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GEOLOGY DEPT.
University of Tasmania
THE GEOLOGICAL SETTING AND FORMATION OF
THE ROSEBERY VOLCANIC-HOSTED MASSIVE
SULPHIDE OREBODY, TASMANIA.

by Geoffrey R. Green, B.Sc.(Hons.)

A thesis submitted in partial fulfillment of the
requirements for the degree of Doctor of Philosophy.

UNIVERSITY OF TASMANIA

HOBART
1983
Except as stated herein, this thesis contains no material which has been accepted for the award of any other degree or diploma in any University, and, to the best of my knowledge and belief, this thesis contains no copy or paraphrase of material previously published or written by another person, except where due reference is made in the text of the thesis.

Geoffrey R. Green.

September, 1983
"Rosebery: A tale of despair."

... Blainey (1954)
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ABSTRACT

The Rosebery pyritic zinc-lead-copper-silver-gold orebody is the major massive sulphide deposit associated with the Cambrian Mount Read Volcanics of western Tasmania. The Mount Read Volcanic belt in the Rosebery area is composed principally of rhyolite and is in fault contact with the Rosebery Group to the west. The latter group, now preserved as a number of fault slices, consists predominantly of marine sedimentary rocks, deposited below wave base, and derived partly from the Mount Read Volcanics and partly from a Precambrian metamorphic terrain (the Tyennan Geanticline) further east. The Rosebery Group sediments were probably deposited contemporaneously with part of the Mount Read Volcanics and are supposed to be correlative with the Middle to Late Cambrian Dundas Group.

Ore formation occurred in a marine environment, following deposition of a thick, mainly welded, ignimbrite and attendant subsidence. The orebody consists of a folded, discontinuous massive sulphide horizon and in the southern part of the mine there is a normal across-layer metal zonation with Fe-Cu-rich ore overlain by Zn-Pb-Ag-rich ore. A complementary lateral metal zonation is present with a lower and central Fe-Cu-rich zone surrounded by a Zn-Pb-Ag-rich zone. Barite-sulphide ore occurs as lenses higher in the sequence separated from the sulphide orebody by unmineralized cleaved siltstone. Overlying rocks include siltstone with a minor component of volcaniclastic sandstone and turbidites derived from the Tyennan region. Succeeding pyroclastic units are unwelded and, in contrast to rocks in the footwall of the orebody, contain quartz phenocrysts and rip-up mudstone clasts. Later feldspar-phyric rhyolite marks a change to less explosive volcanism and probably a return to subaerial conditions.
A broad alteration zone of quartz + sericite ± pyrite underlies the Rosebery ore deposit. Flanking alteration assemblages include partial carbonate - and sericite - replacement of plagioclase pheno­
crysts in both hanging wall and footwall volcanic rocks. Towards the massive sulphide ore chlorite and pyrite increase in abundance in the footwall rocks with chlorite and chalcopyrite being more prominent below Cu-rich sections of the then lenticular orebody. Chemically footwall alteration resulted in a general increase in Fe, Mg, Rb, K, Mn and H_2O and a marked decrease in Na, Sr, and generally, Ca. The Co content of pyrite is highest under Cu-rich sections of the orebody. 

δ^34S values of sulphide minerals show good correlation with the lateral metal zonation, but are largely independent of the across-layer zonation. δ^34S values range from 8.0 to 13.4‰ in the Fe-Cu-rich lateral zones, to 15.1 to 17.2‰ in the major Pb-Zn-rich zone at the southern end of the orebody. This pattern is paralleled by a decrease in the Fe/Fe + Mg + Mn ratios of chlorite and a decrease in the FeS content of sphalerite. The δ^34S values of sulphides in the barite orebody are even higher at 14.5 to 19.8‰ and the FeS content of sphalerite is lower. Barite in the barite orebody has a range of values of 34.6 to 41.2‰. The δ^34S values of barite-sulphide pairs indicate a temperature range of 255^°C to 295^°C for the barite orebody, but deposition of this ore is believed to have occurred at about 250^°C.

The presence of premetamorphic arsenopyrite in the sulphide orebody suggests deposition of this ore was from solutions in which H_2S was the dominant sulphur species. This sulphur was probably derived from a mixture of totally reduced seawater sulphate and sulphur derived from leaching of the underlying rocks. The δ^34S value of the barite orebody may represent partial reduction of seawater sulphate.
The orebody and enclosing rocks were deformed and metamorphosed to the lower greenschist facies in the Devonian, with textural evidence suggesting that the peak of metamorphism post-dated cleavage formation in the rocks. The formation of rare biotite and spessartine, corrosion of arsenopyrite by tetrahedrite-tennantite indicate that the peak metamorphic temperature may have been around 350°C. Chlorite compositions indicate equilibration at temperatures of about 250°C, probably during retrograde metamorphism.

The use of $\delta^{34}\text{S}$ values of sulphides as stratigraphic markers indicates that the sulphide orebody is diachronous. The lateral metal and sulphur isotope zonation may be related to discharge of ore solutions as a plume showing reversing buoyancy during formation of the Pb-Zn-rich ore, or a shift in the locus of solution discharge with time, or a combination of these factors. The ore fluid is believed to have been generated by thermally-driven convective circulation of seawater, the energy source probably being a Cambrian granitic pluton.
The volcanic-hosted massive sulphide deposit at Rosebery, Tasmania (145°33'E, 41°46'S) is a relatively large orebody of its type. At the 30th June, 1981 some 10.9 million tonnes of ore grading 18.0% Zn, 5.5% Pb, 0.8% Cu, 14.3% Fe, 187 g/t Ag and 2.8 g/t Au had been produced (Adams et al., 1976). Proven plus provable reserves at that date amounted to 7.64 million tonnes of ore grading 0.59% Cu, 5.2% Pb, 16.4% Zn, 175 g/t Ag and 3.6 g/t Au were quoted (Electrolytic Zonc Company of Australasia Ltd. Annual Report, 1981; J.H.A. Mill, pers. comm.). Current annual production is about 560,000 tonnes of ore.

The Rosebery ore deposit is one of an important sub-class of volcanic-hosted ores termed "Rosebery-type" by Solomon & Walshe (1979). Among the characteristics of this class are relatively high ore tonnages and Zn+Pb/Cu ratios, fine banding in the massive ore, a well developed metal zonation parallel to the layering (e.g. Lusk, 1969), a thin sheet-like morphology, generally relatively poorly-defined alteration pipes in the pre-ore rocks and a tendency to form vertically-stacked ore sheets. Another significant feature is a mineral assemblage containing arsenopyrite ± pyrrhotite suggestive of ore deposition from a relatively reduced fluid (e.g. Ohmoto, 1972; Green, 1979). Deposits of the type are generally of Palaeozoic age and include Hercules, Tasmania; Captains Flat, New South Wales (Davis, 1976); Woodlawn, New South Wales (Malone, 1979); the deposits of the Bathurst district, New Brunswick (e.g. Lusk, 1969). Some of the above features are shown by the Archaean Kidd Creek deposit, Canada (e.g. Sangster & Scott, 1976). The occurrence of welded tuffs near a number of these deposits (Rosebery, Hercules, Captains Flat, Woodlawn) suggests that cauldron subsidence may have occurred during formation of the ore-forming environment (e.g. Ohmoto, 1978). Consequently, study of the
environment of ore formation at Rosebery should have implications for a significant number of other deposits.

The approach of this study was governed by the fact that Brathwaite (1970, 1972, 1974) had described the mine geology in detail and had thoroughly investigated the structural geology of the Rosebery orebody. His work also established the exhalative origin of the ore, which had been tentatively suggested earlier (Hall & Solomon, 1962) but was still not generally accepted (e.g. Hall et al., 1965).

It was therefore decided to complement Brathwaite's approach by:

1. Detailed geological mapping of the Rosebery area, the aims being to clarify the relationship between the volcanic succession hosting the orebody and the sedimentary units to the west, and to establish the environment of ore deposition.

2. An investigation of the mechanism of ore deposition through a consideration of metal zoning, ore and alteration mineralogy and a sulphur isotope study. The purpose of this work was to gain an insight into the physical and geochemical mechanisms of ore formation. This has necessitated an appraisal of the conditions of greenschist facies metamorphism which affected the orebody subsequent to its formation.

Previous Literature

Early workers (Montgomery, 1895; Waller, 1902; Hills, 1915; Finucane, 1932; and Hall et al., 1953, 1965) all favoured an origin of the orebody by replacement of schist or carbonate rocks following Devonian deformation. Hall & Solomon (1962), Campana & King (1963), Pereira (1963), and Solomon (1964) were the first to tentatively suggest a possible exhalative origin for the ore, but this was by no means generally accepted (e.g. Hall et al., 1965). At about the same time as Brathwaite's work, sulphur isotope studies (Stanton & Rafter, 1966; Solomon et al., 1969) and trace element studies (Loftus-Hills & Solomon, 1967; Loftus-Hills,
1.3 provided supporting evidence for the syn-sedimentary mode of formation of the ore. More recent work has been aimed at refining the genetic concepts and providing guides for exploration. Thus, Eastoe (1973) studied the mineralogy and distribution of some elements in the sedimentary unit immediately hosting the orebody, Fitzgerald (1974) did a similar study of the nearby Hercules mine, emphasizing Hg distribution, and Smith (1975) analysed numerous samples of the sedimentary host rock and surrounding volcanics and studied precious and volatile element distributions in particular. McLeod (1978) as part of a wider study of mineralogical associations in eastern Australian massive sulphide deposits, studied sphalerite, chlorite, white mica and carbonate compositions of the orebody. Dixon (1980) investigated the mineralogy and geological associations of the carbonates in the orebody and carried out some carbon and oxygen isotope analyses. Solomon & Walshe (1979), in a study of the behaviour of hydrothermal solutions following venting onto the seafloor, proposed that the characteristics of Rosebery-type ore deposits could be accounted for by stagnation of a hydrothermal plume at the air-water interface in a shallow water environment, though this model was later superseded (Green et al., 1981). Much of the material covered in this thesis has been published (Green et al., 1981).

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Discussion with staff and members of the Department of Mines, Tasmania, was invaluable throughout the project. Two days in the field with E. Williams on separate occasions provided me with insight into the structural geology of the area. On the first occasion, he pointed out the significance of the deformation associated with the contact between the Rosebery Group and the Primrose Pyroclastics. On a subsequent visit the remark that I understood perhaps 90% of the geology provided the impetus to examine the Pieman River sections once more and led to discovery of the fault contacts within the Rosebery Group. P.W. Baillie, P.L.F. Collins, K.D. Corbett, N.J. Turner, and especially A.V. Brown, who has provided information on Cambrian stratigraphy and volcanic associations prior to publication, have also provided useful discussions.

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of this thesis), S.E. Drummond, Jr., C.S. Eldridge, D. Wesolowski, V. Pisutha-Arnond and other members of the U.S.-Japan-Canada research project on the Kuroko deposits, for much spirited discussion and exchange of ideas.

M. Solomon suggested the project, supervised the thesis and provided much insight, advice, and criticism of some naive ideas. During the latter part of the project, as co-author of the published in Economic Geology, he put considerable effort into the editing and refining of a rambling manuscript and provided most of the input into the discussion of the behaviour of the ore fluid after discharge into the sea. J.L. Walshe provided much discussion of geochemical processes. As co-author of the paper in Economic Geology his studies of chalcopyrite solubility and the chlorite solid-solution model provided a quantitative basis for the discussion of geochemical processes. He has also provided an unpublished modification of the chlorite model, which has enabled me to carry out further calculations on the conditions of chlorite equilibration. Others provided invaluable discussion of the manuscript originally submitted to Economic Geology and aided considerably in its refinement. Unfortunately I am unaware of their individual contributions, but R.A. Both and R.R. Large are especially thanked in this regard. At this time D. Archibald and M.J. McDonald also provided contributions by discussions with M. Solomon.

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(76-324), and W. Jablonski of the University of Tasmania analysed arsenopyrites from the sulphide orebody and sphalerites from the barite orebody. The expertise of P.B. Nankivell, T.R. Bellis, A. Hollick, J. Ładaniwskyj and J. Pongratz in drafting diagrams, and J. Pongratz, A. Taylor, C. Humphries, V. Smith and L. Laughlin in typing is acknowledged. Most of the isotope analyses were carried out at the C.S.I.R.O. Mineral Research Laboratories, Sydney. J.W. Smith and M.S. Burns taught me the techniques of preparation of SO₂ samples from sulphide samples and performed the isotopic analyses of these samples.

Finally, the co-operation, assistance, and moral support of my wife, Lee, and children Melanie and Douglas, have been essential in bringing this project to completion.

Location Description

The localities of specimens collected from surface outcrops in the Rosebery and Hercules areas are given in metres and described in terms of the Australian Map Grid (A.M.G.) in accord with the convention adopted on the Tasmanian Lands Department 1:100 000 topographic sheet Sophia (Sheet 8014, Edition 2, 1973). A map of localities of central western Tasmania used in the text is included in the pocket at the back of the thesis (Fig. 1.1). The localities of specimens collected from drill core and underground workings in the immediate locality of the Rosebery Mine are given in feet and described in terms of the Rosebery Mine Grid. The origin of this grid is at 378 870.55 m E 53 181.69 m N A.M.G. and mine grid north is 348°36' A.M.G. To avoid conclusion, for sample locations described by the mine grid, northings are listed before eastings. For samples described by the A.M.G. the reverse holds.

Footnote

After receipt of examiners' reports, sections of this thesis have been rewritten. Affected sections include pp. 6.11 onwards and Figs. 6.3 - 6.5 inclusive, Sections 7.2(ii) and 7.3, and p. 8.24. Also Appendix 6 has been added. References relating to these sections are listed in the Additional References section and indicated in the text by an asterisk. These include two references (Walshe, 1982; Styrt et al., 1981) cited in the original text, but inadvertently omitted from the original list of references. The modifications to the thesis were done in March-April, 1984. The writer is grateful to his referees for pointing out shortcomings in the original text, but none of the major conclusions has been changed.
Chapter 2

REGIONAL GEOLOGY

The regional geology of western Tasmania has been discussed in a number of recent reviews (Williams et al., 1976; Williams, 1978; Solomon, 1981) and only the salient features will be discussed here. Precambrian terrains separate a number of narrow linear belts of Cambrian rocks and similar troughs occur in Victoria (Vande Berg, 1978) and Antarctica (Laird et al., 1977). The Mount Read Volcanics, which host the Rosebery ore deposit, occupy the eastern part of the Dundas Trough, the largest belt of Cambrian rocks in Tasmania (Fig. 2.1).

2.1 PRECAMBRIAN FRAMEWORK

The Tyennan region, to the east of the Dundas Trough, is underlain by quartzite, quartz-mica schist, phyllite and subordinate amphibolite and eclogite. Two phases of isoclinal folding and upper greenschist facies metamorphism of the Frenchman Orogeny occurred in these rocks at about 800 Ma and a later important fold event is dated at about 600 Ma (Raheim & Compston, 1977).

Lesser deformed and metamorphosed rocks of the Rocky Cape region to the west of the Dundas Trough are composed of two main sequences. The western sequence, the Rocky Cape Group, consists of shallow marine quartz sandstone, siltstone and shale (Gee, 1968), whereas the Burnie Formation to the east is predominantly a quartzwacke-mudstone turbidite sequence with minor limestone and mafic lava (Gee, 1976; Green, 1976b). Dolerite dykes intruding the Burnie Formation have been dated at 725 ± 35 Ma (J.R. Richards in Crook, 1979) and have been interpreted as syntectonic by Gee (1976) or as synsedimentary (Crook, 1979). Folding of the rocks of the Rocky Cape Group is attributed to the Penguin Orogeny. The relative ages of the Burnie Formation and the Rocky Cape Group are...
Fig. 2.1 Geological map of western Tasmania slightly modified after Solomon et al. (1976).
not well known. The boundary between them is occupied by the Arthur Lineament (Gee, 1968) a zone of Penguin deformation some 10 km wide consisting of quartz-mica schist, quartzite and greenschist. Metamorphosed pillow lava and metasediments within the Arthur Lineament host the Savage River iron orebody, a stratiform, apparently exhalative, magnetite-pyrite deposit (Coleman, 1976). The deformed mafic rocks yield K-Ar ages on biotite and hornblende of 744 ± 22 Ma (data of Coleman, 1976 corrected to the decay constants of Steiger & Jager, 1977, D.C. Green, pers. comm.). Correlates of the Burnie Formation occur in the Mount Bischoff and Zeehan areas (Figs 2.1 and 2.2), the correlative sequence in the Zeehan area being known as the Oonah Formation, which contains dolomite horizons and units of metabasalt of alkaline affinity towards the top of the sequence (Blissett, 1962; Varne & Foden, 1980; Brown, 1982).

Rocks of both the Tyennan and Rocky Cape regions are overlain unconformably by dolomite-dominated sequences, the Jane Dolomite (Spry, 1962a) and the Smithton Dolomite (Lennox et al., 1982) respectively. Stromatolitic detritus in a diamictite overlying the Smithton Dolomite has been dated at 1000 to 680 Ma, based on comparison with dated stromatolitic sequences elsewhere (Griffin & Preiss, 1976). The younger age may indicate the maximum time break between the Penguin Orogeny and the earliest preserved Precambrian deposits.

