpreted to have deposited as hydrothermal fluid temperature decreased, and pH and \( f_{O_2} \) increased. These changes would have been brought on through the mixing of a hydrothermal fluid with cooler, more oxidised seawater. An increase in pH may also have been brought on through the reaction and buffering of the hydrothermal fluid with clay-feldspar assemblages and organic carbon in the host sediments (von Damm et al., 1985; Goodfellow and Blaise, 1988). The presence of significant carbonate associated with the sphalerite-galena mineralisation has been interpreted to indicate near neutral pH, and the observation made during the current study of significant amounts of detrital albite preserved within P lens sphalerite-galena-pyrite ore further supports this inference.

The subsequent phase of mineralisation proposed by Huston and Large (1988) was one of replacement of sphalerite-galena-pyrite ore by pyrite-chalcopyrite under slightly acid (pH = 3-5) and higher temperature conditions, with fluids within the pyrite stability field near the pyrite-pyrrhotite buffer. During replacement temperature and \( f_{O_2} \) decreased while pH increased through the reaction:
Figure 9.4 $fO_2$ - pH diagrams for hydrothermal fluid at 250°C and 300°C
Diagrams show the evolution of the Rosebery hydrothermal fluid (A and B) as modelled by Huston and Large (1989). Also shown for the same conditions are the solubility of Zn and Cu as chloro-complexes (C and D). The sphalerite-galena ore field has been extended to account for of albite within Zn-Pb ore. The diagrams have been calculated assuming parameters that lie within the range of volcanogenic fluids:

\[
\begin{align*}
\sum S &= 10^{-2.5}, \\
\sum Cl &= 10^{0.0}, \\
\sum Na &= 10^{0.0}, \\
\sum K &= 10^{-1.0}, \\
\sum Fe &= 10^{-3.0} \text{(as FeCl\textsuperscript{+})}, \\
\sum C &= 10^{-1.0} \text{(as H}_2\text{CO}_3), \\
\sum Ba &= 10^{-4.0} \text{(as BaCl\textsuperscript{+}) and } \\
\sum H_3AsO_3 &= 10^{-4.0}. 
\end{align*}
\]

(modified after Huston and Large, 1989)

Figure 9.5 Temperature vs $a_K/a_{H^+}$ diagram.
Shows Au solubility contours, Au complex predominance fields and pH of hydrothermal fluid.
Fluid evolution pathway shown approximates that of fig.9.4. The diagram has been calculated for a typical high sulphidation volcanogenic fluid:

\[
\begin{align*}
1.0 \sum Cl, \\
0.9 \sum Na, \\
0.1 \sum K, \\
10^{-2.5} \sum S \text{ and } \\
m_{\text{SO}_4}/m_{\text{SO}_3}. 
\end{align*}
\]

(modified after Huston, 2000)


\[
4\text{CuCl}_0 + \text{ZnS} + \text{FeS (in sphalerite)} + 4\text{H}^+ + \text{O}_2 = 4\text{CuFeS}_2 + \text{ZnCl}^+ + 2\text{H}_2\text{O}
\]

Under the conditions modelled in the study by Huston and Large (1988), the low temperature hydrothermal fluid carried Au as $\text{Au(HS)}_2^-$ and Ag, Pb and Zn as chloro-complexes. The predominance of $\text{Au(HS)}_2^-$ under low temperature, near neutral fluid conditions is consistent with recent modelling undertaken by Huston (2000) where the $\text{AuHS}_0$ complex has also been considered (fig.9.5). Precipitation of Ag, Pb and Zn occurred due to cooling upon mixing with seawater, whereas Au precipitation was caused by oxidation of the hydrothermal fluids upon mixing with seawater, and by the precipitation of metallic sulphides.

Subsequent to the initial precipitation of Au with the sphalerite-galena ore, hotter fluids remobilised Au as $\text{Au(HS)}_2^-$. The remobilised Au was transported to the cooler margins of the ore and adjacent zones of barite mineralisation, where it precipitated due to oxidation and coprecipitation with metallic sulphides. With further increases in temperature and decreases in fluid pH, the solubility of Au as $\text{Au(HS)}_2^-$ would have decreased (fig. 9.5). Pyrite-chalcopyrite replacement of the sphalerite-galena ore would have commenced at $\sim 300^\circ\text{C}$ (Eldridge et al., 1983). At this temperature Au was transported primarily as $\text{AuHS}_0^0$, and Au precipitation would again have been initiated by oxidation of the hydrothermal fluids through mixing with seawater and by the precipitation of metallic sulphides. At more elevated temperatures Au may also have been transported as a chloro-complex, with Au precipitation potentially occurring through an increase in pH, or a decrease in temperature and $f\text{O}_2$. As the metalliferous fluids migrated to cooler areas the remobilised sphalerite and galena re-precipitated.

