The geology, timing of mineralisation, and genesis of the Menninnie Dam Zn–Pb–Ag deposit, Eyre Peninsula, South Australia

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This thesis contains no material which has been accepted for the award of any other higher degree or graduate diploma in any tertiary institution and, to the best of the author's knowledge and belief, contains no material previously published or written by another person, except where due reference is made in the text of the thesis.

Michael W Roache

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Frontispiece: View looking south along the 10000 mE base line of the Menninnie Dam Pb - Zn - Ag prospect. The road to riches, or the road to nowhere?
Abstract

The Menninnie Dam Pb - Zn - Ag deposit is located on the northern Eyre Peninsula of South Australia, approximately 1-2 km south of the main mass of the Gawler Range Volcanics. The deposit has no surface expression and was located by drilling of aeromagnetic anomalies in deeply weathered, flat lying terrain. Mineralisation occurs over ~3 km strike and is hosted by upper amphibolite facies (= 700°C and 7 kbars) marbles and diopside-rich calc-silicate lithologies, correlated with the Katunga Dolomite and lower units of the Cook Gap Schist of the Middleback Subgroup, a subdivision of the Hutchison Group (1964 - 1845 Ma.). The Katunga Dolomite is bound to the west by a shear zone (> 50 m wide) formed during the latter stages of the Kimban Orogeny (1740 - 1710 Ma.). Syn-deformation granite dykes intruded into the shear zone, and, to a lesser extent, into the Katunga Dolomite and Cook Gap Schist. Late tectonic pegmatite dykes intruded into the Katunga Dolomite and lower units of the Cook Gap Schist, where they produced zones of K-feldspar - calc-silicate metasomatism.

Previous researchers have incorrectly suggested that the Menninnie Dam Pb - Zn - Ag deposit is a Broken Hill-type. However, Pb - Zn - Ag mineralisation post-dates metamorphism, deformation and metasomatism and has replaced the host marble, and to a lesser extent, diopside-rich calc-silicate lithologies. On the basis of lithologic relationships, most mineralisation formed within 100 - 600 m of the palaeosurface, and resulted in a central zone of stockwork and matrix to marble and calc-silicate breccias, flanked by veins. Sulphide and gangue minerals consist of a simple assemblage of pyrite, quartz, sphalerite, galena and Ca-Mn-Mg-Fe carbonate, with accessory chalcopyrite, chlorite, adularia, sericite, fluorite, rhodonite, talc, phlogopite, dolomite, hematite and matildite. Three paragenetic stages of mineralisation are present with an early pyrite stage, followed by a sphalerite galena, and late pyrite stages. Metal abundances have a bell-shaped distribution along the length of the deposit and the highest values are associated with the zone of stockwork and breccia mineralisation.

Post metamorphic, porphyritic rhyolite intruded the Hutchison Group, and interacted with (heated?) groundwater resulting in explosive fragmentation of the host rocks and formation of polymictic breccia pipes. Some of these erupted onto the palaeosurface and formed layered polymictic breccias. Rhyolite continued to intrude through polymictic breccia pipes, and resulted in formation of peperite at the margins. Some rhyolite intrusions erupted onto the palaeosurface and formed volcanic breccias and rhyolite lavas. Polymictic breccias contain clasts of paragenetically early sulphide and gangue minerals, and have an altered matrix that includes paragenetically late sulphide minerals, indicative of syn-mineralisation emplacement. U-Pb zircon dating of the rhyolite intrusions constrains the timing of mineralisation to 1594 ± 7 Ma, which is indistinguishable from that determined for the Hiltaba Suite granitoids and co-magmatic Gawler Range Volcanics.
Modelling of regional gravity and aeromagnetic data indicates the Menninnie Dam deposit lies near the north-western margin of a 20 km diameter Hiltaba Suite granite that intruded to within 1 - 3 km of the palaeo-surface. Lead was derived from the underlying Hiltaba Suite granite and leached from Cook Gap Schist. Lead isotope ratios have a spatial distribution on a prospect scale, and the least radiogenic ratios correspond with the highest metal values and the central zone of stockwork and breccia style mineralisation, consistent with a single zone of fluid up-flow. Carbonate gangue is interpreted to have precipitated via interaction of the mineralising fluid ($\delta^{18}O = -2.0\%_o; \delta^{13}C = -6.9\%_o$) in equilibrium with $H_2CO_3(aq)$ (> 0.01 molal) and the host marbles ($\delta^{18}O = 15.5 \text{ to } 21.09\%_o; \delta^{13}C = -1.1 \text{ to } 1.6\%_o$) between 200° and 125°C. Hydrous phyllosilicates associated with mineralisation have calculated fluid values of $\delta^{18}O = -0.7 \text{ to } -2.0\%_o$ and $\deltaD = -43 \text{ to } -48\%_o$, indicative of a mixed meteoric - magmatic origin for the mineralising fluids. Sulphide $\delta^{34}S$ values range from -3.0 to 8.2\%o, with most between 4 to 6\%o. The lack of evidence for sulphur isotope fractionation between different sulphide minerals is consistent with non-equilibrium precipitation of sulphides from a reduced fluid, low temperature kinetic effects and / or a $H_2S : \text{metal} \gtrsim 1$. Sulphur was sourced from either the magma, the country rocks, or a combination of both.

Primary fluid inclusions hosted by sphalerite and quartz have a range of trapping temperatures and salinities interpreted to have resulted from mixing of $\approx 140^\circC$ and $\approx 27 \text{ wt.\% NaCl}$ equivalent Na-Ca-K-Cl brine with a $\approx 180^\circC$ dilute chloride water. Thermodynamic modelling has shown that sufficient concentrations of $Pb$ and $Zn$ (> 1 ppm) can be transported together with reduced sulphur ($\sum S = 0.002 \text{ molal}$) in a low temperature ($150^\circC$) saline brine ($\approx 6 \text{ molal}$) to form the Menninnie Dam deposit. The physiochemical attributes of the mineralising fluid at $150^\circC$ are estimated to have been $\log fO_2 = -46$ and $\text{pH} = 4.6$. Dilution through mixing with heated groundwater was a possible base metal depositional mechanism but is predicted to have been less effective than the $\text{pH}$ increase that resulted from dissolution of the host marbles.

Soon after cessation of the mineralising event, the stratigraphy was mantled by a single cooling unit $> 260 \text{ m thick}$ of lithic-rich ($\approx 45\%$ and up to 20 m across) welded ignimbrite (MD ignimbrite). The thickness, abundance and size of lithic clasts in the MD ignimbrite, and shallow intrusion of granite are consistent with an intracaldera setting. Following welding and cooling of the MD ignimbrite, the lower part of the MD ignimbrite and the Hutchison Group near the southern end of the Menninnie Dam deposit were partially altered to a texturally destructive quartz - chlorite - carbonate - calc-silicate assemblage by a hot, low salinity water ($190 - 356^\circC$ and 0 - 3 wt. \% NaCl equiv.). Carbon, hydrogen, and oxygen isotopes are consistent with a meteoric water that had undergone partial isotopic exchange with igneous rocks. Mineral textures, whole rock geochemistry, lead and sulphur isotope data are consistent with the Menninnie Dam Pb - Zn - Ag mineralisation being partially dissolved and reprecipitated by this event, with no addition of metals or sulphur.
My thanks go to my partner, Margaret for her love, patience and support throughout the course of this thesis.

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11.1 Thermodynamic data used to construct models
Australia’s major lead-zinc deposits can be subdivided into four groups:

- **Mississippi Valley type.** Typically 2-10 million tonnes (eg. Cadebut, W.A; Murphy, 1990).
- **Volcanic-hosted massive sulphides.** Typically 5-15 million tonnes (eg. Hellyer, Tasmania; Large, 1992).
- **Sediment hosted stratiform deposits of Middle Proterozoic age.** Typically up to 200 million tonnes. (eg. HYC, McArthur River, N.T; Logan et al., 1990).
- **Sediment hosted stratiform deposits of Lower Proterozoic age.** (eg. Broken Hill; 280 million tonnes; Parr and Plimer, 1993).

Australian Lower to Middle Proterozoic, sediment hosted, metamorphosed, Broken Hill-type (Beeson, 1990a) deposits have accounted for a considerable proportion of the worlds past production of lead, zinc and silver, however, as reserves of this deposit type are depleted, renewed emphasis is being placed on additional discoveries. The genesis and timing of mineralisation in relation to the host rocks has been debated since the discovery of the Broken Hill deposit and is still poorly understood (Walters, 1996). Research and documentation of Broken Hill-type deposits, particularly in the area of genesis and timing of mineralisation, will greatly improve our understanding of the geological environment in which these deposits were formed and contribute significantly to the success of exploration programs.

The Menninnie Dam prospect is a sub-economic Pb-Zn.Ag deposit hosted by Lower Proterozoic rocks of the Eyre Peninsula, South Australia (Fig. 1.1). The deposit was interpreted to have a number of similarities to Broken Hill-type mineralisation including:

- **Lower Proterozoic carbonate and calc-silicate host rock lithologies** (Higgins et al., 1990)
- **Stratabound, and possibly stratiform, mineralisation** (Higgins et al., 1990)
- **High metamorphic grade** (Higgins et al., 1990)
- **Associated carbonate-facies banded iron formations** (Higgins et al., 1990)
- **Broad geochemical similarities** (Breson, 1990b)

Early in the project’s history, Billiton Australia Ltd, and joint venture partner, Aberfoyle Resources Ltd, recognised the need to determine the paragenesis of the host rocks and the mineralisation, and undertook intensive in-house research. However, the genetic relationship between the mineralisation and host rocks remained unresolved.
From a company point of view, the selection of specific mineral belts for the application of advanced exploration techniques requires confidence that economic deposits may exist in that belt. While most exploration investment is directed towards mineral belts known to contain large viable deposits with a recognised style of mineralisation, the mineral potential of Lower Proterozoic rocks (≈ 20,000 km²) of the Eyre Peninsula, South Australia is reliant on the significance of the only known large Pb - Zn - Ag occurrence at Menninnie Dam. The genetic model arrived at for the Menninnie Dam deposit could initiate interest in a major 'new' mineral belt close to existing rail, smelting and shipping facilities at Port Pirie.

To ensure continued investment in this substantial portion of South Australia, the genetic style of the Menninnie Dam mineralisation requires formal research on the geologic setting, mineralogy, paragenesis, geochemical and isotopic signatures, nature of the mineralising fluids, deposit halos, and effect of metamorphism, in order to format genetic concepts that will contribute to useful exploration models. Positive results will enhance the confidence of present and future explorers in the area.

1.2 Location
The Menninnie Dam prospect is one of a number of base metal prospects on the Wilcherry Hill exploration lease, located 50 km north of the small township of Kimba and 130 km west-south-west of Port Augusta on the northern Eyre Peninsula of South Australia (Fig. 1.1). The geographic co-ordinates of Menninnie Dam are; latitude 32°38'S and longitude 136°26'E. The Australian map grid co-ordinates are; 632850E, 6386986N. The local
The first public domain information on the Menninnie Dam project was published by Higgins and Hellsten (1986). They briefly described the local geology as part of the Hutchison Group lithologies and subdivided the host rocks into 4 suites. Higgins et al. (1990) gave a more detailed account of the discovery, mineralisation, host rocks, and recognised a fifth rock suite. An inferred resource, based on widely spaced drilling, was estimated as 1.7 million tonnes grading 5% Pb, 8% Zn and 100 grams per tonne Ag.

Beeson (1990b) analysed 212 host rock samples from three prospects on the Wilcherry Hill exploration lease, including Menninnie Dam. Beeson (1990b) used litho-geochemistry to:

- test for geochemical halos which may exist around known mineralisation.
- to compare the composition of marble and calc-silicate associated with mineralisation with that from non-mineralised parts.
- characterise the deposition environment for the three prospects.

The Menninnie Dam prospect was considered to have many similarities to the Northern Australian SEDEX deposits (Beeson, 1990b).

