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SUMMARY

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The stratiform pyritic Zn-Pb-Cu and barite ore deposit at Rosebery is contained in a small shale lens within pyroclastic rocks (the Primrose Pyroclastics) which form part of the Mt. Read Volcanic Group $-$ an extensive belt of silicic volcanic rocks of Cambrian age.

The Primrose Pyroclastics consist of vitric, crystal and lithic tuffs, pyroclastic breccias, shale·lenses and quartz-sericitechlorite schists derived from the pyroclastic rocks. The Primrose Pyroclastics contain probable ignimbrites and therefore may have been deposited on land. They are overlain to the east by massive lavas of the Mt, Black Volcanics. On the, west the Primrose I Pyroclastics are flanked by and appear to intertongue with the Rosebery Group consisting of shale, quartz-wacke, greywacke conglomerate and dolomitic siltstone. The Cambrian rocks, particularly the volcanic formations, in the Rosebery area have undergone low-grade regional metamorphism that has·produced mineral assemblages of the lower greenschist facies together with schistose fabrics.

In the Rosebery Mine area the detailed stratigraphic sequence is: massive pyroclastics (vitric, crystal and lithic tuff), black slate, host rock (siltstone with crystal tuff bands), Barite orebody, host

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rock, Zn-Pb sulphide orebody with an Fe-Cu-rich basal zone, foota **wall schists, and footwall pyroclastics (crystal tuff).** The contact between the massive pyroclastics and the underlying black slate or host rock appears, in part, to be a disconformity. The footwall schists below the Zn-Pb orebody possibly represent a zone of wallrock alteration. A post-ore basaltic dike is corelated with the Jurassic dolerite and is of interest because it contains probable rheomorphic galena-chalcopyrite veinlets.

The Zn-Pb and Barite orebodies are concordant with the ϵ bedded host rock and all have been deformed into one main system of folds with an associated axial surface cleavage. The folds are tight \approx and overturned with east-dipping axial surfaces. $\,$ Similar fold styles are predominant in the host rock and also in the sulphide ore, although flattened concentric styles occur in layered ore in which mark-ed competence contrasts are present between layers. The trend and plunge of the fold axes is variable, particularly in the Zn²Pb orebody, but axial surfaces are of reasonably constant orientation throughout the mine area. The folds in the Zn-Pb orebody are of limited axial extent, approximating to conical folds, and commonly show en echelon patterns. These folds were probably formed in one period of folding by heterogeneous strain that may have been caused by the presence of a mass of highly incompetent

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Zn-Pb-Fe sulphide ore. Lineations parallel to the ore/host rock contact appear to reflect relative movement between the ore and the " host rock during deformation. The folding, cleavage formation and ,low-grade metamorphism is thought to have occurred during the Middle Devonian Tabberabberan Orogeny.

The Zn-Pb orebody and the Barite (orebody are spatially and .compositionally distinct. In comparison with the Zn-Pb orebody the Barite orebody is markedly Fe-poor and barite-rich. The ore is generally fine grained and massive and the orebodies have sharply defined upper and lower boundaries. Much of the ore shows compositional layering on scales of thickness ranging from millimetres to metres. The layering is considered to have formed by sedimentary deposition. The main ore minerals are sphalerite, pyrite, galena, chalcopyrite, tetrahedrite-te'nnantite, arsenopyrite and gold. Other minerals which have been recorded are magnetite, pyrrhotite, ilmenite, enargite, hematite, rutile, bournonite, meheghinite, .jordanite, electrum and an unidentified mineral that is possibly kobellite. The gangue minerals are chlorite, quartz, Mn carbonate, sericite, barite, sphene, calcite and albite.

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The Zn-Pb orebody is more or less continuous over a strike length of about 3,500 ft and averages 15 to 20 ft in thickness. Some overall trends in metal content occur along strike in the $Zn-Pb$

orebody but no systematic pattern is evident. A pyrite -chalcopyrite- ϵ arich basal zone is present in sections of the orebody. The main ore types in the Zn~Pb orebody are (1) high-grade sphalerite -galenapyrite ore, and (2) pyritic low-grade ore which includes pyritechalcopyrite-rich ore.

> The Barite orebody is discontinuous and consists of a number of separate lenses with a total strike length of the order of 2,000 ft.

Other types of mineralization associated with the Zn -Pb and Barite orebodies are (1) small lenses of hematite-pyrite and hematite-magnetite -pyrite, (2) several discordant bodies of massive pyrrhotite, and (3) patches of coarse-grained quartz and Mn carbon- , ate carrying galena, chalcopyrite, sphalerite and sulphosalt minerals. The footwall schists contain zones of disseminated pyrite-chalcopyrite mineralization.

Relict microsphericular pyrite and "colloform" structures involving pyrite and pyrite-chalcopyrite-sphalerite_have been observed.

Most of the present ore textures are interpreted as metamorphic and comprise granular polygonal sphalerite, galena and chalcopyrite as matrix formers and. pyrite and magnetite in porphyroblastic and poikiloblastic habits. The discordant pyrrhotite bodies, the patches of coarse-grained mineralization and some of the sulphosalt minerals were possibly formed during the metamorphism.

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Frequency distributions for the assays of the different metals contract the contract of the contract of $-$ Zn, Pb, Cu, Fe, Ag, Au $-$ in the Zn-Pb and Barite orebodies range from strongly positively skewed, through symmetrical to slightly negatively skewed. The metals present in minor or trace amounts (Au, Ag, Cuand to a,lesser extent Pb) show strong positive skewness, while the metals of greater abundance (Zn and Fe) show a variety of flat-topped and, polymodal distributions which are usually approximately symmetrical, or rarely slightly negatively skewed.

> Correlation coefficients were calculated for the different combinations of metal pairs using assay data from the Zn-Pb and Barite orebodies. The pairs Pb-Zn, Pb-Ag, Zn-Ag and Ag-Au show moderate positive correlation while $Pb-Au$ and $Zn-Au$ show lower positive correlation. Iron displays moderate negative correlation with Pb, Zn and Ag. In the Barite orebody Cu is positively correlated with Pb, Zn and Ag.

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The stratigraphic zonation from the base upwards of $Fe-Cu$, Zn -Fe-Pb-Cu and barite-rich ores is similar to the stratigraphic sequence that has been recorded from a number of ore deposits of likely exhalative -sedimentary origin. This sequence is correlated with a changing source composition which could have been produced ϵ by differentiation of a subvolcanic granitic magma.

. Manganese carbonates are widespread throughout the ore δ deposit and were probably deposited contemporaneously with the ore. The Mn carbonates are in massive, bedded and oolitic forms comparable with sedimentary Mn carbonate ores. The approximate chemical compositions of the Mn carbonates, determined from measurements of their ω refractive indices and $d(211)$ X-ray lines, mainly fall into the three fields of rhodochrosites, ferroan rhodochrosites and magnesian kutnahorites. The Mn carbonates are most abundant near the strike limits of the $Zn-Pb$ orebody where they are in massive lenses up to 400 ft in,length composed of rhodochrosite and magnesian kutnahorite. The ferroan rhodochrosite appears to be restricted to the Zn-Pb sulphide orebody.

> The Rosebery ore deposit is an example of the world-wide class of stratiform pyritic $Zn-Pb-Cu$ ores associated with silicic pyroclastic rocks. For the Rosebery deposit the geographic association with· pyroclastic rocks, the Co and Ni values in· pyrite, the sulphur isotope values and the Cd contents of the sphalerite indicate that volcanic activity, possibly a subvolcanic magma, was responsible for the generation of the ore-forming solutions and that ore deposition was contemporaneous with volcanism. The ore layering, the stratigraphic distribution of the orebodies, the

association with bedded and oolitic Mn carbonate and the lack of evidence for replacement strongly suggest deposition of the ore as a chemical sediment; it is essentially exhalative-sedimentary in origin.

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1. INTRODUCTION

The pyritic Zn-Pb-Cu ore deposit at Rosebery (145[°] 32'E; 41[°] 47'S) in west Tasmania (Fig. 1) is situated about 20 miles inland from the coast at an altitude of 550 ft. The area is mountainous with· Mount Murchison 4, 183 ft, Mount Read 3,660 ft. and Mount Black 3, 037 ft forming the principal topographic features . . The vegetation is predominantly dense rain forest, although button grass covers some of the .flat-lying areas. The average annual rainfall is 82 inches.

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Although the deposit was discovered in 1893 early mining activity was intermittent because of the difficulty of treating the fine . grained.sulphide;ore. Between 1920 and 1926 the Rosebery and nearby Hercules Mines were purchased by the Electrolytic Zinc Company and following the successful application of the selective flotation process full-scale production commenced in 1936. The ore production to July 1969 of the Rosebery Mine is 5.3 million tons at an average grade of 5.4 per cent Pb, 18. 4 per cent Zn, 0.53 per cent Cu, 6.4 *ozJton* Ag and!. 9 *dwtJ* ton Au. The ore reserve as of July 1969 is 9.2 million tons of a grade similar to that mined. The deposit was originally worked by a series of adits from 1, 300 ft above sea. level down to 8 Level (550 ft above sea. level). At " present stoping is carried out down to 14 Level (200 ft below sea level)

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_jand access is from an inclined shaft. In 1967 work was started on sinking a vertical shaft to 23 Level (1,500 ft below sea level).

1.1 Previous Work

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Montgomery (1895) noted that the bands of sulphide ore were conformable with,layering in the argillaceous schist host rock and although favouring a replacement origin he had earlier (Montgomery, 1893) suggested that the very similar Mount Read (Hercules) orebodies, 4 miles south of Rosebery, might be a bedded syngenetic deposit. In 1902 Waller described the Rosebery orebody as conform able and continuous. He favoured a replacement origin from oreforming fluids of granitic derivation but he also allowed the possibility of derivation from the adjacent volcanic rocks. Hills (19l5a, 19l5b) carried out an extensive survey of the Rosebery and Hercules ore deposits and concluded that the orebodies were replacements of carbonate beds in the host rocks by ore-forming solutions of Devonian granitic parentage. According to Finucane (1932) the Rosebery ore deposit was a replacement localized by a line of weakness between sheared intrusive porphyries and xenoliths of slate. A similar view was held by Dallwitz (1945). $\:$ In Hall et al. (1953) the orebodies were considered to be replacement lenses" parallel to cleavage in the host

rock, but in.Hall et al. (1965) the emphasis had changed to replacement of a favourable bed. More recently, exhalative-volcanic and syngenetic origins have been considered (Hall and Solomon, 1962); Campana and.King1963; Pereira, 1963; Solomon, 1964, 1965a).

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1. 2 Object and Scope of the Investigation

A number of published and unpublished reports have discussed the geology of the Rosebery ore deposit but apart from routine mapping of the orebodies no comprehensive, detailed geological studies have been made in recent years. Because of this' lack it was decided that the present work should be a broad study embracing the geological setting, and the stratigraphy, structure, petrography and mineralogy of the orebodies and the surrounding rocks. This' has included regional geological mapping of the Rosebery-Hercules area on a scale of four inches to one mile (Fig. 3), underground mapping, logging of many thousands of feet of drill core, petrographic studies of the rocks of the mine area, mineragraphic study of the orebodies, and investigation into aspects of the chemistry of the orebodies. The wide scope of the study has meant some unevenness of treatment and it has not been possible to deal with some aspects in depth.

The writer worked as a mine geologist at Rosebery for two and a half years before carrying out two years full-time study at the University of Tasmania.

The ultimate aim of the study has been to determine the origin of the Rosebery ore deposit by synthesis of the geological information along with the available chemical compositional and isotopic data .

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.Most previous workers have favoured an epigenetic replacement origin but more recently syngenetic origins have been suggested. However, the arguments for a syngenetic versus epigenetic origin have not been explicitly stated. The argument has been that since the Rosebery ore deposit looks similar to some other stratiform sulphide ore deposits for which syngenetic origins are favoured it might also be syngenetic. Because of the possibility of a syngenetic origin the approach here has been to examine the orebodies in the context of the geological history of the enclosing rocks. A better knowledge of the origin would help in the exploration for new orebodies which could then be conducted on a sounder scientific basis.

This study is part of a wider project on the metalliferous ores of Tasmania led by Dr. M. Solomon. As parts of wider studies: $\rm G.D.$ Loftus-Hills has examined $\rm Co$ and $\rm Ni$ in ore minerals, mainly

pyrite, from Rosebery; D.1. Groves and G. D. Loftus -Hills have analysed for Se in sulphide minerals at Rosebery; the Cd and Mn $_{\odot}$ contents of some Rosebery sphalerite has been determined by $D.I.$ Groves; sulphur and oxygen isotope ratios in sulphides, barite and Mn carbonate at Rosebery have been studied by M. Solomon, T. A. Rafter and M. L. Jensen; and C. E. Gee is currently investigating the geochemistry of the shales at Ros ebery.

1. 3 Acknowledgements

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2. GEOLOGICAL SETTING

2. 1 Introduction

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The Rosebery ore deposit is the largest of a group of pyritic $Zn+Pb-Cu$ deposits, the Hercules-Rosebery-Pinnacles line (Hall and Solomon, 1962; Solomon, 1965a), that occur over a distance of about 12 miles along the north-south trending western margin of the Mt. Read Volcanics (Fig. 2). The Mt. Read Volcanics, which are rhyolitic to andesitic in composition, are a belt of Cambrian volcanic rocks that extends for over 100 miles around the west and north I margins of the Precambrian Tyennan Ge anticline . They are believed to have formed a volcanic arc (Campana and King, 1963; Solomon, 1965a). Several small subvolcanic bodies, such as the Murchison Granite $6\frac{1}{2}$ miles east of Rosebery, intrude the Mt. Read Volcanics.

Near Rosebery the Mt. Read Volcanics are flanked on the west by the Rosebery Group consisting of unfossiliferous shales, $\,$ quartz sandstones, greywacke conglomerate and tuffs. .Conformably overlying the Rosebery Group are the eugeosynclinal sedimentary rocks of the Crimson Creek Argillite and the Dundas Group, the latter ranging in age from Upper Middle Cambrian to Middle Upper Cambrian.

Fig. 2 Geological Map of West Tasmania (after M. Solomon, 1965a).

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In the late Cambrian sedimentation was interrupted by the Jukesian Orogeny which produced uplift, faulting and gentle folding along north-south trends. The uplift caused the deposition of siliceous conglomerates and sandstones (the Jukes Breccia and Owen Conglomerate) in confined terrestrial basins during the Lower Ordovician. A marine transgression followed and resulted in the deposition of up to 7, 000 ft of sandstone, limestone (the Gordon Limestone) and siltstone, which was succeeded during the Silurian and Lower Devonian by sandstone and mudstone of the Eldon Group.

In the ·Middle Devonian the area was affected by the Tabberabberan Orogeny, a major period of deformation, which produced folding, faulting, cleavage and chlorite-grade metamorphism. The deformation was succeeded by granitic intrusions that are Upper Devonian to Lower Carboniferous in age (McDougall and Leggo, 1965).

Following uplift and erosion, tillite was deposited by glaciation early in the Permian. In the Jurassic, dolerite dikes were intruded through the pre-Permian rocks.

Tertiary uplift and erosion removed most of the post-Ordovician rocks and extensive valley and mountain glaciation took place during the Pleistocene.

2.2 Stratigraphy

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The stratigraphic succession in the Rosebery area as inferred from the geological map (Fig. 3) is shown in Figure 4 (cf. Loftus-Hills, Solomon and R. Hall, 1967, Fig. 4).

The youngest rocks are Pleistocene tillite and fluvioglacial deposits which are extensive along the western side of the area and unfortunately obscure many critical contacts in the Cambrian rocks.

South of the Hercules Mine, Permian tillite and siltstone α , and Ahmad, 1962) on a contribution occur in several small outliers (Banks and Ahmad, 1962) on a plateau-like surface co-existensive with Mt. Read and probably representing a pre-Permian erosion surface.

> The Ordovician is represented by the Jukes Breccia and Lower Owen Conglomerate which forms the Mount Murchison massif. The' Mount Murchison section is described in Campana and King (1963).

> The history of the stratigraphic classification of the Cambrian rocks of the Rosebery area is summarized in Table 1.

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Fig. 3

Geqlogical Map of the Rosebery Area.

Mapped and compiled by R.L. Brathwaite from mapping by K.J. Finucane (1932); D. King; G. Loftus-Hills; M.. Solemon; R. J. Hall; Department of Mines, Tasmania, (1966), Mackintosh Sheet; and geologists of the Electrolytic Zinc Company, Rosebery.

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The detailed stratigraphy and stratigraphc nomenclature of the " Cambrian rocks is given in Appendix 1.

Crimson Creek Argillite

Conformably overlying the Rosebery Group are thin-bedded red and green argillites which have been correlated with the Crimson Creek Argillite (Campana and King, 1963; Loftus-Hills et al., 1967).

Rosebery Group

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The sedimentary rocks of the Rosebery Group near Rosebery dip west and face west (Loftus -Hills et al., 1967; Geological Map, Fig. 3), contrar'y to suggestions made by Hall et al. (1953) and Campana and King (1963) who considered that the west dips were overturned. In descending stratigraphic order the Rosebery Group consists of the following formations: Munro Creek Slate, Westcott Argillite, WHliamsford Volcanics, Salisbury Conglomerate, Natone Volcanics, Stitt Quartzite, and Chamberlain Shale. The total thickness is of the order of 5, 000 ft.

The Munro Creek Slate consists of grey and black shales with-grey quartzose sandstone beds. The sandstone component appears to decrease in relative abundance from south to north. The thickness west of Rosebery is 1,500 ft.

The Westcott Argillite consists of purple argillite and dolomitic siltstones (Campana and King, 1963). The purple argillites are petrographically lithic greywackes with a hematitic cement. The -thickness near Rosebery is about 500 ft. Between the Jupiter Mine and Williamsford the Westcott Argillite appears to cut out against the western margin of the Primrose Pyroclastics of the Mt. Read Volcanic Group. Northwards the lithology appears to change from predominantly purple argillite to red and green argillite although correlation is complicated by folding on the northern side of Bobadil Plain.

The Williamsford Volcanics are a band of poorly exposed, massive rhyolitic tuffs. They pinch out northwards and are inferred to join the Primrose Pyroclastics near Williamsford.

The Salisbury Conglomerate is a fuchsite-bearing, greywacke conglomerate which consists of pebbles of chert, quartzite, dark shale, vitric tuff and discrete masses of green fuchsite in a matrix of quartz, plagioclase, sericite, carbonate, chlorite and pyrite. The thickness is about 150 ft in the Pieman River near Rosebery but it thins southwards to less than 10 ft on the Williamsford Road and cuts out against the Primrose Pyroclastics.

The Natone Volcanics are white to grey lithic, crystal and vitric tuffs and contain embayed quartz crystals, sericitized

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. plagioclase and sericite-quartz lithic fragments in a sericite-quartz groundmass. - Some-of the vitric tuffs show relict vitroclastic texture. Although-sheared they are not noticeably bedded or sorted. The Natone volcanics pinch out northwards and converge southwards with the Primrose Pyroclastics. The area where the Natone Volcanics and the Williamsford Volcanics meet the Primrose P yroclastics is covered by glacial gravels but from the convergence of strikes ·and similarity' in lithology it is inferred that these two bands of volcanic rocks are tongues extending out from the western margin of the Mt. Read Volcanic Group.

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. The Stitt Quartzite can be traced continuously from the Williamsford Road to the Pieman River near the Emu Bay Railway bridge and is the best regional marker unit. The Formation consists of thick-bedded quartz-wacke and micaceous sandstones with thin shale interbeds. The thickness near Rosebery is 1;000 ft but' it thins rapidly southwards and cuts out against the Primr,ose Pyroclastics.

The Chamberlain shale is composed of thin-bedded grey, black and green siltstones and mudstones. The maximum thickness as exposed in the Stitt River is **1,500** ft but the true thickness is probably less because of repetition by folding. The Formation thins

rapidly to the south and north of the Stitt River. The contact with the Primrose Pyroclastics although,poorly exposed appears to be conformable.