2.2 CLASTIC SEQUENCES OF THE DUNDAS TROUGH

The Dundas Trough was initiated along a fracture close to the original boundary between the two contrasting belts of Precambrian rocks. Rocks with a similar deformational history to those of the Tyennan region occur west of the Dundas Trough at Cape Sorell (Baillie et al., 1978) and rocks similar to those of the Burnie Formation crop out within the
Fig. 2.2 Geological map of central western Tasmania showing the divisions of the Mount Read Volcanics (after Corbett, 1981).
Dundas Trough (Turner, 1977). These relationships lend support to the concept that the Dundas Trough developed by rifting of the Precambrian terrain (Williams, 1978).

At the western margin of the Dundas Trough, the Success Creek Group and correlates in the Renison Bell area overlie the Oonah Formation unconformably (Fig. 2.3; Brown, 1980; Williams et al., 1976). These rocks consist of some 500 m of a basal conglomerate, shallow marine (?) quartz sandstone, siltstone, carbonate horizons, and hematitic conglomerate (Brown, 1980; Taylor, 1954; Patterson et al., 1981). The conformably overlying Crimson Creek Formation, approximately 3000 m thick, is composed of mudstone, mafic volcaniclastic turbidite and minor tholeiitic basalt and carbonate horizons (Blissett, 1962; Brown & Waldron, 1982).

A number of ultramafic to mafic igneous complexes are present in the Dundas Trough. The largest and most complete of these, the Heazlewood River Complex, is dismembered by faulting, but probably consisted of basal serpentinized and deformed dunite and orthopyroxenite, overlain by layered cumulate harzburgite and orthopyroxenite which pass transitionally upward into a layered gabbro suite. This unit is intruded by gabbro dykes and is overlain by volcanic rocks including tholeiitic lava and dacitic agglomerate (Rubenach, 1973; Brown et al., 1980). This complex and the similar Serpentine Hill Complex west of Renison Bell were considered to represent oceanic crust flooring the Dundas Trough (Solomon & Griffiths, 1972; Rubenach, 1974; Williams, 1978). Brown & Waldron (1982) disagreed with this view and distinguished three phases of mafic volcanic activity which accompanied development of the Dundas Trough. They believe the first stage involved outpourings of tholeiitic lava and intrusion of gabbro during deposition of the Crimson Creek Formation.
Fig. 2.3 Geological history of northwestern Tasmania, showing principal types of ore deposits and their relationships to rock types (modified after Collins, 1978).

<table>
<thead>
<tr>
<th>Geological Unit</th>
<th>Rock Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Holocene-Tertiary</td>
<td>Glacial, glacio-fluvial, alluvial deposits. alluvial deposits</td>
</tr>
<tr>
<td>Tertiary</td>
<td>Basalt</td>
</tr>
<tr>
<td>Jurassic</td>
<td>Dolerite</td>
</tr>
<tr>
<td>Triassic-Late Carboniferous</td>
<td>Glacial/glacio-marine/fresh water sequences (Parmeener Supergroup).</td>
</tr>
<tr>
<td>Late Devonian</td>
<td>+ + Adamellite-granodiorite.</td>
</tr>
<tr>
<td>Early Devonian-Late Cambrian</td>
<td>Quartz sandstone and mudstone (Eldon Group and correlates).</td>
</tr>
<tr>
<td></td>
<td>Gordon Limestone and correlates.</td>
</tr>
<tr>
<td>Late Cambrian-Precambrian</td>
<td>Siliceous conglomerate &amp; quartz sandstone (Owen Conglomerate and correlates).</td>
</tr>
<tr>
<td></td>
<td>Dominantly greywacke turbidite sequences (Dundas Group and correlates).</td>
</tr>
<tr>
<td></td>
<td>Mudstone/chert/greywacke/limestone (Crimson Creek Formation and correlates).</td>
</tr>
</tbody>
</table>
|                           | Dolomite/limestone/quartzite/mudstone (Success Creek Group and correlates).
|                           | Dominantly felsic-intermediate volcanic rock. (Mount Read Volcanics and correlates). |
|                           | Dominantly mafic volcanic rock.                                           |
|                           | Mafic-ultramafic and associated rocks.                                    |
|                           | Granite                                                                   |
| Precambrian               | Mudstone/orthoquartzite/quartzwacke/basalt/carbonate. (Rocky Cape Group; Burnie and Oonah Formations). |
|                           | Pelitic and quartzite sequences; amphibolite. (Tyennan and Forth regions; Arthur Lineament). |
|                           | Granite                                                                   |

Legend:
- Fissure vein
- Stratiform massive sulphide and/or oxide
- Skarn in contact aureole
- Metasomatic replacement bodies
- Unconformity
- Erosion surface
- Thrust fault
The basalts have relatively high TiO$_2$ (> 0.7%) and rare earth element contents ($\text{La}_{\text{chondrite}}$, or $\text{La}_N$, values of 30-80), and light REE enrichment with chondrite-normalized La/Yb ratios ($(\text{La}/\text{Yb})_N$ values) of about 1.5 to 3.5.

Succeeding lavas, of the high magnesian andesite association are chemically similar to boninites (Varne & Brown, 1978), are depleted in REE ($\text{La}_N$ of 8-12), have concave REE patterns with $(\text{La}/\text{Yb})_N$ values of about 1 and low TiO$_2$ content (< 0.1%). This suite is believed to be the surface representation of the ultramafic intrusive bodies and followed deposition of the Crimson Creek Formation. A final phase of low-TiO$_2$ (0.2 to 0.6%) basalt interfingers with the basal units of the overlying Dundas Group, and has even lower REE contents ($\text{La}_N$ values of 0.5 to 1) and $(\text{La}/\text{Yb})_N$ values of less than 0.5 with negatively sloping REE patterns.

The middle Middle to middle Late Cambrian Dundas Group (Elliston, 1954; Blissett, 1962; Jago, 1979) in its type area consists of some 3800 m of mudstone, shale, lithicwacke, and conglomerate units with chert, mudstone and quartzite clasts. A number of felsic volcaniclastic horizons are present (Blissett, 1962; A.V. Brown, pers. comm.). Uplift of the ultramafic-mafic complexes resulted in spilitic and minor ultramafic detritus being included in the basalt units of the Dundas Group (Rubenach, 1974). Contacts between the Crimson Creek Formation and the Dundas Group are faulted (Brown, 1982). The correlative Huskisson Group (Taylor, 1954; Blissett, 1962) contains a higher proportion of felsic volcaniclastic detritus (A.V. Brown, pers. comm.).

The sedimentary rocks west of Rosebery, the Rosebery Group of Taylor (1954), include sandstone, mudstone, conglomerate and tuff horizons. The Rosebery Group is described in detail in the following chapter.
2.3 MOUNT READ VOLCANICS

The Mount Read Volcanics occupy a meridional belt, at the eastern margin of the Dundas Trough, extending some 160 km from Elliott Bay in the south to the Que River in the north and 10-15 km in width. Similar rocks crop out within belts of Cambrian rocks at the northern margin of the Tyennan Geanticline.

The Mount Read Volcanics are predominantly calcalkaline volcanic rocks (Spry, 1962b), chemically similar to Andean volcanics (Anderson, 1972; White, 1975) of mainly rhyolite to dacite composition with andesite prominent in some areas. They are host to a number of mineral deposits, the most prominent of which are the massive pyritic Zn-Pb-Cu-Ag-Au ore-bodies of Rosebery, Hercules and Que River, and the disseminated to massive pyritic Cu deposits of Mount Lyell.

Corbett (1981) has subdivided the Mount Read Volcanics into three broad units (Fig. 2.2), which have been adopted here. The following discussion is drawn mainly from Corbett's work, with other workers cited as required.

2.3 (i) Central sequence

The central sequence forms the bulk of the Mount Read Volcanics and consists mainly of rhyolite to dacite lava, welded and nonwelded ignimbrite, breccia, agglomerate and tuff. A number of areas of andesite and, more rarely, basalt are present. Clastic sedimentary rocks, most commonly dark grey pyritic siltstone and slate, form a relatively minor component of the succession. Intrusive bodies of granites and adamellite occur near Mt Murchison and Mt Darwin (Fig. 2.2).

The major orebodies occur in the central sequence, albeit in different geological settings. The Mount Lyell deposits occupy a belt 4 km long by 1 km wide of strongly altered felsic pyroclastic rocks and
lavas and intermediate intrusive and extrusive rocks (Corbett et al., 1974; Reid, 1976; Walshe & Solomon, 1981). The disseminated deposits lowest in the sequence form broadly concordant sheets of disseminated pyrite-chalcopyrite in quartz-sericite ± chlorite rocks (Walshe & Solomon, 1981). Stratigraphically higher bornite-chalcopyrite-barite bearing deposits occur within, or adjacent to, irregular bodies of chert which may have been siliceous sister deposits. Massive banded pyrite-chalcopyrite and pyrite-sphalerite-galena ores probably represent exhalative mineralization which accumulated in topographic depressions on the seafloor (Markham, 1968; Green, 1971; Solomon, 1976). Sulphides from the chert-associated, commonly bornite-bearing ores have lower $\delta^{34}$S values than those from the disseminated pyrite-chalcopyrite ores and are believed to have been derived from leaching of these ores by fluids containing both $H_2S$ and sulphate species (Walshe & Solomon, 1981).

The Rosebery and Hercules massive sulphide orebodies occur in siltstone lenses within rhyolitic pyroclastic sequences (Brathwaite, 1974; Green et al., 1981) and will be discussed in detail in later chapters.

The geology of the Que River mine area is dominated by andesitic agglomerate, breccia and lava (Webster & Skey, 1980; Wallace & Green, 1982). The immediate mine sequence includes cleaved and sericitized dacitic lapilli tuff, tuff and massive porphyritic bulbous-shaped dacite bodies of extrusive to shallow intrusive origin. Two main horizons of mineralization are present. The lower zone consists of vein and stringer pyrite with accessory chalcopyrite, sphalerite and galena in porphyritic dacite with a lens of massive pyrite, sphalerite and galena at the top of the dacite. The upper body, separated from the lower by massive dacite and tuff is a massive lens of pyrite, sphalerite, and galena with accessory chalcopyrite, tetrahedrite and arsenopyrite with sericite, chlorite, quartz and minor barite gangue.
Numerous subeconomic volcanic-hosted deposits occur within the Mount Read Volcanics. In general the deposits between South Darwin Peak and Mount Sedgwick (Fig. 2.2) are Cu-rich disseminated deposits, whereas from Hercules northward polymetallic Zn-Pb-Cu-Ag-Au pyritic massive sulphide deposits predominate. Solomon (1979) ascribed this to different levels of erosion along the Mount Read Volcanic belt caused by northward plunging Late Cambrian folds. Alternatively, the boundary between predominantly Cu-rich and Zn-Pb-rich deposits could be drawn along the Henty Fault Zone, an important Devonian structure, which may have been active in the Cambrian (Solomon, 1981).

2.3 (ii) Western sequence

The western sequence (Fig. 2.2) is characterized by a higher proportion of clastic sedimentary units than the central sequence. These are dominated by siltstone and mudstone and a number of Precambrian-derived sandstone units occur. Quartz-phyric tuffaceous rocks, frequently displaying features of cold subaqueous mass-flow emplacement, including 'crystal tuffs', are common as are fine-grained vitric tuffs. Intrusive bodies, particularly of quartz-feldspar porphyry are common, but lava and massive welded ignimbrite units are rare (Corbett, 1981). Mineralization in the western sequence is comparatively minor and the main style developed is disseminated and vein Pb-Zn adjacent to major fault zones (e.g. the Henty River Prospect - Reid & Meares, 1981; Silver Falls).

In the Que River area, the Que River Beds, consisting of black slate and minor volcaniclastic rocks, contain a marine fauna of late Middle Cambrian age (Gee et al., 1970; Jago, 1979). This unit, locally the basal unit of the western sequence, is overlain by quartz-phyric breccia with a vitroclastic matrix containing blocks of the underlying black slate up to 500 mm in diameter (Gee et al., op. cit.). A few kilometres to the south the Que River Beds are overlain with apparent conformity by the
Animal Creek Greywacke which consists of some 500 m of Precambrian-derived quartzwacke turbidite, vitric tuff and mudstone (Collins, 1981b).

The contact between the western and central sequence is regionally discordant (Corbett, 1981; Green et al., 1981; Collins, 1981b) and between Williamsford and north of the Pieman River the central sequence is faulted against the Rosebery Group, composed of a number of formations of clastic and volcaniclastic rocks (Green et al., 1981), which is described in detail in Section 3.2. Considerable diversity of opinion exists as to the relative ages of the western and central sequences. In the Mount Darwin-Queenstown area, Corbett et al. (1974) regarded the central sequence as predominantly older than, but partly coeval with, the western sequence and White (1975) believed the central sequence (his Intercolonial Volcanics) to be the older unit. Corbett (1979, 1981) has reversed this interpretation and has suggested that the western sequence formed the western wall of an elongate volcano-tectonic rift which was filled by volcanic rocks of the central sequence. These relationships, particularly where they affect the Rosebery area, will be discussed in more detail in the next chapter.

2.3 (iii) Tyndall Group

The mineralized volcanic rocks of the Mount Lyell field are overlain disconformably at Lyell Comstock by a bioclastic limestone containing marine fossils of late Middle Cambrian age (Jago et al., 1972; Jago, 1979). The limestone is the basal unit of the Comstock Tuff (Corbett et al., 1974) which consists of quartz-phyric crystal tuff, lava and agglomerate, siltstone and an overlying unit, the Jukes Formation, consisting mainly of volcaniclastic sandstone and conglomerate with an increasing proportion of quartzose Precambrian-derived detritus upward. These formations comprise the Tyndall Group of Corbett et al. (1974).

The Tyndall Group is equivalent in age to the upper part of the Dundas Group and probably to the western sequence in the area between Que River and the Pinnacles (Fig. 2.2), which appears to be a conformable
sequence with the Que River Beds at its base (Collins, 1981b). Common
to these sequences is the presence of quartz-bearing volcaniclastic rocks,
and sedimentary rocks containing an important component of Precambrian-
derived detritus.

2.4 LATE CAMBRIAN TO DEVONIAN SEQUENCES

The Owen Conglomerate consists predominantly of closed-framework
siliceous conglomerate of probable terrestrial origin fining upward into
marine quartz sandstone (Walshe & Solomon, 1981). In the Tyndall Range
area north of Queenstown, the basal unit of the formation consists of a
quartz-wacke turbidite-mudstone sequence of middle Late Cambrian age
(Corbett, 1975; Jago, 1979). Deposition of the Owen Conglomerate and
correlates occurred in fault-bounded troughs and, locally, conformable and
gradational relationships exist between the siliceous conglomerates and
the underlying Jukes Formation and Dundas Group, but elsewhere the base is
disconformable to unconformable on the Cambrian sequences (e.g. Wade &
Solomon, 1958; Campana & King, 1963; Corbett et al., 1974; Solomon, 1969;
Williams, 1978). Local disconformities within the Owen Conglomerate and
rapid lateral thickness variations (e.g. Solomon, 1969) suggest continued
fault movement during deposition. At least locally, significant compres­
sion and uplift occurred prior to deposition of the Owen Conglomerate as
disorientated cleaved clasts of the volcanics are present in the correlate
of the Jukes Conglomerate at Mount Darwin (S.F. Cox in Corbett, 1979).
This break between the Jukes Conglomerate and the central sequence volcanics
is known as the Jukesian Unconformity (Carey & Banks, 1954; Williams, 1979).

The Early Ordovician marked the beginning of a shallow marine trans­
gression which affected the Tyennan Geanticline as well as the trough
sequences. Deposition of the Gordon Limestone throughout most of the
Ordovician (Banks & Burrett, 1979) was followed by deposition of shallow
marine sandstone, mudstone, shale and minor limestone of the Eldon Group
in the Silurian to middle Early Devonian (Banks, 1962; Flood, 1974).

2.5 DEVONIAN DEFORMATION

A major deformational episode halted sedimentation during the Devonian in western Tasmania. The presence of plant fossils of middle Middle Devonian age in unfolded cave sediments within cleaved Gordon Limestone, near Devonport in northern Tasmania, places an upper limit on the age of this deformation (Balme in Burns, 1965).

The earliest phase of Devonian folding in western Tasmania produced flattened parallel folds with steeply dipping axial surfaces of half wavelength between 2 and 15 km. The trend of these folds was governed by the disposition of the competent blocks of Precambrian rocks which were little affected by this phase of deformation. In general folds become tighter towards the Tyennan block with high angle reverse faults important locally (Williams, 1978).

The second phase of deformation produced NW to WNW trending folds with associated thrusts dipping to the NE. The most important structure, the Linda Disturbance (Bradley, 1956; Solomon, 1965) extends in an ESE direction from Trial Harbour (500E, 550N, Fig. 2.2) through the Mount Lyell area and into the Tyennan region. West of Rosebery, the southern part of the Huskisson Syncline (800E, 700N, Fig. 2.2) strikes in a NW direction and has been attributed to this phase of deformation (Williams, 1978).