1.4 Menninnie Dam project history

Billiton Australia Ltd (now Acacia Resources Ltd) began exploring for sedimentary-exhalative Broken Hill-type mineralisation in the Eyre Peninsula of South Australia in the early 1980’s. Metasediments of the Lower Proterozoic Middleback Subgroup were considered to have a number of attributes in common with the host sequences to major base metal deposits including Aggeneys and Gamsberg in South Africa and Broken Hill in Australia (Higgins et al., 1990). The Middleback Subgroup was considered comparable to lithologies in the Broken Hill area on the basis of a Lower Proterozoic age, mixed clastic and chemical sedimentary lithologies and high metamorphic grade.

Aero-magnetics were used to define initial exploration targets. Targets were defined as positive linear magnetic features parallel to stratigraphy, interpreted as banded iron formations (Higgins et al., 1990). In 1981, drill testing of a series of isolated magnetic highs on the Wilcherry Hill exploration lease intersected a number of zones of base metal mineralisation. Disseminated and massive base metal sulphides intersected at a prospect near the Menninnie stock watering dam subsequently became known as the Menninnie Dam
prospect. By 1989, six percussion holes and fourteen diamond holes had outlined an area of anomalous Pb and Zn over a 3 km by 1 km area (Higgins et al., 1990).

In 1989, Aberfoyle Resources Ltd. joint ventured into the Wilcherry Hill exploration lease acquiring a 60% share and project management. Aberfoyle Resources Ltd drilled a further 18 diamond holes and 13 percussion holes between 1989 and 1993, extending the known area of mineralisation. Because of the wide spacing of drill holes and the complex nature of the mineralisation, no new resource figure was calculated.

Given the unusual nature of the Pb - Zn - Ag mineralisation, the joint venture partners decided more detailed research was required to investigate the deposit genesis. It was around this time that I approached the joint venture partners and suggested they offer the Menninnie Dam prospect as a PhD research topic examining the timing and genesis of mineralisation. Mr John Anderson (then; Regional Manager of Aberfoyle Resources Ltd., South Australia) and Dr. Bob Beeson (then; Senior Research Scientist, Billiton Resources Ltd) were instrumental in supporting a PhD research project, and in mid 1992 final approval was given. Since establishment of this research project, negligible additional exploration has been undertaken by the joint venture partners. In the latter half of 1995, Aberfoyle Resources Ltd diluted its interest in the Wilcherry Hill exploration lease below 50% and passed management to Acacia Resources Ltd.

1.5 General statement of aims and methods
This research project was established to unravel the genesis of the Menninnie Dam Pb - Zn - Ag deposit. A thorough study of all aspects of the deposit is considered to be beyond the limits of this PhD, and three topics were selected for in-depth research:

- Geologic setting of mineralisation
- The timing of mineralisation
- The nature of the ore forming fluids

The decision to target these topics was based on the belief that they would contribute significantly to understanding the processes involved in formation of the deposit.

The local geology has been briefly described by Higgins and Hellsten (1986), Higgins et al. (1990) and Beeson (1990). Since publication of their work, additional drilling has been undertaken, enlarging the zone of known mineralisation and contributing significant new information on the local geology. It was therefore necessary to review and revise the prospect geology so that the contribution of the various geologic processes that controlled the mineralising event could be better appreciated.
The following data has been made available by the Joint Venture Partners and by Dr Graham Carr, Dr. Anita Andrew, and Dr. Judy Dean (CSIRO) for the purposes of this research project:

- Core from the 32 diamond drill holes (11,639m total).
- Percussion drill chips.
- Geochemical assay results from drilling (elements analysed: Mn, Fe, Zn, Cu, Pb, Ag and, rarely, W, Sn, As, Hg and Au).
- Whole rock geochemistry for drill holes PD0001-PD0008 as used by Beeson (1990b).
- Collection of 210 thin and polished sections.
- Airborne magnetic survey results for the Menninnie Dam prospect and surrounds.
- Lead isotope database for the Menninnie Dam prospect (228 analyses).
- Sulphur isotope database (14 analyses).

The prolonged period of exploration at Menninnie Dam (1981 - present), with the involvement of a number of companies and numerous geologists, has resulted in a lack of uniformity of drill hole logging techniques. In order to overcome this problem, the diamond drill core has been re-logged, with emphasis on documenting the relationship between the host rocks and mineralisation. The percussion pre-collars to diamond drill holes have not been re-logged.

Establishment of the relationship between the host rocks, various tectonic events and mineralisation is fundamental to determining the timing of mineralisation. The ages of the host rocks and major tectonic events can be constrained within reasonable limits by comparison to similar units elsewhere on the Eyre Peninsula for which there are well-established ages. In this thesis, constraining the timing of mineralisation is attempted using a number of independent techniques including; stratigraphic relationships, model lead isotope ages and U-Pb zircon dating.

The nature and composition of the mineralising fluids have been investigated through examination of; sulphide and gangue mineralogy, mineral paragenesis, fluid inclusions, sulphur, lead, oxygen, deuterium and carbon isotope data and thermodynamic modelling.
Chapter two
Physiography and regional geology

2.1 Introduction
This chapter provides a review of the physiography and regional geologic setting of the Eyre Peninsula of South Australia. Drexel et al. (1993) reviewed and collated publications on the Precambrian geology of South Australia, as well as contributing a significant amount of new data compiled by the Geological Survey of South Australia. Much of the regional geology summary is taken from their publication. The geologic time scale adopted in this study is based on the IUGS, 1989, Global Stratigraphic Chart (Fig. 2.1).

2.2 Physiography
Much of the Eyre Peninsula has low topographic relief, rising gently from the coast to form extended flat plains. On the eastern Eyre Peninsula, the plains are dissected by ranges of low hills, such as those inland from Cowell, north of Cleve and the extensive Gawler Ranges which form the northern margin of the Eyre Peninsula (Fig. 2.2). Elsewhere, the plains are broken by isolated hills and ridges of more resistant granite and sedimentary formations. The Middleback Ranges, Darke Peak, Uno and Corunna are examples of prominent, resistant sedimentary formations while Carappee Hill and Minnipa Hill are examples of resistant granite bodies. The highest peak on the Eyre Peninsula is Mount Kolendo in the Gawler Ranges (488 m A.S.L).

The morphology of the Gawler Ranges was described by Campbell and Twidale (1991) as "ordered rows of bornhardts" formed by differential weathering along fracture systems including orthogonal fractures, sheet fractures and columnar jointing. Bornhardts were interpreted to have been initiated by weathering in the Jurassic or earlier Mesozoic, and then exposed by erosion in the Late Cretaceous to Early Tertiary (Campbell and Twidale, 1991).

Outcrop over much of the Eyre Peninsula is poor, and Quaternary aeolian sands and silts form an extensive cover sequence. Inland, rock exposures are restricted to isolated ridges, creeks, quarries and road cuts. This contrasts with the eastern coast line of the Eyre Peninsula which probably has some of the better coastal exposure in Australia.

Much of the central and southern Eyre Peninsula is cropped for wheat, however some 20 km south of the Gawler Ranges, land use changes from wheat to sheep. The Menninnie Dam prospect is about 20 km north of the northern margin of the wheat belt and immediately south of the Gawler Ranges in an area dominated by mallee scrub and salt bush. The terrain rises gently from the edge of the wheat belt to the low hills of the Gawler Range Volcanics.
Fig. 2.1. Geologic time scale based on the IUGS 1989 Global Stratigraphic Chart. The Neoproterozoic Adelaidean subdivisions follow usage by Preiss (1987).
Fig. 2.2 Tectonic provinces, and summarized stratigraphy of South Australia (after Parker et al., 1993). The approximate margin of the Gawler Craton is indicated by a dotted line.
Chu~ter 2  Phvsiourunhv curd
re~ional geolo~v
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2.4

The Menninnie Dam prospect area is covered by a variable thickness of Quaternary reddish brown and yellowish brown gravelly and sandy clay, silt and sand with soft earthy carbonates and hard calcrete nodules (Blissett et al., 1988). Within a kilometre to the east and south of the prospect, breakaways of terrazzo type silcrete form small scarps a few metres high. Widely scattered outcrop of low lying Gawler Range Volcanics surround the prospect area and increase in abundance and profile to the west and north where they have formed the foothills of the Gawler Ranges. Areas of no outcrop are commonly weathered to depths of > 60 m.

2.3 Regional Geology

Gawler Craton

The Gawler Craton is a stable block of late Archaean and Early Proterozoic gneisses, granites and metasediments, Middle Proterozoic sedimentary rocks and extensive acid volcanics (Fig. 2.2). Parker et al. (1993) defined the Gawler Craton as that region of crystalline basement that has not been substantially deformed or remobilised, except by minor epiorogenic movements since approximately 1450 Ma. The Gawler Craton is bounded to the north-west by the Musgrave Orogenic Domain, although the contact zone is concealed by sediments of the Officer Basin. The western boundary is also indistinct, and has been placed along a distinct gravity and aeromagnetic feature, interpreted to define the western edge of shallow crystalline basement (Parker et al., 1993).

The Torrens Hinge Zone forms the eastern margin of the Gawler Craton, with the boundary mantled by sedimentary lithologies of the Adelaide Geosyncline. The south-eastern margin of the Gawler Craton was interpreted by Belperio and Flint (1993) to extend to the eastern margin of the Torrens Hinge Zone and to the northern margin of Kangaroo Island. The south-western margin of the Gawler Craton is interpreted to be a faulted contact associated with the break up of Gondwana and separation of Antarctica during the Mesozoic (Willcox and Stagg, 1990).

Subdomains

The Gawler Craton has been subdivided into a number of discrete tectonic subdomains based on structural, metamorphic, and stratigraphic differences. Boundaries are separated by major aeromagnetic linears interpreted as significant crustal sutures (Parker et al., 1993; Fig. 2.2). The Nawa Subdomain crops out in the north-west of the craton and is of mixed or complex character incorporating late Archaean - Paleoproterozoic rocks, Mesoproterozoic granites, Neoproterozoic sediments and Cambrian - Ordovician sediments and volcanics. The Christie and Coulta subdomains (Fig. 2.2) contain most of the exposed Archaean rocks, and some of the Paleoproterozoic granitoids, but are largely free of younger Proterozoic metasediments, volcanics, or intrusives. The Wilgena and Nuyts subdomains (Fig. 2.2) contain Archaean metasediments intruded by Mesoproterozoic granites.
The Cleve Subdomain crops out in the south east of the Eyre Peninsula (Fig. 2.2) and is a Paleoproterozoic fold belt containing both clastic and chemical sediments (eg. Hutchison Group 1950-1840 Ma.), and voluminous granitoids (eg. Lincoln Complex) associated with closure of the basin and deformation (eg. Kimban Orogeny, KM - D1 to KM - D3). The Moonta Subdomain crops out in the north-eastern Eyre Peninsula (Fig. 2.2) and is considered to be a younger extension of the Cleve Subdomain (Parker et al., 1993). Much of the deformation and emplacement of granitoids in the Cleve Subdomain ceased by ~1740 Ma., when sequences in the Moonta Subdomain were being deposited or erupted. Later deformation was manifested as shear zones in the Cleve Subdomain and extensive folding in the Moonta Subdomain.

The Mesoproterozoic Gawler Ranges Volcanic Province is the central Gawler Craton formed during voluminous out-pouring of acid to intermediate lavas and pyroclastics. Whereas other sub-domains have north-east trending boundaries, the Gawler Range Volcanics Province has poorly defined boundaries and no clear underlying tectonic control. Unlike other subdomains, the Gawler Range Volcanic Province is relatively undeformed. Co-magmatic Mesoproterozoic, granitoids of the Hiltaba Suite crop out in a broad belt along the southern and western margins of the Gawler Range Volcanic Province (Fig. 2.2).

Mesoproterozoic and late Paleoproterozoic sediments (eg. Tarcoola Formation and Corunna Conglomerate) have not been assigned to basins, except for the Blue Range Beds (Itiledoo Basin of Flint and Rankin, 1989), and the Panduura Formation which was deposited in the Midproterozoic Cariewerloo Basin (Cowley, 1993).