In respect of thickness and the association of quartz-wacke, greywacke conglomerate, volcanics, shale and dolomitic siltstone the Rosebery Group helongs to the unstable shelf association of Krumbein and Sloss '(1963).

The Rosebery Group was correlated by Campana and King $\,$ (1963) with similar lithological sequences in west Tasmania which were later termed the Success Creek phase by Solomon (1965a). The age of the Success Creek phase is thought to be Lower Cambrian.

Mount Read Volcanic Group

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The volcanic rocks in the Rosebery-Hercules area were first ¹ described as extrusive keratophyres-and quartz keratophyres (Twelvetrees and Petterd, 1899). Finucane (1932) believed them to be intrusive porphyries of Devonian age and they became known as the "Porphyroids" (Carey, 1953). In Hall et al. (1953) they were termed the "massive pyroclastics" and their volcanic origin is now well established (Solomon, 19@{;); Spry, 1962; Campana and King, 1963; Solomon, 1964). The name Mount Read Volcanics was introduced by Banks and Solomon (1961) and was defined by

Campana and King (1963).

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In the Rosebery-Hercules area two distinct lithological assemblages have been mapped by the writer (Fig. 3) and consequently it is proposed that the Mt . Read Volcanics be given Group status and the two assemblages be termed the Mount Black Volcanics and the Primrose Pyroclastics.

The Primrose Pyroclastics appear in part to underly and in part to be laterally-equivalent to the Rosebery Group (Fig. 4; Loftus-Hills et aI., 1967), which would suggest that the Primrose Pyroclastics are Lower Cambrian in age. However this does not agree with the Upper Middle Cambrian age of trilobite-bearing black shales-which 'are interbedded with pyroclastic breccias and tuffs of the Mt. R-ead Volcanic Group at Que River, 14 miles north-west of Rosebery (Gee, Jago and Quilty, in press). Possibly the correlation of the Rosebery Group with the Success Creek phase is incorrect.

The type section of the Mount Black Volcanics is exposed along the Murchison Highway from $\frac{1}{4}$ mile east of Rosebery to the western boundary of the "Farrell Slates" near Tullah. The dominant lithology is albite porphyries. They are massive, although locally sheared, homogeneous fine to medium grained and porphyritic with pink albite crystals in a dark green chloritic matrix. Individual

compositional units range from **10** to 240 ft and average 80 ft in thickness. Contacts between the different units appear to be concordant and are commonly graditional. Some of the units contain amygdules. The Mt. Black Volcanics are mineralogically classed as keratophyres and consist of albite and chlorite with lesser amounts of quartz, sericite, epidote and sphene. They contain noticeably less quartz than the Primrose Pyroclastics and the one available chemical analysis (Appendix 2), has been termed a sodi-potassic trachyte by Spry (1962). From the field relations and textures the Mt. Black Volcanics appear to be lavas.

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The Primrose Pyroclastics consist of vitric, crystal and lithic tuffs, pyroclastic breccias, shale lenses (notably including the shale lenses that are hosts to the Rosebery and Hercules deposits), and quartz -sericite -chlorite schists derived from the pyroclastic ${\tt rocks.} \quad$ Crystal, lithic and vitric tuffs are the predominant lithologies. They are massive and unsorted. Mineralogically they are quartz keratophyres and the principal minerals are quartz, albite, sericite and chlorite. Quartz and albite are the crystal components and the quartz shows the embayed forms typical of rhyolites and ignimbrites. The lithic fragments comprise crystal and vitric tuff, altered rhyolite, pumice and slate, Relict vitroclastic textures are not uncommonly preserved (Fig. 6) in the microcrystalline aggregate of

quartz, sericite and albite that makes up the vitric tuffs and the matrix of the crystal tuffs. Many of the massive tuffs of the Prim .rose Pyroclastics appear to be ignimbrites (Spry, 1962; Solomon 1964; this thesis, Section 3.9). From the chemical analyses given in Finucane (1932) (Appendix 2), the tuffs of the Primrose Pyroclastics are chemically comparable to potassic and sodi-potassic rhyolites (Spry, 1962). Lenses of grey and black shale are common throughout the Primrose Pyroclastics. Apart from the large shale lenses of the Rosebery Mine, the Hercules Mine, Barkers o Road, Primrose Road and south of Hercules, there are a number of smaller lenses of the order of 2 to 10 ft in thickness. The Primrose Pyroclastics grade eastwards into the Mt. Black Volcanics.

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2.3 Metamorphism

Metamorphic effects in the Rosebery Group are generally slight. The Chamberlain Shale and shale interbeds in the Stitt Quartzite have a weak slaty cleavage while in the Natone Volcanics, the Salisbury Conglomerate and the Munro Creek Slate a pronounced, near vertical cleavage is commonly present. The mineral assem-. blages consist mainly of quartz-sericite-muscovite with or without chlorite, which would be stable in the lower greenschist facies but

are not diagnostic.

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Both the Primrose Pyroclastics and to a lesser extent the Mount Black Volcanics contain belts of schistose rocks. The schists are most extensive along the footwall and to the west of the Rosebery and Hercules host rock lenses. They grade laterally and vertically through less sheared rocks into crystal tuffs. These schists are - grey to green with a phyllonitic appearance and show a weak to moderate schistosity imparted by alignment of sericite and chlorite flakes and quartz augen.

The most common mineral assemblage in the Primrose Pyroclastics is quartz -albite -sericite -chlorite -sphene, while in the Mt. Black Volcanics albite-chlorite -quartz -sericite -epidote -sphene is most typical. Other less common assemblages contain calcite, pyrite or green biqtite. These mineral assemblages belong to the lower greenschist facies of regional metamorphism and this is consistent with the dynamic metamorphic fabric present in some of the rocks.

2. 4 Structure

Folds

West of Rosebery the sedimentary rocks of the Rosebery Group dip west and face west while the Rosebery Mine shale lens dips east and faces east (see Section 3), and an anticlinal axis is inferred to lie between them (see Loftus -Hills et al., 1967). The other possibilities of an unconformity or a major fault are discounted because of lack of field evidence. There is a suggestion of an anticlinal hinge in a cutting on the Primrose Road about where a change from east to west dips takes place, but the possible anticlinal trace cannot be defined elsewhere because of the absence of marker beds in the Primrose Pyroclastics. The implications of an anticlinaL structure are that part of the Mount Read Volcanic Group is older than the Rosebery Group, as pointed out by Loftus-Hills et al. (1967), and that part of the Mt. Read Volcanic Group is laterally equivalent to the Rosebery Group (Fig. 4). This latter implication requires rapid east-west facies changes, for which there is some evidence in the apparent intertonguing of the Natone and Williamsford Volcanics with the sedimentary formations of the Rosebery Group along the western margin of the Mount Read Volcanic Group (see Appendix 1).

It has been suggested that the Rosebery Mine shales and the west dipping "Farrell Slates" four miles east of Rosebery, are the

same horizon on the opposite flanks of either a syncline (Hills, 1915b) or an anticline (Carey, 1953). However, the "Farrell Slates" probably face east, as do the Rosebery Mine shales, so that the intervening structure must consist of a syncline and an anticline (see Solomon, 1962, Fig. 71), or alternatively they are two different stratigraphic horizons and there is a continuous east facing succession between them. The absence of similar lithologies to the lithic, crystal and vitric tuffs of the Primrose Pyroclastics in association with the "Farrell Slates" favours the view that they are different . horizons .

. In the Rosebery Group along the northern side of Bobadil Plain a syncline and an anticline are inferred from reversals of sedimentary facings (see Loftus-Hills et al.,1967,p. 335). These folds do not appear to extend south.

Faults

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Campana et al. (1958) and Campana and King (1963) thought that the Hercules, Rosebery, Chester and Pinnacles ore deposits lay along a major Cambrian fault zone which also marked the western boundary of the Mt. Read Volcanic Arc.. Certainly an original structural control of this western boundary and the adjacent \hat{c} Hercules -Rosebery-Pinnacles line seems likely (Solomon, 1964)

but the correlation by Campana and King (1963) of the footwall schist ϵ zones of the Rosebery and Hercules Mines with a major fault zone appears mistaken. The schistose zone dies out just north of Rosebery and the volcanic rocks east and west of the supposed fault are lithologically very similar and have been mapped as one formation - the Primrose Pyroclastics.

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A north-east striking fault (the Jupiter Fault) was inferred by Campana and King (1963) to cut off the formations of the Rosebery Group along the western margin of the Mt. Read Volcanic Group. This apparent cut-off can however be explained by sedimentary wedging-out and field evidence for a fault is poor (Appendix 1).

The few faults that can actually be mapped strike northwest.

2.5 Mineralization

The ore deposits of west Tasmania can be classified (Hall $\,$ and Solomon, 1962; Solomon, 1965a) according to their spatial relationships to particular igneous rocks (see Fig. 2) into four main groups :

> (1) Within ultramafic and mafic rocks, pentlanditeheazelwoodite in serpentinite (Trial Harbour);

magnetite-pyrite lenses in amphibolite (Savage River).

(2) A&Sociat-ed with rocks of the Mt. Read Volcanic Group, stratiform lenses of pyrite-chalcopyrite (Mt. Lyel1·) and sphalerite -pyrite -galena -chalcopyrite ..(.Bosebery, Hercules).

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- (3) Within granitic rocks, cassiterite in granite (Heemskirk).
- (4) Adjacent to granitic rocks, cassiterite-pyrrhotitepyrite in :replacement lenses (Renison Bell, Cleveland, . Mt. Bischoff); Ag-bearing galena-sphalerite ores in \hat{c} fissure veins (Zeehan, Magnet).

Campana and King (1963) classified the deposits according to supposed age into Cambrian (Group (2) and Cu-Ni mineralization of $Group (1)$ and Devonian (Groups (3) and (4)) metallogenetic epochs.

The Mt. Lyell deposit, 20 miles south of Rosebery, contains a yariety of pyritic Cu orebodies in sericite-chlorite schists of the Mt. Read Volcanic Group (Solomon and Elms, 1965; Markham, 1968). There is a small occurrence of banded pyritic Pb-Zn-Cu· ore, the Tasman shaft mineralization, which is similar to the Rosebery ore. The presence of possible Lower Ordovician fossil hematite-barite gossans indicates a pre-Ordovician phase of mineralization (Solomon, 1967b) and textural evidence indicates that much of the ore is pre-tectonic, i.e. pre-Middle Devonian in age (Markham, 1968).

The significant deposits of the Hercules-Rosebery-Pinnacles: line from south to north are Hercules (pyritic Zn-Pb-Cu), Jupiter (pyritic Zn-Pb-Cu), KOQyna (pyriticZn-Pb-Cu,), Rosebery (pyritic Zn-Pb-Cu), Chester (pyrite) and Pinnacles (pyritic Zn-Pb). Of these only Rosebery and Hercules _are in, production and of the others only Jupiter and Chester have produced significant amounts of ore. The ores arid associated rock types of the Hercu1es deposit are very similar to the Rosebery deposit (Hills, 1915a, 1915b; G. Hall et al., 1965; R. Hall, 1967).

A number of small fissure-and replacement veins occur within the Primrose Pyroclastics in the Rosebery area. The veins are post-folding. They are composed of quartz, tourmaline, flourite, chlorite, pyrite, siderite and chalcopyrite, with minor amounts of gold (WaIler, 1902). As well, bismuthinite, wolframite, jamesonite, galena, sphalerite and cassiterite have been reported from some of the veins (Finucane, -1932). Waller (1902) pointed to the presence of flourite and tourmaline as indicating a,Devonian granitic source. The nearest deposits of obvious granite parentage are the aXinite-pyrrhotite:chalcopyrite deposit of Colebrook Hill and the Athenic tin mine, respectively 3 and $3\frac{1}{2}$ miles south-west of Rosebery. Other than the Pine Hill complex, 6 miles west of Rosebery, the nearest outcropping Devonian granite is the Granite

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Three to five.miles east and north-east of Rosebery are the Ag-Pb-Zn fissure vein deposits of the Sterling Valley, Murchison and Mt. Farrell Mines (Hall et al.,1953; Solomon, 1965b). The orebodies occur within the "Farrell Slates" and are composed of galena with lesser_amounts of sphalerite, chalcopyrite, pyrite, tetrahedrite and siderite. They have generally been considered to be of Devonian age (e. g.Campana and King, 1963) but largely on the basis of sulphur isotope results Solomon, Rafter and Jensen (in press) suggest the possibility of remobilization of volcanic Cambrian sulphides during the Tabberabberan Orogeny.

3. STRATIGRAPHY AND PETROLOGY OF THE ROCKS OF THE MINE AREA

In the Rosebery Mine area the Primrose Pyroclastics have been subdivided into the following local units:

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Away from the mine area the massive pyroclastics and footwall pyroclastics cannot be lithologically distinguished but they are useful as local names to describe the mine geology. The black slate and host rock form a lens some 7, 000 ft in length and averaging 220 ft in thickness. The lens pinches out at its northern and southern limits (Fig. 3);

As noted by Hall et al. (1953) the massive pyroclastics disconformably overlie the black slate and host rock. The evidence for the disconformity is as follows: \cdots

> (1) The black slate is missing south of co-ordinate 600 S* and the massive pyroclastics are directly in contact with the host rock. The subsurface delineation

'"600S and similar numbers are the mine coordinates in feet \cdot **north or south.** Mine grid north (GN) is 10⁰ west of true north.

of the southern limit of the black slate shows up an , irregular contact that pitches gently north at about R.L.00* from 500S to 1500N. The black slate recurs below R. L. 9500 between 1000S and 500N and as far as subsurface delineation has gone the two occurrences are not connected (Fig. 5). The absence of the black slate between these two occurrences could be explained by displacement along a fault, non-deposition, or erosion before deposition of the massive pyroclastics. Some faulting is present along the massive pyroclastic/black slate contact but an east-side-down throw of about 600 ft would be required. However the drags on the mapped faults near the contact have an east-side-up movement and the projection of these faults intersect the orebodies without significant displacement. The explanation of non-deposition is a possibility but there is no evidence for the deposition of two separate areas of lithologically identical black slate.

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(2) The contact between the massive pyroclastics and the host rock is slightly discordant to the dip of the cut-off black slate and suggests a slight angular unconformity.

"R. L. 00 and similar numbers are reduced levels. Sea level is taken \mathbb{R}^n as zero for reduced levels above sea level. Reduced levels below sea level are obtained by subtraction from sea level as 10, 000 feet.

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(3) The massive pyroclastics contain fragments of black and grey slate which appear to increase in number towards the black slate contact suggesting that the pyroclastics have incorporated fragments of underlying slate during deposition.

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The extent of the disconformity is not known but it may be only a relatively local feature as might be expected in volcanic terrains. The disconformity is particularly important because it shows that the massive pyroclastics are younger than the black slate and the mine sequence is right-way-up.

Where the host rock is overlain by the massive pyroclastics it is commonly considerably thinner than usual, which could be due to erosion before deposition of the massive pyroclastics. Of particular interest in some of these areas is that the Barite and Zn-Pb orebodies are close to the massive pyroclastic/host rock contact. If this is an erosional surface it might be expected, assuming the orebodies are syngenetic, that an unconformable contact between the ore and the massive pyroclastics could occur. Unfortunately the areas where the orebodies and the massive pyroclastics are close together are as yet only known from drill-hole intersections. In one drillhole intersection at R. L. 9575 on 600N less than 2 ft of host rock are present between crystal-lithic tuff of the massive pyroclastics and

and underlying Zn-Pb sulphide ore. Sphalerite occurs as disseminated grains in crystal tuff of the massive pyroclastics at the host rock contact in several drill-core specimens (e.g. 35787* at R.L. 8960 on l500N) and it is possible that this sphalerite was derived from erosion of pre-existing sUlphide ore.

3.1 Massive Pyroclastics

The massive pyroclastics (Hall et aI., 1953) consist of crystal, lithic and vitric tuffs. In the mine area they are fairly well known through about 50 drill-hole intersections. The most detailed drilling has covered the area from 14 Level (R. L. 9820) down to R.L. 8800 between 1000S and l500N. In this area the stratigraphy of the massive pyroclastics is :

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The boundary between these two tuff units is generally gradational. In some places (e. g. Fig. 5) a lens of black slate occurs at this boundary. In drill-hole intersections 3, 000 ft further north around 4000N the vitric tuff unit appears to have changed laterally into crystal and lithic tuffs.

*35787 and similar numbers refer to specimens in the Geology Dept.
University of Tasmania.

Vitric Tuffs

These rocks are fine-grained, very massive, pale yellowgreen and grey, and may show a conchoidal type fracture. They are composed of quartz, plagioclase, sericite and very fine-grained sphene. Calcite and chlorite are present in some thin sections. Crystals of quartz and plagioc1ase 0.05 to 0.2 mm in size are set in a very fine-grained quartz-sericite-plagioclase matrix. Relict vitroc1astic texture as defined by arcuate shard structures (Fig. 6, 35730, 35736, 35738, 35740) indicates that these rocks are vitric tuffs. The matrix of 35737 contains relict collapsed. pumice fragments measuring 1. 0 to 3.0 mm in length. The pumice is identified by discrete patches showing a streaky parallel structure (e.g. Fig. 7). Quartz and subhedral plagioclase crystals up to 2.0 mm in length occur locally and with an increasing. proportion of crystals there is a gradation into vitric-crystal tuffs (35738).

Crystal and Crystal-Lithic Tuffs

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These tuffs are massive with crystals of quartz and plagioc1ase, and lithic fragments in a fine-grained green matrix (Fig. 8). Of the crystal components plagioc1ase is generally more abundant than quartz which is absent from some of the tuffs. The lithic fragments attain a maximum size of 8 cm but are generally less

Fig. 6 - Relict vitroclastic texture in vitric tuff.

 $(35730, x 300,$ uncrossed nicols).

Fig. 7 - Relict collapsed pumice texture in vitric tuff.

(35737, x 180, uncrossednicols).

than 2 cm. They comprise mainly silicic volcanic rocks and grey and black slate. The volcanic fragments are difficult to identify but in 35727 one fragment resembles a devitrified, flow-banded rhyolite. The quartz crystals measure **0.5** to 3.0 mm and are generally angular or rounded but may show some crystal faces. The embayed forms typical of rhyolitic volcanic rocks are common. Effects of. deformation are shown by undulant extinction and breaking of large grains into subgrains. Plagioclase (0.5 to 3.0 mm), though commonly in broken crystals, preserves some crystal faces. Alteration to sericite is marked and ranges from about 20 to 50 per cent replacement. The matrix is composed of quartz, sericite, chlorite, second generation plagioclase, calcite and accessory sphene, pyrite and rare zircon. Relict vitroclastic texture is found in the matrix of some thin sections (35722, 35725, 35747), but it is not common. In 35722 relict shards, some with complete bubble walls, are well preserved and no significant welding of the shards can be observed. Rare tuffaceous sandstone" bands up to **10** ft in thickness occur within the lithic and crystal tuffs. They have a greywacke -type texture (35731, 35742) and are composed of angular quartz fragments up to 1. 5 mm, muscovite **(0.1** to 1. 0 mm in length), calcite, chlorite and plagioclase. Lithic fragments of granular quartz -sericite and quartz-muscovite rock occur in 35742.

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Fig. 8 - Crystal tuff showing embayed quartz crystal and tabular plagioclase crystal in fine-grained quartz-chlorite-sericite matrix. (35743, x **100,** crossednicols).