2.6 DEVONIAN TO CARBONIFEROUS GRANITOID EMBLACEMENT AND MINERALIZATION

The passive emplacement of granitic bodies in western Tasmania post-dated Devonian deformation and occurred at about 340 to 360 Ma (McDougall & Leggo, 1965; Brooks, 1966, ages adjusted to decay constants of Steiger & Jager, 1977). Dominant rock types are porphyritic and non-porphyritic
biotite adamellite and granite (Groves et al., 1972; Klominsky, 1972; Patterson et al., 1981). In general the plutons show discordant contacts with their country rocks and have metamorphic aureoles up to 3 km wide.

Major pyrrhotite-cassiterite replacement deposits of Eocambrian to Cambrian carbonate sequences at Mount Bischoff, Renison Bell and Cleveland are related to the granites (Groves et al., 1972; Patterson et al., 1981; Collins, 1981a). These deposits are commonly surrounded by haloes of Pb-Zn-Ag vein deposits (e.g. Solomon, 1981).

Within the Mount Read Volcanics several Pb-Zn-Ag vein deposits occur within, or adjacent to, major Devonian faults including the Mount Farrell Ag-Pb-Zn field (Burton, 1976), which may include a component of remobilized Cambrian mineralization (Solomon et al., 1969). Gravity modelling by Leaman et al. (1980) indicates that the Mount Read Volcanics between Hercules and Que River occupy an area in which granite might lie at a depth shallower than 1 km below sea level. Consequently it might be expected that Devonian metamorphism could have played a significant role in the post-depositional history of the Rosebery orebody. This will be discussed in Chapter 7.
3.1 MOUNT READ VOLCANICS

The Mount Read Volcanics in the Rosebery-Hercules area have been subdivided into two units by Brathwaite (1970, 1974), namely the Primrose Pyroclastics and the overlying Mount Black Volcanics. This stratigraphic subdivision has been found to be viable in this study, although it masks some complexity particularly in the upper part of the Primrose Pyroclastics (Fig. 3.1).

3.1 (i) Footwall Pyroclastics

The lowest unit of the Primrose Pyroclastics, the footwall pyroclastics, lies beneath the ore-bearing shale lenses in the Rosebery-Williamsford area and consists of a uniform assemblage of vitric-crystal-lapilli tuff with phenocrysts of albite and/or K-feldspar, the latter showing a peculiar mottled texture under crossed polars. The base of the footwall pyroclastics is not exposed because of faulting but the unit is at least 1000 m thick. Flattened discoid chloritic fragments are prominent in most outcrops and contrast sharply with the pink, grey or cream matrix. They define a marked foliation which is parallel to bedding in the overlying clastic sedimentary rocks. These 'fiamme' are believed to represent relict collapsed pumice fragments (Fig. 3.2). The primary foliation is oblique to cleavage.

Apart from variation in phenocryst content from about 5 to 15% and presence or absence of fiamme, the footwall pyroclastics are extremely uniform over a mapped strike length, open to the south and truncated to the north by faulting, of 11 km. Their mineralogy is simple with feldspar phenocrysts in a fine grained felsitic groundmass of quartz, albite and K-feldspar. Chlorite, carbonate and epidote are common secondary minerals, the latter two frequently replacing feldspar.
Fig. 3.2 Hand specimen (natural size) of welded vitric-crystal tuff from footwall pyroclastics (75-665) showing primary foliation due to relict collapsed pumice fragments (darker grey).

Fig. 3.3 Photomicrograph of welded tuff (77-2) from footwall pyroclastics. Collapsed pumice fragment at lower left shows marked internal lineation and is distorted against blocky lithic fragment at left. Strong welded tuff has virtually obliterated shard outlines. Plane polarized light.
phenocrysts. Sericite is almost ubiquitous and increases in abundance towards zones of mineralization in which the rock grades into quartz-sericite schist in which all feldspar has been destroyed. These zones are believed to represent zones of hydrothermal alteration associated with massive sulphide formation (Loftus-Hills et al., 1967) and will be discussed in detail in Chapter 5. Zircon and apatite are accessory minerals.

In many thin sections the finer textural detail in the rocks has been obliterated by sericitization and cleavage development. In some uncleaved specimens (77-2; 77-3; 75-765; 75-771; 75-766), the groundmass of the rock displays welded and deformed relict glass shards, rarely showing ghost axiolitic textures (Figs. 3.3, 3.4). This texture is readily distinguished from the later cleavage foliation observed in both probable welded and unwelded tuffs (Fig. 3.5). Thus the mesoscopic foliation present at the outcrop scale is interpreted to be compatible with the welding and distortion of glassy fragments observable in thin section. Snowflake devitrification textures similar to those produced in the laboratory by high temperature devitrification of rhyolite glass (Lofgren, 1971) occur in a few sections (75-646; 75-650B). These features, together with the homogeneity over the considerable thickness and lateral extent of the footwall pyroclastics, are strongly indicative that the rocks are the deposits of hot ash flows (e.g. Ross and Smith, 1961). This suggestion, for part, or all, of the Primrose Pyroclastics has previously been made by Spry (1962b), Solomon (1964), Hall (1967), Brathwaite (1974) and Eastoe (1973). The presence of both welded and unwelded tuff and minor lithological variations in the footwall pyroclastics indicate a number of flows were responsible for their accumulation.
Fig. 3.4 Welded tuff (75-771) from the footwall pyroclastics. Shards are not strongly deformed, note especially the crescentic shard at left except adjacent to collapsed chloritic pumice fragment (dark area at top). Relict axiolitic texture is present in the larger shards, particularly that at lower centre. Plane polarized light, width of field of view is 1.63 mm.

Fig. 3.5 Cleaved, unwelded vitric-crystal tuff from Rosebery Lodes area (specimen V-10 of Eastoe, 1973; specimen held by E.Z. Company, Rosebery). Fine-grained pyrite euhedra with quartz pressure shadows are present, particularly at top left. Plane polarized light, scale bar is 1 mm long.
Although the assertion of Rankin (1960) that welded ash flow tuff deposits imply a subaerial environment of deposition has been widely accepted, this has been disputed by Mutti (1965), Francis and Howells (1973), Wright and Coward (1977) and Sparks et al. (1980a, b). The papers of Francis and Howells (op.cit.) and Wright and Coward (op. cit.) convincingly demonstrate welding and therefore hot ash-flow emplacement of tuffs in the Ordovician sequences of Wales. In these cases the tuffs were believed to be erupted on land before entering shallow water in which the hot ash flow retained sufficient heat for welding to occur. In this case, however, peculiar structures such as quartz nodules, mega-flames (up to 200 m long), and rootless vent-like structures 20 m in diameter in which originally discoid fiamme are drawn up into elongate ellipsoids in the pipes, are present. These features are interpreted to be due to the escape of steam generated by volatilization of seawater trapped beneath and within tuff units (Francis and Howells, op. cit; Wright and Coward, op. cit.), and are probably diagnostic of subaqueously emplaced hot ash-flow tuff deposits. In addition, the top part of the units shows evidence of reworking by marine currents (Francis and Howells, op. cit.).

None of the features described above have been recognised in the footwall pyroclastics and this evidence, together with a lack of interbedded sedimentary rocks, is strongly indicative of a subaerial depositional environment for the unit. Sparks et al. (1980b) argued on theoretical grounds that welding of a subaerially erupted hot ash-flow which had flowed into water was likely, but that a critical requirement was that the flow had to have a density greater than water. This criterion is most likely to be met by dense flows rich in non-vesicular lithic clasts, quite unlike the pumice- and shard-rich footwall pyroclastics.
This was confirmed by the experimental work of Sheridan (1979) which indicated a flow density of less than one at the onset of fluidization in pyroclastic deposits with a similar grainsize distribution to the footwall pyroclastics.

Mutti (1965) believed that a 5 m thick shard-bearing graded tuffaceous sandstone bed occurring within a deep marine sequence on the island of Rhodes was welded, but it has now been established that the unit was emplaced by a cold turbidity current and is not welded (Wright and Mutti, 1982). Sparks et al. (1980a) described a 13 km long lobe projecting above the seafloor extending seaward from an ignimbrite eruptive centre in Dominica. Welded tuff was dredged from the fan which was therefore inferred to consist largely of welded tuff. The morphology of the deposit is similar to that of submarine fan lobes composed of clastic sediments and it seems equally likely that the deposit was formed by a cold mass flow, initiated perhaps by explosive disruption of a hot ash-flow, containing clasts of welded tuff, entering water (cf. G.P.L. Walker, 1979; J.V. Wright, pers. comm., 1982).

3.1. (ii) Host Rock

The host rock, which is up to 70 m thick in the Rosebery mine area, conformably overlies the footwall pyroclastics and consists predominately of siltstone, slate and lenses of crystal tuff (Brathwaite, 1974). In the mine area cleavage commonly masks bedding in the host rock and the mineralogy of the unit is quartz, sericite, chlorite, with minor carbonate, pyrite and albite and accessory sphene, zircon, rutile and tourmaline (Eastoe, 1973). Locally intense zones of silicification and chloritization occur in the footwall (Adams et al., 1976) and the former may mask the contact with the footwall pyroclastics. Concentrations of carbonate may occur in both the footwall and hanging-wall sections (Adams et al., op. cit.).
Away from zones of strong alteration the predominant rock type is laminated pale grey sericitic siltstone and mudstone with laminae of siltstone rich in quartz, feldspar and opaque minerals (75-767). Chlorite and carbonate are less abundant and zones of silicification are absent. Crystal-rich tuffs are prominent north of Rosebery, particularly towards the top of the host rock. A prominent rock type (75-386; -387; 766) is a coarse grained crystal vitric tuff with outsize curvilinear shards which are replaced by quartz. The shards show some alignment and the rock was previously misidentified as a welded tuff (Green, 1976) but the shards are discrete and do not display evidence of welding. In one specimen (75-385) irregularly intertonguing contacts between crystal tuff and siltstone suggest the former was emplaced by cold subaqueous flow.

3.1 (iii) Black Slate

The host rock is overlain abruptly by dark grey, thinly laminated slate which is up to 30 m thick and contains 0.11 to 0.67 mass % of non-carbonate carbon (Gee, 1970). Pyrite occurs in thin bands of euhedra parallel either to cleavage or bedding. Its Co and Ni content (Loftus-Hills and Solomon, 1967; Loftus-Hills, 1968) and sulphur isotope composition (Solomon et al., 1969; Schwarcz and Burnie, 1973) indicate a biogenic origin. A minor component of crystal tuff occurs in laminae a few millimetres thick up to beds 30 cm thick. Carbonate-rich horizons are present (75-764). Quartz-feldspar porphyry intrusives occur in both the host rock and black slate (e.g. 75-388) and have also been described from the Dallwitz area 6 km south of Rosebery, where the groundmass has a granophyric texture (Eastoe, 1973). An unusual, but significant lithology, is sandstone, in graded beds up to one metre thick which occurs towards the top of the black slate. The sandstone (77-699; -700; Fig. 3.6), is composed of angular sand grains of entirely
Fig. 3.6 Photomicrograph of turbidite sandstone (77-699) from black slate, Rosebery open cut area. Sandstone consists of clasts of quartzite and quartz-mica schist (centre and top) in a silt-grade matrix. Crossed polars, width of field of view is 4 mm.

Fig. 3.7 Hand specimen of crystal-lithic tuff (105188) from the footwall pyroclastics. Black to grey chloritic fragments (flattened pumice ?) define a foliation which is parallel to regional bedding, but the high proportion of crystals and dense lithic fragments and thin section characteristics argue against a hot ash-flow origin. Included in the large chloritic fragment at upper right is a vitric crystal tuff clast, possibly from the footwall pyroclastics. Natural size.
metamorphic origin, including quartzite, phyllite, polydeformed quartz-mica schist and strained quartz in a sericitic matrix. The high proportion, and nature, of the rock fragments, the textural immaturity and the lack of any volcanic detritus are all indications that the sand was derived directly from the deformed Precambrian rocks of the Tyennan Geanticline to the east. The bed contains outsize, up to 10 cm long, discoid, rip-up clasts of slate, identical to the underlying black slate, and is clearly a turbidite deposit.

The black slate and host rock represent a period of relative quiescence in volcanic activity during which the major sulphide mineralisation formed in a marine environment. Also, the nature of the volcanism changed from feldspar- to feldspar-quartz-phyric at this time.

3.1 (iv) Massive Pyroclastics

Local disconformity occurs at the base of the massive pyroclastics with the black slate absent and the host rock thinned in places (Brathwaite, 1974). The disconformity is believed to be due to scour and the basal unit of the massive pyroclastics is rich in black shale clasts (Gee, 1970; Brathwaite, 1974; Adams et al., 1976).

The pyroclastics above the Rosebery ore horizon are a diverse suite of rocks both along and across strike in contrast to the homogeneous footwall pyroclastics. The main distinguishing features are:

(a) The massive pyroclastics contain numerous breccia horizons commonly containing shale clasts, which are absent in the footwall pyroclastics.

(b) Quartz phenocrysts are common in the massive pyroclastics and host rock, but are absent in the footwall pyroclastics.

(c) K-feldspar phenocrysts are restricted to the footwall pyroclastics.
(d) Although some rocks in the massive pyroclastics have sericitized feldspar phenocrysts, the extensive development of quartz-sericite schist is generally restricted to the footwall pyroclastics.

The massive pyroclastics in the area comprise two main facies.

The lower facies, about 450 m thick, is composed of aphyric rhyolite breccia, quartz-plagioclase bearing breccia, quartz-feldspar crystal tuff (105190, 75-382; Fig. 3.7) and banded fine grained felsic tuff. No detailed correlations could be made between sections of this unit, but the presence of breccia is characteristic. The breccia contains clasts of black shale, glassy tuff, quartz porphyry and probable fragments of footwall welded tuff, generally a few cm in diameter, but occasionally up to 30 cm in diameter. A few thin horizons of siltstone and shale occur and in one of these at the Dalmeny Mine minor sulphide mineralization is present. Another characteristic is the local presence of outsize rafts of black shale up to several metres long in breccia and crystal tuff units.

These rocks are well exposed between the contact of the Rosebery Group and the Mount Read Volcanics in the Pieman River (7744E, 7844N) and the Bastyan damsite (7756E, 7833N). At the damsite, when the bed of the Pieman River was exposed in 1979, rafts of black shale up to tens of metres in length were observed in coarse grained quartz-feldspar phryic lithic tuff. A large irregular clast, some 2 m in length, of massive pyrite and minor sphalerite also was exposed in the unit (K.D. Corbett and P.L.F. Collins, pers. comm.).

The lower units of the massive pyroclastics occur on the western margin of the Primrose Pyroclastics south west of Rosebery where they appear west of a faulted anticlinal structure (Fig. 3.1; Section 3.3). The rock succession in this area is inferred to face west and a sericitic
3.11

siltstone, black slate and crystal tuff units correlated with the host rock and black slate are flanked to the west by strongly cleaved quartz-feldspar phryic tuff and breccia with black slate fragments (75-652; -653; 76-371) which is correlated with the massive pyroclastics. Similar quartz-feldspar phryic tuff with slate and felsic volcanic fragments occurs west of the Hercules mine (75-748; -749; -750). Within this sequence graded sandstone horizons up to 20 cm thick with minor interbedded black shale occur (75-747). This rock type is very similar to the Stitt Quartzite (Section 3.2 (ii)).

The presence of a weak primary foliation in the massive pyroclastics (Fig. 3.7) previously led the writer (Green, 1976b) to conclude that the massive pyroclastics were probably hot ash-flow deposits deposited on land, similar to the footwall pyroclastics, as had been suggested previously by Brathwaite (1970, 1974). However, a weak primary foliation of pumice clasts has been noted in submarine pyroclastic rocks by Fiske (1969) and is common, for example, in some of the Miocene pyroclastic rocks in Japan (Tanimura et al., 1974) which are inferred to have been deposited at water depths in excess of 2500 m (Guber and Ohmoto, 1978). Flattened, weakly aligned pumice fragments are therefore not a palaeoenvironmental indicator. Shards are rare in thin sections of the massive pyroclastics but where present are undeformed. The bulk of the evidence, including breccias with shale fragments at several horizons, outsize shale rafts and occasional size grading of pyroclastic debris displayed well in drill hole 71R suggests that the unit was emplaced mainly as a number of submarine density flows or slides which disrupted quiet water sedimentation represented by the rare shale horizons and more common shale rafts.

The abundance and size of both shale fragments and quartz phenocrysts decreases southwards, and east of Hercules the lower 250 m of the massive
pyroclastics are represented by plagioclase-phyric vitric crystal tuff with pumice fragments which are mineralogically similar to the footwall pyroclastics, but are unwelded. The quartz-phyric, shale-fragment bearing pyroclastic suite crops out over 10 km and disappears near Hercules (Fig. 3.1) and a source area to the north is therefore inferred.

The only possible instance of deposition of hot pyroclastic material in the lower part of the massive pyroclastics at Rosebery occurs in specimens of crystal lithic tuff (75-384A, 75-769) which contain fragments of mafic scoria and rhyolite glass in a non-welded vitroclastic matrix. This material is in irregular contact with an altered aphyric rock with peculiar ovoid structures up to 2.5 mm in diameter (75-369); 75-304; 75-763) which occur singly or are interconnected by thin necks. Some of these are partly filled with fine grained cherty material, probably originally glassy dust. The surface of this chert is subparallel between ovoids (geopetal texture?). The remainder of the ovoids are filled with coarse-grained quartz and rarer albite (Fig. 3.8). The origin of these structures is puzzling, but they may be similar to the quartz nodules discussed, but not well described, from the subaqueous welded tuffs of Ordovician age in North Wales (Francis and Howells, 1973; Wright and Coward, 1977).