2.3.1 Stratigraphy

This section focuses on the Archaean and Proterozoic units of the Cleve and Coulta Subdomains and the Gawler Range Volcanics Province. The occurrence of Archaean and Proterozoic rocks is detailed in the relative geological maps produced by the South Australian Department of Mines and Energy and has been summarised by Parker et al. (1988). A more detailed account, which includes recent data, was given by Drexel et al. (1993). The following summary is largely derived from these two publications. The stratigraphy of the Eyre Peninsula is summarised in Figures 2.3. and 2.4.

Archaean

Sleaford Complex

The Sleaford Complex are the basement rocks of the Gawler Craton and are more abundant in the western half of the Eyre Peninsula. Outcrop is excellent along the southern and southwestern coastlines, however, inland exposure is restricted to scattered outcrops surrounded by Cainozoic cover.
Figure 2.3  Stratigraphic column of the Archaean and Proterozoic rocks of the Cleve Subdomain and Gawler Range Volcanic Province. (Modified from Parker et al., 1988; Drexel et al., 1993).
Fig. 2.4  Simplified geology of the eastern Eyre Peninsula of South Australia (modified after Parker, 1993b).
The Sleaford Complex consists of gneissic metasediments deposited in a shallow water platform sequence, possibly with associated contemporaneous tholeiitic basalt sills and flows. The age of sedimentation was interpreted to be older than 2640 Ma., based on U-Pb zircon dating by Fanning et al. (1986).

The Sleaford Complex is divided into four units:
- Dutton Suite granitoids youngest
- unnamed gneiss
- Wangary Gneiss
- Carnot Gneisses oldest

Carnot Gneiss
The Carnot Gneiss has been subdivided into ortho and para-gneisses and consists of compositionally banded metasediment, concordant sheets of tholeiitic metabasalts, (flows, sills or dykes), and synorogenic felsic intrusions.

Fanning et al. (1980) described a compositionally layered garnetiferous quartzo-feldspathic ortho-gneiss cropping out on a wave-cut platform at the southern-most tip of the Eyre Peninsula (Cape Carnot; Fig. 2.4), and applied the name 'Carnot Gneiss' to them. The Carnot Gneiss consists of plagioclase, K-feldspar, quartz, biotite and garnet with accessory spinel, magnetite, ilmenite and rutile and rounded zircons. Geochemically, the gneisses contain normative corundum, indicating excess $\text{Al}_2\text{O}_3$, consistent with a sedimentary origin.

The para-gneisses consist of concordant 1-2 m thick bands of tholeiitic composition interpreted to have been emplaced as sills and / or dykes prior to, or during peak metamorphic conditions (Daly and Fanning, 1993).

The Carnot Gneiss also crops out in the Warramboo-Waddikee area of the central Eyre Peninsula, where it consists of compositionally banded, to homogenous, quartz, garnet, feldspar gneiss. Whole rock Rb-Sr geochronology of felsic gneiss from a North Broken Hill Ltd and CRA Exploration drill hole (Warramboo WD1; Fig. 2.4) gives an age of 2520 $\pm$ 163 Ma. (Webb et al., 1986).

Wangary Gneiss
The Wangary Gneiss is exposed 13 km southwest of Point Sir Isaac on the Coffin Bay Peninsula and along the coast for = 60 km north of Coffin Bay (Fig. 2.4). The Wangary Gneiss is a compositionally layered microcline-plagioclase-quartz-biotite-muscovite gneiss interpreted as a metasediment and interpreted to be a lower grade equivalent of the Carnot Gneiss (Daly and Fanning, 1993). Whole rock Rb-Sr dating by Webb et al. (1986) gives an age of 2380 $\pm$ 60 Ma. A poorly defined $U$-$Pb$ zircon age of 2445 $\pm$ 32 Ma. was obtained by Fanning (1987).
Chapter 2 Physical and regional geology page 29

Sleaford Orogeny
The earliest recorded orogenic episode on the Eyre Peninsula spans the period 2300 - 2637 Ma., and is known as the Sleaford Orogeny (Webb, 1980; Fanning et al., 1986). The Sleaford Orogeny is interpreted to have consisted of three and possibly four deformational events. U-Pb zircon dating of an intrusive hypersthene granulite facies gneiss at Cape Carnot gave a concordant age of 2637 ± 21 Ma. (Fanning et al., 1986), which has been interpreted as the maximum age for the onset of metamorphism (Daly and Fanning, 1993). The minimum age for peak granulite facies metamorphism has been interpreted at 2436 ± 137 Ma. based on whole rock Rb-Sr dating (Fanning et al., 1986). Prograde mineral assemblages for the Sleaford Complex indicate temperatures of 700 - 800°C and pressures of 7 - 9 kb (Bradley, 1980). Two pyroxene geothermometry by Fanning (h Parker et al., 1988) for the basic granulites give comparable calculated temperatures of 800 - 860°C.

Paleoproterozoic
Paleoproterozoic rocks in the Gawler Craton are widespread, forming a broad belt across the eastern Eyre Peninsula (Fig 2.4).

Dutton Suite
The Dutton Suite consists of synorogenic granitoids emplaced during the Sleaford Orogeny, exposed in the southwest of the Eyre Peninsula, and on nearby islands. The timing of the later stages of the Sleaford Orogeny are interpreted from whole rock Rb-Sr dating of the Dutton Suite granitoids at 23 16 ± 71 and 2334 ± 109 Ma. (Webb et al., 1986).

Palaeoproterozoic gneisses
In a number of localities, there is a sequence of gneisses separating the Archaean Sleaford Complex from the overlying Palaeoproterozoic Hutchison Group metasediments. An unnamed gneiss underlying the Warrow Quartzite at Ullabidinie Creek and at 'Minbrie Springs' (= 20 km north of Cleve on the eastern Eyre Peninsula) has a Rb-Sr age of 2315 ± 175 Ma., which is indistinguishable from the age obtained for the Dutton Suite Granites.

The Miltalie Gneiss underlies the Warrow Quartzite at Plug Range (Fig. 2.4), and consists of a medium grained, well foliated quartz-microcline-plagioclase-biotite gneiss with pink pegmatitic segregations. U-Pb geochronology by Fanning et al. (1988), gave a zircon crystallisation age of 2014 ± 28 Ma., with metamorphism at = 1964 Ma. The 1964 Ma. age provides a constraint on the maximum age of the overlying Hutchison Group.

Hutchison Group
Unconformably overlying the Sleaford Complex, or Miltalie Gneiss (and equivalents), is the Paleoproterozoic Hutchison Group. The Hutchison Group consists of high metamorphic grade rocks derived from shallow elastic and chemical marine sediments, and minor acid and
motic volcanics (Parker, 1993b). Corbett (1987) estimated the total thickness of the Hutchison Group to be about 10 km. The Hutchison Group has been intruded by numerous synorogenic granitoids, known as the Lincoln Complex, associated with the Kimban Orogeny. The Hutchison Group is host to Pb-Zn-Ag mineralisation at Menninnie Dam. Parker (1980b) subdivided the Hutchison Group into three main sequences:

- A basal quartzite sequence equated with the Warrow Quartzite of the southern Eyre Peninsula.
- A mixed chemical and clastic sequence equated to the Middleback Group of the Middleback Ranges, re-defined as the Middleback Subgroup.
- An upper pelitic unit known as the Yadnarie Schist.

The Warrow Quartzite consists of a massive flaggy, quartz-feldspar quartzite with rare bedding and cross stratified layering. It overlies the Sleaford Complex in the southern Eyre Peninsula and has a locally developed basal conglomerate and/or pelitic schist unit. In the eastern Eyre Peninsula, complex deformation and intrusion of synorogenic granitoids of the Lincoln Complex has obscured the base and, locally, the top of the Warrow Quartzite. The thickness of the Warrow Quartzite is not known, although quartzite units up to 1000 m thick are exposed at Caralue Bluff in the central Eyre Peninsula (Fig. 2.4).

Trough cross bedding in the Warrow Quartzite at Caralue Bluff has been interpreted to indicate fluviatile to marginal marine deposition. In the Cleve and Lipson regions, the Warrow Quartzite shows an upwards transition from massive flaggy quartzite to interbedded pelitic schist. The transition is interpreted to be a lateral facies change from shallow marine to a more distal marine environment of deposition associated with landward or westerly transgression which immediately preceeded deposition of the Middleback Subgroup (Parker and Lemon, 1982).

The Middleback Subgroup

The Middleback Subgroup conformably overlies the Warrow Quartzite. The thickness of the Middleback Subgroup has been estimated at 2000-2500 m, based on a section northwest of Cleve (Parker, 1993b). Based on the stratigraphy in the Middleback Ranges west of Whyalla (Fig. 2.4), Lemon (1980) subdivided the Middleback Subgroup into a basal dolomite unit overlain by two iron formations that are separated by a schist unit of clastic origin. The Katunga Dolomite lies unconformably on the Warrow Quartzite (Lemon, 1980). Parker (1993b) described the Katunga Dolomite as "a generally massive to poorly layered, white to pale grey dolomitic marble which is mostly medium grained and holocrystalline. Layering (when present) is defined by bands of serpentine (± forsterite relics) and calcite ± diopside ± tremolite alternating with massive dolomite ... pale yellow-green serpentine nodules and diopside characterise high grade metamorphic zones whereas tremolite, actinolite and talc are common in lower grade areas".
The Katuna Dolomite contains minor bands of schist, and becomes more Fe-rich up stratigraphy, and grades into the overlying 'Lower Middleback Jaspilite'. Pyrite occurs at the dolomite / iron formation boundary (Lemon, 1980). Iron silicate minerals in the Lower Middleback Jasperlite include iron-rich talc, grunerite, cummingtonite and anthophyllite, diopside / hedenbergite, with rare hypersthene present at higher metamorphic grades. Magnetite is the dominant iron oxide at depth, with more oxidised minerals such as haematite and iron oxyhydroxides occurring in the near surface environment. Parker (1993b) stated that "the iron formations are well banded with alternating 0.5 - 20 mm thick bands of quartz, carbonate or iron oxides ± iron silicates ± carbonate. Accessory minerals include garnet, biotite, tremolite and occasionally microcline. Original chert has been variably recrystallised to form fine to medium, and locally coarse, grained quartz. Strong crystallographic and grain-shape fabrics are ubiquitous and define axial planar foliations in early fold generations". The prominent banding and local cyclic banding in carbonate facies iron formations were interpreted to be relict primary layering (Parker and Lemon, 1982).

The Lower Middleback Jaspilite increases in clastic content and grades upwards into the Cook Gap Schist. The Cook Gap Schist is estimated to have an apparent thickness of >1500 m, and consists of predominantly semi-pelitic, quartz-veined garnet-mica schist and gneiss with minor calc-silicate gneiss, magnetite-bearing gneiss, and concordant amphibolites (Parker, 1993b). The origin of the amphibolites is uncertain; they may either be concordant volcanics of quartz tholeiite composition, or metamorphosed sills and dykes rotated into structural concordance.

The Cook Gap Schist grades upwards into a mixed carbonate-silicate facies iron formation that forms the basal section of the Upper Middleback Jaspilite. The Upper Middleback Jaspilite is chemically and mineralogically similar to the Lower Middleback Jaspilite, and Yeates (1990) argued tectonic repetition, an interpretation disputed by Parker (1993b). The top of the Middleback Subgroup is defined as the top of the Upper Middleback Jaspilite.

In the Cleve region (Fig. 2.4), the Upper Middleback Jaspilite is overlain by the Yadnarie Schist which consists of a fine to medium grained pelitic to semipelitic, quartz-veined mica schist that contains quartz, muscovite and biotite with minor opaque minerals, plagioclase, sillimanite, garnet and tourmaline. Amphibolites are rare or absent. Near Carappee Hill, in the central Eyre Peninsula (Fig. 2.4), the Middleback Subgroup is overlain by mixed acid volcanics and calc-silicates of the Bosanquet Formation (Rankin et al., 1988) which was considered to be a lateral equivalent of the Yadnarie Schist (Parker, 1993b). The volcanics consist of recrystallised rhyodacite with distinctive relict phenocrysts of microcline and bluish coloured quartz. U-Pb dating of zircons from the Bosanquet Formation give a zircon crystallisation age of 1845 ± 9 Ma. (Rankin et al., 1988), and provides a minimum age for deposition of the Hutchison Group.
Lemon (1980) interpreted the changes in lithofacies within the Middleback Subgroup to have resulted from deposition in a shallow sea with a north-south trending shoreline. Initial transgressive chemical deposition of the dolomite was succeeded by sulphide, carbonate-silicate, and oxide facies iron formations. This was terminated by a period of shoreline progradation and a change to clastic sedimentation (Cook Gap Schist) followed by a return to transgression and chemical sedimentation (Upper Middleback Iron Formation). This was probably terminated once more by clastic sedimentation, (Yadnarie Schist).