Black and grey slate are found as discrete fragments with sharp irregular contacts. The slate has a hornfels appearance (e. g. 35748) and is composed of very fine-grained quartz and sericite. Some of the quartz and sericite is in larger (0.1 mm) aggregates along with calcite. As well, slate occurs in lenses up to 40 ft in thickness, particularly near the boundary of the crystal and vitric tuffs units. This slate is generally contorted and veined with quartz and calcite. $\, {\rm A} \,$ dark grey thin-bedded siltstone, from a slate lens at the boundary of the crystal and vitric tuff units on Cross Section 800S, contains up to 20 per cent of bedded sphalerite. In thin section (Figure 9, 35751) it consists of quartz, chlorite, muscovite, sericite, sphalerite and pyrite with accessory sphene and zircon. As shown in the photograph the sphalerite is in disseminated grains and is richer in the coarser grained beds. In addition to the disseminated sphalerite and pyrite, veinlets of pyrrhotite with rare arsenopyrite can be seen in polished section.

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3.2 Black Slate

The black slate is dark grey to black, thin-bedded, of siltstone or mudstone grain size, and commonly shows a pronounced slaty cleavage. Quartz and quartz-carbonate veins occur throughout and are locally abundant. The grain size ranges up to 0.01 mm

Fig. 9 - Bedded siltstone from slate band in massive pyroclastics, Note sphalerite grains (black) pgri fe ? in coarse-grained bed. \cdot

(35751, x 300, uncrossed nicols).

and averages 0.005 mm. The predominant minerals, as determined from thin section examination and by X-ray diffraction are quartz, sericite and chlorite (e. g. 35754). Fine-grained disseminated carbonate (calcite) is present in some sections (e. g. 35753). The dark colour is due to very fine -grained granular opaque material which is disseminated and in thin (0.05 mm) lenses. The non-carbonate carbon t content of 16 analysed samples of black slate ranges from 0.24 to 0.82 per cent and averages 0.45 per cent (C.E. Gee, pers. comm,.). Zircon and tourmaline are rare accessory minerals.

Pyrite is ubiquitous in minor amounts and occurs in several forms:

- (1) As very fine -grained material in pyrite -rich beds 1. 0 to 3. 0 mm in thickness. Some of this pyrite is framboidal (C.E. Gee, pers. comm.) and is therefore of probable sedimentary-diagenetic origin (Love and Amstutz, 1966).
- (2)- In lenses, measuring about 30 x 5 mm, aligned parallel to the cleavage. The lenses are arranged en echelon and a line through their centres is parallel to bedding. It appears that they were originally lenses parallel to the bedding which have been rotated to parallel the cleavage.
- (3) Relatively coarse grained as disseminated crystals (**1.** 0 to- 2.0 mm) and in veinlets.

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The Co and Ni analyses of the pyrite in the black slate (Loftus-Hills and Solomon, 1967; Loftus-Hills, 1968) are different from the orebody pyrite and fall in the sedimentary field. Sulphur isotope results from the black slate pyrite show a wide spread of s^{34} S values indicating a biogenic sedimentary origin (Solomon et al., in press).

Pyrrhotite veins are not uncommon in the black slate and the pyrrhotite is accompanied by minor amounts of pyrite, arsenopyrite, chalcopyrite and sphalerite (100402).

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In several places a thin band of well-bedded grey siltstone is present between the black slate and the massive pyroclastics. In thin sections (35750) the grey siltstone is composed of quartz and sericite with minor amounts of chlorite, pyrite and sphalerite. The pyrite is in· large ragged masses **1.0** to 2.0 mm in size. Sphalerite. occurs as disseminated grains (0.05 to **1.0** mm) and is amber coloured.

Some coarse-grained sandstone beds (e.g. 35749) occur in the black slate. They have a greywacke-type texture with muscovite, carbonate and angular quartz grains up to **1.5** mm in size in a quartz-. sericite-chlorite matrix. Pyrite, sphalerite are tourmaline are accessory.

Massive coarse-grained (1.5 mm) calcite lenses are well exposed in the black slate in the 12 level 1400N crosscut. The lenses are irregular in shape and range up to 2 ft in thickness and 10 ft in $\,$ length.

3.3 Host Rock

The stratigraphy of the host ${\tt rock}$, black slate and the ore bodies for 12 and 13 levels is shown on a columnar correlation chart (Fig. 10). The main lithology in the host rock is sericitic shale although crystal tuff is locally prominent.

Shale

The sericitic shale is grey-green in colour and locally wellbedded (Figs. 13 and 16), as defined by alternations of relatively coarser grained quartz -rich beds with finer grained sericite-rich beds. The shales are predominantly of siltstone grain size and are composed of quartz, sericite and chlorite with generally minor amounts of pyrite, carbonate and plagioclase. Sphene, tourmaline and zircon are accessory. Some of the shales are cherty in appearance and consist of quartz and sericite.

The slaty cleavage in the shales is generally parallel or subparallel with bedding and imparts a weak schistosity that is marked by parallel alignment of sericite and chlorite. The shales are

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petrographically more correctly termed fine-grained quartz-sericite schists. The texture is typically (Fig. 11, 35721) a granular aggregate of quartz, sericite and chlorite. In the coarser grained beds the larger quartz grains (0.1 mm) are subangular and probably clastic. Generally the quartz forms a fine-grained polygonal mosaic indicative of metamorphic crystallization. As well, some irregular coanse-grained patches, particularly in association with relatively coarse-grained pyrite crystals, are present.

Sericite occurs as minute flakes (0.005 mm) disseminated throughout, in aligned sheaves imparting a weak schistosity, and particularly in the coarser grained beds as elongate flakes about' 0.1 mm in length,

Chlorite is in irregular flakes and aggregates up to 2.0 mm in size but mainly less than 0.05 mm. The colour varies from pale grass-green to almost colourless.

Pyrite is widespread and locally abundant. Bedding is commonly accentuated by pyrite-rich layers 0.3 to 1,0 mm in thickness. This "bedded" pyrite is fine grained and is disseminated or in **.....,** granular masses (e.g. 100973). The Co and Ni values for the "bedded" pyrite are consistent with a sedimentary origin although the spread of values is large and may indicate that some of the metals

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Fig. **11 -** Medium-grained siltstone from host rock composed of quartz, sericite and chlorite, (35721, x 300, crossed nicols).

in the pyrite were contributed by ore-forming solutions (Loftus-Hills, $\dot{\mathbb{P}}$ 1968). Pyrite also occurs as coarser grained (0.1 mm) cubes that are associated with coarse-grained quartz and chlorite.

> Sphene is a constant accessory mineral and is in very finegrained (0.005 mm) granular masses that are commonly associated with pyrite. Tourmaline is widespread in minor amounts and occurs as subangular grains averaging O. 1 mm in size. Zircon in trace amounts occurs as relatively large (0.05 mm) broken crystals and in bundles of minute (O. 02 mm) idiomorphic prismatic crystals. Sphalerite is found in 35711 as scattered grains (0.4 mm) some of which are associated with relatively coarse -grained quartz.

Crystal Tuff

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This rock type constitutes the greater part of the host rock around 2500N on 12 Level and forms a distinct band above the Zn-Pb orebody at the north and south ends of the mine. The band at the south end is about 40 ft in thickness and has so far been traced from 600N to 700S on 13 and 14 Levels.

The crystal tuffs are massive or locally bedded and may show a weak schistosity. They are composed of broken plagioclase and quartz crystals up to 3.0 mm in a matrix of quartz, sericite, chlorite, carbonate, pyrite and sphene, Quartz crystals are less

abundant than plagioclase crystals and are abs ent in some sections \hat{z} , and a set of the state of the quartz-sericite matrix shows a weak schistose fabric. In the bedded tuffs (35717, 35718) bedding is defined by alternations of quartz-plagioclase layers with finer grained quartz -sericite-pyrite layers,

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Plagioclase is in subhedral broken crystals averaging 0.7 mm in size. Alteration is variable with from 20 to 90 per cent replacement by sericite and calcite. A second generation of plagioclase showing chess -board albite twinning is common (e. g. 35706). ,·:i'< Quartz is in broken angular crystals or shows the embayed forms typical of volcanic quartz. The quartz crystals show strong undulant extinction and are commonly fractured into subgrains. Chlorite is variable in amount but generally constitutes less than 10 per cent of the rock. In some sections (e. g. 35706) it is associated with epidote and sphene. The colour ranges from pale green to colourless. Pyrite is widespread and is relatively coarse grained ($0.1\ \mathrm{mm}$). Sphene is in granular masses associated with pyrite. Minor amounts of epidote occur in several thin sections (35706, 35707) where it is . in fine-grained (0.01 mm) granular masses associated with chlorite and sphene. Zircon in subrounded (O. 05 mm) grains and apatite (0.1 mm) are present in trace amounts. Tourmaline is rare or absent. Fine-grained pale yellow sphalerite occurs as rounded

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inclusions in plagioclase in 36075 and 35718.

3.4 Footwall Schists

The footwall schists form an extensive zone stratigraphically underlying the host rock. At the north end of the mine and below 14 Level between lOON and 500N quartz schist or chert equivalent to the footwall schists forms the footwall of the Zn-Pb orebody. Elsewhere up to 60 ft of host rock intervenes between the footwall of the orebody and the footwall schists. The stratigraphically higher part of the footwall schist is commonly highly siliceous and has been referred to as the "quartz schist" by Hall et al. (1953). However, as quartz schist is not always present at this stratigraphic level and because "quartz schist" type lithologies recur at lower stratigraphic levels in the footwall schists this latter name is preferred. The footwall schists grade downwards and laterally into crystal tuffs of the footwall pyroclastics. The maximum thickness of footwall schist is in the central and southern part of the ore deposit. To the north the footwall schist zone is about 50 ft in thickness on the surface at 2500N, while at 4000N black slate is underlain by crystal tuff and both the host rock and footwall schist are absent.

Cherts

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Massive chert equivalent to the "quartz schist" occurs in places along the footwall of the Zn-Pb orebody and at various levels in the footwall schists. In addition, chert lenses up to 2 $\text{x} \frac{1}{2}$ ft and are not uncommon in the Zn-Pb orebody.

The cherts along the footwall of the Zn-Pb orebody at the north end of the mine (e. g. 35811) are composed of a very fine-grained (0.005 mm) aggregate of quartz and sericite. Chert (35812)forming the footwall of the orebody at 600N on 17 Level consists of an aggregate , of quartz and aligned sericite with disseminated pyrite. Some relativelycoarse-grained (0.2 mm).quartz grains are present. One occurrence of chert (35716) from the footwall schists immediately • below the host rock consists of a fine-grained granoblastic quartz mosaic with sparsely disseminated sericite and barite.

Schists

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In the field the schists are fine grained and grey-green with a distinct schistosity. Petrographically they could be described as phyllonites. They are composed of quartz, sericite and chlorite with minor amounts of pyrite and sphene. Carbonate occurs: locally and zircon and apatite are rare accessory minerals. Textures vary from weakly to strongly schistose. The schistosity is imparted by aligned sericite and chlorite aggregates and in some rocks by quartz lenses consisting of an aggregate of quartz grains. Generally no trace of the original texture can be seen, but exceptionally, as in 35769, areas

showing a parallel streaky structure and irregular arcuate shapes .. indicate relict collapsed pumice and vitroclastic textures. Some of the relatively coarse-grained quartz grains may represent original volcanic quartz crystals.

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Most of the quartz is in a fine-grained $(0.005 \text{ to } 0.1 \text{ mm})$ mosaic of polygonal grains but some is in relatively coarse-grained patches (0.3 mm), the component grains of which may show sutured boundaries. 'Sericite is in irregular flaky aggregates and also disseminated in fine-grained (0.01 mm) flakes. Chlorite is found in irregular streaky aggregates measuring about 4.0 x 1. 0 mm and less commonly as scattered flakes (0.02 mm). The pleochroism is grassgreen to almost colourless and the interference colours are very dark or in some instances the anomalous berlin blue colour indicative of penninite. In the chlorite-rich schists (e. g. 35764, 35767) large plates of chlorite aggregates contain scattered patches of quartz and sericite in a sieve type texture. Pyrite occurs in disseminations, in massive bands and in veinlets. $\;$ It is generally idiomorphic and the grain size ranges from 0.02 to 0.7 mm.. The larger grains display pressure fringes of quartz, chlorite and sericite. Carbonate occurs sporadically in irregular granular clusters measuring about **0.2** mm. Mainly it appears to be calcite, but in 100876 and possible 35770 blebs of dark coloured ferroan rhodochrosite are abundant. Sphene is in

loose granular clusters and is associated with pyrite and chlorite. An epidote type mineral occurs in 35767 where it is associated with $sphene.$ Sphalerite is found as rare veinlets (35771) and in aggregates (2.0 mm) surrounded by an envelope of relatively coarse-· grained (0.2 mm) quartz, carbonate and chlorite.

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3.5 Footwall Pyroclastics

The footwall pyroclastics as a local term refers to that part of the Primrose Pyroclastics that underlies the footwall schists in the mine area. They grade upwards into the footwall schists and within themselves contain zones of schistose rock.

The footwall pyroclastics are predominantly crystal tuffs. Lithic fragments are rare while pumice fragments are common, which is the reverse of the situation in the massive pyroclastics . . The pumice-bearing tuffs (e. g, 35774, 35775, 35776) contain patches up to 4.0 mm in length displaying a parallel streaky structure. The patches are now composed of a fine-grained aggregate of quartz, second generation plagioclase, sericite and less commonly chlorite. Some of the pumice fragments are crumpled and folded by the schistosity (35776).

Mineralogically the footwall pyroclastics are very similar to the massive pyroclastics. They are composed of plagioclase and

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quartz crystals in a matrix of quartz, plagioclase, chlorite, sericite, sphene and pyrite. $\;$ Zircon and apatite are present in trace amounts: while carbonate and epidote occur in some of the tuffs.

3.6 Plagioclase in the Tuffs

Determination of the composition of the plagioclase in the tuffs of the Primrose Pyroclastics is difficult because of the marked alteration of the original volcanic crystals. From extinction angles and positive optic sign the composition is inferred to be albite. Measurements on extinction angles perpendicular to *x* indicate a composition close to An_{o} . X-ray diffraction values for 20(131) - 20(121) on 5 samples plot close to a composition of An_{0} on the low-temperature structural state curve of Smith and Yoder (1956). Solomon (1964) and R. Hall (1967) found compositions close to An_{o} with low-temperature optics for the plagioclase in the Primrose Pyroclastics.

Plagioclase occurs in two forms in the Primrose Pyroclastics. The first is the original volcanic plagioclase crystals which except for deformation effects have suffered little change in form. They are generally considerably altered and replaced by sericite and presumably were originally more calcic in composition. Plagioclase in non-metamorphosed rhyolitic tuffs is calcic oligoclase or andesine.

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changed interface crystals that are found as matrix formers, as over-The second form is in relatively fine-grained, unaltered, growths around original plagioclase and rarely as· replacements of the original plagioclase. This second generation of plagioclase shows { either simple albite twinning or, very commonly, chess-board albite twinning. The feldspar with chess-board albite twinning appears to \mathfrak{g} be what Solomon (1964) described as potassium feldspar* showing an uneven grid-iron or checkerboard twinning. However, as shown by negative stain reactions with sodium cobaltinitrite and positive optic sign this feldspar is plagioclase and matches descriptions of chess-board albite (e,g. Starkey, 1959). The matrix plagioclase (e. g. 35706, 35781) is in aggregates of irregular interlocking grains, averaging 0.02 mm, and is generally coarser grained than quartz in the matrix. These plagioclase grains show either patchy extinction, discontinuous spindle and wedge-shaped twins or vague cross-hatch twinning typical of chess-board albite twinning. Starkey (1959) considered chess -b'oard albite twinning to be deformation twinning and certainly in the Primrose Pyroclastics it is best developed in the more sheared tuffs. Chess -board albite is believed to form from either potassium feldspar or plagioclase under low-grade metamorphic conditions of the chlorite and biotite zones (Starkey, 1959; Brown, 1965). "

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 $\%$ Solomon (1964, p.130) reported that ".... most of the Primrose Volcanics contain individual crystals of K-feldspar ". Although the presence of K-feldspar in some of the Primrose Pyroclastics is not " disputed it was not identified in any of the writer's thin sections or in the thin sections examined by R . Hall (1967).

3.7 Metamorphism

The rocks in the mine area have been affected by a low-grade regional metamorphism that has produced mineral assemblages of the lower greenschist facies together with deformafional fabrics.

The characteristic mineral assemblages are quartz-sericitechlorite in the black slate, host rock shale and footwall schists, and quartz-albite-sericite-chlorite in the pyroclastic rocks. Sphene and pyrite are ubiquitous in minor amounts.

On a macroscopic scale the obvious deformational effect is zones of schistose rocks. As well as the footwall schists there are. a number of relatively narrow schistose zones within the host rock and in one area in the massive pyroclastics overlying the host rock. These zones grade laterally and across the section into less sheared rocks.

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Microscopically the deformational effects are shown by parallel alignment of sericite and chlorite; by granulation, by undulant extinction and deformation lamellae in quartz crystals; by granulation and rotation of plagioclase crystals; by the formation of chess-board albite; and by the growth of pressure fringes on pyrite and arsenopyrite crystals. The larger (**O.** 3 mm) pyrite grains in the footwall schists, host rock and black slate are bordered by small

pressure fringes of quartz, chlorite and less commonly sericite and carbonate. The quartz shows dimensional orientation with elongate crystals perpendicular to the crystal face of the pyrite. In 35786 from the host rock, arsenopyrite crystals up to $2.0 \:\rm mm$ in length display large pressure fringes up to 3.0 mm in length (Fig. 12). The fringes are composed of quartz and chlorite showing pronounced dimensional orientation perpendicular to the arsenopyrite crystal faces. Some of the fringes are curved towards the ends which indicates that the arsenopyrite crystals have been rotated during the growth. of the pressure fringe. Many of the arsenopyrite crystals are fractured. The implication of the pressure fringes (see Spry, 1963) is that the pyrite and ars enopyrite crystals were present in the rock before the main metamorphic crystallization.

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3.8 Origin of the Footwall Schists

The footwall schists grade both laterally and downwards into crystal tuffs of the footwall pyroclastics. Some of the footwall schists contain relict pumiceous and vitroclastic textures and it therefore appears that they were derived from crystal tuffs of the footwall pyroclastics. The reason why only parts of the Primrose Pyroclastics have been converted into schists is a problem of considerable interest. The favoured explanation is that zones within the

Fig. 12 - Arsenopyrite crystal with pressure fringe of quartz and chlorite showing dimensional orientation.

(35786, x 100, crossed nicols).

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Primrose Pyroclastics underwent alteration by hydrothermal volcanic solutions, ...perhaps the source solutions for the ore, which made them more responsive to later deformation and metamorphism (Loftus -Hills et aL, 1967). Similar explanations have been proposed for the formation of the "Lyell Schists" (Loftus -Hills et aL, 1967; Markham, 1968) and of schistose zones enclosing pyritic deposits in the Urals (Zavaritsky, 1950).

Finucane (1932) described the gradational alteration from massive porphyries (crystal tuffs) to sheared porphyries and quartzsericite-chlorite schists of the Primrose Pyroclastics. He presented chemical analyses (Appendix 2) of the altered and unaltered rocks and tried to distinguish chemical differences between them. Except for an apparent enrichment in potassium in two of the analyses of altered rocks the compositions of both altered and unaltered rocks are variable and probably reflect the rather heterogeneous composition of the Primrose Pyroclastics, as pointed out by Solomon (1964).

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3.9 Origin of the Pyroclastic Rocks

The characteristics of the Primrose Pyroclastics relevant to their origin are summarized as follows:

> (1) They are pyroclastic rocks and contain crystal, lithic and vitric material.

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- (2) Some of the Primrose Pyroclastics contain pumice fragments.
- (3) They are rhyolitic in composition.
- (4) They consist mainly of thick units of non-sorted and unstratified tuffs.

These characteristics satisfy some of the criteria for the recognition of ignimbrites or ash-flow tuffs (Ross and Smith, 1960; Cook, 1966), but the other important criteria of the presence of welded zones and the wide areal extent of individual flow units have not yet bee n demonstrated. It is possible that welded textures may not have been preserved especially if they are more susceptible to obliteration by deformation and recrystallization such as the Primrose Pyroclastics have suffered. R. Hall (1967) figures glass shard textures from tuffs of the Primrose Pyroclastics which look similar to welded textures. The areal extent of individual pyroclastic units is uncertain because of poor exposures but from subsurface information individual lithologic units in the massive pyroclastics can only be traced for about 3,000 ft along strike.