The upper unit of the massive pyroclastics, which is at least 120 m thick, is composed of green pumice tuff and agglomerate containing sparse albite phenocrysts (75-380A, B). Pumice fragments, up to 5 cm in diameter, are represented by equant blebs of spongy chlorite with microcrystalline quartz inclusions in a felsitic matrix (Fig. 3.9). These rocks are best developed 2 km north of the Rosebery orebody, but are also found as far south as the Dalmeny mine area. These rocks are intruded by quartz feldspar porphyry with chlorite alteration and abundant disseminated
Fig. 3.8  Ovoids of quartz and albite in vitric tuff of the massive pyroclastics (specimen 75-769). Ovoids at left are connected by a thin neck of quartz. Ovoid at right shows geopetal structure, right hand side of ovoid consists of fine cherty (vitric ?) material; left hand section consists of coarse-grained quartz. Crossed nicols, width of field is 4 mm.

Fig. 3.9  Pumice tuff from upper part of massive pyroclastics. Equant chloritic pumice clasts with vesicles filled with quartz in glassy matrix. Glass shards are absent. Plane polarized light, width of field is 4 mm.
magnetite (75-605; -606; -607; 75-381). The porphyry has a snowflake texture groundmass.

The first appearances of possible welded tuff in the massive pyroclastics are near the top of the unit in the Mount Read area and north of Rosebery, the latter occurrence being about 670 m stratigraphically above the ore horizon at Rosebery (specimen 83-6).

In the Chester Mine area 6 km north of Rosebery there is a marked change in the lithology of the Primrose Pyroclastics and the volcanic succession is dominated by quartz-feldspar phryic rhyolite lava and welded tuff and dacitic to andesitic crystal tuff and lava (Stevens, 1974). Just north of the Pieman River the volcanic succession contains doubly-graded quartz-feldspar bearing deposits of the type described by Fiske and Matsuda (1963) and is cut by numerous quartz-feldspar porphyry and fine-grained mafic dykes. A noteworthy rock type in this area is a strongly welded tuff with thin relict collapsed pumice fragments up to several metres long (Fig. 3.10), identical in lithology to the footwall pyroclastics. In contrast to the near vertical attitude of the surrounding rocks, the primary foliation in this tuff has a shallow dip and it may be an exotic block of the footwall pyroclastics which slid into the depositional basin from an adjacent fault scarp.

3.1 (v) Mount Black Volcanics

The Mount Black Volcanics generally overlie the Primrose Pyroclastics with a conformable gradational contact and are composed of two units, a lower unit of monotonous rhyolitic and subordinate dacitic flow-banded to autobreciated lava and an upper unit of dacitic, andesitic and rare basaltic crystal tuff and lava (based on rock analyses in Anderson, 1972, and Collins, 1981).
Fig. 3.10  Welded tuff with pronounced eutaxitic foliation (parallel to stick) at 7983E, 8058N. Fiamme are up to a few metres long. Cleavage, visible in weathered rock at top of road cutting, is subvertical. The outcrop may be part of a megablock of the footwall pyroclastics which slid into the area of deposition of the massive pyroclastics.

Fig. 3.11  Breccia at base of Mount Black Volcanics at 7962E, 8027N consisting of angular clasts of flow-banded and massive rhyolite in tuffaceous matrix. Irregular pyrite clast (?) is present just to top right of lens cap.
The unit is probably some 2500 m thick (Green et al., 1981), but lack of structural reference surfaces make this estimate uncertain. Intrusions of dolerite, quartz-feldspar- and feldspar-porphyry are common. In the area north of the Pieman River, just to the east of the possible exotic block of welded tuff previously described, the base of the Mount Black Volcanics is marked by a coarse rhyolite breccia with lava blocks up to 1 metre in diameter set in a glassy matrix. Rare zones containing contorted pumiceous fragments are present and one pyrite clast was noted (Fig. 3.11).

3.1 (vi) Cauldron Subsidence

Several factors point to rapid subsidence before deposition of the host rock and black slate at Rosebery, notably the preceding accumulation of a thick sheet of ash-flow tuff. Sufficient subsidence must have occurred to allow transport of detritus from the Precambrian craton to the east by turbidity currents without intermixing of volcanic detritus as indicated by the sandstone horizons in the black slate at Rosebery and in the massive pyroclastics west of Hercules. The thickness of undoubtedly submarine pyroclastic rocks in the massive pyroclastics north of Rosebery of about 700 m gives some indication of the magnitude of water depth during mineralization, although fluctuations in the elevation of the basin may have occurred during eruption of these rocks. If the footwall pyroclastics were deposited entirely in a subaerial environment, as argued previously, then a minimum amount of subsidence of the order of 1700 m may be inferred. This figure, although large, is comparable with the amount of subsidence (up to 3000 m) deduced for the Long Valley Caldera, California, caused by eruption of the Bishop Tuff (Bailey et al., 1976).

The extent of the inferred volcanic depression is unknown at present, but its northern margin may lie just north of the Pieman River, and
a minimum diameter of about 11 km may be inferred. It is difficult
to establish whether resurgent doming was an important process in the
cauldron cycle, as suggested by Smith and Bailey (1968), as the best
candidates for domal rhyolites, the lower Mount Black Volcanics, form a
very extensive unit and have been traced for over 10 km north of the
suggested northern cauldron margin (Collins, 1981b). It is clear, however,
that major hydrothermal activity, best exemplified by the Rosebery orebody,
ocurred during pre-resurgence sedimentation, not long after cauldron
subsidence, and not during a later post-resurgence doming stage as is
common in subaerial settings (Smith and Bailey, op. cit.).

3.2 ROSEBERY GROUP

The sedimentary rocks west of Rosebery have been the subject of much
discussion and confusion in the literature and subdivision into formations
has generally been made without detailed consideration of facing evidence
or contact relationships between formations.

Hills (1914) correlated these rocks with slates at Dundas before the
Dundas Group was formally defined (Elliston, 1954). Finucane (1932)
introduced the term Rosebery Series. Taylor (1954) coined the term
Rosebery Group and correlated it with the Dundas Group. Campana and King
(1963) introduced a number of formation names, but considered the Rosebery
Group to face east and included the footwall pyroclastics as the upper part
of the group. In this they followed the practice of Electrolytic Zinc
Company geologists (Ball et al., 1953; Taylor, 1954) who used cleavage/
bedding relationships as the primary criterion of facing. Loftus-Hills
et al. (1967), employing sedimentary structures as facing indicators,
showed that the rocks dominantly faced west and correlated them with the
Success Creek Group, an opinion shared by Solomon (1965). Loftus-Hills
et al. also recognised the complexity of the Pieman River gorge section
north of Rosebery, noting that the stratigraphy could not be matched across
their inferred fold hinges. Brathwaite (1970) redefined the formations of the Rosebery Group and distinguished a shale unit, the Chamberlain Shale, from the Primrose Pyroclastics and nominated it as the basal member of the Rosebery Group. Williams et al. (1976) suggested that the Dundas Group and Rosebery Group might be correlative, mainly on the basis that fuchsite clasts in a conglomerate horizon of the Rosebery Group might be of deformed ultramafic origin. Green and R. Williams (1975; in Adams et al., 1976) and Green et al. (1981) recognised that the contact between the Mount Read Volcanics and the Rosebery Group was faulted between Williamsford and the Pieman River (at 3775E, 3785N) because both formations of the Rosebery Group and the host rock horizon of the Primrose Pyroclastics were truncated by the contact, the fact that the contact was a locus for Devonian mineralisation, and by the deformation associated with the contact in the Pieman River (E. Williams, pers. comm. 1975).

Corbett (1981) published without acknowledgement Brathwaite's (1970) stratigraphic names and accepted the correlation of the Rosebery Group with the Dundas Group. Corbett considered that the Rosebery Group-Primrose Pyroclastics contact formed a scarp against which the feldspar-phyric ignimbrite of the Primrose Pyroclastics was ponded, that is, he inferred that the Rosebery Group was the older sequence.

The purpose of this section is to collate information based on field mapping and petrography by the writer, to point out the reasons for some of the confusion in the literature and to suggest the most feasible internal stratigraphy and regional correlation for the Rosebery Group. Part of the prior confusion apparently stems from a lack of appreciation of the style of deformation in these rocks, and consequently the following treatment will emphasise demonstrably conformable lithologic sequences rather than formal stratigraphic formations. However, the formation names of Brathwaite (1970) have been retained as an aid to nomenclature.
3.2 (1) Chamberlain Shale

The Mount Read Volcanics in the Rosebery area are flanked to the west by the Chamberlain Shale (Brathwaite, 1970), a unit of tightly folded grey slate, mudstone, and volcaniclastic sandstone best exposed along a road from 7743E, 7477N to 7808E, 7439N (Fig. 3.1). The eastern part of the unit consists of massive volcaniclastic sandstone in beds up to some 20 m thick separated by thin units of slate. The sandstone (75-602) consists of slightly abraded B-quartz and albite crystals, up to 2 mm in diameter, subordinate felsic volcanic and rare phyllite and quartzite rock fragments in a sparse matrix of fine-grained quartz and feldspar (Fig. 3.12). Rare irregular shale clasts and graded bedding in one unit suggest the sandstone may be a mass-flow deposit, a suggestion made for 'crystal tuffs' in general by Cas (1980) and discussed in some detail by Cas and Wright (1982, Part 8). Similar rocks occur in the Baker Creek area south of Williamsford (around 7550E, 6750N; 75-746). The western part of the unit is composed of slate and subordinate mudstone. A coarse-grained siltstone (75-594) 120 m east of the western boundary of the Chamberlain Shale contains quartz, sericitized relict feldspar and euhedral zircon and apatite grains as do other mudstone beds in the sequence (75-599). Although tight folding, strong cleavage development (in some sections demonstrably two cleavages), and the fine grain size of the rocks in the western part of the Chamberlain Shale makes interpretation difficult, it may be inferred that much of the silt-to-clay-grade material in the unit is probably of volcanic derivation.

Few reliable facings have been obtained from the Chamberlain Shale except within some 80 m of its western margin. Here an increasing frequency westward of laminated, cross- and convolute-laminated sandstone horizons demonstrate a gradational contact with the overlying Stitt Quartzite.
Fig. 3.12 Volcaniclastic sandstone (75-602) consisting of partly abraded volcanic quartz and albite crystals and rarer lithic clasts (e.g. phyllite clast at centre of photograph). Crossed nicols, width of field is 4 mm.

Fig. 3.13 Amalgamated sandstone beds from thick-bedded sandstone facies of Stitt Quartzite, Pieman River gorge. At left is laminated sandstone (mid grey) overlain by massive paler grey sandstone beds with minor shale partings and rip-up shale clast in second bed from right. Irregular amalgamation surface between sandstone beds to right of pick. Facing of sequence is to right.
3.2 (ii) **Stitt Quartzite**

The Stitt Quartzite is a prominent marker unit, consisting of some 350 m of quartzwacke, quartzarenite, siltstone, dark grey slate, and a few horizons of conglomerate with clasts up to 5 cm in diameter. Two facies are present, a dominantly medium to thickly bedded sandstone-dominated suite and a thinly bedded to laminated suite of fine to very fine grained siltstone and slate.

The thickly bedded sandstone facies occurs in units from 4 to some 100 m thick. The basal contact of the association is a scour (prefolding dip 15°) at one locality (7717E, 7826N) although in general evidence for channelling is absent. Beds, up to 1 m thick of coarse- to medium-grained sandstone and, exceptionally, granule conglomerate, are normally massive and ungraded (76-297, 76-298) or display coarse tail grading (Blatt et al., 1972, Fig. 3.2b). Development of a weak planar lamination occurs towards the top of some graded beds and is in some cases succeeded by a thin division of cross lamination (A, AB or ABC Bouma sequence). Amalgamated sandstone beds and occasional rip-up clasts of mudstone are present (Fig. 3.13). Thin to very thin interbeds of sandstone, mudstone and laminated siltstone shale occur, but generally form only a minor proportion (< 10%), in terms of thickness, of the facies.

Sole structures are rare, but possible flute and groove casts at the base of thick sandstone beds have been noted at one locality (Fig. 3.14). Load structures at the base of sandstone beds are more common and include ball and pillow structures and pseudonodules.

Detrital grains are predominantly quartz-rich but comprise a variety of types, notably strained or unstrained monocrystalline quartz, polycrystalline quartz with straight, annealed grain boundaries, quartzite, ribbon quartz with sutured subgrain boundaries and quartz-mica schist.
Fig. 3.14  Groove casts at base of massive sandstone bed in Stitt Quartzite (at 7729E, 7848N).

Fig. 3.15  Photomicrograph of granule sandstone (76-283) from Stitt Quartzite containing grains of quartz-mica-garnet (?) schist (top), quartzite and monocrystalline quartz. The range of clast types is identical to that in the sandstone from the black slate at Rosebery (77-699; Fig. 3.6) but a higher proportion of grains of monocrystalline quartz is present in this specimen. Width of field is 4 mm, crossed nicols.
Phyllite clasts and large flakes of muscovite and rare biotite, locally partly retrograded to chlorite, are common (76-282, -283, -284, -285, -286, 291, -292, 293, -294, -295, -296, -297, 298). Carbonate grains are present in some rocks (e.g. 76-295). Accessory minerals include greenish-brown to brown tourmaline, rounded zircon and rutile. The greatest diversity of clast types is seen in the conglomerate (76-283). In this thin section one grain of volcanic provenance, a rounded grain of aphyric rhyolite 1 mm in diameter with snowflake devitrification texture has been recognized. In the same thin section a quartz-mica schist clast with a well developed crenulation cleavage is present. This grain contains equant grains of chlorite which are probably pseudomorphs of garnet. Sand grains, where their shape has not been affected by partial dissolution during cleavage development, are generally subangular and sorting is poor to moderate (Fig. 3.15).

The thinly bedded sandstone facies consists of beds of medium- to very fine grained sandstone (Fig. 3.16), 1 to 20 cm thick, which show parallel lamination, ripple cross lamination and water escape structures in which the upward-pointing cusp of convolution lamination is commonly truncated by the next bed (Fig. 3.17). These lithologies are organised into Bouma BC, BCD and CD sequences. Ball and pillow structures are seen occasionally at the base of the sandstone beds (Fig. 3.18). Cross bedding has only been noted at one locality (7728E, 7487N), where it occurs throughout the entire thickness of a bed 10 cm thick.

Across layer thickness variation of sandstone beds is poorly developed. Figure 3.19 shows the thickness of 77 sandstone beds, which were part of a 102-bed section through a minor example of the thickly bedded sandstone facies (beds 38 to 58). The 5-bed moving average curve shows a thinning-upwards trend followed by three symmetrical trends, which overall show a thickening-upwards trend. The 20-bed moving average curve shows a
Fig. 3.16  Photomicrograph of typical fine-grained sandstone from Stitt Quartzite (76-284). Detrital grains are predominantly monocrystalline quartz and mica (centre and top right). Original fabric of rock has been substantially modified by recrystallization and grain boundary migration of quartz and formation of wispy micas parallel to cleavage (bottom left to top right). Crossed nicols, field of view is 0.65 mm wide.

Fig. 3.17  Typical Bouma B-C turbidite from thinly-bedded facies of Stitt Quartzite. Zone of parallel lamination (Bouma B) is succeeded upwards by zone of cross-lamination (Bouma C) modified by water escape structures truncated at the top of the bed. Laminated siltstone with thin sandy laminae are also shown.
Fig. 3.18  Load casts at base of thin, laminated sandstone bed, thinly bedded facies of Stitt Quartzite.
Fig. 3.19 Sandstone bed-thickness diagram for Stitt Quartzite in Pieman Gorge at 7718E, 7824N. 5- and 20-bed moving averages shown on right.
uniform trend followed by a symmetrical thickening- to thinning-upwards sequence. Discussion of this feature is in Section 3.2.

3.2 (iii) **Westcott Argillite**

Sections of complete outcrop in the Pieman Gorge and the Flume Road (around 77°13'E, 75°31'N) demonstrate a gradational change between the Stitt Quartzite and the overlying Westcott Argillite. This transition is marked by a decrease in the thickness of fine-grained sandstone to coarse-grained siltstone beds from 1 - 20 cm to generally less than 2 cm, a change in the colour of pelite interbeds from dark grey to pale green-blue, and decreasing frequency of cross lamination in sandy horizons. Thin sections (eg. 76-300) demonstrate that these changes are paralleled by a marked increase in the carbonate/quartz ratio of the sandstone horizons. The Westcott Argillite is at least 200 m thick. In no single section of the formation are both the top and base of the formation exposed.

The sandstone beds and laminae form a minor component of the unit and are intercalated with laminae and beds of massive mudstone up to 10 cm thick. Thin laminae of fine-grained sandstone to coarse-grained siltstone are also present (76-290, 83-3). The sandstone beds are commonly poorly sorted and graded with flame structures at the base, and locally contain rip-up siltstone clasts (e.g. 76-300, Fig. 3.20). Detrital grains consist principally of carbonate, quartz, mica, chert and in one specimen (76-290A) albite is present.