The association of Au with the low-temperature regions of K lens (fig.6.1i) and P lens (fig.6.2i) is consistent with Au transport within the hydrothermal fluids as $\text{Au(HS)}_2^-$. The strong association of Au with pyrite throughout much of the sphalerite-rich ore is consistent with the coprecipitation of pyrite and Au.

9.8 Other sub-seafloor replacement deposits

Ancient volcanic successions that include sulphide mineralisation attributed to a sub-seafloor formation are not uncommon, although many ore deposits include components that can be attributed to formation both at and below the seafloor. Doyle and Allen (2003) have compiled and reviewed many examples of sub-seafloor replacement deposits within ancient volcanic successions, and relevant aspects of the review are summarised below. Sub-seafloor replacement deposits may be located within a diverse range of host rocks, including massive to variably bedded clastic deposits, autoclastic facies associated with lavas and intrusions, and within coherent lavas and intrusions. Many replacement deposits, including Rosebery, are located within more than one volcanic facies. Most ancient replacement deposits appear to
have formed within ~200m of the seafloor, as the host strata to this depth were wet, porous and poorly consolidated, making them more amenable to large-scale replacement and infiltration by hydrothermal fluids than deeper strata that were less permeable due to increased compaction and alteration (Doyle and Allen, 2003).

Ancient replacement deposits wholly or partially hosted within pumiceous gravity flow sediments, as is the case for the stratigraphically deeper Rosebery ore lenses, include the Hercules, Lointown, Gossan Hill and Sulfur Springs deposits in Australia; the Långdal, Långsele, and parts of the Renström and Kyrkvägen deposits in Sweden; some lenses of the Hanaoka-Shakanai cluster in Japan; San Platón and Concepción in Portugal; and the Mattabi and other deposits in the Sturgeon Lake region of Canada (Doyle and Allen, 2003). Several features are commonly associated with replacement deposits hosted within pumiceous gravity flow deposits. Each ore deposit comprises one or more massive sulphide lenses with adjacent zones of disseminated sulphides, and the ore lenses within multi-lens ore deposits typically occur at more than one stratigraphic level. Zones of strong hydrothermal alteration, disseminated sulphides and/or veining link separate ore lenses within a multi-lens system. Alteration envelopes around most deposits are asymmetric, with the strongest and most extensive alteration in the footwall and at the lateral margins of sulphide lenses. Carbonate alteration is a common feature, and the carbonate may display a compositional zonation relative to distance from features such as alteration pipes and ore. The massive to semi-massive sulphides and carbonate-rich alteration zones contain relics of host rock and they grade laterally into host rock. Spotty and blebby sulphide and carbonate textures are common; probably indicating that replacement commenced at scattered nucleation sites within the host sediments (Doyle and Allen, 2003).

Ancient replacement deposits wholly or partially hosted within syn-eruptive lithic breccia-conglomerate-sandstone deposits, as is the case for the stratigraphically shallower Rosebery ore lenses, include the Ansil deposit in Canada; the Los Frailes-Aznacollar deposit in Spain; the Wilga-Currawong, Gossan Hill and Hercules deposits in Australia; and the Matsuki deposits in Japan (Doyle and Allen, 2003). Several features are commonly associated with replacement deposits hosted within syn-eruptive lithic breccia-conglomerate-sandstone deposits. The massive sulphide lenses consist of massive, semi-massive and banded sulphides, interleaved with strongly hydrothermally altered host rocks that contain disseminated, spotty and vein sulphides. Sulphide stringer veins may extend below and locally above the ore lenses (e.g. Ansil, Currawong, Gossan Hill, Matsuki). Parts of some of the ore lenses contain domains of partially altered host rocks or high matrix gangue contents (e.g. Rosebery, Currawaong). Partially replaced beds are common and primary bedforms (e.g. grading and flame textures) of the replaced host rocks are locally preserved. Within interbedded sandstone-siltstone facies, infiltration and replacement occur preferentially along coarser-grained, rapidly deposited beds, rather than within the intervening siltstones.
Replacement fronts occur at the margins and tops of some ore lenses, and strong hydrothermal alteration extends for distances of ~8m to 500m above the deposits. Hydrothermal alteration zones may include carbonate and siliceous nodule/lenses that formed by replacement (e.g. Matsuki, Gossan Hill, Hercules, Rosebery) (Doyle and Allen, 2003).