**Myola Volcanics and Broadview Schist**

The Myola Volcanics and Broadview Schist crop out in the north-east Eyre Peninsula (Fig. 2.4), east of the Middleback Ranges (Moonta Subdomain). While their relationship to the Hutchison Group is uncertain, U-Pb zircon dates for the Myola Volcanics of 1791 ± 4 Ma. indicate they are younger than both the Bosanquet Formation and the Hutchison Group (Fanning et al., 1988).

**Kimbian Orogeny**

The Hutchison Group metasediments underwent a period of orogenic activity between ca. 1845 and 1710 Ma. known as the Kimban Orogeny. Parker (1978) recognised three major tectonic events preserved in rocks of the Cowell-Cleve area, north-eastern Eyre Peninsula that he ascribed to the Kimban Orogeny. The metamorphic events and the associated deformations are referred to as KM-D1, KM-D2 and KM-D3 respectively.

Peak metamorphic conditions for KM-D1 and KM-D2 attained >8 kbar and 800 - 900°C (granulite facies) in the southern Eyre Peninsula (Mortimer et al., 1980; Bradley, 1980) and 5 - 7 kbar and 600 - 675°C (amphibolite facies) in the central Eyre Peninsula (Parker et al., 1988). KM-D1 produced a layer parallel fabric and KM-D2 is characterised by tight to isoclinal folding and an axial planar fabric (Parker, 1993a).

KM-D3 was a retrograde event, with pressures of ~ 1.5 - 2 kbar and temperatures of ~ 450 - 550°C corresponding to the greenschist - amphibolite boundary (Parker, 1993a). KM-D3 resulted in formation of broad, open folds and major north to northeast-trending mylonitic shear zones. The Kalinjala Mylonite Zone is a major linear zone of intense ductile deformation (3 km wide at Port Neill) extending the length of the eastern Eyre Peninsula from Sleaford Bay in the south to the Middleback Ranges in the north (Parker, 1980a; Fig.2.4). The Kalinjala Mylonite Zone consists of mylonite with minor ultramylonite, remnants of protomylonite and scattered boudins of amphibolite. The mylonites at Port Neill were interpreted by Parker (1993a) to have a major component of dextral shear, which is consistent with their regional sigmoidal distribution. Rb-Sr dating of a weakly deformed pegmatite intruding the mylonite at Port Neill places an upper age limit on KM-D3 of 1710 Ma. (Fanning, 1984).
Lincoln Complex

Each deformation phase of the Kimbrian Orogeny was accompanied by the intrusion of (syn-orogenic) granitoids collectively known as the Lincoln Complex (Fig. 2.4). Radiometric age determinations of the granites has helped to constrain the timing of each deformation event (Table 2.1).

Table 2.1 Syn-tectonic granitoids of the Lincoln Complex.

<table>
<thead>
<tr>
<th>Deformation event</th>
<th>Granite Suite</th>
<th>Age</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>KD3</td>
<td>Moody Suite</td>
<td>~1740-1710 Ma.**</td>
<td>5 and 6</td>
</tr>
<tr>
<td>KD2</td>
<td>Middle Camp Granite</td>
<td>1738 ± 68 Ma.*</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>Colbert Suite (iso KD2 to pre KD3)</td>
<td>1757 ± 14 Ma.**</td>
<td>2</td>
</tr>
<tr>
<td>KD1</td>
<td>Minsite Gneiss</td>
<td>&gt;1800 Ma.**</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td>Donington Suite</td>
<td>1818 ± 13 Ma.* and 2</td>
<td>1</td>
</tr>
</tbody>
</table>

*U-Pb zircon; **Rb-Sr


The Donington Suite Granites are I-type, Sr-depleted, Y-undepleted, fractionated granites low in incompatible elements (Wyborn et al., 1992). The Moody Suite Granites are considered to be a sub-group of this granite suite, with F enrichment and variable Y, Zr and Nb. Moody Suite Granites with high SiO2 contents are enriched in K2O, F, Th, U, Rb and K/Rb (Wyborn et al., 1992). Detailed petrochemistry for the Colbert Suite and MiddleCamp Granite has not been published.

Multiply deformed mafic dolerite to gabbronorite dykes of the Tournefort Dyke Swarm comprise up to 25% of the Donington Granitoid Suite (Parker, 1993a), and were interpreted to have been emplaced between 1845 - 1750 Ma. (Parker, 1989).

Tarcoola Formation

The Tarcoola Formation consists of fluvial to marginal-marine clastic sediments deposited in grabens and half grabers (Daly, 1993a) along the north-western margin of the GRV (Fig 2.2). Daly (1985) subdivided the Tarcoola Formation into three members. The basal Peela Conglomerate Member comprises conglomerate, arkosic quartzite and green lithic and volcaniclastic sandstone. The Fabian Quartzite Member overlies the Peela Conglomerate Member and consists of micaceous quartzite. At Tarcoola, quartzite is interbedded with laminated carbonaceous shale and pyritic siltstone of the Sullivan Shale Member which includes minor water-laid tuffs. U - Pb zircon dating of the tuffs indicates and age of 1656 ± 7 Ma. for volcanism and sedimentation (Fanning, 1990). The Tarcoola Formation has undergone lower greenschist facies metamorphism and is folded about broad, open, fold axis (Daly, 1993a).

Wartakan Event

Throughout most of the Eyre Peninsula, there is evidence of a later deformation event named the Wartakan Event by Thomson (1969) and abbreviated here to WD1. Deformation is more
pronounced in the Cowell-Cleve area and the southern Eyre Peninsula where late stage crenulations, kink bands and fractures associated with WD1 occur. Emplacement of granitoids of the St. Peters Suite in the Streaky Bay region are interpreted to have been associated with WD1. Granitoids of the St. Peters Suite are deformed, and characterised by intermingling of felsic and mafic magma phases. The granitoids have U-Pb zircon dates of 1630 - 1620 Ma. (Flint et al., 1990), which constrains the timing of WD1.

Mesoproterozoic

Gawler Range Volcanics (GRV)

Thomson (1966) introduced the name Gawler Range Volcanics (GRV) for the ellipsoidal mass of acid to intermediate volcanics exposed in the north-central Eyre Peninsula (Fig. 2.2; Fig. 2.5). The GRV Province covers an area of at least 25,000 km², and was presumably more extensive prior to being reduced to their current extent by weathering. When scattered outcrops and drill intersections are taken into account, Campbell and Twidale (1991) estimated the original extent of the GRV to be of the order of 60,000 km².

Most of the mapping and interpretation of the GRV has been undertaken by the Mines and Energy Department of South Australia and postgraduate students of the University of Adelaide and La Trobe University. Blissett (1975; 1986) formally described and named many of the units of the GRV. Detailed studies of the Kokatha area (Fig. 2.5) were undertaken by Branch (1978) and Robertson (1989). Giles (1977) undertook a detailed study of the Lake Everard area. Jagodzinski (1985) described the geology of the Toondulaa Bluff area (Fig. 2.5). Creaser and White (1991) described the geology, geochemistry, geothermometry and possible origins of the Yardea Dacite (Fig. 2.5). Giles (1980), Stewart (1992) and Creaser (1995) undertook detailed studies of the petrogenesis of the GRV. A summary compilation of previous work up until 1993 (excluding Stewart, 1992) was presented by Blissett et al. (1993).

Blissett and Radke (1980) subdivided the GRV into 'older' and 'younger' sequences based partly on the attitude of more steeply dipping older beds and flat lying younger beds. Blissett et al. (1993) adopted the terms 'lower' and 'upper', and put the division at the base of the Yardea dacite and equivalents. Stewart (1992) divided the GRV into an earlier 'developmental phase' and a later 'mature phase' on the basis of geochemical and isotopic differences, and placed the boundary at the base of the Nonning Rhyodacite. The developmental phase includes all volcanics in the Lake Everard, Tarcoola, Kokatha areas in the west, and units which lie below and including the Bittali Rhyolite in the south of the GRV province. Subdivision of the GRV into the developmental and mature phases of Stewart (1992) is presented in Figures 2.5 and 2.6.
Fig. 2.5 Distribution of the Gawler Range Volcanics with subdivision into developmental phase and mature phase. (Modified from Stewart, 1992)
Fig. 2.6 Subdivision of the Developmental phase and mature phase of the Gawler Range Volcanics (after Stewart, 1992)
The developmental phase occurs as smaller more scattered volcanic centres along the western and southern margins of the GRV Province (e.g. the Tarcoola, Kokatha, Lake Everard centres in the west and the Bittali Rhyolite along the southern margin; Stewart, 1992). The later, more voluminous units of the mature phase cannot be attributed to any particular volcanic centre.

Units of the GRV comprise variably welded ignimbrites with minor localised lava flows, domes and dykes. Volcanic breccias and air fall tuffs form only a small component. Restricted flows of basalts and andesites are present in the lower part of the succession. Many of the units resemble lavas and show little evidence of a pyroclastic origin. However, Turner (1975), Blissett (1975), Blissett and Radke (1980), Giles (1977 and 1980), Giles et al. (1980), Creaser and White (1991), Stewart (1992) and others, have demonstrated that the GRV consists of composite layers of welded ignimbrite and rheoignimbrite. The Yardea Dacite is the most voluminous unit, covering an area of 12000 km² to an estimated thickness of > 250 m with little textural or compositional variation (Blissett, 1975). The lava-like appearance of the Yardea Dacite may be due to the high emplacement temperature (900 - 1100°C) and rheomorphic behaviour (Creaser and White, 1991; Stewart, 1992).

U-Pb zircon dating of oldest unit, the Waganny Dacite, and youngest unit, the Yardea Dacite, give ages of 1591 ± 3Ma. and 1592 ± 3 Ma. These ages are statistically indistinguishable (Fanning et al., 1988), and are interpreted to indicate eruption of a large volume of felsic magMa. in a geologically short space of time.

The GRV comprise a calc-alkaline assemblage of dacites, rhyodacites and rhyolites with minor potassic andesites and tholeiitic basalts erupted in a subaerial continental environment (Blissett et al., 1993). The entire Gawler Range Volcanic Province is high in K and enriched in Ba, Th, Zr, Y, Rh, and light rare earth elements (Stewart and Foden, 1990; Stewart, 1992). In a study of the geochemistry of the Lake Everard area, Giles (1980) demonstrated a compositional break in the SiO₂ content from 52 to 58 wt%, which could not be satisfactorily explained by crystal fractionation. Giles (1980) demonstrated that the parental andesites and low silica dacites can be closely matched by invoking 20-30% partial melting of a lower crustal, basic granulitic source in which plagioclase, clinopyroxene, orthopyroxene and magnetite are residual. Giles (1980) suggests primitive basalts in the province originated from a relatively shallow (< 60 km) mantle source, and separated from a rising mantle diapir or plume, ultimately providing the heat source for melting of the crust, from which the acid magmas were generated.

By contrast, Stewart (1992) suggested the developmental phase was formed by fractional crystallisation of a mantle-derived magma, and assimilation of up to 43% Archaean and / or Proterozoic crustal material in a series of isolated magma chambers. These isolated magma.
chambers later coalesced to form a single large chamber which gave rise to the mature phase volcanics. The gravity high underlying much of the GRV was interpreted by Stewart (1992) as a major mafic intrusive whose upper surface is at approximately 5 km depth. The GRV have undergone little deformation since emplacement, except for epiorogenic faulting. Low temperature oxidation of iron, associated with devitrification, has resulted in a characteristic brick red colour of the GRV (Blissett et al., 1993).