The formation in the Pieman Gorge is interrupted by two horizons of pebbly mudstone up to 4 m thick. The conglomerates (83-4, 83-5, 76-301, 76-302) contain clasts up to 10 cm in diameter of chert, dolomite, quartz, quartzite and quartzwacke similar to those in the Stitt Quartzite, and rarer grains of mafic volcanics, embayed β-quartz and quartz-chlorite rock of probable altered volcanic origin (Fig. 3.21). Of particular
Fig. 3.20 Photomicrograph of typical laminated siltstone in Westcott Argillite (76-300). Facing is left to right as indicated by flames at base of bed of coarse-grained siltstone at right. Plane polarized light, width of field is 8.5 mm.

Fig. 3.21 Photomicrograph of slide conglomerate in Westcott Argillite (76-301) consisting of clasts of well-rounded quartz-wacke (at right) and subangular chert in quartzwacke matrix. Width of field is 8.5 mm, plane polarized light.
significance is a grain in one section (83-5) of Fe-rich chlorite (strongly pleochroic and relatively high birefringence) containing fine grains of pyrite and reddish-brown sphalerite (Fig. 3.22). Presence of spherulitic chert is common, and intraclast carbonate fragments indicate an intertidal to shallow marine provenance for some of the detritus in the conglomerates (C.P. Rao, pers. comm.). Clast morphologies vary from very well rounded to subangular and high to low sphericity. The matrix of the conglomerates is muddy. Contorted, probably locally derived, wispy fragments of sandstone and mudstone up to 20 cm in diameter abound. The conglomerate bodies display discordant contacts to the surrounding laminated rocks (Fig. 3.23), and these features suggest emplacement as cohesive, viscous debris flows (e.g. Middleton and Hampton, 1976; Lowe, 1982). Near the upper pebbly mudstone horizon are a number of graded very coarse grained sandstone to siltstone beds up to 30 cm thick with discoid shale clasts up to 15 cm long and, in one case, a rounded pebble of chert towards the middle of the bed (Fig. 3.24). The sandstone horizons contain a similar overall clast assemblage to the pebbly mudstone horizons, including a 8-quartz clast in a thin section of Brathwaite (1970, specimen 35692) collected from a bed some 2.5 km to the south of the Pieman River gorge section.

3.2 (iv) **Central Section**

(a) Mudstone-dominated sequence

In the Pieman River gorge the central section of the Rosebery Group consist of disrupted east-facing sequences of siltstone, mudstone, dolomite, slate, lithicwacke and the fuchsite-bearing polymict Salisbury Conglomerate. Interbedded impure stylolitic dolomicrite (76-307; 76-349) and slate are succeeded by mauve-coloured graded dolomitic sandstone and mudstone with some beds (<30 cm thick) of open framework conglomerate horizons and, at the top of this section, the 6 m thick Salisbury Conglomerate. The top of the conglomerate is truncated by a fault.
Fig. 3.22 Photomicrograph of portion of sphalerite-chlorite-pyrite clast in conglomerate of Westcott Argillite (83-5) showing euhedral sphalerite grains (dark grey) in chlorite with portion of chert clast to left and siltstone matrix to right. Plane polarized light, field of view is 0.65 mm wide.

Fig. 3.24 Graded sandstone bed in Westcott Argillite with outsize rounded chert clast (to right of lens cap) and rip-up mudstone clasts.
3.31

The lowest conglomerate horizon contains well rounded clasts of tholeiitic dolerite. One clast (76-324) contains unaltered euhedral of pigeonite with scattered exsolution blebs of orthopyroxene in some zones of the crystals, orthopyroxene euhedra, augite crystals commonly with overgrowths of orthopyroxene, and large zoned plagioclase crystals in a slightly finer grained groundmass of augite, plagioclase and mesostasis (Fig. 3.25).

Electron microprobe analysis of minerals from this rock by D.C. Green shows the composition of pigeonite to be wo\textsuperscript{10} en\textsuperscript{57} fs\textsuperscript{33}; augite shows a range of composition from wo\textsuperscript{37} en\textsuperscript{43} fs\textsuperscript{20} to wo\textsuperscript{36} en\textsuperscript{52} fs\textsuperscript{12}, and hypersthene has a composition from wo\textsuperscript{4} en\textsuperscript{64} Fs\textsuperscript{32} to wo\textsuperscript{4} en\textsuperscript{67} Fs\textsuperscript{29}. The plagioclase is labradorite (ab\textsuperscript{41} an\textsuperscript{59} to ab\textsuperscript{43} an\textsuperscript{57}). The analyses confirm the tholeiitic nature of the rock.

The sandstone horizons are up to 30 cm thick, are graded and commonly contain rip-up clasts of mudstone. In thin section (76-306; 76-322; 76-348) the chief clast types are carbonate, quartz, quartzite and rutilated quartz. In addition 76-306 contains grains of \delta-quartz, albite, and felsic volcanic rocks including carbonate-sericite-chlorite altered rocks. In terms of composition and texture the mudstone-lithicwacke suite is similar to the Westcott Argillite, and was included in the Westcott Argillite by Brathwaite (1970). The rocks are considered to be correlates of this formation, but probably represent a higher stratigraphic level than the Westcott Argillite of the eastern belt of the Rosebery Group.

(b) Salisbury Conglomerate

In the Pieman River gorge section the Salisbury Conglomerate is a closed framework conglomerate with clasts up to 30 cm in length well aligned parallel to the cleavage (Fig. 3.26). In thin section (83-1) clasts are of rounded to subangular chert, quartzwacke, carbonate, siltstone, shale, phyllite, fuchsite and vein quartz in a carbonate bearing lithicwacke
Fig. 3.25 Photomicrograph of dolerite clast in conglomerate of Westcott Argillite correlate of central units of Rosebery Group. Grain of unaltered pigeonite with augite rim (at extinction) at left centre. Crossed nicols, width of field is 4 mm.

Fig. 3.26 Salisbury Conglomerate in Pieman River Gorge, showing clasts of fuchsite strongly aligned parallel to cleavage and subangular, equant quartzite clasts (e.g. to top left of lens cap).
matrix. The western contact of the Salisbury Conglomerate is a fault at this location where a thickness of some 6 m is exposed.

The Salisbury Conglomerate is well exposed in the Natone Creek area and a complete section through the formation is provided by the Tasmanian Department of Mines drill hole, Rosebery 1, collared at 76415E, 71082N. Numerous indications of easterly facing have been obtained from this drill hole, which is the reverse of that for this part of the sequence inferred by Loftus-Hills et al. (1967), Brathwaite (1970) and Corbett (1981). Facing indicators include the presence of a clast of the Salisbury Conglomerate in the basal part of the Natone Volcanics (also seen in outcrop at 7652E, 7235N), graded bedding in coarse to medium sandy horizons in the Salisbury Conglomerate, graded bedding in a 1.5 m thick felsic tuff (or volcaniclastic sandstone) horizon within the Salisbury Conglomerate and by graded bedding in the lithicwacke-mudstone-pebbly mudstone sequence below the Salisbury Conglomerate. The latter sequence is similar to the units underlying the Salisbury Conglomerate in the Pieman gorge section (77-132, -133, -134, -135, -136, -137, -138, -319, -140, -141, -142, -143), but as well as the clast assemblage noted for these rocks quartz-mica schist is present in one specimen (77-137). In common with the eastern sequence in the Pieman Gorge, dolerite clasts have been noted in one specimen (77-132), but these are more strongly altered.

The Salisbury Conglomerate in the drill hole is about 90 m thick. Ten metres below the base of the formation there is another polymict fuchsitic conglomerate horizon 10 m thick which might be included within the Salisbury Conglomerate. In the 90 m thickness there are some 32 beds of conglomerate up to 11 m thick with some beds showing symmetrical grading. Interbeds of coarse- to medium-grained sandstone 4 to 88 cm thick commonly show normal grading. Intercalated mudstone horizons are rare. Towards the base of the formation three beds of strongly cleaved volcaniclastic
sandstone or quartz-phyric lithic crystal tuff with an aggregate thickness of 4.4 m occur. The thickest of these shows normal grading. Compared with the Pieman River Section to the north the conglomerate is finer grained and maximum clast size does not generally exceed 5 cm and the framework is open, the matrix being lithicwacke. No significant trend of bedding thickness variation of conglomerate beds is present. In thin section, clast composition of the conglomerate is similar to that of the Pieman gorge locality (77-156, -164, -165) with the addition of felsic volcanic clasts and spherulitic chert in 77-165.

(c) Natone Volcanics

This formation consists of strongly cleaved felsic vitric-crystal-lithic tuff approximately 120 m thick. Zones of quartz-sericite-, quartz-chlorite-, and carbonate-rich-schist are present in the unit (77-145, -153, and -159). In general, (specimens 77-144 to -155 inclusive; 77-167 to -169) phenocrysts of quartz and sericitized plagioclase up to 3.5 mm long are ubiquitous, clasts of black shale up to several cm long are common and rare granophyre, quartzwacke and chert clasts are present. Two cleavages are apparent in most thin sections and disseminated pyrite, in some cases with pressure shadows of quartz, occurs sporadically. Relict cuspate shard outlines are present in one specimen (77-157) and demonstrate the originally vitroclastic nature of the groundmass, but in general sericitization and strong cleavage development have obliterated original textures.

By virtue of the phenocryst mineralogy and the widespread shale fragments, there appears to be a striking similarity between the massive pyroclastics at Rosebery and the Natone Volcanics. The fragments of sedimentary rocks suggest emplacement by a mass flow mechanism and perhaps the Natone Volcanics might be classed as being of epiclastic rather than pyroclastic origin. However, the poor textural preservation of these rocks prevents a definitive judgement on this matter.
The west-facing Stitt Quartzite flanks the Natone Volcanics to the east (at 7667E, 7235N chert-bearing sandstone which may be the lowest unit of the Westcott Argillite abuts the volcanics). A fault contact is therefore inferred to exist between the Natone Volcanics and the sedimentary units to the east. This may be a southward continuation of the fault which truncates the eastern boundary of the Salisbury Conglomerate in the Pieman Gorge section and which may be responsible for the absence of both the conglomerate and the Natone Volcanics in the Pieman River near 75 000mN (Fig. 3.1), although alluvial cover at the latter location might also be postulated.

3.2 (iii) Munro Creek Formation

The Munro Creek Formation in the Pieman River gorge is bounded by faults at both its eastern and western margins.

It consists of at least 250 m of sandstone, dark grey slate, a conglomerate horizon and minor fine-grained dolomite beds. Sandstone beds vary from massive to graded with laminated tops, commonly with large detrital micas parallel to the laminae (AB Bouma sequence) to thinner beds with basal plane lamination passing upward to a thin zone of cross lamination (BC Bouma sequence). The sandstone horizons are interbedded with dark grey, in places pyritic, slate. A 5 m thick bed of inversely to normally graded, closed framework conglomerate at 37604E, 37719N, has a channelled base. Clasts, of cobble to boulder size, consist of quartzite, quartz schist and vein quartz. The most common lithologies of the formation closely resemble the Stitt Quartzite (76-325; -338; -330; -342).

3.3 CRIMSON CREEK FORMATION

In the Pieman River gorge the Munro Creek Formation is abutted on its western side by a monotonous green to maroon mudstone sequence
Occasionally interrupted by beds of graded lithicwackes, commonly with outsize rip-up clasts of mudstone, or by finer grained sandstone beds displaying climbing ripple cross lamination. The sandstone units have an open framework (Fig. 3.27) and consist almost entirely of angular weakly sericitized plagioclase, mafic lava fragments, clinopyroxene grains and miscellaneous chloritic fragments in a chlorite-rich matrix (76-326; -327; -328; -333; -334). In one of these sections (76-334) is a minor component of felsic volcanic detritus, quartz, carbonate and chert grains.

Recent mapping by A.V. Brown (pers. comm.) has indicated that beds south of 7450N, not mapped by the writer but included with the Crimson Creek Formation, should more properly be equated with the Dundas and Huskisson Groups. However, on the basis of lithology, the rocks north of this parallel discussed above are considered correlates of the Crimson Creek Formation. This formation marks the western boundary of the Rosebery Group.

3.4 STRUCTURE

The rocks of the Rosebery area have undergone at least one period of major deformation and cleavage development (Cottle, 1958; Brathwaite, 1972). This has traditionally been associated with the major period of Devonian deformation correlated with the Tabberabberan Orogeny of Victoria (Carey, 1953; Solomon, 1962; Brathwaite, 1972; Williams et al., 1976; Williams, 1978). This concept is difficult to prove near Rosebery because of the lack of post-Cambrian rocks, but is supported by K-Ar ages of cleaved black slate and host rock siltstone (five dates ranging from 409±6 to 428±6 Ma) and Rb-Sr ages of muscovite and whole rock samples from the footwall schist of 394±50 Ma (Adams et al., in preparation; Black & Adams, 1980), which suggest a high degree of resetting of the original Cambrian ages during the Middle Devonian.
Fig. 3.27 Photomicrograph of sandstone of Crimson Creek Formation (76-329) showing grains of mafic volcanic rocks, dolerite and plagioclase in a muddy matrix. Plane polarized light, width of field is 1.63 mm.
Evidence of post-cleavage deformation in the Mount Read Volcanics west of Rosebery is provided by the abrupt change in cleavage attitude from moderate easterly dips east of 7832E, 7416N (Fig. 3.1) to steep dips to the west of this point. Because of repetition of correlates of the host rock, black slate and massive pyroclastics to the west, this point is considered to mark a faulted anticlinal closure within the footwall pyroclastics. This interpretation is similar to that of Loftus-Hills et al. (1967), but differs from that of Brathwaite (1974) who placed the anticlinal closure within the slate lens in the Primrose-Barkers Crossing area (around 7820E, 7400N). This faulted anticlinal axis has been traced from a point 1 km north of Rosebery, where it is truncated by a second major fault some 3.7 km to the south. Further south the location of the fault is difficult to establish with certainty because of fairly uniform steep dips to the cleavage.

The second major structure, traceable from north of the Pieman River to Williamsford, is demonstrably a fault because it truncates lithologic units both in the Primrose Pyroclastics and the Rosebery Groups, but it was regarded as an interfingerling sedimentary contact by Brathwaite (1974). West of Rosebery there is abundant evidence for post-cleavage silification and alteration of the rocks adjacent to the fault in the form of unstrained quartz veins, galena veins and carbonate-tourmaline-quartz-fluorite veins hosting Pb-Zn-Bi-Sn-Cu mineralisation at the Black P.A., Chamberlain and Salisbury mines (Montgomery, 1893; Hills, 1915; Green & Williams, 1975).

At the Pieman River north of Rosebery (7744E, 7844N) the contact between the Rosebery Group and the Mount Read Volcanics is a north-striking fault dipping at 40° to the east (Fig. 3.28). Post-cleavage hydrothermal alteration, apart from minor secondary carbonate, is absent and two phases of deformation have affected the rocks, particularly the shaley facies of the Stitt Quartzite in the footwall of the fault.
Fig. 3.28  Thrust contact exposed between Primrose Pyroclastics, to right, and slate of Stitt Quartzite at 7744E, 7844N.

Fig. 3.29  Isolated lenticles of sandstone in crenulated black slate 50 m west of contact shown in Fig. 3.28.
For some 80 m west of the contact the thinly bedded sandstone-shale sequence is deformed into a pervasively sheared tectonic breccia, in which commonly refolded isolated lenticles of sandstone, up to 14 x 9 cm in plan are incorporated in a black slate matrix (Fig. 3.29). This tectonic breccia has a well developed crenulation cleavage (76-281; 76-316) which strikes N and dips west at about 85°. The lenticles appear to be bounded by shear planes post-dating the crenulation cleavage. The orientation of the crenulation cleavage is consistent with that of the dominant, and normally only mesoscopically observable, cleavage throughout the area, but to the west of the zone of entrainment a different sense of asymmetry of minor folds to that normally developed is observable. Away from the fault zone, folding is uncommon in the Stitt Quartzite and minor folds verge to the east (e.g. around 7733E, 7526N) (Fig. 3.30). In the area west of the zone of entrainment minor folds climb to the west in the overturned, easterly dipping, Stitt Quartzite and plunge at moderate angles to the south. The enveloping surface to these folds appears to dip at a moderate angle to the east, subparallel to the fault contact. In this area, downward facing folds are common. The fact that the later cleavage is steeper than bedding in this zone of overturned rocks may suggest that the Stitt Quartzite was locally overturned prior to westward thrusting of the Mount Read Volcanics over the Stitt Quartzite, or possibly that the development of cleavage was a late event.

In the Pieman River section through the Rosebery Group, cleavage has a steep dip of generally greater than 70° with a maximum near 345/90, but with considerable scatter in strike direction of up to 30° from this maximum (Fig. 3.21 b). Fold plunges are variable and define a crude planar girdle around the cleavage direction. Minor folds tend to occur in zones with more consistent fold plunges. For example, around 7677E, 7770N folds are tight with moderate to steep plunges to the north.
Fig. 3.30  E-W cross section through Rosebery Group in Pieman River Gorge projected to 7800N. Thickness of some conglomerate units is exaggerated for clarity and folds are somewhat generalized.
This distribution results in a poorly defined girdle on the π-diagram of bedding orientation (Fig. 3.31 a). Fig. 3.31 is a π-diagram of cleavage orientation in the Chamberlain Shale between 7745E, 7471N and 7807E, 7740N, and shows a maximum at 010/88W, distinctly different from that in the Pieman River. The tight folds and bedding-cleavage intersections in the Chamberlain Shale display gentle plunges (Fig. 3.31 d) and a resulting well-defined girdle appears on the π-diagram of bedding orientations (Fig. 3.31 c).