9.9. Synthesis

The pumiceous footwall volcanics at Rosebery were deposited as a series of large-scale syn-eruptive mass flows in a below-wave base marine environment as the result of a nearby rhyolitic eruptive event (fig. 9.6). Emplacement of thick comagmatic peperitic sills occurred shortly after deposition of the mass flow units. Sediments of the TSV were subsequently deposited as a series of turbidites and suspension settled sediments. They initially comprised material reworked from the footwall volcanics, and later received an increasing quartz crystal- and lithic-rich input from more distal volcanic sources. Rare intervals of reduced volcaniclastic input permitted deposition of thin carbonaceous mudstones. Rhyolitic to dacitic sills were emplaced within unconsolidated sediments of the footwall volcanics and the TSV, with commensurate formation of peperitic and autocomlastic textures within the sills. Basement faulting at the time of TSV sedimentation resulted in the formation of a fault-bounded basin and sub-basins, with fault complexity locally increased due to the presence of a large peperitic sill within the footwall volcanics. Subsequent to significant movement along basement faults, a quartz-feldspar-phyric peperitic sill was emplaced near the contact between the footwall volcanics and the TSV. This sill may have breached the sediment-seawater interface during emplacement, resulting in local sedimentary reworking and deposition of thin quartz-feldspar crystal-rich turbiditic sediments. Seafloor topography at the time of TSV deposition was primarily controlled by basement faulting and by the doming of sediments above peperitic intrusions. Sedimentation of carbonaceous mudstone (Black Slate) commenced under anoxic marine conditions as the influx of volcaniclastic material decreased. A series of thick mass flow units were subsequently deposited as the result of increased volcaniclastic input derived from a more distal source that included volcaniclastic, sedimentary and metamorphosed basement materials.

Low-temperature diagenetic alteration commenced immediately after deposition of the volcaniclastic sediments that comprise the footwall volcanics and TSV, producing clay and possibly zeolite alteration. This low-temperature alteration assemblage was rapidly overprinted by higher temperature feldspathic alteration generated as the result of an elevated geothermal gradient. Low temperature diagenetic pyrite framboids and fine-grained overgrowths on clastic material developed within the volcaniclastic sediments shortly after deposition.

Significant hydrothermal alteration and mineralisation commenced prior to compaction and lithification of the volcaniclastic succession, and subsequent to emplacement of the large
(i) Rhyolitic eruption and proximal submarine emplacement of feldspar-phyric pumiceous mass flows (~800m) onto coherent basement at depths below ocean wave-base, and onset of diagenetic alteration. Coherent basement in the vicinity of Rosebery comprising ~500m of Mt Black Formation above ~2000m of Crimson Creek Formation and underlying Precambrian metasediments.

(ii) Emplacement of peperitic sills into unconsolidated footwall volcanics with commensurate doming of overlying sediments. Confining pressure is sufficient to prevent significant vesiculation of lava.

(iii) Block faulting and development of mine-scale basins. Discrete, well defined faults in coherent basement propagating upward into broad zones of disruption within unconsolidated sediments.

(iv) Reworking of unconsolidated footwall volcanics as turbidite flows and water settled deposits with an upward increase in quartz-crystal and lithic components as volcanic provenance becomes more distal.

(v) Emplacement of quartz-feldspar-phyric peperitic sill into unconsolidated volcanioclastic sediments of the footwall volcanics and TSV. Confining pressure is sufficient to prevent vesiculation of lava.

(vi) Development of large-scale hydrothermal circulation with seawater recharge at regional-scale basin margins, with seawater recharge along faults and in areas lacking substantial sediment cover. Thick overlying sediments promote significant lateral hydrothermal flow through basement rocks. As the seawater is heated, contained sulphate precipitates as anhydrite or is reduced to aqueous sulphide with commensurate sulphur isotope fractionation. Hot hydrothermal fluids leach metals and rock sulphur from basement rocks, imparting Precambrian signature to Pb and Sr isotopes in solution. Hydrothermal upflow controlled by mine-scale faulting. At the mine-scale lateral flow is promoted by peperitic sills (A), fine-grained volcanioclastics (B), and the development of stratiform alteration zones and ores (in red).