**Hiltaba Suite**

The Hiltaba Suite comprises anorogenic granitoids that have formed large batholiths and smaller plutons prevalent around the north-western, western and southern margins of the Gawler Ranges (Figs. 2.2 and 2.4). Granites are the most abundant rock type but the suite is bimodal and more mafic lithologies (quartz monzonite and granodiorite) occur as scattered plutons in the north-east of the craton. Granitoids of the Hiltaba Suite commonly have a distinctive brick red colour due to abundant minute inclusions of haematite in plagioclase and K-feldspar. The Hiltaba Suite was co-magmatic with the GRV, and is both overlain by, and intruded into the GRV. Granitoids of the Hiltaba Suite have intruded the Corunna Conglomerate and the Tarcoola Formation but not the Pandurra Formation. A mean weighted age for all U-Pb zircon dates of the GRV and Hiltaba Suite is 1589 ± 2 Ma., with a peak at 1590 Ma. (Creaser and Cooper, 1993).

The Hiltaba Granite and Charleston Granite of the Hiltaba Suite are considered to be a sub-group of I-type, Sr-depleted, Y-undepleted granites, (anorogenic), enriched in incompatible elements and characterised by high FeO / (FeO+Fe2O3) levels compared to all other Australian Proterozoic granites (Wyborn et al., 1992).

**Dykes**

Blissett et al. (1993) described the occurrence of numerous dykes around the margin of the central GRV mass. Dykes vary in length up to 18 km (Moonamby Dyke Suite; Giles, 1977), and vary in width from a few metres up to 1 km northwest of Corunna (Blissett et al., 1993). The dykes vary in composition from rhyolitic to andesitic. Rhyolitic dykes are commonly porphyritic. Blissett et al. (1993) suggested that the dykes in the Kokatha and Lake Everard areas may have been either feeder dykes to overlying volcanics, or slightly younger and co-magmatic with the Hiltaba Suite.

**Sediments associated with the GRV**

The Corunna Conglomerate occurs as scattered outcrops around the south and north-west margin of the GRV and east of Whyalla (Fig. 2.4). It consists of an interbedded thick sequences of coarse conglomerate, fine siltstone and sandstone (Daly, 1993b). Conglomerate clasts include Hutchison Group metasediments and GRV. Finer sediments include siltstones, sandstones, carbonaceous siltstone and silicified dolomite with common
tuffaceous horizons. The age of the Corunna Conglomerate is estimated to be between 1740 - 1585 Ma. and deposition was probably synchronous with the GRV (Daly, 1993b). The Corunna Conglomerate is interpreted to be of fluvial origin although the fine grained carbonaceous siltstone and sandstone at the top of the sequence were probably deposited in a shallow to moderately restricted marine basin (Lemon, 1972; Lemon and Goatin, 1983).

Pandurra Formation
The Pandurra Formation is a thick (>950 m) monotonous sequence of flat lying arenaceous redbed sediments confined to the Cariewerloo Basin on the north-east margin of the Gawler Ranges (Fig 2.2). It consists of medium to coarse grained, poorly sorted quartz and lithic sandstone with minor medium grained sandstone, pebble conglomerate and shale. The Pandurra Formation is interpreted as a continental, dominantly fluvial sequence (Cowley, 1993) although alluvial fan, aeolian and possible deltaic environments have been suggested for some of the formation (Lemon and Goatin, 1983; Lemon, 1972) and a shallow marine environment has also been interpreted (Curtis, 1977; O'Shea, 1982). The age of the Pandurra Formation is uncertain, however it clearly postdates the GRV. Rb - Sr geochronology of interbedded shale and siltstone indicate the unit is pre-Adelaidian, with a minimum age of 1424 ± 51Ma. (Fanning et al., 1983).

Blue Range Beds
The Itiledoo Basin is an elongate, east-west basin in the central Eyre Peninsula (Fig. 2.2). Up to 2500 m of sediments have filled the basin and are assigned to the Blue Range Beds (Flint, 1993a). Sediments were deposited in a braided stream - alluvial fan environment, and are dominated by sandy conglomerates with clasts of Hutchison Group lithologies and GRV (Flint, 1993a).

Cratonisation
The Gawler Craton was cratonised at about 1450 Ma. and has not been subject to significant deformation or remobilisation since, except for minor epirogenic movement and intrusion of the Gairdner Dyke Swarm.

Gairdner Dyke Swarm
Much of the north-eastern Gawler Craton is cut by a suite of north - north-west and north-west orientated mafic dykes known as the Gairdner Dyke Swarm (Fig. 2.2). Widths of composite dykes probably exceed 50 m and are commonly 30 km or more in length (Cowley and Flint, 1993). Dykes were emplaced around 1100 Ma., at approximately the same time as rifting of the Adelaide Geosyncline (Parker, 1989). The Gairdner Dyke Swarm produces a strong, positive, magnetic response, clearly visible on aeromagnetic maps. These magnetic features crosscut all older units of the Gawler Craton.
2.3.2 Mineralisation

Iron ore
Iron ore has been mined from the Middleback Ranges for nearly 100 years (Fig. 2.4). The following summary has been taken from various informal publications produced by the Public Affairs Department of BHP Steel. The explorer credited with discovering iron ore in the Eyre Peninsula was Edward John Eyre, and an entry in Eyre’s journal, dated 18th September 1840, reports the presence of an ironstone formation atop a hill at the northern end of the Middleback Ranges (probably what is now the Iron Baron or Iron Princess mine). However, it was not until 1897 that the leases were acquired by BHP. The iron ore was initially used as a flux in the smelters across the Spencer Gulf at Port Pirie before being shipped to the Newcastle steel works in 1915, signifying the start of the nation's steel production. Iron ore is still being mined some 80 years later with a number of producing mines along a > 40 km strike length of the Middleback Ranges. Total production of iron ore for each of the principal mining areas up to the end of 1990 stood at:

<table>
<thead>
<tr>
<th>Mine</th>
<th>Production (Mt)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Iron Knob</td>
<td>130</td>
</tr>
<tr>
<td>Iron Baron</td>
<td>56</td>
</tr>
<tr>
<td>Iron Duke</td>
<td>&lt;1</td>
</tr>
</tbody>
</table>

Uranium - copper - gold mineralisation
The Olympic Dam U - Cu - Au - Ag deposit was discovered in 1975 by Western Mining Corporation and is one of the largest copper and uranium mines in the world. It is located 520 km north-northwest of Adelaide at the northern end of the Stuart Shelf (Fig. 2.2). In 1987 the measured and indicated resource stood at 32 Mt contained copper, 1.2 Mt contained U3O8, 1200 t contained gold and 7000 t contained silver (Scott, 1987). The orebody occurs beneath 300 m of younger, flat lying sediments, within the Olympic Dam Breccia Complex, a large hydrothermal breccia system. The breccia complex is fully enclosed within the Roxby Downs Granite of the Hiltaba Suite, dated at 1598 ± 2 Ma. to 1588 ± 4 Ma. (Mortimer et al., 1988; Creaser and Cooper, 1993). Detailed descriptions of the history, mineralisation and geology are contained within publications by Scott, (1987), Lalor, (1986), Reeve, (1990a and b), Reeve et al., (1990), Smith, (1993), Creaser, (1989), Johnson and Cross, (1991), Roberts and Hudson, (1983, 1984), Conan-Davies, (1987), Oreskes and Einaudi, (1990, 1992), Bradbury, (1988), Esdale et al., (1987) and Johnson and McCulloch, (1995).

Sub-economic deposits of Fe-Cu-(U-Au) occur 20 km southwest of Olympic Dam at the Acropolis and Wirrda Well prospects (Cross, 1993) and at Oak Dam East, 39 km west of Lake Torrens (Davidson and Paterson, 1993).
Gold mineralisation
Quartz-vein hosted gold mineralisation has been mined at the Earea Dam, Glenloth and Tarcoola Goldfields in the north and north-west of the Gawler Range Volcanics Province (Fig. 2.2). Quartz veins in the Earea Dam and Glenloth Goldfields are hosted by Archaean gneiss, whereas those of the Tarcoola Goldfields are hosted by siltstones and quartzites of the Tarcoola Formation (1656 ± 7 Ma., Fanning, 1990). Mineralisation in all three goldfields is interpreted to be related to granitoids of the Hiltaba Suite (Daly, 1993c). Detailed accounts of the mineralisation and geology are presented by Hein et al. (1994), Daly et al. (1990), Blissett, (1985), Fradd, (1988), Circosta and Gum, (1988).

Base metal mineralisation
The Menninnie Dam Pb-Zn-Ag deposit is the most significant discovery of base metal mineralisation in the Gawler Craton (Fig. 2.4). Smaller deposits within Hutchison Group sediments have been mined at the Atkinson's Silver Mine and Miltalie Mine 20 km northwest of Cowell (Parker, 1993b), however their relationship to the host metasediments has not been resolved.

2.4 Summary
The Gawler Craton is a stable block of late Archaean and Early Proterozoic gneisses, granites and metasediments, Middle Proterozoic sedimentary rocks and extensive acid volcanics. Archaean metasedimentary and volcanic rocks of the Sleaford Complex are more prevalent in the north and west of the craton whereas Palaeoproterozoic rocks are more prevalent in the south-eastern Gawler Craton. The earliest recorded orogenic episode on the Eyre Peninsula, the Sleaford Orogeny, spans the period 2300 - 2637 Ma., and attained upper amphibolite to granulite facies metamorphism. Dutton Suite granitoids were emplaced during the Sleaford Orogeny in the southwest of the Eyre Peninsula.

The Miltali Gneiss and other unnamed gneisses locally separate the Archaean Sleaford Complex from the overlying Palaeoproterozoic Hutchison Group. The Hutchison Group (= 1964 - 1845 Ma.) has been divided into three main sequences; the basal Warrow Quartzite, a mixed chemical and clastic sequence (Middleback Subgroup), and an upper pelitic unit (Yadnarie Schist). The depositional environment is interpreted to have been a shallow sea with a north-south trending shoreline with the periods of shoreline progradation and marine transgression.

The Kimban Orogeny (= 1845 to 1710) Ma. consisted of three major tectonic events. Peak metamorphic conditions during KM-D1,2 attained upper amphibolite facies followed by retrograde upper greenschist facies during KM-D3 (Parker, 1993a). The Wartakan Event (Thomson, 1969) is a later unrelated minor deformation event interpreted to have been related to emplacement of granitoids of the St. Peters Suite (1630 - 1620 Ma., Flint et al.,
1990). The Tarcoola formation (1656 ± 7 Ma.) in the north-west of the GRV province consists of a fluvial to marginal-marine clastic sequence. The Gawler range Volcanics (GRV) consists of an ellipsoidal mass of acid to intermediate variably welded ignimbrites with minor localised lava flows, domes and dykes exposed in the north-central Eyre Peninsula. Two phases of volcanic activity are recognised termed the 'developmental phase' and 'mature phase' by Stewart (1992). The Hiltaba Suite comprises anorogenic granitoids that have formed large batholiths and smaller plutons prevalent around the north-western, western and southern margins of the Gawler Ranges. A mean weighted age of 1589 ± 2 Ma. is obtained for the GRV and Hiltaba Suite (Creaser and Cooper, 1993).

The Corunna Conglomerate occurs as scattered outcrops around the margin of the GRV and is considered to have been deposited in a fluvial to shallow marine environment synchronous with emplacement of the GRV (Daly, 1993b). The Pandurra Formation, Blue range beds, and Gairdner dyke swarm all post date the GRV. The Pandurra Formation consists of redbed sediments of fluvial origin and is confined to the Cariewerloo Basin on the northern side of the Gawler Ranges (Cowley, 1993). The Blue Range Beds are confined to the Ikedoo Basin in the central Eyre Peninsula and consist of sandy conglomerates interpreted to have been deposited in a braided stream - alluvial fan environment (Flint, 1993a). The Gawler Craton was cratonised at about 1450 Ma. and has not been subject to significant deformation or remobilisation since, except for minor epirogenic movement and intrusion of the Gairdner Dyke Swarm around 1100 Ma.