Features of the geology of the Rosebery Group such as disappearance and re-appearance of certain units along strike (e.g. Salisbury Conglomerate), facing changes at contacts between units (e.g. the contact of the Natone Volcanics and the Stitt Quartzite (Fig. 3.32), lack of correlation of rock types across fold hinges (Fig. 3.30), and development of strike-parallel zones of tectonic schist are characteristic of a terrain affected by tectonic slides (Hutton, 1979).

Hutton defines a tectonic slide as "...a fault which forms in metamorphic rocks prior to or during a metamorphic event. It occurs within a zone of coeval penetrative (i.e. microscopic) deformation that represents intensification of a more widespread, often regionally developed, deformation phase. Within this zone of high strain slides may lie along and be subparallel to (although they will cross-cut on a large scale) the boundaries of lithological, tectonic and tectonic-metamorphic units."

It should be noted that 'metamorphic' in the context of slides covers a wide range of pressure-temperature conditions, and Hutton discusses examples in terrains ranging from the prehnite-pumpellyite to granulite facies of metamorphism.

The only fold hinge of regional importance recognised in the Pieman River gorge section is at 7676E, 7740N, where the fold closure is an overturned syncline plunging at a moderate angle to the north (Fig. 3.23).
Fig. 3.31  Lower hemisphere equal-area projection of poles to bedding and cleavage from the Rosebery Group.
(a) 133 poles to bedding in Rosebery Group and Crimson Creek Formation in Pieman River north of 7450N. Contours 1, 5, 10%, per 1% area.
(b) 50 poles to cleavage from same area as in (a). Contours 1, 10, 20, 30%, per 1% area.
(c) 95 poles to bedding Chamberlain Slate between 7743E, 7477N and 7807E, 7439N. Contours 1, 5, 10, 20%, per 1% area.
(d) 31 poles to cleavage from same area as in (c). Contours 1, 10, 20%, per 1% area.
133 Poles to bedding
Rosebery Group & Crimson Creek Formation Pieman River N of 7450N

50 poles to cleavage
Rosebery Group, Pieman River
- Plunge of fold

95 Poles to bedding
Chamberlain Shale

31 Poles to cleavage
Chamberlain Shale
- Plunge of minor fold

N = Grid North

Fig. 3.24
Fig. 3.32  E-W cross section through Rosebery area at 7350mN. Fault contact between Stitt Quartzite and Natone Volcanics (inverted "V" symbol) inferred from Mines Department drill hole Rosebery 1 (Appendix 2). Other faults may exist in section.
To the east the Westcott Argillite and underlying Stitt Quartzite form a continuous stratigraphic sequence. To the west an additional unexposed slide may exist as a considerably diminished thickness of the Westcott Argillite is present before beds identical to the Stitt Quartzite are met (Figs. 3.1; 3.30). This zone may extend some 2.5 km to the south as an east-facing wedge of the Stitt Quartzite is exposed in Chasm Creek at 7696E, 7525N (76-352, -353, -354, -355) and at this northing the exposed thickness of the eastern section of the Westcott Argillite is much diminished, consistent with the northerly plunges of the folds in this area.

The next tectonic contact to the west is adjacent to the upper contact of the Salisbury Conglomerate. As the contact is approached, bedding in the sandstone mudstone sequence is disrupted and isolated lenses of sandstone occur in strongly cleaved mudstone (Fig. 3.33). The slide zone is marked by a 5 m width of tectonic breccia with rounded to elongate orange sandy fragments and irregular wisps of green mica (fuchsite?) in a grey-green slatey matrix (Fig. 3.34). The Salisbury Conglomerate contains well-aligned clasts and in particular the fuchsite clasts are strongly flattened (Fig. 3.26). A marked change in bedding orientation across the slide exists from 359/53E to the east to 335/63E to the west.

The contact between the Munro Creek Slate and the correlate (or upper part) of the Westcott Argillite exhibits similar characteristics with a change in cleavage direction at the sharp contact (Fig. 3.35) and pseudo-lenticular bedding in the surrounding rocks. Similarly a change in bedding orientation exists from 346/63E in the Westcott correlate to 010/84E in the Munro Creek Formation. The adjacent rocks are tightly folded with folds plunging at moderate angles to the north. Quartz veining occurs in the fault zone.

The contact between the Munro Creek Formation and the Crimson Creek Formation is marked by quartz veining and a change in bedding attitude from 353/83E in the former formation to 330/65E in the Crimson Creek Formation. The southward continuation of these slides in the Pieman River
Fig. 3.33 Disrupted bedding in very thinly bedded sandstone-mudstone sequence of Stitt Quartzite (?) adjacent to slide zone with the Salisbury Conglomerate in the Pieman Gorge.

Fig. 3.34 Tectonic breccia in slide in Pieman River Gorge at upper contact of Salisbury Conglomerate. Wispy green mica clasts (fuchsite ?) are present, for example, above lens cap.
Fig. 3.35  Features of slide contact between Westcott Argillite correlate (foreground) and Munro Creek Slate (background) in Pieman River Gorge. Strong cleavage development (compare for example Figs 3.18 and 3.33) is apparent and change in cleavage orientation across contact (running in an approximately horizontal direction through lens cap). Minor development of tectonic breccia is apparent to the left of the lens cap.
south of 7650N is not apparent in outcrop but a zone of moderately northerly plunging folds has been recognised in the Munro Creek Formation. The slides are probably masked by river gravels.

Recognition of the slides in the Rosebery Group provides an explanation for the diverse opinions regarding the stratigraphic succession, and possibly for the lateral impersistence of some units. The age of the deformation cannot be established with any degree of confidence, but the deformation style is different from that hitherto recognised to be of Devonian age (Section 2.5; see Williams, 1978). Three K-Ar dates of 475±8 to 487±7 Ma have been obtained from slates of the Chamberlain Shale (Adams et al., in prep.; Black & Adams, 1980), and these dates may represent partial resetting of ages established in a Late Cambrian deformation event, possibly correlated with the Delamerian Orogeny. Clearly, more dating, particularly of rocks in the slide zones, is required to settle this question. Devonian deformation may have been important in the Mount Read Volcanics near Rosebery and Devonian rejuvenation of the fault separating the Rosebery Group and the Mount Read Volcanics is probable. The unresolved difficulties probably stem from the fact that Cambrian palaeogeographic elements and fold trends parallel the earliest Devonian structures (Williams, 1978). Several thin sections examined bear testimony to this in the form of weakly developed crenulation cleavages. These must reflect nearly parallel axial surfaces of the superimposed deformational events because in the field no grossly cross-cutting cleavages have been observed, and the few crenulation cleavages recognised (Fig. 3.1) have a submeridional trend. This is in contrast with areas in which the second phase of Devonian deformation is well developed, such as the Linda Disturbance where NW to W trending cleavages fold axes and faults abound (Solomon, 1962; 1965; Baillie & Williams, 1975; Williams, 1978).
3.5 ENVIRONMENT OF DEPOSITION AND PROVENANCE OF THE ROSEBERY GROUP

A number of features indicate that the Rosebery Group was deposited in a subaqueous environment at depths exceeding storm water wave base (about 70 to 100 m; Walker, 1981). Sandstone beds more than a few cm thick lack cross-bedding in all but one bed, and thick beds (>30 cm) are either graded or massive. Similarly the open-framework slide conglomerates of the Westcott Argillite lack evidence of reworking. Although there is no macrofossil evidence, a marine environment is most probable. This argument may be extended to the Natone Volcanics, which, despite poor textural preservation, contain clasts of shale and other sedimentary rocks.

The sandstones of the Chamberlain Shale are composed of angular to subrounded volcanioclastic detritus, showing no evidence of reworking by currents or waves, and a mass flow origin was interpreted for these rocks. The high proportion of monocrystalline quartz in sandstones of the Stitt Quartzite may suggest mineralogical maturity (cf. Pettijohn et al., 1973), but the rocks are texturally immature and clast composition is also a function of available sediment in the source area and, particularly for quartz-rich material, its grain size. In this context it is important to note the higher proportion and variety of clasts in the conglomerate (76-283). Although no detailed palaeocurrent studies have been carried out, cross lamination in the Stitt Quartzite in the Flume Road area suggests a source to the east. This is reinforced by the rhyolite clast in 76-283, and by the common occurrence of clasts of quartzite, quartz-mica-schist and quartz-mica-garnet schist, which were almost certainly derived from the Tyennan Geanticline.

The thickly bedded facies of the Stitt Quartzite shows some similarity with the facies C turbidites of Walker (1979) and were possibly deposited in the mid-fan region of a submarine fan (cf. Nelson & Nilsen, 1974).
The thinly bedded facies with Bouma B-C and C division turbidites occur as parallel beds over some 50 m (limit of exposure) in the Flume Road area and may have been deposited in the lower fan region (cf. Walker, 1979).

No similarly convenient model exists for the Westcott Argillite. In the Walker (1981) model a mudstone-dominated sequence with very thin sandstone interbeds might be related to a lower fan or abyssal plain environment. However, such sequences are commonly characterised by beds dominated by Bouma CDE sequences (Walker, 1981; i.e. a basal division of cross-lamination) rather than the very thin graded beds (cf. Bouma A division) of fine grained sandstone observed. Further, the occurrence of the pebbly mudstone debris flow deposits and graded coarse grained sandstone beds would not be expected in such an environment. A problem with the Walker submarine fan model is that it has largely been constructed from present-day fans developed from deeply incised submarine canyons disgorging sediment into a deep, broad ocean basin (e.g. western United States).

Many of the features of the Stitt Quartzite and Westcott Argillite are also seen in the sequence of the Late Carboniferous Pesaguero Fan, northern Spain, described by Rupke (1977). This sequence, believed to have been deposited in a small, fault-bounded, intracratonic basin consists of a series of facies triplets. Each of these consists of, in ascending order, mudstone, sandstone and conglomerate, and represents progradation of sand and gravel lobes over the inactive, muddy part of the fan (Rupke, op.cit.). The basal mudstone unit, 250 m thick, has very thinly bedded, fine-grained sandstone units towards the top of the unit and a general increase in grain size upward from basal claystone. The bulk of the mudstone sequence appears to have been deposited from low density turbidity currents (Rupke, op.cit.). This facies has marked similarity to the bulk of the Westcott Argillite.
The mudstone is abruptly overlain by the sandstone unit, 180 m thick, consisting of some 800 sandstone beds of medium- to fine-grained sandstone ranging in thickness from 1 cm to 1.5 m, but only some 10% of the beds are thick to very thick (> 30 cm). A number of thickening and thinning upward cycles are present. Interbeds of laminated siltstone and mudstone are present. The sandstone beds display many similarities to the Stitt Quartzite, e.g. (a) paucity of sole marks and channelling, (b) massive beds with rip-up clasts and poor grading, (c) amalgamated beds, and (d) numerous beds which may be described by the Bouma sequence. An abrupt contact exists between the sandstone and the conglomerate unit, which comprises 60 thick beds in a section of 80 m of basal very-coarse grained wackes passing upward to pebbly sandstones and open-framework conglomerates with a sandy matrix.

Rupke (op.cit.) interpreted the facies triplet sequence in terms of fan progradation and avulsion, the mudstone sequence representing deposits from low density turbidity current overflow from the active part of the fan, but not necessarily further removed from the feeder channel than the depocentre of the contemporaneous sandstone lobe. A number of models was suggested for the rapid change in position of the sand lobe, including the increase in elevation of the older lobe providing a potential energy advantage for formation of a new lobe elsewhere. Rupke was able to demonstrate the existence of a single feeder channel during construction of the fan by a radiating pattern of conglomerate tongues from a single apex and outward palaeocurrent directions from that apex.

No evidence for a point source of sediment dispersal exists within the Stitt or Westcott Formations, and the lack of development of a thick conglomerate sequence suggests these formations may represent a more distal analogue of the Pesaguero Fan. The transition from the Stitt Quartzite to the Westcott Argillite might represent abandonment of a fan lobe, but the change in sediment composition with the influx of carbonate and chert
detritus in addition to the quartzose detritus of the Stitt Quartzite suggests a different feeder channel may have been involved. The conglomerate horizon in the Munro Creek Formation, a possible correlate of the Stitt Quartzite (see later) might represent the distal part of a channel. The inferred correlate of the Westcott Argillite in the central units of the Rosebery Group and the Salisbury Conglomerate represent a further diversification in the type of clastic detritus supplied to the trough, specifically in the form of fuchsite and tholeiite clasts. These probably represent detritus from the deformed ultramafic complexes and mafic units in the Crimson Creek Formation, which are now exposed only to the west of the Rosebery Group. This change in the nature of the detritus was probably due to uplift on the western side of the depositional basin of the Rosebery Group. The coarsening upward sequence from the mudstone-lithicwacke units of the Westcott correlate to the Salisbury Conglomerate might thus represent progradation of a separate fan from the west. Apparently at about this time the felsic volcaniclastic detritus of the Natone Volcanics, the Williamsford Volcanics and the thin felsic horizons in the Salisbury Conglomerate, was being deposited in the basin, but was presumably derived from a source to the east.

The occurrence of length-slow spherulitic chert in the Westcott Argillite and the Salisbury Conglomerate is of interest since Folk & Pittman (1971) have shown that this type of chalcedony (quartzine) develops as a replacement product of evaporite minerals. Several possible sources exist for the carbonate-chert detritus, including the Precambrian Jane Dolomite or a Cambrian carbonate unit to the east, or dolomite units at the top of the Oonah Formation or the Success Creek Group to the west. Collins (1972) described spherulitic chert from the Red Rock and an overlying dolomite unit at the top of the Success Group Group at Renison Bell.
Such a depositional model has a number of important corollaries. First, a relatively narrow depositional trough is required to explain the intermixing of detritus from both east and west, and second, active faulting is implied during accumulation of the Rosebery Group. The range of clast types in the Rosebery Group is only matched by the Dundas and Huskisson Groups, again supporting the correlation of the Rosebery and Dundas Groups.

3.6 REGIONAL EXTENT OF THE ROSEBERY GROUP AND CORRELATES

D.J. Jennings (in Collins et al., 1981) mapped a greywacke-mudstone-minor limestone sequence in the Coldstream River about 19 km north of the Pieman River Gorge, which he considered to be of Cambrian age. On the basis of deformational style, A.V. Brown (pers. comm.) now considers this to be an equivalent of the Oonah Formation. The correlative of the Oonah Formation is faulted against an east-facing sequence of conglomerate, lithicwacke and mudstone (Jennings, op. cit.).

Re-examination of the thin sections of Jennings shows that the clast assemblage in these conglomerates is identical to that in the conglomerates of the Westcott Argillite. Open framework conglomerate with clasts of microsparite occur in a silty mudstone matrix with grains of carbonate, quartz and mica in specimen 65-101, and closed framework conglomerate with a sparse quartzwacke matrix (65-79) contains rounded clasts of carbonate, chert and quartzite. Conglomerate and lithicwacke higher in the sequence are polymict with clasts of chert, including length-slow spherulitic chalcedony, carbonate, quartz, albite, quartz-mica schist, quartzite, quartzwacke, chlorite, dolerite, mafic lava, fuchsite, muscovite, biotite and embayed ß-quartz in a muddy matrix (65-76, -77, -82, -83). The transition upward from carbonate-chert-dominated to polymict detritus is similar to that inferred for the Westcott Argillite and supports the
assignment of the mudstone-lithicwacke of the central fault slice of the Rosebery Group to the Westcott Argillite. The thickness of the unit is about 2000 m.

The conglomerate-lithicwacke-mudstone sequence is apparently overlain to the east by the strongly folded greywacke, slitstone and mudstone sequence of the Middle Hatfield Valley (Fig. 1.1; Jennings, op.cit.). Examination of thin sections of this unit (65-67, -69, -89, -90) indicates lithological identity to the Stitt Quartzite. Jennings did not determine sedimentary facings in this unit, but his cross-section (Fig. 3.36) suggests structural discordance across the contact, similar to that seen across the tectonic slides in the Pieman River gorge. It appears highly possible, therefore, that the contact is faulted.

The next unit to the east consists of blue to black slate, sandstone and minor quartz-rich conglomerate (mudstone, quartzite, greywacke sequence of Collins et al. (1981). Specimens collected by D.J. Jennings and G. Urquhart have been re-examined. The 'greywackes' (65-134, -135, -137) consist predominantly of angular volcanic quartz and albite grains with a variety of rock fragments including quartz-feldspar porphyry, intermediate volcanics, quartzwacke, quartz-mica schist, quartzite and wispy mudstone clasts, and one grain of spherulitic chert in a silty chlorite-rich matrix. The conglomerate (65-102) consists of rounded clasts of quartzite, quartz-mica schist, quartzwacke, quartz-feldspar porphyry with a snowflake textured groundmass, and porphyritic vesicular andesite with plagioclase and chloritized ferromagnesian phenocrysts in a hyalopilitic groundmass, and embayed β-quartz. This rock is similar to the andesite at Que River described by Collins (1981b).