Thus while an ash-flow tuff origin appears likely, the features of large thickness of unsorted material, apparent lack of welding and lateral impersistence could be produced by air-fall deposition near eruptive vents or by dry avalanches of loose pyroclastic material in

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the vicminty of eruptive vents (Smith, 1960). The latter mechanism probably grades into true ash-flow eruptions and possibly all three mechanisms contributed to the deposition of the tuffs of the Primrose Pyroclastics.

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3.10 Environment of Deposition of the Primrose Pyroclastics

The presence of possible ignimbrites and the absence of stratification and sorting in the Primrose Pyroclastics strongly indicate that they were deposited on.land (e. g. Rankin, 1960; S olomon 1964). $\;$ The shale lenses in the Primrose Pyroclastics $\;$ ${\tt would \ have \ required \ quite \ water \ depositional conditions \ and \ of \ the}$ poss ible subaerial environments either a lagoonal or a lacustrine environment seem most likely.

Some of the Pleistocene and Recent volcanic rocks of the Taupo volcanic'zone, New Zealand, are possibly younger analogues of the Primrose Pyroclastics. In-particular there are similarities to the Huka Group in the Wairakei area (Grindley, 1965). The Huka Group consists of up to 4, 000 ft of lacustrine mudstones and siltstones, pumice sandstones, pumice breccias, lap'illi tuffs, vitric-crystal tuffs. ignimbrites and rhyolites. The lacustrine sediments are well-bedded dark to medium grey mudstones, with thin siltstone bands, that are

laterally quite extensive and range in thickness from 10 to 100 ft. Compared to the sediments of the Huka Group the shales of the Primrose Pyroclastics are less extensive and pumice sandstones are lacking but otherwise the thickness and lithologic associations are comparable.

3.11 Dolerite Dikes

Dikes of basaltic composition are found in the mine area and west of Rosebery. In the mine there is one main dike about 3 ft in thickness that has been traced for 4,000 ft from 200N to 4000N. Around 1400N on 12 Level another thinner dike is located about 20 ft east of the main dike and may be a bifurcation from it. The dike trends at 350° and dips at 50° to 60° east more or less parallel to the cleavage'direction. In places the dike steps up to 10 ft from one cleavage plane to another. The dike cuts through both the Zn-Pb orebody and the enclosing rocks and has chilled margins against the ore and the host rock.

In hand specimen the dike rock is fine-grained, massive and dark grey to black in colour with scattered phenocrysts of a dark green mineral. The textures are intersertal with patches of dark brown glass, and plagioclase and augite crystals in an intergranular arrangement. Larger orthopyroxene phenocrysts (0.2 to 1. 0 mm) which are commonly altered to chlorite (6898), are scattered

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throughout. The chilled contact phases (35783, 35784) are microchcc porphyritic with phenocrysts of plagioclase, augite, and orthopyroxene altered to chlorite in a dark isotropic glassy matrix. Carbonate and sericite are locally common as alteration products. In the occurrence of orthopyroxene with finer grained plagioclase and augite and the absence of olivine these rocks are very similar to the fine-grained marginal phases of the Jurassic dolerite (Spry, 1962). The other possible basaltic source is the Tertiary basalts which crop out north of Rosebery. The absence of olivine makes this source unlikely because almost all the Tertiary basalts of Tasmania are olivine bearing (F. L. Sutherland, pers. comm.). Also, similar dikes at Renison Bell are chemically similar to the Jurassic dolerite (D.l. Groves, pers. comm.).

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The dolerite dike is of particular interest because where it cuts through the Zn-Pb orebody it contains galena-chalcopyrite veinlets. In Hall et al. (1965) the presence of ore sulphides in the dikes was regarded as evidence that the dikes were pre-ore and the ore was therefore epigenetic. However, the chilled contacts against the ore and layering in the ore displaced by the dikes are clear evidence that the dikes are post-ore. The galena-chalcopyrite veinlets (e. g. 35783) occupy what appear to be tensional fractures approximately perpendicular to the walls of the dike. As seen in polished section $\pounds 7-29)$ *

^{&#}x27;~7-29 and similar numbers refer to polished sections in the Geology Dept. University of Tasmania.

one of the veinlets is composed of about 50 per cent chalcopyrite, ^r40 per cent galena, **10** per cent pyrite and less than one per cent sphalerite. Some of the pyrite and chalcopyrite is in elongate rodlike bodies which are parallel and are oriented approximately at right angles to the cleavage planes of the galena. This texture has similarities to some eutectic textures figured in Edwards (1954). One sphalerite grain contains elongate sphalerite blebs that are -, oriented in a grid pattern.

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The most likely explanation for the origin of the galena chalcopyrite veinlets in that they are a remobilized fraction formed during the emplacement of the dikes (e.g. Mookherjee, 1964). Experimental work on the Cu-Fe-Pb-S system (Craig and Kullerud, 1965-66, 1967) and the Cu-Pb-S system (Craig and Kullerud, 1968) has shown that a Cu-Pb-Fe-S melt from which pyrite can crystallize is formed at 547° C and first melting of chalcopyrite-galena assemblages takes place at 630° C. The temperature of basaltic magma is of the order of 1200° C and even taking into account rapid cooling at the dike contact an adequate heat source is available to produce fractional melting of a Cu-Pb-Fe-S liquid from the adjacent ore. While a comprehensive study has not been made, pyrrhqtite appears to absent in pyrite ore adjacent to the dikes which indicates that the temperature of ore at the dike contacts did not reach 743° C.
The maximum temperature reached in the ore at the dike contact was thus between the limits of 630° C to 743° C.

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4, . STRUCTURE OF THE ORE DEPOSIT

This section is mainly concerned with the detailed fold structure of the ore deposit as exposed on the **11,** 12, 13 and 14 Levels of the Rosebery Mine. The fold structure has been investigated by analysis of the geometry of the mesoscopic structural elements and the macroscopic folds in the Zn-Pb orebody and the host rock and biack slate.

Hills (1915b) recognized that the orebodies were folded and he postulated two directions of folding at right angles on north and east trends. Hall et al. (1953) observed small scale folds in the ore and host rocks but they thought that folding was not important in the overall structure. However in 1963, Hall et al. (1965) suggested that the orebodies and the host rocks had been involved in a large fold and that an en echelon fold pattern was present. Since then the overall fold structure has been confirmed by V. M. Cottle, R. D. Pratten, J.:G. Druett and others of the geological staff of the Electrolytic Zinc Company, Rosebery.

4.1 Mesoscopic Structures and their Macroscopic Geometry

Bedding in the Host Rock and Banding in the Ore

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Bedding in the host rock and black slate is generally easily

seen. The host rock and slate varies from thin-bedded to massive, poorly-bedded types and in consequence bedding is rare in some areas. Locally, bedding has been partly transposed by cleavage but this does not usually obliterate the bedding.

The ore-host rock contacts are sharply defined and are concordant with bedding in the host rock (Fig. 13). They provide a useful structural reference surface. Much of the ore displays compositional banding (Figs. 18, 19 and 21) which is concordant with ore/host rock contacts and rare shale bands in the ore.

Orientation: - Figure 14 shows equal-area projections $(p^{j}$ -diagrams) of poles of bedding, ore/host rock contacts, and banding in ore. The pi-diagram for bedding in the host rocks includes all available measurements from 12, 13 and 14 Levels. While sampling is of necessity uneven $-$ bedding is better developed in some areas and mine openings are mainly confined to the orebodies and their immediate vicinity $-$ the position of the point maximum is well defined and probably satisfies the requirement of reproducibility. The data for the ore/host rock contacts and banding in ore represent a complete cover of the orebody as exposed between 1000N and 1000S on 14 Level. Other pi-diagrams for ore/host rock contacts and banding in ore were prepared for 13 Level, 12 Level and subareas of 13 and 14 Levels. They all showed similar point distributions and

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Fig. 13 - Concordant contact between bedded shale of host rock and banded $Zn-Pb$ -Fe sulphide ore.

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are consistent with tight folds in which fold limbs are dominant over fold hinges (see Ramsay, 1964). There are a few points scattered in an east-west direction and a weak north-south elongation is present, but no girdles are developed. The lack of girdles indicates that folding is inhomogeneous over the area covered by the sampling for the pi-diagrams (Ramsay, 1964). Attempts to divide 13 and 14 Levels into homogeneous subareas were unsuccessful because of variable fold plunges.

Cleavage

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A slaty cleavage is present in the host rock and black slate. It is parallel to the axial surfaces of all observed folds. Because folds are tight and tend to be isoclinal the cleavage is parallel to bedding on the limbs of the folds and results in a weak bedding schistosity. Cleavage is generally absent from the sulphide ore, but as the ore contains very little mica or other platy minerals the formation of a slaty c'leavage is hardly to be expected. Exceptionally, as illustrated in Figure 20, a faint axial-plane cleavage can be seen in the ore.

Orientation: - Figure 14 shows the orientation of 50 poles of cleavage planes in the hest rock and black slate from 12, 13 and 14 Levels. There is a strong point maximum corresponding to a strike of 350°

and dip of $5\overset{1}{8}$ ^O east. The spread of points around the maximum is small and illustrates the uniform orientation of the cleavage. Cottle (1958) reported that the average strike and dip of 250 cleavage measurements in the host rock and black slate was $350^{\sf o}$ at $55^{\sf o}$ east. and that 95 per cent of the measurements were within 20 $^{\circ}$ of this figure. The much higher degree of preferred orientation of the cleavage compared with bedding, ore/host rock contacts and banding in ore is emphasised.

Folds

Orientation of Fold Hinges: - The attitudes of 52 measured fold hinges are shown in Figure 15 in which fold hinges in host rock, in ore/host rock contacts and in banded ore are distinguished separately. The sampling probably gives a fair cover of the Zn-Pb orebody and its vicinity from 12 Level to 14 Level between 800S and 1500N. The distribution of all three types of fold hinges appears to be similar but the number of measurements is too small to be conclusive. The mesoscopic folds are of variable trend and plunge ranging from flat north and south· plunges to north-east, east and south-east trending folds of moderate plunge. The distribution spreads in a girdle that coincides closely with the mean cleavage plane (Fig. 15).

The mesoscopic folds are generally congruous with nearby

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macroscopic folds. They are symmetrically related to the macroscopic folds in the Zn-Pb orebody, and the change in shape of the mesoscopic folds is used at Rosebery, in the sense of "drag folds", to distinguish the opposite limbs of the macroscopic folds.

Fold Style in Host Rock and Black Slate: - In profile the folds are tight with inclined axial surfaces. . Slaty cleavage parallel to the axial surface is generally prominent (Figs. 16 and 17). The fold style conforms very closely to similar folds. Bed thickness measured parallel to the axial plane is constant throughout the folds and orthogonal thickness measurements plot in the field of similar folds on the orthogonal thickness ratio graph of Ramsay (1962a). Competence differences between beds are small or absent and bedding has behaved passively during deformation. Folds with either strongly sheared limbs or completely detached hinges occur locally.

Fold Style in Ore: - Folds in ore tend to have rather irregular geometry (Fig. 18). Fold hinges in one compositional band are commonly disharmonic with adjacent bands. Fold style ranges from similar to concentric. Similar styles are more common in the examples studied (Figs. 19 and 20). They show constant thickness parallel to the axial planes and plot in the field of similar folds on the orthogonal thickness ratio graph. Some other profiles tend

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Fig. 16 - Similar style folds in thin-bedded shale of host rock. Note sub-vertical cleavage.

Fig. **17 -** Similar style fold in black slate showing cleavage parallel to axial surface.

- Fig. 18a Folds in layered Zn-Fe-Pb sulphide ore. Note concordant contact with bedded shale in top left corner .
- Fig. l8b Traced from photograph.

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Fig. 19 - Similar style fold in layered sulphide ore. Sphalerite-galena-rich layers dark grey, pyrite-chalcopyrite-rich layers light grey. Fig. 20 - Similar style fold in layered sulphide ore showing cleavage parallel to axial surface. Sphalerite-galena-rich layers (dark grey), pyrite-rich layers (light grey.). Also shows discordant replacement by pyrrhotite (yery light grey),

towards concentric forms (Figs. 21 and 22). One example as illustrated in Figure 22 is .a concentric fold with 50 to 60 per cent flattening on the orthogonal thickness ratio graph. The folds showing concentric geometry occur in the more competent layers $-$ pyrite layers in sphalerite -galena rich ore, or cherty shale layers in sulphide ore. The apparent contradiction of the sulphide exhibiting both similar and concentric fold styles can be explained by the sulphide in bulk having a low mean competence but with locally high internal competence contrasts as between pyrite-rich and sphaleritegalena-rich layers (e. g. Donath and Parker, 1964). The mechanism of folding is governed by the behaviour in bulk and as the sulphide is of low mean competence similar style folding predominates. Where high internal competence contrasts are present the layering controls the fold mechanism and concentric fold styles form in the most competent layers, This is the mechanism of quasi-flexural folding of Donath and Parker (1964).

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Folds Involving Ore/Host Rock Contacts: - Folds in the ore/host rock contacts are generally congruous with folds in the adjacent host rock (Fig. 23). A feature of the hanging wall contacts, particularly in the sphalerite -galena-rich southern part of the orebody, is thin prongs of massive ore extending for up to several tens of feet into the host rock. The form of the prongs suggests that they are tight,

Fig. 21 - Concentric style folds in finely layered sulphide ore. Pyrite-rich layers light grey, sphaleritegalena-rich layers medium grey, chert layers dark grey.

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Concentric style folds in layered sulphide ore. Fig. $22 -$

attenuated anticlines. Commonly a larger anticline terminates by breaking down into a number of smaller prongs as illustrated in Figure 23.

One locally discordant contact between bedding and an ore/hangingwall contact (Fig. 24) has been observed. Bedding dipping gently east can be followed to within about two inches of a near vertical contact. The ore transgresses bedding for about three feet along this contact and then closes off in an anticlinal prong.. The discordant contact is interpreted as having been formed by diapiric intrusion of sulphide ore into an anticline in the overlying host rock. It is inferred that more mobile SUlphide would flow at a greater velocity when deformed and would diapirically intrude its less mobile host rock in the manner of rheid material (Carey, 1954). Diapiric -like movement has probably occurred elsewhere along anticlines involving the hangingwall contact, as might be inferred on a larger scale from Figure 26. At the Hercules Mine the tops of some of the orebodies appear to diapirically intrude anticlinal folds in the overlying black slate.

Lineations

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Lineations formed by the intersection of bedding and cleavage (L1) are rare because of the common parallelism of bedding and

Fig. 23a - Folds in ore/host rock contact along hangingwall of Zn-Pb orebody.

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Fig. 23b - Traced from photograph with additions from field sketch. $\sim 10^{11}$ km

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cleavage. In the hinges of folds where cleavage cuts bedding at a high angle the lineations can be measured and they are parallel with the adjacent fold hinges.

A second lineation (L2) is locally developed near the footwall contacts of the Zn-Pb orebody and less commonly near the hangingwall contacts. The lineations mainly occur north of 00 and where the orebody dips more steeply. As illustrated in Figure 25 they consist of discontinuous, parallel striations and are confined to prominent surfaces close to the contacts and have a polished appearance. The pitch of the lineations generally ranges from 70[°] to 90[°]. The orientation of 50 L2 lineations is shown in Figure 14. There is a strong point maximum and the mean strike and dip of the surfaces on which the lineations were measured passes through the highest contour, indicating that the majority of the lineations are parallel to the direction of dip of the contacts.

The lineations are thought to have formed during relative movement between the ore and the host rock. It is suggested that as the more mobile sulphide moved under stress, surfaces of slip were formed in the adjacent less mobile host rock to take up the relative movement between them. The lineations would be parallel to the direction of movement and are analogous to slickensides formed by bedding plane slip in concentric folding.

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Fig. 25 - L2 lineations on surface parallel to and about 6 inches below footwall of Zn-Pb orebody.

Miscellaneous Structures

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Other structural features which occur sporadically are kink folds and quartz and carbonate veins. The kink folds have quartz filled joints through their axial surfaces and appear to have formed by drag on the joints. The cleavage is folded by the kink folds.

Extensive quartz and carbonate veining is present in the black slate exposed in the 1400N cross-cut east on 12 Level. The veins have a wide variety of attitudes and may be concordant with bedding or cleavage, or transgressive. The concordant veins commonly display boudinage structure. Some veins are folded into irregular ptygmatic-type folds. Considerable mobility of quartz and carbonate during deformation is indicated.

4. 2 Macroscopic Folds

The macroscopic folds have been outlined by using the Zn-Pb orebody as a marker bed. Some additional information is provided by using the contacts of the host rock with the black slate and the footwall schist as marker horizons.

In cross section (Fig. 26 and Fig. 27) the macroscopic folds are overturned with axial surface dips varying from 45° to 85° east, but mostly in the range of 55° to 60° east. Each fold consists of a

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fold-pair comprising a long west limb, a synclinal hinge, a relatively short overturned limb, an anticlinal hinge, and a long eastern limb. These fold-pairs are "drag folds" but this term is not favoured because it is not a satisfactory descriptive term and its genetic implications may be incorrect (Ramsay 1967, p. 397).

The shape of the folds in vertical longitudinal projection is illustrated in Figure 28. The crest and trough lines in longitudinal projection (Fig. 27 and Fig. 28).are not true fold plunges. That would be so only if the axial surfaces were vertical. In the present case where axial surfaces dip east at moderate angles only horizontal crest and trough lines are true fold plunges, all others are inclined at steeper angles than the plunges. Folds above 10 Level are not shown in Figure 28 because available information on them is incomplete.

The main features of the fold pattern as shown in longitudinal projection (Fig. 28) are:

(1) Southerly plunges south of 200S.

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(2) A change from southerly to northerly plunging folds between 200S and 00.

 (3) Predominantly northerly plunges from 00 to 1500N.

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 $\Delta \phi$ $\sim 10^{11}$ and $\sim 12^6$

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(4) A plunge reversal from southerly to northerly plunges near 2200N.

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(5) Predominantly northerly plunges north of 3000N.

A plunge change from northerly to southerly plunges is implied between 1500N and 2200N but this cannot be confirmed because of lack of data.

The fold structure of the central and southern part of the Zn-Pb orebody between 11 Level and 14 Level is shown in greater detail on a composite plan combined with cross sections and a longitudinal projection $(Fig. 27)$. Each fold-pair is numbered so that folds can be followed from plan to section and longitudinal projection. Individual folds generally show greatest amplitude in the hangingwall contact of the orebody. The same fold in the footwall contact is either of much smaller amplitude or is just a gentle roll in the contact. Exceptions to this are the synclines of the three largest folds in the orebody 84, 89 and 817. The smaller folds in the orebody are not reflected in the black slate and footwall schist contacts and the large fold-pairs A4-84, A9-89 and A17-8l7 are of smaller amplitude in these contacts.

A characteristic feature of the folds in the orebody is that individual folds are of limited axial extent; they wax and wane.

The crest and trough lines are curved and they approximate to .:... conical folds (Wilson, 1967) .

Two types of folds are present in the orebody. The predominant type are fold-pairs 'which in logitudinal projection have cresttrough outlines that close at both ends. This type corresponds to the pod folds of Mendelsohn (195 9a) or, when the fold-pairs are . arranged en echelon, to the en echelon elliptical folds of Campbell (1958). However, as pointed out by Mendelsohn (1959b), the cresttrough outlines are commonly not elliptical and the term pod fold is preferred because it has a more general usage, The other fold type at Rosebery is that of anticlines linked by smaller shared synclines. This is the echelon type of Mendelsohn (195 9a) or the zig-zag en echelon type of Campbell (1958). En echelon fold patterns are a natural consequence of conical folds (Wilson, 1967).