Figure 9.6 Development of the Rosebery mine stratigraphy and ore mineralisation continues on next page....
Mineralisation develops as cool, locally circulating seawater mixes with hot hydrothermal fluid moving laterally through coarse-grained horizons within the sediments. The initial phase of Zn-Pb-rich sulphide mineralisation forms at lower temperatures (~200-250°C), and is overprinted locally by high-temperature Fe-Cu mineralisation (>280-300°C) as the system heats up. Minor barite and pyrite vein set develops as the hydrothermal system wanes.

Ambient anoxic carbonaceous sedimentation with minor volcanioclastic input.

Significant increase in volcanioclastic influx from distal source results in emplacement of thick mass flow unit of the Hangingwall Volcanioclastics.
quartz-feldspar-phyric peperitic sill. The precise timing of mineralisation relative to deposition of the hanging-wall succession is unclear, but lack of significant alteration or elemental enrichment suggests that mineralisation occurred prior to deposition of the basal hanging-wall carbonaceous mudstone (Black Slate). Hydrothermal fluids were derived from modified seawater, with seawater recharge occurring through large-scale bounding faults and regions of exposed basement rocks distal to Rosebery. As the seawater was heated the contained sulphate was reduced, with commensurate isotopic fractionation of sulphur. A thick volcaniclastic sediment cover led to significant lateral movement of the fluids through the basement rocks from which metals and rock sulphur were leached. The hot hydrothermal fluids initially flowed up along discrete basement faults, with a more diffuse fluid upflow loosely focussed along syn-sedimentary faults within the overlying volcaniclastic succession. Lateral flow was promoted by stratiform barriers to upwelling fluids in the form of sedimentary stratification, fine-grained horizons within the volcaniclastics, peperitic sills, and by stratiform zones of altered sediment produced during diagenesis and initial low temperature hydrothermal circulation.

Mineralisation occurred as metalliferous fluids passed laterally along more permeable horizons within the volcaniclastic succession. Permeability was controlled by primary sedimentary features such as grain size, and by changes brought on through the dissolution and precipitation of primary and secondary mineral components. Initial mineralisation occurred under hot (200-250°C), neutral to slightly acid conditions, as Zn-Pb-bearing hydrothermal fluids mixed with cooler seawater circulating locally within the unconsolidated sediments. Sulphide precipitation would also have been initiated through the reaction and buffering of hydrothermal fluids with clay-feldspar assemblages in the host sediments, and through a pH increase related to the reaction of hydrothermal fluids with organic carbon contained within minor carbonaceous sediments. Zn-Pb-rich sulphide mineralisation formed proximal to the upwelling fluid conduit, and barite and carbonate precipitated at the cooler, more oxidised margins. Initial low temperature mineralisation, characterised by primitive pyrite morphologies and a low temperature trace element suite preserved within the sulphide minerals, also locally preserved albitised volcanic plagioclase phenocrysts. As the mineralising system become hotter the hydrothermal fluids became more Fe-Cu-rich and more acidic, permeating laterally beneath and through mineralised host sediments. Initial low-temperature sphalerite-galena mineralisation was overprinted by a higher temperature sulphide phase characterised by more crystalline pyrite morphologies, a high-temperature trace element assemblage within the sulphide minerals, an increased chalcopyrite-pyrite content, and the destruction of albitised plagioclase phenocrysts. Significant chalcopyrite-pyrite mineralisation occurred at the base of sphalerite-galena lenses and proximal to hydrothermal upflow conduits, due to mixing with cooler seawater and through the replacement of low-temperature sulphide and gangue minerals. The lateral flow of hot hydrothermal fluids along permeable horizons within the ore and adjacent sediments probably led to formation of strong compositional banding observed in
the Cu-Fe rich southern ore lenses, and locally to an apparent reversal in the vertical metal zonation (e.g. G lens). The remobilised low temperature sulphide and barite components were re-precipitated at the upper and lateral margins of the ore lenses. As the hydrothermal system cooled, a minor baritic and pyrite vein set developed proximal to the ore. Although no diagnostic evidence exists, it is possible that some sulphide and barite mineralisation may have formed at the seafloor, as the result of hydrothermal fluids venting and mixing with cooler seawater or modified seawater within a brine pool.