Within the greywacke-mudstone sequence west of Que River, Collins (1981b) has indicated an upward transition from a thickly bedded volcanic arenite sequence to a black slate-mudstone sequence similar to that seen in the Chamberlain Shale.
Fig. 3.36 Cross sections through the Rosebery Group correlates in the northern part of the Mackintosh map sheet (after D.J. Jennings in Collins et al., 1981). Sections A-B and C-D are orientated approximately NW-SE, and are located about 19.5 and 22.5 km north of the Rosebery Group-Primrose Pyroclastics contact in the Pieman River Gorge respectively. Approximate locations are:

A: 3759E 53987N
B: 3809E 53952N
C: 3766E 53012N
D: 3823E 53967N (Fig. 1.1)

Discordant structural patterns between greywacke-conglomerate, sandstone and mudstone sequence (Westcott Argillite correlate) and mudstone and greywacke sequence of the Hatfield River (Stitt Quartzite correlate) is apparent in section A-B. (Co-ordinates for points A, B, C, and D in Fig. 3.36 refer to the superseded Australian National Grid.)

Fig. 3.37 Cross section, orientated approximately E-W from Moores Pimple (Fig. 1.1) to Mount Read showing southern correlates of the Rosebery Group included in the western sequence of the Mount Read Volcanics (after Corbett, 1981).
Folding is diagrammatic but includes all possible data. Way-up established in greywacke-conglomerate only.

**CAMBRIAN**
- Greywacke-conglomerate, sandstone and mudstone sequence.
- Mudstone and greywacke sequence of the Hatfield River.
- Black shale and quartzite sequence with quartz conglomerate horizons.

**TERTIARY**
- Apsait/non-marine sediment.
- Greywacke, mudstone sequence of the Coldstream River
- Limestone conglomerate.

**MOORES PIMPLE - MT READ**

**WESTERN SEQUENCE & RELATED ROCKS**
- Mudstone-greywacke sequences.
- Quartz-phryic tuffs and other volcanic.
- Quartz sandstones and shale.
- Polymict conglomerate.
- Limestone-dolomite.

**CENTRAL MOLT VOLCANICS**
- Ashfall-phryic ash-flow tuffs.
- Polymict-phryic lavas and intrusives.
- Quartz-phryic volcanics.
- Ashend ash-flow tuff quartz-phryic sediments.
- Shale and fine silt.
- Basic intrusive.
The upward transition to a quartzwacke-quartz arenite-siltstone sequence is further evidence of similarity to the basal units of the Rosebery Group.

In the Moores Pimple area about 2 km south of the Hercules Mine, Corbett (1981) mapped an apparently conformable sequence (Fig. 3.37) of quartz-feldspar-phyric crystal tuff, shale and greywacke (Chamberlain Shale correlate?) followed by a thin quartzose sandstone unit (Stitt Quartzite correlate?), shale and greywacke (Westcott Argillite correlate?), carbonate-fuchsite-chert conglomerate with carbonate horizons (Salisbury Conglomerate correlate) and, at the top of the sequence, quartz-feldspar-phyric tuff and autobrecciated lava (Natone Volcanics correlate).

In summary, it is apparent that the Rosebery Group is not an isolated unit, but that correlates extend over a distance of some 35 km from Moores Pimple in the south to the Que River area in the north. From the evidence given in this section and sections 3.2 and 3.4, it is possible to reconstruct the most probable composite stratigraphic column for the Rosebery Group (Table 3.1). It is clear that the quartz-phyric volcanic rocks formed part of the source region for the Rosebery Group throughout its history.

3.7 RELATIONSHIP BETWEEN THE ROSEBERY GROUP AND THE MOUNT READ VOLCANICS

In the area north of the Pieman River, rocks considered here to be correlates of the Rosebery Group, have been regarded as the upper members of the western sequence of the Mount Read Volcanics by Collins (1981b), but were regarded as separate units by Corbett (1981), who included them with other 'undifferentiated' sedimentary sequences (Fig. 2.1). These rocks are underlain conformably by a mixed sequence of volcanic and sedimentary rocks and quartz-feldspar porphyry sills of the lower part of the western sequence of Collins (op. cit.). The basal unit of this sequence, the late Middle Cambrian Que River Beds, has been described in Section 2.3.2. There is regional discordance between the central and western sequences in the area south of Que River and a faulted boundary has been inferred.
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<td>Westcott Argillite</td>
<td>Salisbury Conglomerate</td>
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<td>Salisbury Conglomerate</td>
<td>Natone Volcanics</td>
<td>Stitt Quartzite =?</td>
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<td>Natone Volcanics</td>
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<td>Munro Creek Formation</td>
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<td>Stitt Quartzite</td>
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<td>(in part equivalent to the Rosebery Group)</td>
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(Collins, 1981b). However, work by Aberfoyle Limited geologists west of the Que River Mine suggests that the Que River Beds overlie andesite of the central sequence conformably (C.H. Young, pers. comm. in Green & Wallace, 1982).

Both the western sequence in the Que River area and the massive pyroclastics at Rosebery contain a preponderance of quartz-feldsparphyric volcaniclastic units. This suggests a possible correlation as does the occurrence of Precambrian-derived detritus in the black slate at Rosebery, in the massive pyroclastics south of Williamsford, in the Rosebery Group and in the western sequence of The Pinnacles-Que River area. In addition to the chlorite-pyrite-sphalerite clast in the pebbly mudstone of the Westcott Argillite, Corbett (1981) has described a 50 cm boulder and smaller clasts of massive pyrite in quartz-phyric agglomerate of the western sequence about 3.5 km SW of Mount Read. The association of mineralisation with the central sequence and the presence of sulphide detritus in the western sequences suggests that the latter is probably younger than the major ore horizons in the area north of Hercules, a suggestion reinforced by the andesite clast in the correlate of the Chamberlain Shale.

Corbett (1979, 1981) held the opposite view and regarded the central sequence as occupying a volcanic rift structure, developed within a marine basin in which the western sequence had largely accumulated. Corbett (1981) described two occurrences of possible collapse blocks of western sequence rocks occurring within the central sequence. At one of these localities, SW of Mount Read, a block of quartz-rich tuff approximately 100 x 50 m in size, occurs within massive glass-rich feldspar-phyric tuff of the central sequence (Corbett, 1981; pers. comm.) but it is not known if the host tuff belongs to the footwall or massive pyroclastics.

The other locality, the Bastyan Dam site on the Pieman River (7756E, 7833n), where outsize rafts of black shale and a large clast of massive
pyrite were included in a matrix of coarse grained quartz-feldspar bearing crystal lithic tuff, has been described in section 3 (iv).

Corbett's (1981) interpretation of this unit as a collapse breccia, presumably derived from the western sequence, developed near the wall of the volcanic rift, is dubious for a number of reasons:
1. The occurrence of the pyrite block favours derivation of the unit from the central sequence.
2. Collapse breccias would be expected to be developed earlier in the history of the volcanic rift, that is, prior to eruption of the footwall pyroclastics or the unknown thickness of volcanic rocks presumably under-lying the footwall pyroclastics.

The unit at the Bastyan dams site is regarded here as probably associated with cauldron subsidence, but the subsidence is believed to be associated with the ignimbrite eruptions which produced the footwall pyroclastics. This interpretation obviates the necessity for uplift of the deep marine western sequence required by Corbett's model to account for the fact that many of the rocks in the volcanic rift were apparently deposited in a subaerial environment. It is suggested that the western sequence, including the Rosebery Group and the massive pyroclastics, were deposited over a similar time interval. It is not known where the Mount Black Volcanics fit into this scenario and much further work is required to clarify the time relationships between the Mount Read Volcanics and the sedimentary units of the Dundas Trough.

A palaeogeographic sketch of the eastern part of the Dundas Trough showing the postulated relationships between the Mount Read Volcanics and the sedimentary sequences to the west in the late Middle to early Late Cambrian is shown in Fig. 3.38. Aspects of this reconstruction will be considered in more detail in the following section.
Fig. 3.38 Palaeogeographic sketch showing inferred relationships between the Mount Read Volcanics, Rosebery and Dundas Groups in early Late Cambrian time. Precambrian continental basement shown (double dash symbol), basement of Crimson Creek Formation and ultramafic-mafic complexes (wavy pattern) and Late Cambrian deposits (stipple). Diagram shows the diverse sources of detritus in the Rosebery and Dundas Groups, including Precambrian quartzite and carbonate, the Mount Read Volcanics (from the east) and intratrough chert, mudstone and ultramafic-mafic clasts (from the ridges). Abbreviations: Q: Que River, R: Rosebery, H: Hercules, L: Mt. Lyell.
3.8 TECTONIC HISTORY

3.8 (i) Previous Models

In the last 11 years, a number of attempts have been made to place the geology of the Dundas Trough in a plate tectonic framework. Solomon and Griffiths (1972) suggested that a west-dipping subduction zone existed between the Mount Read Volcanics and the Tyennan Geanticline, with the sedimentary and ultramafic rocks of the western Dundas Trough forming a back arc basin. Corbett et al. (1972) criticized this model on a number of grounds, principally the lack of possible trench deposits between the Mount Read Volcanics and the Tyennan Geanticline and the presence of feldspar porphyry dykes intrusive into Tyennan quartzite near the western margin of the geanticline. Corbett et al. suggested a number of possible alternatives, but could see no compelling reason to abandon the earlier rift model of Campana and King (1963), which was based on geosynclinal theory.

Solomon and Griffiths (1974) summarized and extended the evidence for the ensialic (Andean) nature of the Mount Read Volcanics. They considered that the volcanics once extended across the Tyennan region, a width of some 150 km, and that the Adamsfield Trough [the belt of ?Vendian to Cambrian sedimentary and ultramafic rocks west of Adamsfield (Fig. 2.1)] was underlain by the Mount Read Volcanics. Recent mapping in the Adamsfield area (Brown et al., 1983; Brown and Corbett in Cooper and Grindley, 1982) suggests that the Adamsfield Trough lacks acid volcanics and had a separate history from the Dundas Trough. Williams (1978) rejected the Solomon and Griffiths model, pointing out that fragments of the Oonah Formation existed on the eastern side of the Dundas Trough precluding its ever being a broad ocean basin, and that the complex
deformation expected at a subduction zone was lacking in the Dundas Trough. Most recent models have suggested that the ultramafic-mafic complexes of the Dundas Trough represent either a minor development of oceanic crust or that they formed in intracontinental cumulate magma chambers (Brown et al., 1980; Brown and Waldron, 1982; Brown in Cooper and Grindley, 1982). The only contrary opinion has been that of Crook (1980a,b) who viewed the sedimentary deposits of the Dundas Trough and the Burnie and Oonah Formations as part of a forearc complex to the Mount Read Volcanics deformed during ultimate collision between the Rocky Cape and Tyennan blocks.

3.8 (ii) Constraints on a Tectonic Model

In this section a number of important considerations for the formulation of a tectonic model will be discussed and evaluated.

1. The Dundas Trough is bounded by multiply deformed schist-quartzite sequences similar to those of the Tyennan region on its western margin at Cape Sorell, south of Macquarie Harbour (Fig. 2.1, Baillie et al., 1975; Williams, 1978) and correlates of the Oonah Formation occur towards the eastern margin of the trough at Mount Bischoff (Groves et al., 1972) and Dundas (Turner, 1979; Williams, 1978). This suggests that the Rocky Cape and Tyennan region were originally conjoined and that the Dundas Trough was initiated along a fracture slightly oblique to this boundary. It further suggests that collision between originally discrete terrains was not a feature of its development (Williams, 1978; Corbett in Cooper and Grindley, 1982; contra Solomon and Griffiths, 1974; Crook, 1980).

2. The presence of a landscape and structural unconformity between the Oonah Formation and Success Creek Groups (Brown, 1980) clearly shows that the Penguin Orogeny is unrelated to the closure of the Dundas

3. The accumulation of the substantial thickness (about 1000 m, Taylor, 1954) of the shallow marine conglomerate-sandstone-carbonate-siltstone sequence of the Success Creek Group (Patterson et al., 1981; Brown, 1980; Collins, 1972) is typical of sedimentation during an early stage of rift development (e.g. Hoffman et al., 1974). The Lake Beatrice sequence, consisting of Precambrian-derived siliceous conglomerate, quartz, sandstone and mudstone, unconformably overlies the rocks of the Tyennan region (location 390E, 5346 – 5383N, Fig. 212) and is faulted against the Mount Read Volcanics (Corbett, 1981). This sequence may be a correlate of the Success Creek Group (Solomon, 1965, Fig. 3) and may thus also represent sedimentation during this early phase of rifting. Although the rocks hosting the Cleveland deposit have been correlated with the Crimson Creek Formation (e.g. Groves et al., 1972; Collins, 1981a), there is a possibility that this sequence is a correlative of the Success Creek Group. The Cleveland mine sequence consists (Collins, 1981a) of basal, locally pillowed tholeiitic basalt, chemically similar to ocean floor basalt (Foden, 1973) overlain by the host sequence to the Cleveland mineralization which consists of shale with limestone, chert and basaltic flows and pyroclastic rocks. A succeeding unit, in excess of 350 m thick, consists of a quartzwacke turbidite-mudstone sequence with minor chert and spilitic horizons (Collins, op. cit.). It is tempting to suggest that the clastic units are a deeper marine equivalent of the Success Creek Group developed over oceanic crust.

4. Acritarch (algal microfossil) assemblages suggest a Vendian age
for the Dalcoath Quartzite, a thick quartz arenite unit in the Success Creek Group, and an Early Cambrian age for the base of the Crimson Creek Formation at Renison Bell (Vidal in Cooper and Grindley, 1982). If reliable, these palaeontological data place important limits on the time framework of the development of the Dundas Trough. Odin et al. (1983) suggest that the base of the Cambrian should be fixed at 530 ± 10 Ma and the Cambrian-Ordovician boundary at 495 ± 10 Ma.

5. Olivine tholeiite volcanism accompanied deposition of the Crimson Creek Formation (Brown and Waldron, 1982). The ultramafic-mafic complexes are tectonically emplaced into the Crimson Creek Formation. They include high-magnesian andesite lavas (HMAS) with a severely depleted content of incompatible elements, Ti, Zr and which have rare earth element patterns similar to the Eocene boninites of the Marianas arc (Brown and Waldron, 1982; Hickey and Frey, 1982). Probably cogenetic, cumulate ultramafic-mafic intrusive rocks differ from typical ophiolites (Penrose Conference Participants, 1972) in their high abundance of othopyroxene and lack of a sheeted dolerite dyke complex (e.g. Brown et al., 1980; Rubenach, 1973). This association is similar to that of rocks dredged from the Mariana forearc (Bloomer, 1981).

The analogy between the ultramafic-mafic complexes and the rocks of the Marianas arc is strengthened by the dredging of orthopyroxene-rich cumulate gabbros from the inner wall of the Mariana Trench (Bloomer, 1981).

6. The better preserved ultramafic-mafic complexes, the Serpentine Hill Complex (Rubenach, 1974), the Heazlewood River Complex (Rubenach, 1973) and the Hibbs ultramafic belt (Hall et al., 1969), although internally faulted, display a gross eastward facing with ultramafic
rocks succeeded to the east by gabbro and basalt in the first two cases. Amphibolite lenses are exposed at the faulted base of the Heazlewood River and Serpentine Hill complexes (Rubenach, 1973; 1974).

7. A major constraint on the nature of the Dundas Trough is provided by the Crimson Creek Formation. The contact between this unit and the Success Creek Group in the Renison Bell area marks a dramatic change in terms of both sediment provenance and depositional environment. The thickness of the formation of some 3000 m (Blisset, 1962), if not a result of thrust stacking, the presence of mafic volcaniclastic turbidites and the paucity of terrigenous detritus are features:

(a) not consistent with typical ocean basin deposits, which are characterized by relatively thin sequences dominated by pelagic sediments, and

(b) not easily reconciled with a narrow rift in which a major component of terrigenous detritus might be expected.

8. The Mount Read Volcanics are calcalkaline in nature and display a wide range in compositions from basalt through andesite and dacite to rhyolite and formed over sialic crust (Anderson, 1982; Solomon and Griffiths, 1974; White, 1975). Some of the Mount Read Volcanics, and the ore deposits of Que River and Mount Lyell had formed before the late Middle Cambrian, but there is no evidence to suggest that Mount Read Volcanism was coeval with deposition of the Crimson Creek Formation.

9. The ultramafic-mafic complexes were tectonically emplaced on the seafloor and were eroded into basal sediments of the Dundas Group, the earliest fossils of which are of middle Middle Cambrian age (Jago, 1979). The older units of the Dundas Group contain a predominance of detritus cannibalistically derived from older trough sequences including clasts of mudstone, chert and mafic and ultramafic rocks. The first felsic volcanic detritus appears in the stratigraphically
higher Brewery Junction Formation, the lower part of which is of late Middle Cambrian age (Jago, 1979). Quartzite detritus, of presumed Precambrian derivation, becomes abundant in stratigraphically higher units (Blissett, 1962; Brown in Cooper and Grindley, 1982).