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At the south end of the orebody a group of pod folds A1-S1 to A6-S6 form a right hand en echelon pattern. Each pod fold develops down dip and to the right of the previous one (Fig. 27, Section 500S). The region of change in fold plunge around 00 is complicated and is not well understood, but there appear to be three overlapping pod folds A7-S7, A8-S8 and A9-S9.

The largest fold A9-S9 in the orebody begins near 00, grows

to a maximum near 600N (see Fig. 28) and then appears to rapidly die out. The syncline S9 shows a marked flattening in plunge near 500N on 16 and 17 Levels and this, together with drill hole data, suggests that the crest and trough lines would meet near 800N at R. L. 9400. The fOld-pair A9-S9 has an overall moderate to steep northerly plunge although there are some local flat and southerly plunges, particularly along the anticlinal crest. In detail the anticline A9 consists of four overlapping anticlines each linked by a small syncline, and corresponding to the anticlinal echelon fold type. A number of small pod folds $-$ A10-S10, A11-S11, A12-S12 and $A13-S13 - occur$ on the west flank of the syncline S9. They all have steep northerly plunges except for A11-S11 which plunges southeast, at 45° to 50° . The northerly plunging pod folds show a left hand en echelon arrangement as is to be expected for congruous "drag folds" on the west flank of a north plunging syncline .

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The major fold A9-S9 is also reflected in the black slate and footwall schist contacts (Fig. 26 and Fig. 28). In the black slate contact A9-S9 begins as a gentle roll on 9 Level and has been traced . for over 200 ft down its axis. It maintains a fairly constant amplitude and· plunge over this distance. In comparison with the same fold in the orebody, the crest-trough outline in the black slate contact is much more regular. The anticline A9 in the black slate

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has been intersected by the 1400N cross-cut east on 12 Level and , consists of numerous asymmetric mesoscopic folds with axial surfaces dipping east at about 60° .

None of macroscopic folds persist into the overlying massive pyroclastics. For example, the major fold A9-S9 is represented by only a slight undulation in the massive pyroclastic/black slate contact.

4.3- Interpretation of the Fold .Geometry

Mesoscopic folds in the ore and in the host rock-black slate are generally similar in orientation and style and therefore belong to the same phase of deformation. Departures from the general similarity are attributed to the greater tectonic mobility of the sulphide ore as compared to the host rock-black slate. Likewise the. mesoscopic folds are generally congruous with and therefore syn'chronouswith nearby macroscopic folds.

The mesoscopic and macroscopic folds show variable trend and plunge, particularly in the $\operatorname{\mathsf{Z}}\nolimits{\mathsf n}$ -Pb orebody, while axial surfaces are of reasonably constant orientation throughout the mine area. Folds having that geometry could be produced by (see Hobbs, 1965 p. 20) either superposition on an already folded surface and implying

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two periods of folding, or by· heterogeneous strain in the axial surface in one period of folding. At Rosebery, the latter mechanism seems more likely (see Ramsay, 1962b, **p.472;** Turner and Weiss, 1963, **p.** 508; Stauffer and Rickard, 1966, p. 429) because:

- (a)-Individual folds are of limited axial extent.
- (b)- Change in fold plunge is accompanied by regular -changes in profile.
- (c)-Change in fold plunge is not usually large.
- (d)-Only one axial surface cleavage is present.

The heterogeneous strain in the axial surface is envisaged as taking place by differential flow parallel to the strike of the axial plane (Woll, 1960). By this mechanism individual folds can rapidly grow and die out along their axes.

The predominance of similar fold styles suggests that the principal fold mechanism in both the sulphide ore and the host rockblack slate was simple shear by laminar flow parallel to the axial surfaces of the folds. An east-side-up simple shear couple is indicated by the sense of asymmetry of the "drag folds" in the Zn-Pb orebody.

Compared to the folds in the host rock and black slate the folds in the Zn-Pb orebody show less regular fold geometry, greater

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fold amplitude and greater intensity of folding. These differences .together with evidence of local diapiric-like movement indicate that the sulphide ore was highly incompetent with respect to the enclosing rocks. It has apparently behaved like a rheid material (Carey, 1954) during the folding. This leads to the suggestion that the presence of a mass of highly incompetent sUlphide may have modified the external stress system and caused the heterogeneous strain pattern.

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Implicit in the foregoing analysis is the inference that the ore was present before the folding took place. This is justified because if ore deposition was post -folding it is difficult to find a good explanation for the differences in fold style between the ore and the host rock.

4. 4. Faulting

Faulting is of minor importance. The majority of some 30 faults mapped in the vicinity of the Zn-Pb orebody on 11, 12, 13 and. 14 Levels strike approximately north-south and dip east at from 40° to 80° . The faults have narrow crush zones and displacements of up to 5 ft at the most.

Some faulting has occurred near the contact of the massive pyroclastics with the black slate or host rock, but where these faults

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intersect the orebodies no significant displacements occur. One of these faults is well- exposed in the 1400N cross -cut east on 12 Level. This fault is located in a sheared dark green rock about 5 ft below the first massive pyroclastic rock. There is a crush zone of white fault pug about 9 inches wide and well developed drags against the fault plane indicate relative east-side-up movement. This faulting is thought to reflect movement localized near the contact of the very competent massive pyroclastics and the relatively incompetent slate during deformation.

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4.5. Structural History

Two orogenies have affected the Cambrian rocks in west Tasmania. The first is the late Cambrian to early Ordovician Jukesian Orogeny and the second is the Devonian Tabberabberan Orogeny. Solomon (1962, 1965a) suggested that there were two phases of folding in the Tabberabberan Orogeny: an early phase of long-wavelength folds on mainly north-south trends controlled by preexisting Precambrian and Cambrian trends and a later phase of north-north-west folding. This later phase was associated with a well developed axial surface cleavage.

The structural history at Rosebery is tentatively outlined as follows.

- (1) Emplacement of the orebodies as concordant lenses, possibly by sedimentary deposition, during the Cambrian volcanism.
- (2) Eastward tilting of the orebodies and enclosing rocks. This could have taken place during either the Jukesian Orogeny or the first phase of the Tabberabberan Orogeny.
- (3) Subsequently, the present fold pattern and associated axial surface cleavage were produced during the second phase of the Tabberabberan Orogeny. Recrystallization of the ore and metamorphism of the enclosing rocks were probably associated with this event.

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5. ORE MINERALIZATION

Prior to mining activity Pleistocene glacial till covered much of the mine area and fresh sulphide ore occurred close to the base of the Pleistocene cover. The ore deposit was in fact discovered from the presence of Zn-Pb sulphide boulders in Rosebery Creek. One small outcrop of limonitic gossan containing significant amounts of manganese oxides occurs on the surface.

There are two spatially and compositionally distinct orebodies here termed the Zn-Pb orebody and the Barite orebody. The Barite orebody, which is the smaller of the two, overlies the Zn-Pb orebody (Fig. 26 and Fig. 29). The orebodies are in the form of folded sheets, concordant with the bedded host rock (Section **4.** 1. The ore is generally fine grained and massive with sulphide or sulphate content in the range of 50 to 90 per cent by volume. . Semimassive and disseminated ore is minor in amount and mainly occurs near the strike limits of the orebodies. The upper and lower bounding surfaces of the orebodies are sharply defined by a cut-off from ore to barren or pyritic host rock. The ore commonly shows a compositional layering.

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5.1 The Zinc-lead Orebody

The strike limits of the Zn-Pb orebody are shown in vertical longitudinal projection on Figure 28. At the northern, southern and lower limits the orebody fades out through disseminated sphalerite, pyrite and galena into barren host rock which may carry disseminated pyrite. A gap of about 100 ft in the continuity of the orebody near 2200N between 2 Level and 6 Level separates the orebody into a northern and central-southern sections. The orebody on either side of the gap is thin and of low grade .

In the upper levels of the mine the orebody was considered to consist of a eeries of discrete lenses (Finucane, 1932; Hall et al., 1953). Examination of the old mine plans shows, however, that the orebody is more or less continuous and, as suggested in Hall et al. (1965), the lenses join up via an en echelon fold pattern. In the lower levels of the mine geological information is much more complete and the orebody is continuous within the northern and central-southern sections, as shown for example on 13 Level (Fig. 29).

-The thickness of the Zn-Pb orebody on the flanks of the folds ranges up to-about 40 ft and averages 15 to 20 ft. In the fold noses repetition has produced widths up to 120 ft. The original thickness distribution of the orebody is difficult to reconstruct because of the

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folding, but the thickest part of the orebody appears to be around 500N between 15 and 17 Levels.

The Zn-Pb orebody is c0mposed of f0ur main ore types. They are (1) high-grade sphalerite-galena-pyrite 0re, (2) pyritic low-grade ore, (3) pyrite-chalcopyrite ore and (4) disseminated to semimassive low-grade ore consisting mainly of sphalerite. The high-grade and pyritic 10w-grade ore types are intergradational and together make up most of the orebody.

The gangue minerals in the Zn-Pb orebody, in approximate order of abundance, are chlorite, quartz, manganiferous carbonate, sericite, barite, sphene and albite.

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The high-grade 0re consists mainly of sphalerite and galena with less than 10 per cent pyrite and min0r amounts of chalcopyrite and tetrahedrite-tennantite. Some modal mineral analyses for this ore type are given in Table 5 (p. 136). In appearance it is massive and homogeneous with layering present only locally. It is common throughout the strike extent of the oreb0dy but is most abundant in the southern section. With an increasing proportion of pyrite and pyrite-rich layers the high-grade ore grades into pyritic low-grade ore.

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The pyritic ore is composed of pyrite, up to 80 per cent, and sphalerite with minor amounts of galena and chalcopyrite. Interlayering of pyrite and sphalerite-rich material is present on all scales. This ore type is most abundant in the central part of the orebody.

The pyrite-chalcopyrite ore is composed of pyrite and chalcopyrite with very minor amounts of tetrahedrite-tennantite, sphalerite, galena and an unidentified mineral that is possibly kobellite. Modal analyses of two samples are given in Table. 5. This ore type commonly forms a distinct zone along the stratigraphic footwall side of the orebody, particularly in the central and southern part below 14 Level (e. g. Fig. 5). In places this footwall zone is represented by semimassive or massive pyrite with little or no chalcopyrite. At the north end of the Zn-Pb orebody between 3000N and 4000N on 10 and 11 Levels the footwall zone is represented by a discontinuous zone of relatively coarse-grained chalcopyrite in a chlorite schist matrix. On 14 and 13 Levels part of the Zn-Pb orebody changes laterally to massive and semimassive pyrite- \mathbb{Z}_{p^2} chalcopyrite ore. On 13 Level (Fig. 29) this pyrite-cha1copyrite zone has a strike length of about 250 ft and appears to interfinger with sphalerite-galena-pyrite Ore at its strike limits. This zone fades out down-dip into barren host rock.

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The, low-grade spha1erite ore consists of streaky layers and blebs of sphalerite in a schistose and commonly chloritic matrix. It occurs mainly towards the extremities of the orebody where the host rocks pass into footwall schist type rocks.

5.2 The Barite Orebody

The Barite orebody is discontinuous and consists of a number of separate lenses (Figs. 26, 29). The total strike extent is of the order of 2, 000 ft while individual lenses range up to 500 ft in length by 40 ft in thickness. On 13 Level around the nose of the anticline $\mathtt{A9}$ and in a number of other places two separate and in part overlapping barite layers have been outlined. Towards their strike limits the barite lenses commonly pass into hematite and pyrite with or without barite.

The Barite orebody consists predominantly of fine-grained massive or semimassive barite (up to 80 per cent) with generally minor amounts of sphalerite, galena, tetrahedrite, chalcopyrite and pyrite. Compared to the Zn-Pb orebody the Barite orebody is relatively enriched in galena and tetrahedrite and relatively much poorer in pyrite, which is commonly under one per cent. Gangue min@rals are rhodochrosite, sericite, albite and quartz. In places the Barite orebody carries enough spha1erite and galena

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to make it worth mining. The barite ore shows both fine and coarse \sim layering (Fig. 30) defined by variation in sulphide content.

> In three areas the Barite orebody appears to grade laterally into massive sphalerite-galena-pyrite-chalcopyrite ore. One of these lateral changes has been exposed on 14 Level and takes place over about 10 ft by interfingering of sulphide and barite ore. On 14 Level a further small lens of sphalerite-galena-pyrite ore occurs just above the Barite orebody. This ore appears to be unusually rich in tetrahedrite -tennantite.

5.3 Other types of Mineralization

Hematite -Pyrite

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Immediately overlying the Barite orebody on the east flank of the anticline A9 are several small concordant lenses about 1 ft in thickness of massive hematite-pyrite mineralization. It is associated with barite and also contains noteworthy amounts of green tourmaline as well as quartz, albite and sericite (100887, 100893). Layering of pyrite-rich and hematite-rich material is present.

A hematite-tourmaline-quartz body (100978), 3 ft in thickness, crops out within the host rock on the surface at $2400S$. It possibly correlates with hematite intersections obtained in drill holes near

Fig. 30 - Fine-layering in Barite orebody. Also shows rhodochrosite bodies in barite.

10008 and about 500 ft south of the southern limit of the Zn-Pb orebody.

Hematite-Magnetite-Pyrite

Rare occurrences of concordant, layered hematite-magnetite-. pyrite have been noted within and adjacent to the Zn-Pb orebody. ,.. The best known occurrence is a lens 3 ft in thickness immediately overlying sphalerite-galena-pyrite ore at the southern limit of the Zn-Pb orebody on 13 Level. Figure 31 is a photomicrograph of this ore.

Pyrrhotite

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Massive pyrrhotite with minor amounts of chalcopyrite, pyrite and sphalerite occurs in several places at the southern end of the Zn-Pb orebody. The two occurrences exposed by mine workings are around 2508 and 7608 on 14 Level. Thepyrrhotite at 2508 forms an irregular body up to 50 ft long by 15 ft wide that cuts across and replaces the sphalerite-galena-pyrite ore. In Figure 20 the pyrrhotite can be seen cutting across compositional banding in sphalerite-galenapyrite ore. The pyrrhotite completely replaces the sphalerite-galena bands but pyrite-chalcopyrite bands persist into the pyrrhotite. The contact between the pyrrhotite and the sphalerite-galena ore is \mathbf{r}_{c} . commonly delineated by a thin (1.0 to 3.0 mm) zone of very dark and presumably high Fe sphalerite. As seen in polished section (15-93),

Fig. 31 $-$ Photomicrograph of hematite (light grey elongate crystals), magnetite (medium grey), pyrite (white) mineralization.

(**1-25,** x 170, air, reflected light).

the contact between the pyrrhotite and sphalerite-galena ore is irregular and is marked by a relative abundance of chalcopyrite and locally of arsenopyrite.

Pyrrhotite-rich mineralization in an equivalent position to the • Zn-Pb orebody has been intersected in drill holes R1440 at R. L. 8800 on 400S and in R1527 at R. L. 8500 on 1000S, In drill-core from $R1440$ massive Δ ipyrrhotite is associated with chlorite, patches of very dark sphalerite and scattered chalcopyrite and pyrite. Drillcore from R1527 contains occurrences of (1) massive pyrrhotite with chalcopyrite and no pyrite, and (2) interlayered pyrrhotiterich, pyrite -rich and chlorite -rich material.

The pyrrhotite consists of mixtures of the hexagonal and monoclinic phases. The ratio of the monoclinic to hexagonal phases and the metal contents of the hexagonal phase for four specimens were determined by X-ray diffraction (Table 2). The presence of both monoclinic and hexagonal phases was confirmed by the formation of lamellar intergrowths of the two phases on etching with chromic acid (Arnold, 1966),

The determined phase assemblages and metal contents fall within the two phase field of hexagonal low-temperature pyrrhotite and monoclinic pyrrhotite that is present below 310° C in the Fe-S

system as presented by Kullerud (1967).

The pyrrhotite bodies on 14 Level appear to plunge parallel to the local macroscopic fold plunge and their emplacement was possibly structurally controlled.

No.	Mineral Assemblage	Monoclinic Hexagonal	2 At. % metals Hexagonal Phase
100619a	Pyrrhotite-chalcopyrite	1.2/1	47.2 ± 0.2
106619b	Pyrrhotite-pyrite- chalcopyrite		$4/1$ 47.1 $\stackrel{+}{_{0}}$ 0.2
100620	Pyrrhotite-chalcopyrite- sphalerite	1/1.2	$47.4 \div 0.2$
100968	Pyrrhotite-chalcopyrite	1/1.2	$47.2 \div 0.2$

TABLE 2 PYRRHOTITE COMPOSITIONS

- 1. Ratio of monoclinic/hexagonal phases. Determined by X-ray diffraction charts according to graph of Arnold (1966).
- 2. Atomic per cent metals in hexagonal phase. Determined by X-ray diffraction from d(102) vs. metal content graph of Toulmin and Barton (1966). Location of d(102) peak obtained by counting for fixed time with external silicon standard. Samples crushed under kerosene in porcelain mortar to prevent oxidation.

However, as shown in the specimen illustrated in figure 20, the folding evidently occurred before the pyrrhotite was emplaced.

Irregular, transgressive bodies of pyrrhotite-chalcopyrite have been recorded from a number of pyritic Zn-Pb ores (Stanton, 1960; Vokes, 1968) and Vokes has suggested that they may be a result of mobilization of a pyrrhotite-chalcopyrite fraction during metamorphism.

Coarse-Orained Gangue

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Adjacent to and within the orebodies there are irregular blind veins and patches of coarse-grained quartz and carbonate carrying galena, chalcopyrite, sphalerite, pyrite and less commonly tetrahedrite-tennantite and other sulphosalt minerals. The volume of this type of mineralization is very small in proportion to that of the fine-grained ore.

The patches and veins are particularly common at changes in strike or dip along the footwall contact of the Zn-Pb orebody; and in general in dilatant zones. The largest masses, up to 30 ft wide, occur near the ends of folds in the Zn-Pb orebody. Such an area is from 10 Level at 200S down to 14 Level at 200N. Elsewhere the veins and patches range in width from one inch or less up to about 10 ft.

The carbonate in these patches is most commonly pale pink magnesian kutnahorite although rhodochrosite and calcite are also present. Coarse-grained barite alone or barite with carbonate occurs loc ally.

The type of sulphide or sulphosalt appears to be related to the composition of the adjacent fine-grained ore, which suggests that the components of the coarse grained mineralization are of local derivation. Quartz-carbonate patches in chalcopyrite-pyrite ore contain mainly coarse-grained chalcopyrite with minor amounts of pyrite, while galena and sphalerite are the dominant ore minerals where the patches are adjacent to sphalerite-galena ore. The rare sub sulphosalt m inerals - meneghinite, pyrargyrite, jordanite have only been recorded from the coarse-grained quartz-carbonate mineralization.

It is inferred that the sulphides and sulphosalts like quartz and carbonate were mobilized wia a "dispersed" fluid phase and deposited in dilatant zones formed in response to deformation during the metamorphism.

Disseminated Pyrite-Chalcopyrite in the Footwall Schists

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 τ \sim The footwall schists carry disseminated and semimassive pyrite which in places is accompanied by chalcopyrite. Some

lenses of disseminated pyrite-chalcopyrite up to 300 ft in strike length and 40 ft in thickness have been outlined by diamond drilling. This mineralization is in relatively coarse-grained disseminations, veinlets and bands in quartz-rich and chlorite-rich schists below the Zn-Pb orebody.

5.4 Layering in the Ore

One of the most prominent features of the sulphide and barite ore, and also the hematite-pyrite-magnetite material, is the compositional layering or banding. Most obviously this is a fine compositional layering, principally of pyrite-rich and sphaleriterich layers, the thickness of which can be measured in millimetres or centimetres. On a larger scale the layering is reflected by overall sphalerite-galena-rich and pyrite-rich zones.