Units of the Middleback Subgroup in the north-eastern Eyre Peninsula hosts iron ore bodies on which the Australian steel industry was founded over eighty years ago. In the northeast of the Gawler Craton the world class Olympic Dam Cu-Ag-U-Ag deposit is hosted by a hydrothermal breccia developed in a granite of the Hiltaba Suite. The Menninnie Dam Pb-Zn-Ag deposit is the most significant discovery of lead and zinc mineralisation in the Gawler Craton.
3.1 Introduction

For the purposes of this thesis, descriptions of the Menninnie Dam prospect geology have been subdivided into the Palaeoproterozoic and the Mesoproterozoic and are reviewed in Chapters three and four respectively. Chapter three incorporates descriptions of:

- Palaeoproterozoic stratigraphy
- Metamorphism
- Deformation
- Intrusions
- Pegmatite veins
- Metasomatism

A local north-south, east-west metric grid was established over the Menninnie Dam prospect by Billiton Australia Ltd. and all local features are referenced to this grid throughout this thesis (Fig. 3.1).

The prospect geology has been interpreted from relogging of all diamond drill holes drilled prior to 1996 (drill holes PD0001 to PD0032), and investigation of more than 300 petrographic sections. Unfortunately, attempts to orientate the drill core during drilling were mostly unsuccessful, and a detailed structural interpretation of the Menninnie Dam prospect is not possible. The orientation of individual rock units has been interpreted from their intersection in two or more drill holes, and on two or more cross sections. Summary drill hole logs and cross sections are presented in Appendix 3.1. An interpretative geological plan, and the 9400 mN cross section are presented as Figures 3.1 and 3.2 respectively. The geology could only be accurately constrained in the immediate vicinity of drill holes and where drill holes are widely spaced, interpretation of the geology has been based on interpolation between drill holes and adjacent east-west sections.

3.2 Palaeoproterozoic stratigraphy

The Menninnie Dam prospect occurs within a NNE-oriented linear band of magnetic rocks (approximately 5 km long by < 1 km wide), flanked by less magnetic rocks to the east and west (Fig. 3.3). Examination of drill core from the prospect has revealed that the magnetic response is due to magnetite-bearing marbles, calc-silicate lithologies and iron formations. The non-magnetic rocks to the east and west comprise schist and leucogranite.
Figure 3.1 Menninnie Dam prospect geology plan (schematic interpretation)
plan view at -65m R.L (approx -280m below ground level). Structure, alteration and mineralisation are not shown.
Local grid scale is in metres
Figure 3.2. Section 9400 mN. Schematic geological interpretation. Maximum forward and back extrapolation ± 100m. Drill hole ticks are 100m intervals.
Key to interpretative geology plan and 9400 mN cross section

Quaternary
- Surficial, transported clays and sand. Minor ferricrete and silcrete.

Mesoproterozoic
- Lithic rich Menninnie Dam ignimbrite.
- Discordant polymictic breccia
- Rhyolitic intrusions.

Palaeoproterozoic
- Granite dykes.
- Central suite. Carbonate - calc-silicate facies, and silicate facies iron formations.
- Central suite, dolomitic and calcitic marbles. Minor disseminated and chaotically banded magnetite.

Approximate maximum lateral extent of known mineralisation

Western, central and eastern suite boundaries

Horizontal projection of drill holes
- Prefix 'PD' = diamond drill hole with percussion precollar
- Prefix 'PE' = reverse circulation percussion drill hole
- Prefix 'RP' = rotary air blast percussion drill hole
Fig. 3.3. False colour, enhanced, 1:300,000 scale aeromagnetic image of the Wilcherry Hill exploration lease and surrounds. False sun angle is 090°. Units of highest magnetic intensity are coloured white; lowest are coloured purple. Prospects within the Wilcherry Hill exploration lease were intended to test small magnetic highs for Broken Hill-type Pb-Zn-Ag mineralisation (Higgins et al., 1990).
Subdivision of the schists, marble and calc-silicate lithologies into five suites by Higgins and Hellsten (1986) and Higgins et al. (1990) was based on distinct mineralogical and textural characteristics. Since the publication of Higgins et al. (1990), an additional 18 diamond drill holes have been drilled into the prospect, enabling further refinement of the stratigraphy. As a consequence, the original subdivisions have been discarded, and a new subdivision is proposed which divides the prospect geology into western, central and eastern suites. Correlation between the proposed subdivisions and those of Higgins and Hellsten (1986) and Higgins et al. (1990) is presented in Table 3.1.

**Table 3.1. Correlation between the proposed subdivisions and those of Higgins and Hellsten (1986) and Higgins et al. (1990)**

<table>
<thead>
<tr>
<th>Proposed subdivision (this study)</th>
<th>Subdivision of Higgins and Hellsten (1986) and Higgins et al. (1990)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Western suite</td>
<td>Suite 5</td>
</tr>
<tr>
<td>Central suite</td>
<td>Suites 3 and 4</td>
</tr>
<tr>
<td>Eastern suite</td>
<td>Suites 1 and 2</td>
</tr>
</tbody>
</table>

The Warrow Quartzite was interpreted by previous authors to occur to the immediate east of the prospect, directly underlying ‘suite 1’ rocks (Higgins and Hellsten, 1986; Higgins et al., 1990; Beeson, 1990). Warrow Quartzite is also shown in this region on the 1:250,000 scale Yardea geological sheet (Blissett et al., 1988). However, exposures that could be unequivocally assigned to the Warrow Quartzite could not be located within the vicinity of the prospect, and have not been intersected by drilling. Hellsten (1985) observed garnet in weathered surficial rubble to the east of the prospect that had previously been interpreted as Warrow Quartzite sub-crop. Hellsten (1985) suggested the rubble may have been weathered, silicified, impure marble or calc-silicate lithologies rather than Warrow Quartzite.

### 3.2.1 Western suite

The western suite correlates with ‘suite 5’ of Higgins et al. (1990). The western suite consists of variably sheared and retrograde schist that has been intruded by north-south trending granite dykes. Detailed descriptions of the metamorphism, granite dykes, and deformation are presented in Sections 3.3, 3.4 and 3.5 respectively.

Unaltered schist is preserved as clasts in a protocataclasite (Fig. 3.4a). Clasts consist of dark brown, medium to fine grained, quartz-feldspar-biotite-muscovite-sillimanite-garnet schist. Garnet is rare, constituting less than one percent of the rock. Garnet-rich layers are uncommon, although a one metre wide band with ~12% disseminated, 1-2 mm diameter, garnets was observed in drill hole PD0018. The garnet-rich band is orientated parallel to the metamorphic foliation.

A single 5 m wide zone of migmatisite occurs near the end of drill hole PD0023. Pale quartzofeldspathic segregations (‘neosome’; Mehnert, 1968) comprise ~30% of the rock and are separated by medium grey-green schist relics (‘mesosome’; Mehnert, 1968).
Fig. 3.4a. Drill hole PD0018, 256.65 m. Deformed western suite schist. Augen, so elongated clasts of schist (1) are separated by thin zones (2) of sericite and chlorite-altered, brecciated, schist matrix (10 - 15%). The schistose fabric of protolith clasts is variably orientated indicating clasts have been rotated during deformation. Grain size reduction resulted from brittle fragmentation, and the resultant texture is that of a protocataclasite.

Fig. 3.4b. Drill hole PD0027, 510.6 m. Quartz - diopside rock of the central suite. Quartz (Q) has formed bands and isolated domains in pale green diopside (D). The dark green fine bands near the right hand side of Fig. 3.4b is a fine grained mix of quartz and rod shaped diopside. At the left hand side of Fig. 3.4b, a 3 mm wide, contorted, dark green band of talc and tremolite (T) separates diopside - quartz rock from marble (M).

Fig. 3.4c. Drill hole PD0026, 599.1 m (upper) and 433.9 m (lower). Massive diopside of the central suite. 
Upper: light green diopside with dispersed inclusions of magnetite (M). Epidote (E) is interpreted to have formed during retrograde metamorphism.
Lower: light green diopside with minor retrograde minerals including chlorite (C) and garnet (G).

Fig. 3.4d. Drill hole PD0013, 153.5 m. Diopside - marble rock of the central suite. Chaotically banded light green diopside (D) and white marble (M) separated by thin, dark green rinds (T) of tremolite, talc ± serpentine, actinolite and magnetite. Dark flecks of magnetite are dispersed throughout the marble.

Fig. 3.4e. Drill hole PD0008, 277 m. Diopside - marble rock of the central suite. Whitish quartz (Q) is separated from magnetite and serpentine-bearing marble (M) by a halo of light green diopside (D) and dark green intergrown tremolite, talc and serpentine (T).

Fig. 3.4f. Drill hole PD0027, 483.4 m. Massive, white, dolomitic marble.

Fig. 3.4g. Drill hole PD0015, 509.6 m. Calcite marble of the central suite with thin biotite-rich metapelitic bands and dispersed flecks of biotite. Some of the metapelitic bands adjacent to the white and pink carbonate-bearing vein (V) have been partially altered to K-feldspar by metasomatic processes (eg. M).

Fig. 3.4h. Drill hole PD0007, 255.2 m. Rock type 5. Calcite marble with dispersed flecks of dark brown biotite, some of which is altered to dark green chlorite.
Most of the schist has been altered, with biotite altered to chlorite, and plagioclase feldspar altered to calcite and sericite.

A fine grained amphibolite unit was encountered near the end of drill hole PD0028. The amphibolite contains hornblende (= 38%), plagioclase (= 30%), biotite (= 15%), quartz (= 15%), apatite (= 1%) and magnetite (= 1%). Minerals have textures typical of high grade metamorphism (Turner, 1980), and a schistose fabric parallel to that of the enclosing schist. The amphibolite does not appear to have been affected by later tectonic deformation (Section 3.5) or by mineralogical changes associated with retrograde alteration (Section 3.3).

3.2.2 The central suite

The contact between the central and western suites has a north-south strike and dips steeply to the east. The central suite is characterised by marble and calc-silicate lithologies, iron formations and minor graphitic and meta-pelitic schists. Central suite rocks strike north-northeast and dip between 65° and 80° to the west (Figs. 3.1 and 3.2). The orientation of units has been deduced from drill hole intersections of probable primary sedimentary lithologies, such as iron formations. Iron formations (and by inference, other lithologies) have been deformed into open folds with a = 400 m wavelength and ENE trending fold axes (Fig. 3.1). The central suite is wedge-shaped in plan view and has a minimum present thickness of = 650 m in the north of the prospect, and is absent in the south of the prospect over a distance of = 4 km (Fig. 3.1).

Diopside is the most abundant mineral in calc-silicate lithologies. Tremolite is a minor component (< 5% of the total calc-silicate minerals). Grossular garnet and epidote (< 1%) locally occur in marble and calc-silicate lithologies. Marbles and calc-silicate lithologies can be subdivided into five rock types, based on variations in the ratios of dolomite or calcite, calc-silicate minerals, quartz, and other silicate minerals. These rock types are described in the following sections. Rock types 1 - 4, including the iron formations and the graphitic schists, correspond to suite 4 of Higgins et al. (1990).

1) Quartz - diopside rock

Quartz - diopside rock constitutes a few percent of the central suite. The rock is characterised by the presence of quartz with undulose extinction and diopside, and the absence of dolomite or calcite (Fig. 3.4b). Interlayered quartz and diopside bands are a few centimetres to several tens of centimetres thick and have irregular contacts. Dispersed magnetite (< 5%) occurs in some diopside bands. Retrograde mineral assemblages are rare or absent. Small rod-shaped diopside aggregates occur within many of the quartz-rich layers and commonly have long axes that are parallel to quartz - diopside banding, but appear to be discordant to nearby iron formations. Where quartz - diopside rock is in contact with
marble-bearing rock types, quartz-rich layers not occur in contact with marble and are invariably separated by an intervening layer of diopside ± tremolite ± talc (Fig. 3.4b).

The quartz - diopside rock cannot be easily correlated between drill holes or between adjacent sections in a manner that conforms to the interpreted orientation of stratigraphy (Fig. 3.1).