The stratigraphic succession of the Rosebery Group and correlates inferred in this thesis suggests that a broadly reversed sequence of clast types were deposited closer to the Mount Read Volcanic belt. Felsic volcaniclastic sediments interfinger with the late Middle Cambrian Que River Beds (Gee et al., 1970) and were succeeded by units with Precambrian-derived quartzite and schist detritus, and by later units with clasts of fuchsite and mafic volcanics and intrusives, some of which are well rounded.

10. In many areas structural conformity exists between Cambrian and Ordovician rocks (Williams, 1978) and the parallelism of early Tabberabberan structures to Cambrian tectonic elements has prevented definitive evaluation of the extent of Cambrian deformation.

Disoriented cleaved felsic volcanic clasts within Late Cambrian Jukes Conglomerates near Mount Darwin (White, 1975; Corbett, 1981) show that, at least locally, Cambrian deformation was sufficiently intense to produce a cleavage. Also the pattern of deformation in the Rosebery Group, with closely spaced tectonic slides subparallel to strike is quite different from the broad open N-S folds typical of the early Tabberabberan structures. The K-Ar dates from slates within the Rosebery Group of 475 ± 8 Ma to 487 ± 7 Ma (3 samples; Black and Adams, 1980; Adams et al., in preparation) could represent cooling ages following Late Cambrian deformation; they certainly do not suggest Devonian cleavage development. The shallow east dip of the fault contact between the Rosebery Group and the Mount Read Volcanics suggests an eastward direction of tectonic transport, the opposite to that associated with the N-trending Tabberabberan folds (e.g. at Mount Lyell, Williams, 1978).
Deep marine sedimentation in the Dundas Trough ceased around middle to late Late Cambrian (Jago, 1979) when the accumulation of quartz conglomerate in fault-bounded troughs marked the start of molasse-type deposition (Campana and King, 1963). Subsequent sheet-like shallow marine quartz sandstone and limestone of Early to Late Ordovician spread over the Dundas Trough and much of the Tyennan region.

3.8 (iii) Discussion and Tectonic Model

The simple ensialic rift model for the Dundas Trough (Brown et al., 1980; Brown and Waldron, 1982) cannot account for the asymmetry of the Dundas Trough with the Mount Read Volcanics located on its eastern side. Such a configuration is atypical of a rift or aulacogene where the lithological components are commonly symmetrically disposed (e.g. Hoffman et al., 1974).

Comparison of the Mount Read Volcanic with the Mesozoic rocks of the Lemombo monocline in southern Africa (Brown et al., 1980), which formed in an intracontinental rift is not strictly valid. The rocks of the Lemombo monocline are bimodal with a distinct compositional gap in the range 56 to 61% SiO₂ (Cox, 1972) and include alkali basalts, phonolites and nephelinites. Both basalts and rhyolites of the Lemombo monocline are richer in Nb than those of the Mount Read Volcanics by about one order of magnitude (comparing analyses in Cox et al., 1965, 1967 with those in White, 1975 and this thesis).

Also, it is difficult to envisage the Crimson Creek Formation forming in an ensialic trough some 40 km wide (allowing some 100% extension from the presently exposed width, Williams, 1980; constraint 7).

It is therefore necessary to reconsider a model involving subduction within the Dundas Trough. This will involve application of the constraints listed in the previous section, and analogy with the structure of modern arc-trench complexes and some orogenic belts.
Objections to the existence of a broad Dundas Trough on the grounds of deformational style (Williams, 1978) are not borne out by recent studies of active forearc regions and some orogenic belts. Forearc regions consist of two elements:

(1) An accretionary wedge comprising sediments scraped off the down-going oceanic plate in discrete thrust packets has now been demonstrated by deep sea drilling (e.g. Karig and Sharman, 1975; Moore et al., 1982a,b; von Huene et al., 1982).

(2) The accretionary wedge may be overlain by slope deposits derived from the continent or island arc. These may be difficult to distinguish from the underlying wedge deposits after deformation, particularly where the forearc region has a relatively smooth profile allowing terrigenous detritus to accumulate directly in the trench (e.g. Moore et al., 1982a).

The width and morphology of the forearc region can vary widely. The accretionary wedge off Acapulco, Mexico, is some 20 km wide, has a smooth profile and has accumulated in some 10 Ma (Moore et al., 1982a). The forearc complex east of northern Honshu has been forming since the Cretaceous and is some 240 km wide with a well developed, sediment-filled, forearc basin some 10 km thick (von Huene et al., 1982). The subduction complex off the North Island of New Zealand, where the Pacific plate is descending obliquely under the landmass is even more complex (Fig. 3.39). The accretionary prism consists of a number of well-defined ridges with a morphology controlled by active reverse faults (Cole and Lewis, 1981). The highest accretionary ridge and the forearc basin are on land. The accretionary ridges are separated by basins from 5 to 30 km wide and 10 to 60 km long. The difference between the rate of uplift of the ridges and subsidence of the basin axes is some 3 km per million years. Separating the accretionary complex from the Taupo volcanic zone is a region of active uplift and
Fig. 3.39 Diagrammatic model of the major structural elements of the Taupo-Hikurangi oblique-subduction margin. Stippling represents over-riding plate, denser stippling indicates pre-Cenozoic "basement" of over-riding plate. (From Cole and Lewis, 1981)
strike slip faulting. The Taupo volcanic zone is a region of active extension, and rhyolite-andesite volcanism has been active for about the last million years (Cole and Lewis, op. cit.).

Active deformation in accretionary complexes does not necessarily appear to be marked by pervasive cleavage development and folding. For example, drilling off Barbados has demonstrated fault repetition of sequences and local overturning of strata (Moore et al., 1982b). An intense scaley foliation is developed in mudstone within 20 cm above the fault and slickensided surfaces occur at approximately one cm intervals for some 50 m above the fault. Further downhole, drilling was terminated in a zone 50 m thick with intense, shallowly dipping scaley foliation, believed to occur above a décollement, which might represent a surface of sediments scraped off the down-going plate and incorporated into the accretionary complex (ibid.). A drill hole collared near the crest of an accretionary ridge close to the Nankai trench, southeast of Shikoku, established 525 m of an upward-facing, coarsening upward, mudstone-turbidite sequence structurally overlying 86 m of inverted beds (Moore and Karig, 1976). The overturned anticlinal fold closure dips 9-14° landward and the bottom of the hole is believed to shallowly overlie a reverse fault. Within 200 m of the fold hinge an axial surface spaced fracture cleavage is present (ibid.).

Likewise, the intensity of deformation in ancient accretionary and collisional terrains is quite variable. For example, Leggett et al. (1979) describe 10 or more distinct deep-marine stratigraphic sequences occurring over an outcrop width of 60 km in the Southern Uplands of Scotland. The sequences, which are separated by strike faults, are believed to represent the accretionary prism on the margin of the Iapetus Ocean which formed an important faunal barrier in the Ordovician. Deformation between the fault zones is relatively minor and
the rocks face predominantly to the NW, that is away from the ocean and towards an Ordovician-Silurian volcanic arc. This mode of deformation contrasts sharply with that of Franciscan mélange (e.g. Hsu, 1974), which is generally believed to be typical of subduction terrains.

This brief review suggests that a lack of intense regional deformation is not, in itself, a sufficient condition to negate the possible existence of a former subduction zone. Rather, faulting parallel to strike with localized zones of deformation near the faults appears to be characteristic. This is typical of the style of deformation in the Rosebery Group and further study might show it is common in the sedimentary sequences of the Dundas Trough.

The intensity of deformation in forearc regions appears to be independent of such factors as the rate of convergence between plates. Rather, ductile deformation, characterized by common moderate to steep, and locally inconsistent, dips, stratal disruption, faulting and tectonically kneaded slope deposits occurs in areas with copious sediment supply and prisms of actively accreting sediment. Brittle deformation, typified by low dips, locally intense fracturing and localized discrete zones of deformation occurs in regions with a rigid basement such as igneous rocks or rigidified older accreted material. In such areas the slope stratigraphy may be well preserved (Lundberg and Moore, 1982). Intraoceanic forearc regions, characterized by sediment-starved forearcs and trenches, commonly display brittle deformation (Lundberg, 1981).

The asymmetry of the Dundas Trough suggests that the Mount Read Volcanics were related to an east-to-southeast-dipping subduction zone located within the Dundas Trough. An alternative suggestion that the trench dipped to the west and was located in the Adamsfield Trough to the east of the Tyennan region (Corbett et al., 1972) is not favoured because it does not readily account for the restriction of the
Mount Read Volcanics to the western and northern margin of the Tyennan region and the lack of felsic intrusive rocks within all but the extreme western margin of the region.

Among ophiolitic complexes, the ultramafic-mafic bodies of the Dundas Trough show strongest similarity with the Betts Cove Complex, Newfoundland, where ultramafic cumulates, massive gabbro and a dolerite dyke complex are overlain by basaltic to andesitic pillow lavas (Coish and Church, 1979). The lowest pillow lava unit is very similar in terms of major, trace and rare earth element compositions to the high magnesian andesites associated with the Heazlewood River Complex (Coish et al., 1982). The two succeeding lava units are progressively less depleted (higher TiO₂, Zr, Y, P₂O₅) and the upper unit is similar in composition to oceanic tholeiite (Coish and Church, op. cit.). An overlying dominantly sedimentary sequence of mafic volcanic rocks, argillite and chert is up to 4-12 m thick (Williams, 1979). Kean and Strong (1975, fig. 10) believed the Betts Cove Complex occupied a forearc region in the Early Ordovician, a similar regime to that of the boninites of the Bonin Islands with respect to the present day Marianas arc. Similarly, Upadhyay and Neale (1979) considered that the petrological diversity of many ophiolites, including the Betts Cove Complex, and the presence of a thick overlying sedimentary sequence were consistent with formation in a marginal basin environment near an island arc. Williams (1979) included the Betts Cove Complex in the Dunnage Zone, a longitudinal belt believed to represent preserved vestiges of the Iapetus Ocean.

Uncertainty exists regarding the tectonic significance of the type boninites. For example, Cameron et al., 1979; 1980) pointed out their occurrence in a forearc environment. Crawford et al., (1981) suggested that boninites were formed after early arc tholeiite volcanism during the initial stages of island arc rifting, but Hawkins et al., (1982) believed them to represent the initial stages of arc volcanism
in an intra-oceanic setting.

Clearly, a marginal basin environment characterized by formation and splitting of ensimatic island arcs, provides an attractive framework to account for the diverse rock types of the western Dundas Trough. Such a setting is compatible with the thick mafic volcaniclastic-bearing Crimson Creek Formation. The andesite-dominated Noddy Creek Volcanics which occur within sediments of the Dundas Trough south of Macquarie Harbour (White, 1975) are also compatible with such a terrain. However, it is not yet possible to assign a specific environment to all of the intra-trough volcanic sequences with any degree of confidence.

The Dundas Trough was initiated by rifting of Precambrian basement some time after the cessation of Penguin deformation (ca 670 Ma) with deposition of shallow marine siliciclastic sequences (the Success Creek Group and the Lake Beatrice sequence) probably accompanying the early stages of rifting and subsidence (Fig. 3.40).

Further rifting is believed to have led to the formation of oceanic crust in the Dundas Trough. At some stage subduction commenced and it is suggested that this subduction was directed westwards with a trench located to the east of a proposed ensimatic arc. This is because the structural conformity between the Crimson Creek Formation and the underlying Success Creek Group in the Renison Bell area is not consistent with a trench intervening between the western margin of the trough and the source of mafic volcaniclastic detritus of the Crimson Creek Formation (Constraint 3, Fig. 3.40).

Little evidence of the forearc region need survive because of its ensimatic nature (e.g. Lundberg, 1981). This phase of subduction is presumed to have terminated by collision of the continental crust at the eastern side of the trough with the trench. This Precambrian crust is now preserved as inliers at Mount Bischoff (Fig. 2.1) and west of Mount Dundas (Fig. 2.2). The collision event is believed to
Late Proterozoic: Rift stage. Siliceous clasts, mudstone, carbonate and possible evaporite sequences are developed on continental crust (Success Creek Group, Lake Beatrice sequence?) and possibly extend onto oceanic crust (Cleveland area).

Early Cambrian: Intraoceanic arc phase. Westward subduction in the trough is responsible for ensimatic arc volcanism. Crimson Creek Formation is partly derived from eroded arc material.

Late Cambrian: Collision of continental crust with the subduction zone results in initial thrust emplacement of ultramafic-mafic complexes and reversal of subduction polarity. Mount Read Volcanics erupted and separated depocentres of the Rosebery and Dundas Groups are established in the forearc region (see also Fig. 3.38).
I  LATE PROTEROZOIC (POST 670 Ma)

Success Ck. Jane Group Dolomite

"Rocky Cape" crust

Oceanic crust

"Tyennan" crust

Jane Dolomite

II  EARLY CAMBRIAN

Crimson Ck. Fm.

Arc vulcanism

III  EARLY LATE CAMBRIAN

Accretionary wedge (Crimson Ck. Fm.)

Mount Read Volcanics

Rosebery Gp.

Dundas Gp.
have been responsible for the initial thrust emplacement of the ultramafic-mafic complexes. Buoyant island arc material would be likely to be preserved following such a collision (e.g. Ben-Avraham et al., 1982).

As is common following trench-continent collisions (e.g. Hoffman, 1981; Ben-Avraham et al., 1982) a reversal of subduction polarity is postulated to have occurred at this time with eastward underthrusting of oceanic crust initiated to the west of the old arc. During this stage, which may have commenced in the early Middle Cambrian (the age of the earliest Dundas Group fossils), or earlier, the Mount Read Volcanics were erupted. The rocks of the Crimson Creek Formation are believed to represent the accretionary prism. At the latitude of Rosebery at least two distinct forearc basins, which received sediments in part synchronously, were developed (Figs. 3.38, 3.40). The eastern basin was the depocentre for the Rosebery Group, which was formed partly over sialic crust. The western basin was the site of accumulation of the Dundas Group and was probably formed on the accretionary wedge which overlay oceanic crust. The increase in quartzitic detritus in Dundas Group sediments upwards in the stratigraphic sequence and the late appearance of fuchsitic detritus in the Rosebery Group (Constraint 8) is consistent with progressive closure between the Precambrian blocks on either side of the Dundas Trough, a compressional regime in the forearc region and continued, possibly episodic, uplift of the ultramafic-mafic complexes. Despite this localized uplift, it appears that subsidence and accumulation of sediments was continuous in the forearc basins, a situation paralleled in the Pleistocene to Recent history east of the North Island of New Zealand (Cole and Lewis, 1981). Other apparent similarities lie in the comparison between the Mount Read Volcanics and the rocks of the Taupo region. Both are rhyolite-dominated calcalkaline sequences and the evidence of cauldron subsidence at Rosebery agrees with the presence of a number of calderas in the Taupo
volcanic zone (e.g. Healy, 1962).

The analogy between the New Zealand and Tasmanian cases breaks down when the early history of the Dundas Trough is taken into account. The trough apparently developed by the creation and destruction of a small ocean basin, a situation paralleled by the Sistan Suture Zone, Iran where a similar cycle occurred over a time span of some 50 Ma in Cretaceous to Palaeogene times (Tirrul et al., 1983).

The maximum width the Dundas Trough achieved during its development is not easy to estimate. The active volcanoes of the Taupo Graben are some 80 to 100 km above the subduction zone. The dip of the seismic zone under most of the accretionary prism is about 12°, but it steepens to 50° some 40 to 50 km east of the volcanic region. Flattening of the seismic zone towards the trench is common, particularly where the accretionary prism is well developed (Karig and Sharman, 1975). A fairly narrow forearc basin probably existed west of the Mount Read arc (Fig. 3.38). Assuming an average seismic zone dip of 45° and a convergence rate of 5 cm/year some 3 Ma would have been necessary between the initiation of eastward subduction and the onset of Mount Read volcanism. If the volcanism lasted for say 5 Ma, then consumption of a width of some 400 km of oceanic crust is indicated. To this figure an additional amount must be added to account for the postulated ensimatic arc. There is no need to involve a maximum width of the Dundas Trough of more than 1000 km, a figure comparable to the width of the present Sea of Japan, to satisfy the proposed model. A total time span of 50 Ma, well within the constraints of the K-Ar and palaeontological data, would be adequate to account for the proposed evolution of the trough.

The parts of the model dealing with the early history of the Dundas Trough are clearly speculative, but the later history is better controlled, both bio- and litho-stratigraphically. The model can account for the nature and position of the major lithological units.
No attempt has been made to include the history of the other Vendian to Cambrian troughs in Tasmania. However, dating of the mafic-ultramafic complexes, detailed study of the structure and metamorphism of the Dundas Trough sequences, further K-Ar dating of the sedimentary rocks and in particular, deep crustal seismic reflection studies (as proposed by Lewis and Williams, 1983) would provide critical tests of the model and permit its extension to other areas.

Finally, it is probable that evaporites developed early in the history of the Dundas Trough. Palaeomagnetic studies elsewhere in Australia (Klootwijk, 1980) indicate that Tasmania lay within 30° of the equator in the Cambrian. In such an environment, formulation of evaporites both on the platform, and within the rift basin, is likely. Evidence of evaporite-derived detritus in the Rosebery Group is compatible with the conclusion of Kinsman (1975) that there is an 80% probability of development of brackish water in low latitude proto-oceans within the first 10 to 20 Ma of their history. The possible extent of the early rift deposits under the Mount Read Volcanic has important ramifications for ore genesis.