The fine layering consists of three main compositional types: sphalerite-rich, pyrite-rich and chalcopyrite-rich. Where sphalerite-rich and pyrite-rich. layers alternate the sphalerite layers range in thickness from 0.3 to **10.0** mm and consist mainly of sphalerite with generally less than 10 per cent galena. In contrast the pyrite-rich. layers are generally thicker $(0.5 \text{ to } 20.0 \text{ mm})$ and contain 50 to 80 per cent pyrite with variable amounts of sphalerite,

galena, chalcopyrite and silicate gangue. Chalcopyrite-pyrite-rich layers contain 20 to 60 per cent chalcopyrite and 20 to 40 per cent pyrite, and they vary from 0.5 to 8.0 mm in thickness. Less obvious than these three types are galena-rich layers in sphalerite-rich ore. They contain around 15 per cent galena compared with 5 per cent galena in the adjacent ore. Magnetite-rich layers (0.2 mm) and arsenopyriterich layers (0.5 mm) have been recorded in several polished sections (8-30, 8-51).

In the Barite orebody the fine layering consists of layers, 1. 0 to 10.0 mm in thickness, that are relatively enriched in sulphides, alternating with sulphide-poor layers. The sulphides are mainly sphalerite and galena.

Another type of layering occurs in massive pyrite and consists of alternations of fine and relatively coarse-grained layers, 1. 0 to 10.0 mm in thickness.

Layers marked by varying proportions of silicate and carbonate gangue constituents and layers of cherty, chloritic or sericitic host rock occur sporadically throughout the orebody. Some examples are illustrated in Figures 22 and 32.

On a larger scale the layering, which is important for ore grades consists of sphalerite-rich and pyrite-rich zones. Some of

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Fig. 32 - Interlayering of pyrite-chalcopyrite and chloritic host rock in Cu-rich section of Zn-Pb orebody.

the pyrite-rich zones are relatively rich in chalcopyrite. This layering is defined either by individual layers similar to the fine layers but of greater thickness, or by zones of finely layered ore which, depending on the relative proportions of the pyrite-rich and sphalerite-rich layers, can be separated from adjacent zones. The thickness of these layers is reflected in the length of the mine and drill-core samples taken to determine ore grades. The samples range from 1 to 10 ft in length, although the majority are 5 to 7 ft. Histograms for Zn, representing sphalerite, and Fe, largely representing pyrite, for groups of these samples throughout the Zn-Pb orebody are given in Figures 43 and 48. They show the frequency of occurrence of the layers of varying sphalerite and pyrite content.

The lateral extent of the layers is difficult to determine because of incomplete exposures but some of the thicker (say 5 cm) layers can be traced for strike lengths of greater than 60 ft. On the other hand many of the thin layers are lenticular and even the thicker layers are lenticular over 100 ft or more.

Origin of the Layering

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Possible explanations that may be considered for the origin. of *r* the layering in the Rosebery ore are:

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- (a) Metamorphic differentiation either by deformation concurrent with emplacement of the ore, or by post-ore deformation.
- (b) Selective replacement of a bedded host rock.
- (c) Sedimentary deposition of the ore minerals.

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Mechanism (a) is not favoured because the layering is folded and is pre-cleavage and therefore its formation. pre-dates the main period of deformation. Also the layering is parallel to bedding in the adjacent host rock. Further, the shape of the layers, the range in compositional types and the wide variation in thickness of the layers are more akin to sedimentary stratification than to metamorphic layering, which in any given rock type is generally plate like in shape and of fairly uniform thickness and composition.

Selective replacement (b) would have to be pre-folding since the fold style in the sulphide ore is different from that in the host rocks and indicates that a highly incompetent mass of material such as sulphide ore was present before the folding. Selective replacement is not considered likely because (I) the thickness of the layers is both . lesser and greater than bedding in the host rock, and (2) there is no unequivocal evidence of replacement on either a microscopic or macroscopic scale. At least six different compositional types of layers are present and it is difficult to conceive how they all could

be produced by selective replacement of the host rock which consists of only two or three compositional types.

Therefore, as indicated by the above arguments the layering is considered to be sedimentary bedding". To recapitulate: the layering is parallel to bedding in the adjacent host rock, the shape of the layers is like lenticular beds and is unlike metamorphic layering, the ore layers are in places "interbedded" with host rock, the composition is fairly uniform within layers but is sharp between layers, and there is no evidence for selective replacement of host rock.

The possible mechanism of formation and deposition of the ore as a chemical sediment is discussed later under "Ore Genesis" (Section 8).

5.5. Ore Mineralogy

The ore mineralogy has been previously described by Stillwell (1934) and some of the rarer minerals have been described by WiUiams (1960). Extensive exploration of the ore deposit in recent years has enabled the present study to cover a much wider area and range of ore types.

*The Baas Becking Geobiological Laboratory has recently produced a compositional layering of base-metal sulphides by diffusion of basemetals through gels in a biological system supplying H_2S . The feasibility of this mechanism in producing the layering in natural orebodies remains to be evaluated however.

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Description of the ore mineralogy is based on examination of about 120 polished sections and 30 thin sections. This suite covers a range of the different ore types but the bulk of the specimens were collected systematically from drill-hole intersections that were selected to cover the lateral extent of the Zn-Pb and Barite orebodies.

Ore Minerals

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The ore minerals recorded in the polished sections are sphalerite, pyrite, galena, chalcopyrite, tetrahedrite-tennantite arsenopyrite, magnetite, gold, electrum, pyrrhotite, ilmenite, ." enargite, hematite, rutile, meneghinite, jordanite and an unidentified mineral that is possibly kobellite. Other minerals which have been recorded from Rosebery but which are not described in the present study are pyrargyrite and freilsbenite, as noted in Ramdhor (1960). $\frac{7}{6}$ $%$ $\frac{1}{2}$ $\frac{1}{2}$

Sphalerite: - Sphalerite is overall the most abundant mineral in the Zn-Pb orebody and commonly constitutes greater than 60 per cent of the ore. The main forms of occurrence are almost massive aggregates, sphalerite-galena mosaics, and intergranular to pyrite in pyrite-rich layers. The sphalerite, as revealed by etching, is in aggregates of polygonal twinned grains (Fig. 37) and the grain size ranges from 0.005 to 0.7 mm but is generally in the range of 0.3 to 0.15 mm.

The colour of the sphalerite in thin section and from internal reflections in polished section varies from pale yellow to deep redbrown and shows considerable colour differences between and within individual aggregates and bands. The average Fe content from chemical analyses of HCl soluble Fe in zinc concentrates is about 2.5 per cent. Sphalerite in the Barite orebody is uniformly pale yellow and noticeably lighter in colour than that in the Zn-Pb orebody.

Some of the sphalerite contains minute blebs of chalcopyrite that are commonly arranged in chains along sphalerite grain boundaries and twin planes (Fig. 33). The sphalerite appears to be richer in chalcopyrite blebs in the vicinity of chalcopyrite-rich bands or where the sphalerite is a minor phase in chalcopyrite-rich ore. By comparison with the Zn-Pb orebody sphalerite in the Barite orebody is almost free of chalcopyrite blebs.

An unusual occurrence of p phalerite (100918) is as boatshaped bodies, measuring up to 3.5 by 1.5 mm and averaging 1.5 mm in length, in a pyrite-sericite matrix. Sphalerite in this form occurs at Hercules and is similar to the description by Markham (1961) of sphalerite "pellets" in chalcopyrite-rich ore from Peelwood, New $\rm{South~Wales.}$

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Pyrite: Pyrite is present in almost all the polished sections examined and in the pyritic ore constitutes up to 80 per cent.

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In pyritic ore and pyrite-rich bands in sphalerite-galena ore the pyrite occurs as granular, loose-packed or compact aggregates (Fig. 38). It is in a subidiomorphic to idiomorphic cube habit and the grain size varies from 0.01 to 0.5 mm or rarely to 1. 0 mm and averages 0.07 mm.

In contrast, pyrite in sphalerite-galena ore is coarser grained, averaging 0.2 mm and ranging from 0.05 to 0.8 mm. It occurs as discrete crystals scattered or disseminated in a sphalerite-galena matrix. The-shape of the crystals is variable. Many crystals are irregular in shape and are commonly roughly circular or appear as embayed cubes (Figs. 39 and 40). The circular forms generally have delicately fretted margins and are usually larger than the other forms. Idiomorphic or subidiomorphic cube habits are rare. In many specimens the larger pyrite' crystals are fractured and even brecciated; the fractures being infilled with galena and less commonly sphalerite and quartz. Inclusions of sphalerite, galena and chalcopyrite occur in many of the larger pyrite crystals and in several specimens (e.g. $5-55$, Fig. 41) the inclusions show a roughly concentric arrangement.

Pyrite in chalcopyrite-rieh ore is similar in form to pyrite in sphalerite-galena ore. It is relatively coarse grained (up to 1, 0 mm) and is in subidiomorphic and circular habits. Fracturing and brecciation with infilling by chalcopyrite, and inclusions of chalcopyrite and sphalerite are similar features to the pyrite in sphalerite - galena ore.

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Pyrite occurring in quartz, sericite, chlorite or barite gangue is generally subidiomorphic and relatively fine grained (0.01 to 0.07 mm).

As previously noted by Stillwell (1934) the pyrite is distinctly anisotropic (see also Stanton, 1959a).

Galena: - Galena is disseminated throughout the sphalerite-galenapyrite ore and constitutes up to 18.0 per cent but is generally between 5.0 and 10.0 per cent. In pyrite-rich ore it is intergranular to the pyrite. In sphalerite, chalcopyrite, tetrahedrite and barite matrices the galena is typically in cuspate grains or cuspate aggregates of grains. In galena-rich ore the cuspate aggregates join to form a discontinuous network and they enclose rounded bodies of sphalerite. The grain aggregates, as revealed by etching, show galena grains meeting at approximately 120° triple point junctions. The grain size of the galena ranges from 0.02 to 0.3 mm and averages 0.05 mm.

Chalcopyrite: - The main forms of occurrence of chalcopyrite are as scattered and disseminated grains in sphalerite-galena-pyrite ore, as intergranular material in pyrite-rich bands, as very fine-grained blebs in sphalerite, and as xenomorphic granular aggregates in chalcopyrite-rich ore. Chalcopyrite in sphalerite-galena-pyrite ore is finer grained than the other minerals, averaging 0.03 mm. The form is irregular but tends to be cuspate in sphalerite. In massive chalcopyrite-pyrite ore the grains range in size from 0.02 to 0.3 mm, average 0.1 mm, and show approximately 120⁰ triple junctions. Some of this chalcopyrite contains broad twin lamellae. The very fine-grained blebs in sphalerite range in size from 0.001 to 0.005 mm and are located mainly along grain boundaries and twin planes in the sphalerite (Fig. 33). It is not clear whether or not these are true exsolution blebs.

Tetrahedrite-Tennantite: - Tetrahedrite-tennantite is a minor but widespread constituent in the Zn-Pb and Barite orebodies. It generally makes up less than 1. 0 per cent, although locally reaching 2.0 to 3.0 per cent, and occurs as xenomorphic scattered grains or aggregates averaging 0.04 mm. Although total sulphide content is much lower in the Barite orebody than in the Zn-Pb orebody the tetrahedrite-tennantite content is relatively higher.

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Fig, 33 - Chalcopyrite blebs **in** sphalerite outlining grain boundaries.

(5-53, x 750, oil, reflected light).

From the chemical analysis quoted in Stillwell (1934) and X-ray diffractions on two samples of coarse-grained material the compositions lie towards the tetrahedrite end of the tetrahedritetennantite series. This is confirmed by the assay figures for the Cu concentrate in which the Sb: As ratio, both of which are probably contained almost entirely in tetrahedrite -tennantite, is about 4 to **1.** The tetrahedrite-tennantite is silver-bearing as indicated by the analysis of 795 oz. per ton given in Stillwell (1934) and also by the sympathetic enrichment of Ag and Sb in the Cu concentrate relative to the crude ore. The main associations of the tetrahedritetennantite are galena, chalcopyrite and arsenopyrite. A common mode of occurrence is as relatively small grains or aggregates containing fine-grained (O. 04 mm) inclusions of chalcopyrite and arsenopyrite. One interesting occurrence (5-53) consists of a finegrained spongy aggregate of tetrahedrite -tennantite, chalcopyrite and arsenopyrite. The association of tetrahedrite-tennantite and arsenopyrite in the Rosebery ore was noted by Stillwell (1934) and is common in other ores of this type $-$ Captains Flat, N.S.W. (Edwards, 1943), and Peelwood, N.S. W. (Markham, 1961).

Arsenopyrite: - Arsenopyrite is a widespread but generally minor (under 1. 0 per cent) constituent of the Zn-Pb orebody.

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Although in sections $5-62$ and $8-57$ from drillhole $51R$ at the north end of the orebody, arsenopyrite is relatively abundant, comprising up to 20 per cent.

The main modes of occurrence are as scattered, relatively coarse-grained iqiomorphic crystals, intergrown with granular pyrite aggregates and as small (**O.** 005 to **0.05** mm) embayed crystals in tetrahedrite-tennantite. The coarse -grained crystals range from **0.04 to 0.2 mm in length and some show symmetrical zoning** parallel to the crystal faces and fine lamellar twinning (e.g. 8-57). Another mode of occurrence is as very fine-grained **(0.005** mm) spongy aggregates that are generally arranged in lenses parallel to the layering (8-30, 5-53).

Kobellite: **-** A mineral which has not been positively identified occurs in minor amounts in chalcopyrite-rich ore (7-26, 7-27, 7-28, 8-62) from the chalcopyrite-pyrite footwall zone. In reflected light the mineral is white with a blue-grey tint and the reflectivity is slightly less than chalcopyrite. It shows weak bireflectance and is strongly anisotropic from straw-brown to dark brown. The habit is prismatic crystals up to **0.03** mm in length. Elongate inclusions of tetrahedrite-tennantite are common. Unfortunately grains dug out of one section by the method described by Williams (1962) were

too heavily contaminated with cha1copyrite and pyrite to give a distinguishable X-ray powder pattern.

The properties of this mineral best fit those of kobellite \cdot (Pb₆ Fe Bi₄ Sb₂ S₁₆) but the match is not very good.

Magnetite: - Magnetite is sporadically distributed in minor amounts throughout sphalerite-galena-pyrite ore in the Zn-Pb orebody. It occurs as subidiomorphic grains ranging in size from 0.01 to 0.3 mm. Locally it forms lenticular aggregates parallel to the layering (8-51b). The colour in reflected light is distinctly brown and it commonly has a pinkish tint indicating a Ti-rich variety. In some sections the magnetite is porphyroblastic in appearance and contains inclusions of pyrite, spha1erite, chalcopyrite and hematite (Fig. 42, 8-5lb).

Gold: - As shown by the assay values gold is a widespread but rare mineral in the Zn~Pb orebody, averaging about 2.0 dwt per ton, and is even rarer in the Barite orebody. It is most frequently observed in coarse-grained tetrahedrite-tennantite where it occurs as blind veinlets and. patches about 0.05 mm in width (8-64, 34-5, 35 -4). The association of gold with coarse-grained tetrahedrite at Rosebery has previously been noted by 8tillwell (1934) and Williams (1960). Gold has also been observed in a chalcopyrite-arsenopyrite-

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tetrahedrite aggregate (5-53) and in sphalerite-galena-pyrite ore . $(5-51)$. In these occurrences the gold is present as groups of small grains averaging about 0.005 mm in diameter.

Electrum: - Electrum is relatively abundant in massive pyrrhotite from drill holes R.1527 and R.1440 (15-89, 15-90, 15-91). The colour is creamy-white with a very high reflectivity and under crossed nicols the characteristic green colour is apparent. It occurs as ragged grains ranging from 0.4 to 0.8 mm. In section 15 -89 the electrum is associated with several grains of a white mineral with very high reflectivity which is possibly native silver. These occurrences of electrum are probably explained by the absence of tetrahedrite-tennantite, galena or other silver-bearing minerals in the pyrrhotite mineralization.

Pyrrhotite: **-** Apart from the massive pyrrhotite mineralization, which is described in Sections 5.3 and 5.7, the only other occurrence is in polished sections from drill hole 51R at the north end of the Zn-Pb orebody (e. g. 8-54, 8-56, 5-62). This pyrrhotite is present in very minor amounts as small (0.03 mm) inclusions in galena, pyrite, chalcopyrite and tetrahedrite.

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llmenite: **-** llmenite occurs in a number of polished sections $(5-64, 5-65, 8-66)$ as scattered, tabular crystals (0.03 mm in) λ length) and aggregates that are associated with silicate gangue minerals and particularly sphene. In section 5-55 it is present as inclusions (0.01 mm) in magnetite and shows lamellar twinning. Stillwell (1934) described ilmenite from low-grade schistose ore at Rosebery.

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Enargite: **-** A mineral which has been tentatively identified as enargite occurs as minute $(0.001$ to 0.005 mm) inclusions in pyrite in some sections $(5-64, 7-26, 8-62, 8-63)$. The mineral is pearlgrey with a pinkish tint and displays strong anisotropism in blue and red tints.

Hematite: **-** As-As well as in the massive hematite-pyrite and hematite-magnetite-pyrite mineralization, hematite has been identified in three other polished sections where it occurs as prismatic crystals $(0.09 \times 0.02 \text{ mm})$ in silicate gangue $(5-74)$ and as inclusions (0.005 mm) in magnetite, pyrite and quartz (8-52, 5-57). In the massive hematite-bearing mineralization (1-24, 1-27) the hematite is in acicular crystals up to 0.4 mm in length.

Rutile: **-** Minute (0. 002 mm) inclusions of a mineral that matches

the properties of rutile are conta:ined in ilmenite in section 5-62 ,.. and in hematite in section 5-74.

<u>Meneghinite:</u>- Meneghinite (Pb₁₃ Sb₇ S₂₃) has been recorded in two specimens of coarse-grained quartz-carbonate gangue. In section 64-1 it occurs as granular (0.2 mm) aggregates, with .carbonate, pyrite and rare arsenopyrite. The meneghinite shows strong bireflectance and anistotropism with red internal reflections. Another occurrence (11750), identified by X-ray diffraction, is as fine needle-like crystals in cavities in a quartz vein cutting the Zn-Pb orebody in 8/J-37N stope. Williams (1960)also described meneghinite in a quartz gangue from 8 Level.

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Jordanite:- Jordanite (Pb₄ As₂ S₇) has been recorded from a specimen (83 **-1)** of coarse-grained sphalerite and tetrahedrite from 9/K-37N Stope. It was identified from polished section and X-ray powder photograph by R.A. Both. In the polished section the j jordanite is associated with sphalerite, tetrahedrite, galena and rare chalcopyrite. The jordanite is white with a greenish tint, is distinctly bireflectant and strongly anisotropic in shades of grey, blue and brown. Jordanite has also been observed by the writer in sphalerite-galena-pyrite ore (5 -20) from the Hercules Mine.

Bournonite: **-** Both Stillwell (1934) and Williams (1960) recorded bournonite in the Rosebery ore but in this study it has been observed in only one section (9-7). It occurs as distinct grains (0.15 mm), showing the characteristic parquet twinning and as very fine-grained (0.03 mm) inclusions in galena and tetrahedrite.

Gangue Minerals

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The gangue minerals recorded from the Zn-Pb and barite orebodies are chlorite, quartz, manganiferous carbonate, sericite, barite, sphene, calcite and albite. Stillwell (1934) described spessartite in association with the manganese carbonate but it has not been recorded in this study.

Chlorite: - Overall, chlorite is the most common gangue mineral in the Zn-Pb orebody where it occurs in tabular and prismatic crystals averaging from 0.05 to 0.4 mm in length. This chlorite, in respects of grain size and appearance in thin section, is very similar to chlorite in the adjacent host rock.

Quartz: - In the Zn-Pb and Barite orebodies quartz occurs as scattered grains or aggregates and is locally present as chert nodules and lenses, Where it is abundant (100959, 100979, 35788) the quartz forms the fine-grained polygonal mosaics typical of recrystallized quartz. Some of the larger quartz grains show undulant extinction.