2) Massive diopside
Massive diopside is characterised by greater than 70% equigranular, medium to coarsely crystalline diopside (Fig. 3.4c). Magnetite (< 5%), quartz (< 10%), Ca - Mg carbonate (< 10%) and tremolite (< 5%) are accessory minerals in massive diopside units. Where present, magnetite occurs dispersed throughout diopside crystals (Fig. 3.4c), and as thin, highly contorted bands. Accessory quartz and Ca - Mg carbonate are invariably separated by a layer of tremolite or diopside.

Massive diopside units are commonly several metres, to several tens of metres thick, and are commonly discordant to stratigraphic layering. Massive diopside locally has an intense fracture pattern or jig-saw fit breccia texture with light brown garnet, epidote and hematite - dusted albite or microcline as the breccia matrix.

3) Diopside-marble
Diopside - marble rock is characterised by variable proportions of equigranular, medium to coarsely crystalline diopside and calcitic or dolomitic marble. Accessory minerals include tremolite, talc, serpentine, actinolite, magnetite and rare quartz. Diopside and marble bands of variable thickness (commonly < 5 cm) are complexly interlayered, resulting in a chaotic banded texture with orientations that vary over distances of a few centimetres (Fig. 3.4d, e). Diopside - marble rock is distributed throughout the central suite and appears to be discordant to stratigraphic layering (Fig. 3.1).

Diopside bands occur as irregularly - shaped lenses that vary in width from tens of centimetres to less than 1 centimetre. Diopside bands of ≥ one metre wide are classified as massive diopside. The contact between diopside and calcitic or dolomitic marble bands typically has an intervening dark green rind of silicate minerals, including tremolite, talc, actinolite, chlorite and serpentine (Fig. 3.4d). Magnetite is locally concentrated in this zone, and also occurs dispersed throughout marble bands, and to a lesser extent, diopside bands. Rock types 2 and 3 together comprise between 50 and 70% of the central suite.

4) Marble
This rock type consists of white, off - white, and pale grey, massive, coarsely crystalline dolomitic, and calcitic marble devoid of quartz (Fig. 3.4f). Drill intersections of marble range from 1 m to > 100 m. Accessory minerals constitute less than 10% and include
diopside, serpentine, talc, graphite and magnetite. When present, trace amounts of dispersed graphite impart a pale grey colour to the marble. Accessory minerals are more common in calcitic marbles compared to dolomitic marbles and calcitic marble commonly contain between 2 - 10% disseminated (Fig. 3.4g) or chaotically-banded magnetite.

5) Impure marble

The impure marble unit correlates with suite 3 of Higgins et al. (1990). This rock type occurs near the eastern extent of the central suite, bounded by iron formations* to the west, and by schists to the east. Impure marble is characterised by calcite with minor biotite, the common presence of thin pelitic schist bands, and little, or no diopside (Fig. 3.4h). Small aligned, biotite flakes (typically < 0.5 mm) occur dispersed throughout the impure marble, resulting in a mottled brownish-green colour. Schist bands comprise less than 10% of impure marble units and are typically 0.3 - 3 cm thick. Biotite (partially altered to chlorite) is the most common silicate mineral in schist bands with minor plagioclase, hornblende, quartz and accessory apatite.

Impure marble units are interpreted to be gradational between the marble, calc-silicate lithologies, and iron formations of the central suite and the meta-pelitic schists of the eastern suite. To the west, the impure marble unit is locally separated from iron formations by several metres of massive diopside rock or diopside-marble rock. However, where these rock types are absent (eg. drill holes PD0001 and PD0007; Appendix 3.1), the contact between the iron formations and impure marble is gradational over several metres (Fig. 3.5a). The gradation from iron formations to impure marble is marked by a progressive decrease in the thickness of iron formation bands from several tens of centimetres, down to less than one centimetre, and the common occurrence of a 1 - 5 mm wide hedenbergite + diopside + tremolite band between marble and iron formation bands.

Within 20 m of the eastern suite contact, the amount of dispersed biotite increases gradationally to around 20%, and schist bands become progressively more common and thicker, eventually grading into the eastern suite schists. Near the eastern suite contact, the quartz and feldspar content of schist bands increases, and bands > 0.2 m thick are mineralogically zoned with quartz and minor feldspar occurring in the western portions and biotite in the eastern portions (Fig. 3.5b). This is interpreted to be the metamorphosed equivalent of graded argillaceous sandstone beds and is used to propose an eastward younging direction.

A significant proportion of impure marble rock has a mottled pale pinkish and pale apple green colour interpreted to be a result of metasomatism. Mineral changes associated with metasomatism are described in Section 3.7.

* Iron formations are described in more detail on Page 3.6
**Iron formations**

In this study, iron formations are considered to be sedimentary units containing greater than 15 wt.% total iron in accordance with the definitions of James (1954) and Gross (1991). Packets of banded, iron-rich metasediments (iron formations) are a minor component of the central suite. Iron formations appear to be lenticular, with individual lenses ranging from several hundred metres to several kilometres in length (Fig. 3.1). Iron formations in the western half of the central suite are less than a few metres thick, whereas those in the eastern part of the central suite reach 20m in thickness.

Iron formations consist of 0.5 cm to 10 cm thick bands rich in iron silicate and oxide minerals including hastingsite / ferrohastingsite, grunerite, hedenbergite, biotite, magnetite ± almandine garnet, alternating with 0.3 to 5 cm thick layers of either quartz, marble, or diopside (Fig. 3.5c). The proportions of amphibole, pyroxene, biotite, magnetite and garnet vary widely, but not systematically, between iron formations, and in many iron formations, garnet and quartz are absent. Iron formations and the intervening quartz, diopside or marble bands have granoblastic textures (Turner, 1980), and a foliation formed by the preferred orientation of amphibole and biotite (Fig. 3.5d). Iron formations hosted by marble are commonly surrounded by a halo of calc-silicate minerals including hedenbergite, diopside, tremolite, garnet and epidote.

**Graphitic Units**

Minor graphitic schist units occur in the central suite (eg. drill holes PD0008 and PD0031; Appendix 3.1). Graphitic schists range from 0.5 - 10 m thick and are probably lenticular in shape, rarely exceeding a few hundred metres in maximum lateral extent. They consist of dark grey to black meta-pelitic schist with a graphite content between 5 and 10% (Fig. 3.5e). Traces amounts of pyrite, sphalerite and galena are locally present, and typically occur in thin veins and segregations that crosscut the metamorphic fabric (Fig. 3.5e). Sulphides textures are described in Chapter 5.

### 3.2.3 Eastern suite

The contact between the carbonate and calc-silicate lithologies of the central suite and the schists of the eastern suite is gradational over several metres. The boundary between the central and eastern suites is arbitrarily placed at the first occurrence of a schist band more than 5 m thick. The eastern suite consists of dark brown, biotite - quartz - feldspar - muscovite - sillimanite - almandine garnet schist with trace amounts of graphite and pyrite (Fig. 3.5f). Graphite and pyrite are part of the prograde mineral assemblage and commonly occur intergrown with garnet and biotite. Plagioclase has been partially altered to sericite and fine grained calcite and biotite has been partially altered to chlorite.
Fig. 3.5a. Drill hole PD0006, 382 m. The contact between iron formations and marble is gradational over several metres. Iron formation bands (IF) comprise intergrown hastingsite, biotite and magnetite, commonly with an outer rind of hedenbergite or diopside ± tremolite. Metapelitic bands (MP) comprise disseminated biotite with accessory magnetite, hornblende, and apatite. Biotite is commonly partially altered to dark green chlorite.

Fig. 3.5b. Drill hole PD0016, 400.6 m. The contact between impure marble and eastern suite schist is gradational and the metapelitic component increases towards the east. Some metapelitic bands are mineralogically zoned with quartz and feldspar-rich bases (B) and biotite-rich tops (T), interpreted as metamorphosed graded beds. Metapelitic bands are separated by calcitic marble with disseminated biotite (C).

Fig. 3.5c. Upper: drill hole PD0026, 524 m. Silicate facies iron formation, with bands of greyish quartz (Q) separated by darker bands (IF) of hastingsite / ferrohastingsite, grunerite, hedenbergite, biotite ± magnetite ± almandine garnet. Lower: Drill hole PD0006, 179 m. Carbonate facies iron formation with bands of whitish marble, disseminated biotite and hedenbergite (I), separated by dark bands of hastingsite / ferrohastingsite, grunerite, hedenbergite, biotite almandine garnet ± diopside ± magnetite (2).

Fig. 3.5d. Drill hole PD0016, 254.5 m. Transmitted light photomicrograph of a silicate facies iron formation band consisting of intergrown, aligned, biotite flakes (B), olivine green hastingsite (H), magnetite (M), garnet (G), and quartz (Q).

Fig. 3.5e. Drill hole PD0008, 192 m. Graphitic schist with cross-cutting veinlets of sphalerite, galena and pyrite (S). Graphitic schist has a metamorphic schistosity, whereas sulphide veinlets do not appear to have been metamorphosed.

Fig. 3.5f. Drill hole PD0006, 410.6 m. Transmitted light photomicrograph (x nicols) of eastern suite quartz - K-feldspar (K) - biotite (B) - plagioclase (P) - sericite - muscovite (M) - sillimanite - calcite - garnet (G) schist.
A 2 - 3 m thick, medium grained, hornblende amphibolite with trace amounts of quartz was intersected in the south of the prospect (ie. drill hole PD0031 at 525 m; Appendix 3.1). Lithologic contacts between the amphibolite and schist are sharp and parallel to the schistose fabric. The mineralogy and texture of the amphibolite is similar to the amphibolite intersected in the western suite schists (Section 3.2.1).

Calc-silicate, marble and graphitic schist units several metres thick occur as lenticular, to continuous, units within the eastern suite schist, particularly near the contact of the eastern and central suites and is interpreted as a transitional assemblage between the central and eastern suites. The calc-silicate and marble units within the eastern suite equate to 'suite 1' of Higgins and Hellsten (1986) and Higgins et al. (1990). Suite 1 was previously interpreted to directly overly the Warrow Quartzite (Higgins and Hellsten, 1986). However, subsequent drilling has shown that meta-pelitic schist of the eastern suite lie to the east of 'suite 1'. The thickness of the eastern suite schist is not known. A minimum thickness of 50 m has been defined by drilling although aero-magnetic data is interpreted to indicate a substantially greater thickness (Chapter 6).

3.3 Determination of the metamorphic history at Menninnie Dam

Higgins and Hellsten (1986) and Lee (1989) used mineral assemblages to interpret the host rocks and mineralisation at Menninnie Dam to have been subjected to amphibolite facies regional metamorphism (650 - 700°C; 3 - 5 kbars). A retrograde metamorphic event was also recognised. Parker (1993a) estimated similar peak metamorphic temperatures of 600 - 700°C, but higher pressures (5 - 8 kbars) for the central Eyre Peninsula.

In this section, the peak and retrograde metamorphic conditions that the Menninnie Dam host rocks were subjected have been established using the compositions of biotite, garnet and plagioclase for thermobarometric calculations. Comprehensive reviews of mineral thermobarometry have been presented by Ferry, (1982), Spear (1993) and Kretz, (1994).

3.3.1 Methods

A sample of schist from 410.6 m in drill hole PD0006 comprised the least altered or deformed* portion of eastern suite schist encountered during drilling at Menninnie Dam. Mineral abundances for this sample have been visually estimated at: 29% quartz, 27% K-feldspar, 22% biotite, 7% plagioclase, 5% sericite, 4% muscovite, 3% fibrous sillimanite, 2% calcite, 1% garnet, and trace amounts of chlorite, graphite, Fe-Ti oxides (limenite?), pyrite, apatite and zircon (Fig. 3.5f). Retrograde alteration is minimal, with partial replacement of plagioclase by a fine grained mixture of sericite, calcite and quartz, local alteration of biotite to chlorite and minor alteration of sillimanite margins.

* Deformation and alteration of lithologies at Menninnie Dam are introduced in Section 3.5.
The bulk rock composition was determined by XRF analysis at the Geology Department, University of Tasmania. Mineral compositions were determined using a Cameca SX50 electron microprobe in the Central Science Laboratory, University of Tasmania under the supervision of W. Jablonski. Details of the procedures and machine settings for XRF analysis and microprobe analysis are included in Appendices 3.2.