Devonian metamorphism and deformation led to recrystallisation of the less competent gangue minerals, and physical and chemical remobilisation of the ore components. Cleavage development was related to regional deformation and reverse movement along the Rosebery and Mount Black Faults. The ore and host succession were moderately disrupted by reverse faulting and associated meso-scale folding. The ore was physically remobilised into dilational zones associated with fractures and fold axes. Chemical remobilisation of ore components also occurred, with subsequent re-precipitation of the components within dilational zones associated with rotated and boudinaged material, and within syn-tectonic quartz-carbonate veins. Precious metals were moderately enriched within the quartz-carbonate vein set due to preferential remobilisation.

Intrusion of late syn- to post-deformation granites during the Devonian produced minor oblique faulting, further vein development and metasomatic alteration of the ore and host rocks. Metasomatic fluids permeated along brittle-ductile faults and through the footwall schists, generating new mineral assemblages that included magnetite, haematite, biotite, chlorite, pyrrhotite, tourmaline, garnet, helvite and cassiterite. The metasomatic fluids remobilised sphalerite-galena-tetrahedrite out of the ore, replacing them with massive pyrite-pyrrhotite and magnetite-biotite ± chalcopyrite assemblages. Sulphide annealing occurred during or after the final stages of metasomatism. Au remobilised from the magnetite-biotite ± chalcopyrite alteration zone was precipitated in the pyrite-pyrrhotite zone, resulting in enriched gold grades relative to the precursor sulphide ore.

A minor phase of post-Devonian brittle faulting subsequently cut the mine sequence. Minor dolerite dyke emplacement occurred within the mine sequence, probably during the Jurassic.
References


Alt, J. C., and Chausiodon, M., 1988, Ion microprobe analyses of the sulphur isotopic composition of sulfides in the hydrothermally altered rocks, DSDP/ODP Hole 504B: Proceedings of the Ocean


Barton, P. B., Jr., 1978, Some ore textures involving sphalerite from the Furutobe mine, Akita Prefecture, Japan: Mining Geology, v. 28, p. 293-300.


Blevin, P., and Jackson, S. E., 1998, Potential applications of LA-ICP-MS technology in economic


Brill, B. A., 1989, Trace-element partitioning of elements in ore minerals from the CSA Cu-Pb-Zn deposit, Australia: Canadian Mineralogist, v. 27, p. 263-274.


References


Gifkins, C. C., Allen, R. L., and McPhie, J., 2000, Textural characteristics of diagenetic and hydrothermal...


Hannington, M. D., Poulsen, K. H., Thompson, J. F. H., and Sillitoe, R. H., 1999b, Volcanogenic gold in

References


Harcourt-Smith, J., 1898, Report on the Mineral Fields in the Neighbourhood of Mt Black, Ringville, Mt Read and Lake Dore.


Huston, D. L., and Large, R. R., 1987, Genetic and exploration significance of the Zinc Ratio (100Zn/References
(Zn+Pb)) in massive sulfide systems: Economic Geology, v. 82, p. 1521-1539.


Jackson, S. E., 1996b. LAMTRACE: Sydney, Macquarie University.


References


References


References
composition of lead from stratiform deposits: Canadian Journal of Earth Science, v. 4, p. 245-269.

Paulick, H., Herrmann, W., and Gemmell, J. B., 2001, Alteration of felsic volcanics hosting the Thalanga massive sulfide deposit (northern Queensland, Australia) and geochemical proximity indicators to ore: Economic Geology, v. 96, p. 1175-1200.


Perkins, C., 1996, $^{40}$Ar/$^{39}$Ar age constraints on deformation and mineralization, Rosebery Zn-Pb-Cu and Mount Lyell Cu deposits, Tasmania, Australia: Mineralium Deposita, v. 31, p. 71-83.


Plimer, I. R., and Lees, T. C., 1988, Tourmaline-rich rocks associated with the submarine hydrothermal Rosebery Zn-Pb-Cu-Ag-Au deposit and granites in western Tasmania, Australia: Mineralogy and Petrology, v. 38, p. 81-103.


Sato, T., 1972, Behaviours of ore-forming solutions in seawater: Mining Geology, v. 22, p. 31-42.


References
References
References


Waller, G. A., 1902, Report on the Ore Deposits (other than Tin) of North Dundas: Report of the
Secretary of Mines for 1902, p. 1-66.