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Quartz also forms pressure fringes around the larger pyrite crystals.

Carbonate: - Manganese carbonate is widespread and in places abundant in the Zn-Pb and Barite orebodies. The carbonate mineralization is described in detail in Section 7.

In the sulphide ore carbonate occurs as scattered, commonly rhomb-shaped crystals (0.1 to 0.4 mm) and as irregular granular aggregates. The habit is commonly poikiloblastic (e.g. 100959.)1(with inclusions of sphalerite, pyrite, galena and quartz.

Sericite: - Sericite is widespread but generally minor in amount and occurs as elongate flakes (0.05 mm in length) or fibrous aggregates.

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Barite: - It occurs in minor amounts in the Zn-Pb orebody, noticeably as coarse-grained patches, In the Barite orebody it is the major constituent and forms a granoblastic aggregate of average grain size 0.3 mm. Many of the triple point grain junctions approximate 120 $^{\circ}$. Where the barite is in a carbonate matrix (100894,) its habit is commonly prismatic.

Sphene: - Bphene is a widespread accessory constituent in the gangue of the Zn-Pb and Barite orebodies. Its mode of occurrence is identical to that in the host rock, forming clusters of fine-grained

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prismatic crystals. In polished section it can readily be identified $^{\mathrm{!}}$ by its reflectivity, which is slightly greater than sphalerite, and by the very strong internal reflections,

Albite: - Albite is a constant accessory mineral in the Barite orebody where it occurs in small (0.05 to 0.2 mm) xenormorphic grains. It is clear and unaltered and shows only a few broad albite twin lamellae.

5.6 Primary Ore Textures

Although the ore textures in the Rosebery and other similar *^r*deposits in low-grade metamorphic rocks are probably largely metamorphic in origin (Section 5.7; Zavaritsky, 1950; Stanton, 1964; Vokes, 1968) some primary textures are occasionally preserved. Such primary textures are the "colloform" and microsphericular forms of pyrite described from certain non-metamorphosed and slightly metamorphosed pyritic Pb-Zn deposits - McArthur River (Croxford, 1968), Mount Isa (Love and Zimmerman, 1961), Rammelsberg (Love and Amstutz, 1966) and various deposits in the Urals (Zavaritsky, 1960). The microsphericular pyrite is thought to have formed during early diagenesis (Love and Amstutz, 1966).

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Microsphericular pyrite has been observed in three polished $_{\circ}$ sections (8-59a, 8-59b, 9-7) from the Zn-Pb orebody at Rosebery. In 8-59 the microsphericular pyrite occurs in a'chert matrix adjacent to massive pyritic ore. The pyrite spherules (Fig. 34) range in diameter from 0.01 to 0.05 mm and consist of an aggregate of minute pyrite cubes giving the typical framboidal appearance. In section 9-7 the microsphericular pyrite (0.04 mm) is in a sphalerite matrix in sphalerite-galena-pyrite ore. Here the microsphericular pyrite is rare relative to other forms of pyrite and recrystallization appears to be further advanced compared with the pyrite in sections 8-59a and 8-59b.

> Several "colloform" structures have been observed. As illustrated in Figure 35 (8-5 9b) one of the structures consists of a central zone showing fine concentric banding of minute pyrite blebs which is partly enclosed by an outer rim of massive pyrite plus a minor amount of sphalerite. Figure 36 shows a "colloform" structure with crudely concentric layers of sphalerite, cha1copyrite, galena and pyrite, Similar relict "colloform" structures involving pyrite, chalcopyrite, bornite and galena have recently been described from various ores of the Mt. Lyell field by Markham (1968).

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Other possible primary structures, apart from the compositional

Fig. 34 - Microsphericular pyrite in chert matrix with sphalerite (dark grey).

(8-5 gb, x 1300, oil, reflected light).

Fig. 35 - "Colloform" pyrite structure in chert matrix.

(8-59b, x 450, oil, reflected light).

Fig. 36 - Relict "colloform" structure composed of sphalerite (dark grey), chalcopyrite (very light grey) and pyrite (white).

(9-7, x 625, oil, reflected light).

layering, are the extremely fine-grained spongy aggregates of chalcopyrite-pyrite -galena (9-7), chalcopyrite-arsenopyritetetrahedrite (5.53) and arsenopyrite-pyrite (8-55, 8-69).

5.7 Metamorphic Features

There is considerable evidence that the Rosebery orebodies were formed during the Cambrian volcanic activity and therefore they have been subjected to the deformation and $\operatorname{low}\nolimits$ -grade metamorphism. of the Devonian Tabberabberan Orogeny that has affected the enclosing rocks. Indeed a number of aspects of the textures and mineralogy are most reasonably explained in terms of metamorphism. The metamorphic features are described under the headings of textural, mineralogical, deformational and remobilization.

${\tt Textural}$

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Stanton (1964) has demonstrated that some of the ore textures in the Rosebery and other similar ore deposits are best explained by metamorphic crystallization rather than the older interpretation of a sequence of sulphide deposition and replacement (e. g. Stillwell, 1934). Application of the concept of interfacial energy developed by Smith (1948) to explain grain boundary relations in annealed metals, suggests that the grain boundary configurations in some sulphide ores are the

 ${\tt result}$ of equilibrium of interfacial energies during metamorphic growth (Stanton, 1964). The method used by Stanton was to measure the angle between grain boundaries at triple junctions between the same or different minerals. In monomineralic aggregates in which textural equilibrium has been attained the grain shape becomes polygonal and boundaries meet at angles close to 120° . This is typical of monomineralic patches of quartz, feldspar and carbonate in metamorphic rocks (Voll;, 1960). In two or three phase aggregates such as sphalerite-galena or sphalerite-galenachalcopyrite, the interfacial angles are relatively constant for any given combination of phases, suggesting that interfacial equilibrium has been reached (Stanton, 1964).

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In the present study, etching of sphalerite, galena and chalcopyrite aggregates in fine-grained ore shows a general tendency for the grains to be polygonal and meet at angles close to 120° . Although, for sphalerite it can be seen (e. g. Fig. 37) that the grain shape is rather variable and the spread of triple point angles around the mean of 120° is considerable. The standard deviations of this spread for 150 triple point angles in each of four specimens ranged from 16.8 to 22.8 which compares with a standard deviation of about , 10 for randomly sectioned triple point junctions in fully annealed ores (Stanton and Gorman, 1968). Some of the sphalerite shows

Fig. 37 - Sphalerite-galena aggregate showing twinning in sphalerite. Etched with HI. (5-60, x **170,** air, reflected light).

deformational phenomena such as deformation twinning, bent and kinked twin lamellae (e. g. 5 -60) and patches of much finer grained sphalerite than usual. While some of the deformation may be due to late fault movements (e.g. Richards, 1966) it is not obvious in the specimens examined and possibly the deformational effects are related to the Tabberabberan Orogeny. However the latter suggestion is difficult to reconcile with the experimental work of Stanton and Gorman (1968) which has shown that the temperatures required to anneal the common sulphides $-$ galena, sphalerite, chalcopyrite $$ are very low. Clearly there is a need for a more detailed study of the grain boundary configurations and deformational features of the different sulphide minerals throughout the orebody.

Another approach to metamorphic ore textures is by analogy with metamorphic textures in silicate rocks (e.g. Ramdhor, 1953, 1960; Stanton, 1964; Vokes, 1968; Markham, 1968). Ramdhor (1960) has emphasized the porphyroblastic habit of pyrite, magnetite and other ore minerals in metamorphosed orebodies. Stanton (1964) has placed the common or e minerals in a crystalloblastic series with the silicate and carbonate minerals.

Particular interest in this study has been focuss ed on pyrite and magnetite. The development of pyrite porphyroblasts can be

followed by comparing the distribution, grain sizes and shape of pyrite crystals in different matrices, specifically in pyrite -rich and in sphalerite-rich matrices. In the pyrite-rich layers the habit is subidioblastic or idioblastic cubes, generally in the range of 0.01 to O. 1 mm in size, but tending to be fairly uniform in anyone layer (Fig. 38). Pyrite crystals in sphalerite -rich layers in layered ^r pyritic ore are on average coarser grained (0.15 mm) than in adjacent pyrite -rich layers. Some of the larger crystals are irregular; others show cube faces on two or three sides while the remaining sides are concave and apparently embayed by sphalerite (Fig. 39). This is the caries texture that has generally been interpreted as resulting from replacement of early pyrite by later sphalerite. Other interpretations are possible however. It seems odd that the "embayed" pyrite crystals are larger than adjacent idioblastic cubes in the pyrite-rich layers. Also the end stages in the replacement, of, for instance, wedge -shaped crystals representing one side of a replaced cube, are not observed. The preferred interpretation is that these are metamorphic growth forms in which growth is faster along certain directions, or put in another way, growth on certain faces is inhibited, as suggested by Stanton (1964, p.69).

The coarsest grained (averaging O. 3 mm) pyrite crystals occur in the massive sphalerite-galena ore. Here the shape is

Fig. 38 - Pyrite-rich ore with granular pyrite in sphalerite matrix. Also shows quartz -rich chert layer.

(8-61, x 170, air, reflected light).

Fig. 39 - Embayed pyrite crystal or pyrite porphyroblast? In sphalerite matrix. Also shows chalcopyrite (light grey) and chlorite (elongate, black crystals). (7-25, x **170,** air, reflected light).

irregular (Fig. 40), although it is commonly crudely rounded. These shapes are similar to the pyrite "metacrysts" figured by Vokes (1968, p. 56) from the metamorphos ed pyritic deposits of Norway. Many of the,larger pyrite crystals contain inclusions of sphalerite, galena and chalcopyrite. The variety and the form of the inclusions suggests that the pyrite is poikiloblastic, the inclusions having been incorporated during growth of the pyrite crystals. This is also indicated by the roughly concentric arrangement of the inclusions in a number of specimens (e. g. Fig. 41, 5-55).

The phenomena of some pyrite crystals showing idiomorphic habits while other crystals are xenomorphic suggests, by analogy with garnets in metamorphic rocks (Rast, 1965), that the more regular idiomorphic crystals are products of slow crystallization, whereas the larger xenomorphic crystals grew faster and thereby incorporated inclusions into the crystal.

Magnetite also shows a tendency to develop porphyroblastic and poikiloblastic forms (Fig. 42, 8-5lb.

The gangue minerals $-$ quartz, barite, carbonate $-$ where they are monimineralic, or nearly so, display approximately equigranular mosaics of polygonal grains indicative of metamorphic *r* crystallization.

Fig. 40 - Irregularly shaped pyrite crystals in sphalerite-galena matrix. Also contains tetrahedrite -tennantite (medium grey). (5-45, x 100, air, reflected light).

Fig. $41 -$ Pyrite poikiloblast showing roughly concentric arrangement of sphalerite inclusions. In sphalerite matrix. Etched with HI. (5-55, x 250, oil, reflected light).

Fig. $42 -$ Magnetite poikiloblast with pyrite inclusions in pyrite matrix. (8-5lb, x 500, oil, reflected light).

The Mn carbonate in the sulphide ore is commonly idioblastic' or poikiloblastic with inclusions of sphalerite, pyrite, galena and quartz.

Mineralogical

As pointed out by, among others, Vokes (1968), the succession of mineralogical changes found in silicate and carbonate rocks with increasing metamorphic grade does not occur to the same extent in the sulphides.

One possible mineralogical change with increasing metamorphic grade is the development of pyrrhotite from pyrite, as is suggested by the abundance of pyrrhotite relative to pyrite in sulphide orebodies in high-grade metamorphic rocks (McDonald, 1967; Vokes; 1968). In this connection the origin and emplacement of the pyrrhotite at Rosebery was possibly related to the metamorphism (Section 5.3).

Another mineralogical change that was possibly produced during metamorphism is the formation of the sulphosalt minerals. Minerals such as meneghinite, jordanite and pyrargyrite have not been reported from sedimentary environments and it seems unlikely that such complex minerals could form at low temperatures. Antimony and Bi sulphosalt minerals of Cu and Ag have been formed in the solid state from the simpler minerals covellite, argentite,

stibnite and bismuthinite at temperatures around 100° C (Ross, 1954). As suggested by Ross the prevalence of sulphosalt minerals and the absence of such minerals as stibnite, orpiment and bismuthinite in the same deposit may be due to formation of the more stable sulphosalt minerals by solid state diffusion during, for instance, lowtemperature metamorphism. Support for this idea at Rosebery is provided by the occurrence of most of the sulphosalt minerals in the coarse-grained gangue mineralization which was probably formed during the metamorphism.

McDonald (1967) in comparing the mineralogy of the McArthur River, Mt. Isa and Broken Hill deposits, noted an increasingly complex ore mineralogy concomit ant with increasing metamorphic grade of the host rocks. He suggested that this might be a reflection of mineralogical changes with increasing. metamorphic grade. Although the effects of differences in initial chemical composition are an unknown factor in such comparisons, the Rosebery deposit has a slightly more complex mineralogy than Mt. Isa which is consistent with the more intense deformation of the host rocks at Rosebery.

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Deformational and Remobilization

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A number of field and experimental studies have demonstrated the ready m9bility of many of the sulphide minerals during deformation (Buerger, 1928; Roberts, 1965; P.J. Solomon, 1965; Gill, 1965; Richards, 1966; Yokes, 1968).

In the Rosebery ore the larger scale deformational features include: (1) folding (as described in Section 4.1), (2) local diapiric intrusion of ore along anticlines (Section 4), (3) transposition and disruption of the ore layering, and (4) mobilization of chalcopyrite into tensional veins .

. Disruption of the ore layering is shown by boudinage structure of pyrite layers in sphalerite-galena ore. This structure is well displayed in an ore specimen from the Hercules Mine illustrated in Solomon (1964).

The chalcopyrite veinlets transgress the ore layering and extend outwards for up to several centimetres from the chalcopyriterich layers from which they are probably derived (e. g. 5-47). One such veinlet (5-58) displaces the ore layering in the sense of a normal fault.

On a smaller scale the effects of cataclastic type deformation

are shown by the larger pyrite crystals, which in sphalerite, galena and chalcopyrite matrices are fractured and brecciated. The fractures . are infilled with galena, chalcopyrite and less commonly sphalerite and quartz. As previously noted Some of the sphalerite shows deformational features such as bent and kinked twin lamellae.

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Some semimassive sulphide ore, particularly in the footwall ore at the south and north ends of the $Zn-Pb$ orebody, shows a schistose fabric in the gangue component and the sulphide minerals are coarser grained than usual.

As discussed in Section 5.3 the coarse-grained gangue mineralization is thought to have been derived via a "dispersed" phase from the ore and deposited in nearby dilatant zones. $\;$ The $\;$ coarse-grained mineralization is mostly post-folding and evidently deposition took place after the deformation had ceased.

6. CHEMICAL CHARACTERISTICS OF THE ORE MINERALIZATION

6.1 Metal Abundances and Distribution Patterns

A large amount of quantitative chemical data on the valuable metal content of the Rosebery ore deposit has been accumulated during routine testing. The purpose of the present investigation has been to use this data to study (1) frequency distributions of the different metals, (2) the differences in chemical composition between the Zn-Pb and Barite orebodies, (3) trends in metal content along strike within the Zn-Pb orebody, and (4) relationships between the different metals.

Sampling

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The assay data from the Zn-Pb orebody were collated into four groups covering the strike extent of the orebody and corresponding to the A and B lens, the D and E lens, the F lens, and the G lens groups (Fig. 29). The A and B lenses are not shown on Figure 29 but on 10, **11** and 12 Levels they occur as a separate ore development about **1,800** ft north of D lens (Fig. 28). The lens terminology dates back to the concept of the ore deposit as a series of discrete lenses and now serves only to geographically subdivide the Zn-Pb

orebody. Some of the divisions between lenses coincide with fold hinges while others are arbitrary. The lenses are now regarded as segments of a largely continuous ore "bed. "

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The assay samples consist of diamond drill core and face chip samples. The samples range from 1 to 10 ft in length corresponding to lengths of fairly uniform Zn content as estimated visually.

The assays used were from 10, 11 and 12 Levels for the ${\rm A}$ and B lens group and from 12, 13 and 14 Levels for the D and E, F, and G lens groups. These assays provide a fairly uniform cover over the strike extent of each lens group between two or three levels. A fifth group from D, E, F and G lenses on 12, 13 and 14 Levels consisted of face chip samples of 6 to 7 ft in length so as to keep sample size approximately constant. For the Barite orebody all available samples between 12 and 14 Levels were used.

The number of assay samples in each group ranges from 138 to 313 and was largely determined by availability. $\,$

All assays were carried out by the Assay Laboratory at Rosebery. Lead, Zn, Cu and Fe were determined by wet chemical methods while Ag and Au were determined by fire assay. Iron was determined as aqua regia-soluble Fe and a minor contribution from

non-sulphide Fe occurring in carbonate and chlorite gangue minerals may be included.

Metal Frequency Distributions

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Frequency distributions of metal assays in orebodies are commonly unimodal and range from slightly negatively skewed, through symmetrical (normal), to strongly positively skewed distributions. Empirically the type of distribution seems to depend on whether the distribution mode is low, medium or high with respect to the maximum occurring assay values (Becker and Hazen, 1961). Positively skewed distributions arise when the mode is low compared with the maximum assay values. This lognormal distribution is typical of low-grade and precious metal ores (Agterberg, 1961; Becker and Hazen, 1961) and is also characteristic of many trace and minor element distributions in rocks (Ahrens, 1953).

Negatively skewed frequency distributions arise when the mode is high compared with the maximum assay values and they are typical of many high-grade ore deposits such as Fe and Mn ores.

The frequency distributions for Pb, Zn and Cu in nine stratiform orebodies studied by Stanton (1958) were mainly either positively skewed or approximately symmetricaL The distribution type was similar for all three metals in each orebody and was

considered by Stanton to indicate that the distribution type was characteristic of locality rather than of mean abundance, or of particular metals. In the Broken Hill deposit frequency distributions for Pb, Zn, Cu and Ag all show positive skewness which decreases with increasing mean abundance (Stanton and Richards, 1961). Becker and Hazen (1961) studied a considerable number of assay frequency distributions from different ore deposits and concluded that there was apparently no correlation of distribution type with ore deposit genetic types. It would appear, however, that high-grade stratiform and sedimentary deposits commonly have negatively skewed metal frequency distributions, while in fissure vein deposits all the metals show positively skewed distributions (e. g. Richards, 1963).

The Rosebery assay data provide,an unusual opportunity to examine the frequency distributions of six different mefals in the one ore deposit.

Histograms for each of the six metals for each lens or orebody group are shown in Figures 43 to 48. For the metals occurring in minor or trace amounts in each lens group or orebody the histograms are characteristically unimodal and strongly positively skewed. This applies to Au, Ag and Cu in all groups, to Fe in the Barite orebody,

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and to a lesser extent to Pb in all groups. The metals of greater abundance show a variety of flat-topped and polymodal distribution but they are usually approximately symmetrical. This applies to Zn in the A and B, D and E, G, and D to G lens groups, and to Fe in the A and B, G, and D to G lens groups. Slight negative skewness is displayed in the histograms for Z n in the F lens and Fe in the D and E lens groups.. These two histograms are those in which the modal value is highest relative to the maximum occurring value.

The polymodal distributions probably indicate mixed populations, while the flat-toppedand skewed distributions suggest the presence of considerable trend in assay values within sample groups. Another effect is due to non-uniform sample size. Smaller samples tend to depart more widely from the mean value than larger samples and thereby the skewness of the distributions is increased. In the D to G lens group the samples chosen were restricted to face chip samples of 6 to 7 ft in length so as to give approximately uniform sample size. The histograms of the D to G lens group do in fact approach normal distributions more closely than in the other groups.

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Mean Metal Abundances

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The ·means and standard deviations for the metals in each group have been calculated (Table 3). In the calculations for the Barite orebody Au was omitted and was replaced by barite. These are not weighted means and are not true estimates of the average ore grade. Calculation of the means, standard deviations and correlation matrices was carried out on an Elliot 503 digital computer .

. A clear difference in mean abundances between the Barite and Zn-Pb orebodies is shown by the markedly lower Fe and to a lesser extent Au, Cu and Pb in the Barite orebody. Also there are statistically significant differences (at $P < 0.01$)* for the mean abundance of some or all metals between the different lens groups taken successively along strike (Fig. 29). These differences which indicate trends along strike are summarized as follows:

- (1) Lower Fe, Cu and Au in the A and B lens group compared to the D and E lens group.
- (2) Higher Fe but lower Pb, Zn, Cu, Ag and Au in the D and E lens group compared to the G lens group.

 $\rm*Group$ means were compared using the t-test which requires equal· variances and equal numbers of samples in each group. The data do not satisfy this exactly but the t-test probably gives sufficiently good results for the present purposes (e. g. Koch and Link, 1963), especially considering the large number of samples in each group.

TABLE 3 MEANS AND STANDARD DEVIATIONS

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no. of samples. M, mean in per cent.

S, standard deviation. (3) Higher Fe and Cu but lower Pb, Zn, Ag and Au in the G lens group compared to the F lens group.

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The central part of the orebody, comprising the D and E, and G lens groups, is enriched in Fe and Cu relative to F lens and the A and B lenses, which is probably a reflection of the greater abundance of pyritic ore in the central part of the orebody. The F lens group is of generally higher average grade with respect to Zn and Pb compared to the rest of the orebody.

6.2 Relationships between Metals

As a part of the study relationships between different metals, in particular correlation between pairs of metals, have been investigated. Two-variable correlation can be looked at qualitatively on scatter diagrams (e. g. Stanton, 1958) but a numerical measure of correlation is- given by the correlation coefficient, which is a number varying from +1, through 0 to -1. The sign shows whether the correlation is positive or negative and the magnitude of the coefficient gives the degree of correlation.

- Correlation analysis as applied to geochemical data is now a well established technique (e. g. Vistelius, 1958; Koch and Link, 1963; Vine, 1966).

Results and Discussion

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• Correlation coefficients were calculated for all combinations of metal pairs for all the sample groups (Table 4). A number of the metals showed lognormal type frequency distributions (Figs. 43 to 48). This was checked by plotting the assay data on logarithmic probability paper which gave reasonable straight line plots for Pb, Cu, Ag and Au in the sample groups from the Zn-Pb orebody. For meaningful correlation coefficients the data used should be normally distributed and accordingly the original assay figures for Pb, Cu, Ag and Au in all the lens groups in the Zn-Pb orebody and for Pb, Zn, Cu, Ag and Fe in the Barite orebody were transformed to logarithms prior to the calculation of the correlation matrices.

Using the z-test (e. g. Vistelius, 1958), correlation coefficients of greater than **0.20** are sta~istically significant at , **P<** 01.01 for the number of assays in each sample group.

In the five sample groups from the Zn-Pb orebody corelations between the different metal pairs are reasonably consistent throughout (Table 4). The pairs Pb-Zn, Pb-Ag, Zn-Ag and Ag-Au show lower positive correlation. The metals Pb-Zn-Ag-Au evidently constitute a distinct association in the Zn-Pb orebody.
TABLE 4 CORRELATION COEFFICIENTS

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Low to moderate positive correlation of Pb and Zn is characteristic in the stratiform pyritic Pb-Zn deposits investigated by Stanton (1955b, 1958). Of the other metals Fe displays moderate negative correlation with both Pb, Zn and to a lesser extent Ag, while Cu shows weak positive correlation with Fe and Au, and weak negative correlation with Pb. In the remaining pairs Zn-Cu, Cu-Ag and Au-Fe, the correlation coefficients although variable are generally not statistically significant at $P < 0.01$.

. The association of Pb, Zn, Ag and Au and their antipathy with Fe is reflected in the ore mineralogy by the compositional differentiation into sphalerite-galena-rich and pyrite-rich ore. Silver probably occurs in solid solution in both the galena and tetrahedrite-tennantite; hence the moderate positive correlation of Pb-Ag. Gold is commonly associated with the tetrahedritetennantite which is relatively richer in the sphalerite-galena ore.

In the Barite orebody the pairs Pb-Zn, Pb-Ag and Zn-Ag exhibit moderate positive correlation but, unlike the Zn-Pb orebody, Cu is positively correlated with Pb, Zn and Ag. This is probably a reflection of the mineragraphic observation of the greater relative abundance of tetrahedrite-tennantite in the Barite orebody compared to the Zn-Pb orebody. In the Zn-Pb orebody a larger part of the Cu :ls contained in chalcopyrite which does not tend to be associated with sphalerite- galena-rich ore.

Chayes (1960) has shown that correlation coefficients derived from percentage data are subject to restraints imposed by their forming a-closed array. This does not affect correlation among the variables occurring in minor amounts but the variables forming the major proportions will show negative correlation when in fact they may be positively correlated. In the present case the correlation coefficients for Zn-Fe are likely to be most affected. Chayes and Kruskall (1966) have developed a test for spurious negative correlations in closed arrays. However, the Rosebery data do not allow application of this test because only partial chemical analyses of the ore are available.

6. 3 Minor and Trace Elements

Iron in Sphalerite

The average Fe content of the sphalerite in the Rosebery ore is indicated by the values for Fe soluble in dilute $HC¹$ in the Zn concentrate. These values range from 2.0 to 2.3 per cent Fe for 12 monthly production-periods. Recalculating to give 100 per cent sphalerite brings these values up to about 2.5 per cent Fe in the pure sphalerite. Microscopic observations of sphalerite in thin and polished sections and grain mounts of Zn concentrate show a considerable variation in colour from pale yellow to red-brown reflecting a wide range in Fe content. Even adjacent bands in thin sections commonly show distinct colour differences.

Sphalerite in the Barite orebody is noticeably light coloured and along with the very low pyrite content points to a low Fe content in the source solutions for the Barite orebody.

Cadmium in Sphalerite

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The average Cd content of the sphalerite, as indicated by the Cd content of the Zn concentrate, is 0.17 per cent. Groves, in Groves and Loftus-Hills (1968), determined the Cd content of eight sphalerites from Rosebery and showed a range from 0.09 to 0.20 with an average of 0.16 per cent Cd. These values are distinctly lower than Cd in sphalerites from the Devonian cassiterite-sulphide and Pb-Zn-Ag fissure vein deposits of west Tasmania. Sphalerites from· pyritic Cu-Zn and Zn-Pb ores in volcanic rocks generally have lower than average Cd contents (Ivanov, 1964), and this may indicate a genetic relationship with volcanism (Groves. and Loftus-Hills, 1968).

Other Elements in the Sphalerite

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A number of trace element determinations have been made by the Electrolytic Zinc Company, Risdon, on Rosebery Zn concentrate. The results are Hg 100 ppm, T^{\perp} 16 ppm, In 12 ppm, Sn 10 ppm, Ge 2 ppm and Ga 1 ppm. Compared with the values compiled by Fleischer (1955) the.Bosebery sphalerite appears to be moderately enriched in Hg, Tl, In and Sn,' and low in Ge and Ga. In fact Hg is recovered as a by-product during refining. In view of the small number of analyses (1 to 3 for each element) these results are only an indication and a more comprehensive study is needed.

The Rosebery sphalerite is also low in Se (Loftus-Hills, 1968; Loftus-Hills, Groves and Solomon, in press).

Cobalt and Nickel in Pyrite

Loftus-Hills (1968) as part of a wider study has determined Co and Ni in pyrite from the Zn-Pb orebody and the enclosing rocks. Preliminary results are presented in Loftus-Hills and Solomon (1967), and in Groves and Loftus-Hills (1968).

In pyrite from the Zn-Pb orebody Co ranges from 0 to 650 ppm while Ni is very low, generally \leq 20 ppm. These values define a trend towards enrichment in Co, which correlates with the trends of Co enrichment in pyrite in the Mt. Read Volcanic Group and

associated granitic rocks, and in the Mt. Lyell ores. This trend is quite different from that of the Co and Ni in the pyrite from the Devonian Pb-Zn deposits and strongly suggests that the source of the metals in the Rosebery pyrite was from the Mt. Read Volcanic Group (Loftus -Hills, 1968).

6.4-Sulphur and Oxygen Isotopes

Stanton and Rafter (1966) presented $\,$ s $\,$ $\rm s^34$ S values for 26 samples of mixed sulphide ore from Rosebery for which the mean value is $+11.2$ per mil. Combining these results with a more comprehensive group of analyses* by Rafter and Jensen, reported in Solomon, Rafter and Jensen (in press), gives a mean value of +10. Q per mil., with a range of $+7.6$ to $+15.5$ per mil. and a standard deviation of 1.8. Sulphides in the Barite orebody average +17.0 per mll. but the lower end of the range overlaps with the range in values for the Zn-Pb orebody.

Enrichment in 34 S and low standard deviations are characteristic of pyritic Cu-Pb-Zn ores associated with volcanic rocks (Stanton and Rafter, 1966; Solomon et a1., in press). The low standard deviation implies that the sulphur is derived from a source of fairly constant composition and has not undergone any

'~Most of the samples on which these analyses were carried out were collected by the writer.

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appreciable contamination or fraclionation during transport and deposition. Magmatic or metamorphic processes are most likely to give a source of constant composition (Jensen, 1967), but this does not explain why Rosebery and other similar deposits show a distinct enrichment in 34° S. Possibly the enrichment may be due to volcanic : processes (Stanton and Rafter, 1966), but this has not yet been : demonstrated.

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The \int_{0}^{34} S values in barite from both the sulphide and Barite orebodies range from $+34.6$ to $+41.2$ per mil., and averages +38.4 per mil. with a standard deviation of 1. 7. These values together with oxygen isotope data could be compatible with derivation of the sulphate in the barite from sea water, connate water, meteoric water or magmatic water (Solomon et a1., in press).

Oxygen isotope values for the carbonate in the rhodochrosite from the Barite orebody indicate that the carbonate is of subsurface derivation (Solomon et a1., in press).

6.5 Lead Isotopes

Unlike many other stratiform orebodies the Pb isotope ratios of galena from Rosebery are slightly displaced from the primary ore-lead curve and are therefore anomalous (Ostic, Russell

and Stanton, 1967). The ratios are, however, very uniform which is characteristic of stratiform orebodies. Ostic et al. suggest that " the discrepancy between the Rosebery values and the primary orelead curve can be explained by the Rosebery Pb being either of multistage origin or due to a small heterogeneity in the primary system. .However, J. Richards (1967) pointed out that the primary ore-lead curve may be an over-simplified model and he indicated the need for comparative isotopic data from Pb in silicate minerals from rocks of the Mt. Read Volcanic Group.

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6. 6. Test for Mode of Deposition of Sulphides

Stanton (1966) has developed a method to distinguish between the deposition of sulphides as replacements or as sediments. He has applied the method to a number of stratiform sulphide orebodies including Rosebery. The published results for Rosebery are inconclusive. They tend to show an increase in the proportion of CO_2 to SiO₂ with increasing sulphide content. The increase in the proportion of CO₂ is very likely due to the contribution of Mn carbonate which was probably deposited from the same ore-forming solutions as the sulphides; a possibility which was hinted at by Stanton.

The method as applied to the Rosebery orebodies is unsatisfactory for a number of reasons.

- (1) The method requires fairly constant composition of the host rock as evidenced by constant proportions of $\mathtt{Al}_2\mathtt{O}_3$, $\mathtt{CO}_2^{}$ and $\mathtt{SiO}_2^{}$ representing the non-sulphide minerals. This has not yet been demonstrated for the Rosebery host rock and petrographic work indicates some variation in the proportions of the 'major components quartz, sericite and chlorite.
- (2)-The probability of addition of SiO₂ and CO₂ as well as sulphur during ore deposition invalidates a ternary plot in terms of SiO₂, CO₂ and A1₂O₃ as representing the sedimentary rock components.
- (3) Ordinary mine samples as used by Stanton are probably not suitable because they vary in size, come from widely separated parts -of the orebody and take no account of the spatial context of the samples. More specific sampling such as, at regular intervals along strike in one ore layer or group of layers is necessary. In practice, however, such sampling is very difficult because of inaccessibility $_{\rm problems}$.

6.7- Compositional Zoning Within and Between the Orebodies

Compositional zoning is present both in a stratigraphic sense across the deposit and laterally within the Barite and Zn-Pb orebodies.

Stratigraphic Zoning

Considering the deposit as a whole, the general sequence from footwall to hangingwall consists of an Fe-Cu sulphide zone in the footwall schist grading up to the Fe-Zn-Pb-Cu sulphide orebody which in places has an Fe-Cu basal zone. The Zn-Pb orebody is overlain by the sulphide-poor Barite orebody. In places hematitepyrite mineralisation is associated with·or immediately overlies the Barite orebody. This zonal sequence is reconstructed in its prefolding form in Figure 49.

The Fe-Cu zone in the footwall schist is composed of pyrite and chalcopyrite as disseminations, small massive lenses and veinlets. This zone is confined to the central and southern part of the ore deposit.

Within the Zn-Pb orebody there is an internal zonation with the basal part tending to be pyrite or pyrite-chalcopyrite-rich.

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 $*$ Minerals counted by automatic point counter with grid spacing of 0.2 mm.
Variance estimated at $\quad \sigma$ < 0.1 from Solomon and Green (1966), (P = mineral per cent). $rac{\sigma}{p}$

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This basal zone is not present throughout and is thickest in the central part of the orebody. It is well displayed on a cross section through the orebody at 400S (Fig. 5). The dimensions of this zone are of the order of 1,000 ft down-dip by 10 ft in thickness. Modal mineral analyses and chemical analyses for Zn , Pb , Cu and Fe through the orebody as intersected in drill hole R. 1418 on 400S are given in Table 5. The change from Fe-Cu-rich to Zn-Pb-rich ore is very sharp.

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As shown by the mean abundance data (Table 3) the Barite orebody is appreciably different in composition from the Zn-Pb orebody. The thickness of host rock between the Zn-Pb and Barite orebodies ranges from 0 to 60 ft and averages about 30 ft, which indicates a time break between the deposition of the Zn-Pb and Barite orebodies. Overlying the Barite orebody are several small concordant lenses of banded hematite-pyrite mineralization.

A sequence of stratigraphic zoning has been recorded from a number of stratiform. ore deposits that can be interpreted as exhalative-sedimentary in origin. Such deposits are Captains Flat (Edwards and Baker, 1953), Rammelsberg (Kraume et al., 1955; Anger et al., 1966) and Meggen (Ehrenberg et al., 1954) in Germany, Buchans in Newfoundland (Swanson and Brown, 1962; Anger, 1963),

the Kuroko ores of Japan (Maruyama, 1967), and a number of deposits in Russia listed by Qerbikov and Hurav'yeva (1967). Table 6 summarizes the zoning sequence in a number of these deposits. In some the zoning takes place within a single orebody, in others the different zones form adjacent but separate orebodies . In all the deposits the stratigraphic sequence from Fe-Cu-rich to Zn-Pb-rich is broadly similar. The barite zone is absent in many of the deposits, but where it occurs it invariably overlies the Zn-Pb and Fe-Cu ore. The zoning clearly represents a time sequence and consistently faces towards the direction of the younger rocks, even when the orebodies in their present position occupy overturned limbs of folds, e. g. Rammelsberg and Captains Flat. This provides independent confirmation that the Rosebery Mine sequence faces east and is right-way-up.

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Derbikov and Hurav'yeva (1967) included a number of Fe ore deposits in their·list of zoned sedimentary-volcanic deposits. This zoning goes from pyrite, with or without magnetite or chalcopyrite, at the base to hematite at the top and was correlated with a change from reducing to oxidising conditions. The hematitemagnetite-pyrite and hematite-pyrite lenses overlying the Zn-Pb and Barite orebodies at Rosebery may be a reflection of such a change to increasingly oxidised Fe minerals.

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TABLE 6 Stratigraphic Zoning in some Exhalative-Sedimentary Ore Deposits, (cL Derbikov & Hurav'yeva, **1967).**

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The repetition of a similar sequence from Fe-Cu to Zn-Pb and Ba-rich zones in a number of deposits of varying geological age suggests that a common process such as differentiation of a subvolcanic granitic magma has produced the ore-forming solutions. This process could be expected to give rise to a changing composition of the ore-forming solutions with time.

As pointed out by Anger et al. (1966) with reference to the Rammelsberg deposit, the zoning from Fe-Cu to Zn-Pb and barite ores is comparable to the classical zoning sequence in hydrothermal vein deposits. An example is Butte, Montana (Sales, 1914; Hewett, 1964), where there are three zones both in plan and depth:

- (1) Peripheral zone with abundant Zn, Pb and Mn, and inclUding minor amounts of barite.
- (2) Intermediate zone of Cu and Fe with minor amounts of Zn, Pb and Mn.
- (3) Central zone with abundant Cu and Fe, and lacking Zn, Pb and Mn .

The control of such a zonal sequence is generally held to be a temperature gradient. The resemblance to the sequence in the exhalative-sedimentary ores is clear and points to a similar kind of source for the ore-forming solutions.

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Lateral Zoning

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Lateral zoning from Zn-Pb-rich to Fe-Cu-rich ore (Section 5.1 J and from Zn-Pb-Fe sulphide ore to barite-rich ore is present in the Zn-Pb orebody. As well, the Barite orebody shows lateral changes in sulphide content and in places it passes into Zn-Pb-Fe-Cu sulphide ore (Section 5.2 J. Some of the lateral changes may be due to stratigraphic overlap of lower and higher zones; but as yet the field relations of the different *:wnes* are not well exposed.

6.8 Wall-Rock Alteration

As discussed in Section 3.7 the Cambrian rocks in the Rosebery area have undergone low-grade dynamothermal metamorphism which tends to obscure the signs of any earlier wallrock alteration. Nevertheless it might be expected that some of the chemical and fabric changes of wall-rock alteration would survive a later metamorphism. In fact the footwall schists possibly represent a zone of alteration formed during the deposition of the orebodies. The maximum thickness of the schist zone coincides with the central part of the Zn-Pb orebody. Disseminated pyrite with minor amounts of chalcopyrite is prevalent in the footwall schists. Other schist zones are present within the Primrose Pyroclastics and

it has been suggested (Section 3.8: Loftus-Hills, et al., 1967) that they represent zones which were more responsive to later deformation and metamorphism than the enclosing.pyroclastic rocks because they had undergone an earlier alteration by hydrothermal volcanic solutions. However, this hypothesis remains to be tested by chemical analyses. to.see if there are differences between the schists and the tuffs of the footwall pyroclastics that could be ascribed to wall-rock alteration.

In contrast to the footwall schists, the host rock overlying the Zn-Pb orebody is relatively unaltered. Bedding and other primary features are reasonably well preserved in the host rock. Other workers (Hall et aI., 1953; Solomon, 1964 and 1967a) have noted sericitization in the host rock. The sericite is, however, present right through the sequence, thousands of feet east and west of the orebodies, and is probably mainly metamorphic in origin. From X -ray modal analyses $(C.E. Gee, pers. comm.)$ the black slate contains comparable amounts of sericite to the host rock. Although; the footwall schists below the orebodies are certainly richer in sericite than the tuffs of the Primrose Pyroc1astics and this may be a. consequence of wall-rock alteration.

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The apparent lack of wall-rock alteration in the host rock overlying the Zn-Pb orebody suggests that these rocks were deposited after the emplacement of the orebody.

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The pattern of altered, mineralized and schisted volcanics overlain by stratiform sulphide mineralization which is in turn overlain by relatively unaltered shale, tuff or lava occurs in a number of stratiform pyritic.orebodies in volcanic terrains, such as the orebodies of the Noranda and Matagami areas of Quebec (Roscoe, 1965; Sharpe, 1965; Boldy, 1968) and the Rio Tinto ores of Spain. The Kurpko ores of Japan (Maruyama, 1967) represent a similar pattern in a non-metamorphosed terrain.