Garnet porphyroblasts were analysed at spacings of $=50 \mu\text{m}$ along a line bisecting the garnet using a beam diameter of 10 $\mu\text{m}$. The outer-most spot analysed was situated as close as possible to the edge of the garnet. Eight additional analyses from within 20 $\mu\text{m}$ of the margins of garnet and biotite crystals in mutual contact were also obtained.

Only visually unaltered biotite crystals were selected for microprobe analysis. Biotite was analysed with a beam current of 10 nA, to avoid volatilisation of lighter elements. In addition to the eight biotite microprobe analyses mentioned above, three analyses were obtained from the cores of large biotites crystals well away from contacts with other Fe or Mg bearing minerals.

Plagioclase compositions are required for a number of geobarometers. Unaltered plagioclase crystals were analysed from one margin to the next at $=50 \mu\text{m}$ intervals in order to examine element zoning. K-feldspar was analysed in order to compare the composition with that of plagioclase.

### 3.3.2 Analytical results

Results of XRF analyses are presented in Table 3.2. This sample of the eastern suite schist has a high potassium content, which is consistent with the high potassium content for most rock suites at Menninnie Dam (Beeson, 1996b).

<table>
<thead>
<tr>
<th>Major elements (wt.%)</th>
<th>Trace elements (ppm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO$_2$</td>
<td>57.46</td>
</tr>
<tr>
<td>TiO$_2$</td>
<td>0.78</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>17.83</td>
</tr>
<tr>
<td>Fe$_2$O$_3^*$</td>
<td>9.51</td>
</tr>
<tr>
<td>MnO</td>
<td>0.39</td>
</tr>
<tr>
<td>MgO</td>
<td>3.08</td>
</tr>
<tr>
<td>CaO</td>
<td>0.79</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>0.89</td>
</tr>
<tr>
<td>K$_2$O</td>
<td>5.55</td>
</tr>
<tr>
<td>FeO</td>
<td>0.14</td>
</tr>
<tr>
<td>LOI</td>
<td>2.58</td>
</tr>
<tr>
<td>Total</td>
<td>99.05</td>
</tr>
</tbody>
</table>

*total iron is calculated as Fe$_2$O$_3$*
Results of garnet, biotite, K-feldspar and plagioclase electron microprobe analyses are summarised in Tables 3.3 - 3.7 and detailed in Appendix 3.3. Element variations associated with zoning in garnets are illustrated in Figures 3.6a - d.

Table 3.3 Compositional range of 75 garnet analyses (wt. %). Complete analyses are listed in Appendix 3.3.

<table>
<thead>
<tr>
<th>Element oxide</th>
<th>Range (wt.%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CaO</td>
<td>0.912 - 4.843</td>
</tr>
<tr>
<td>MgO</td>
<td>1.125 - 2.764</td>
</tr>
<tr>
<td>FeO</td>
<td>25.667 - 29.270</td>
</tr>
<tr>
<td>MnO</td>
<td>7.632 - 12.814</td>
</tr>
</tbody>
</table>

*Garnet compositions have been recalculated on the basis of 8 cations and 12 oxygens using MinTab version 1.1 (Rock and Coroll, 1990), which uses the procedures of Droop (1987). Iron, analysed as FeO, was not recalculated as FeO + Fe₂O₃ as Fe³⁺ is unlikely to have been significant in this graphite-bearing assemblage. The mean garnet end member composition was calculated as: 8.5% andradite, 61.6% almandine, 8.5% pyrope, 21.4% spessartine.

Manganese concentrations are elevated in the margins and weakly elevated in the core of garnets (Fig. 3.6d). Almandine garnets with elevated Mn concentrations in their cores are common in many metapelites (Turner, 1980; Spear, 1993) due to the high affinity of garnet for Mn²⁺ (Sturt, 1962). Later overgrowths of garnet resulted in isolation of the garnet core, and the higher Mn signature of the core was retained due to unfavourable diffusion kinetics (Turner, 1980). Decreased Fe, Ca and Mg concentrations in garnet margins (Figs. 3.6a - c) are interpreted to have resulted from depletion of Fe, Ca and Mg and their preferential inclusion in other minerals such as biotite and plagioclase during a decompressional cooling path. The corresponding rise in Mn content at garnet margins (Fig. 3.6d) is a result of Fe, Ca and Mg consumption during this stage.

Table 3.4. Average biotite analyses (wt. %). N = number of analyses. The composition of biotite cores is similar to that of biotite rims. Complete analyses are listed in Appendix 3.3.

<table>
<thead>
<tr>
<th></th>
<th>MgO</th>
<th>Al₂O₃</th>
<th>SiO₂</th>
<th>K₂O</th>
<th>TiO₂</th>
<th>MnO</th>
<th>FeO</th>
</tr>
</thead>
<tbody>
<tr>
<td>Biotite margins</td>
<td>7.89</td>
<td>19.74</td>
<td>33.85</td>
<td>9.36</td>
<td>2.52</td>
<td>0.31</td>
<td>19.65</td>
</tr>
<tr>
<td>Biotite cores</td>
<td>7.68</td>
<td>19.65</td>
<td>34.00</td>
<td>9.64</td>
<td>2.34</td>
<td>0.36</td>
<td>19.77</td>
</tr>
</tbody>
</table>

Table 3.5 Average analyses (wt.%) of K-feldspar and plagioclase. N = number of analyses. Complete analyses are listed in Appendix 3.3.

<table>
<thead>
<tr>
<th></th>
<th>Na₂O</th>
<th>Al₂O₃</th>
<th>SiO₂</th>
<th>K₂O</th>
<th>CaO</th>
</tr>
</thead>
<tbody>
<tr>
<td>K-feldspar</td>
<td>9.12</td>
<td>19.00</td>
<td>62.94</td>
<td>14.86</td>
<td>0.03</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>8.75</td>
<td>25.24</td>
<td>59.1</td>
<td>0.31</td>
<td>6.32</td>
</tr>
</tbody>
</table>

Element zoning in K-feldspar, plagioclase or biotite was not detected and may have resulted from metamorphic homogenisation.
Fig. 3.6. Oxide variations across a zoned almandine garnet (garnet 1A; Appendix 3.3). Refer to the text for interpretation.
3.3.3 Thermobarometric calculations

Frost and Tracy (1991) outline minimum criteria for calculation of P-T-time paths from zoned garnets, which include:

1) The equilibrium mineral assemblage or assemblages present at all stages of garnet growth must be known.
2) The changes in composition of phases participating in the garnet-producing reaction, particularly biotite and plagioclase, must be known.
3) Any modification of growth zoning by post-crystallisation diffusive processes must be understood.
4) Any P-T changes must be corroborated by textural or structural evidence.

Garnets in the eastern suite schist commonly contain inclusions of other minerals (Figs. 3.7a, b). Quartz and biotite are present as inclusions throughout garnet crystals. Muscovite inclusions have only been observed in garnet cores. Fe-Ti-oxides (ilmenite?) are disseminated throughout garnet crystals, commonly in association with an unknown fine grained, high relief, mineral with high interference colours (sphene?). Trace amounts of graphite are also present. Rare, fine-grained mica ± chlorite pseudomorphs of an unknown mineral are present in the mid zones of some garnets. Rare inclusions of feldspar have been observed but are not present in sufficient quantities to determine their distribution. Sillimanite inclusions have not been observed in garnet.

As biotite inclusions are present at all stages of garnet crystal growth, determination of the temperature range over which metamorphism occurred can be attempted using biotite - garnet geothermometers (eg. Ferry and Spear, 1978; Berman, 1990) providing the composition of biotite is known at all stages of garnet growth. However, due to the small number of biotite inclusions in garnet crystals, their compositions have not been systematically analysed. An approximation of the biotite composition can be estimated based on the following argument:

The amount of garnet in eastern suite schists (= 1 wt.%) is significantly less than that of biotite (= 25 wt.%) and the diffusion rate in garnet is slow (Turner 1980). Therefore, assuming a closed system in which the abundance of other Fe-Mg minerals was insignificant (probably reasonable for peak and post-peak metamorphic conditions), any changes in the bulk iron or magnesium content of biotite due to changes in P-T conditions are likely to have been insignificant compared to changes preserved in garnet. Analyses of biotite cores are similar to those of biotite in contact with garnet (Table 3.4), probably indicative of homogenisation during peak and retrograde metamorphism. As there is little variation in biotite compositions, an average composition must be assumed.

The composition of garnets in the eastern suite schist are significantly higher in Mn and Ca than those used by Ferry and Spear (1978) to develop their garnet - biotite geothermometer.
Fig. 3.7a, b. Drill hole PD0006, 140 m. Fig 3.7a Transmitted light photomicrograph, Fig. 3.7b reflected light photomicrograph. Garnets contain inclusions of quartz, biotite, muscovite, rare plagioclase, oxides (ilmenite?), sphene, mica ± chlorite pseudomorphs of an unknown precursor, and trace amounts of graphite. Not all mineral inclusions are present in all garnets. The garnet shown, contains inclusions of quartz (Q), biotite (B), oxides (FT), and partially altered plagioclase (P).
Berman (1990) developed a mixing model for quaternary Ca-Mg-Fe-Mn garnets intended to compensate for the effects of Mn and Ca. The calibrations of Ferry and Spear (1978) and Berman (1990) have been combined in the computer program by Spear and Peacock (1990) and used for this study.

Quartz is present as inclusions throughout garnet crystals and is interpreted to have been stable throughout the metamorphic event. Preservation of muscovite inclusions in the cores of garnet crystals and the common occurrence of K-feldspar throughout the rock is consistent with reaction between muscovite and quartz to form sillimanite and K-feldspar at higher metamorphic grades (Turner, 1980). Geobarometers involving muscovite are therefore unlikely to result in realistic pressure estimates. Geobarometers involving sillimanite are likely to be applicable to peak metamorphism only.

Plagioclase is likely to have been a stable phase above upper greenschist facies (Turner, 1980) although its composition is likely to have varied throughout the metamorphic event. As plagioclase comprises only ~7% of the rock and has an average composition of ~An31, use of an average composition as representative of all stages of metamorphic growth is unlikely to be valid. Application of average plagioclase compositions for geobarometric calculations commonly results in an over estimate of the pressure conditions for all but the final stages of metamorphism. This is due to an increase in the calcium content of garnet and a corresponding decrease in the calcium content of plagioclase at higher metamorphic conditions (Frost and Tracey, 1991; Figs. 3.8a, b). If the plagioclase is assumed to comprise An100 (ie. retained all of its calcium) then a lower limit of peak pressure and temperature conditions can be calculated (Figs. 3.8c, d).

Garnet, biotite, quartz and plagioclase are likely to have been stable phases throughout the period of garnet growth, and with the caveat of assumed compositions for biotite and plagioclase, criteria 1 - 3 of Frost and Tracey (1991) are satisfied within the limitations of these data. A garnet which had the most symmetrical element zonation pattern was chosen for thermobarometric calculations (Figs. 3.6a - d; Appendix 3.3, 'garnet 1A'). Garnet compositions with the highest calcium contents have been used for determination of peak metamorphic conditions using the biotite - garnet geothermometer of Ferry and Spear (1978), and Berman (1990) in combination with the garnet - plagioclase - sillimanite - quartz geobarometer of Koziol (1989) and the garnet - plagioclase - quartz - biotite geobarometer of Hoisch (1990). Garnet and biotite margins in mutual contact have been used to calculate the final metamorphic conditions at which these minerals attained equilibrium using an average biotite composition and the biotite - garnet geothermometer of Ferry and Spear (1978), and Berman (1990), and the garnet - plagioclase - quartz - biotite geobarometer of Hoisch (1990). Results are illustrated in Figure 3.8e and summarised in Table 3.6.
Fig. 3.8. Pressure vs temperature plots for calculated peak (A - D) and final (E) metamorphic conditions at Menninnie Dam using the garnet biotite geothermometer calibrations of Ferry and Spear (1978) and Berman (1990) and various geobarometers and plagioclase compositions. A) Garnet - biotite - quartz - plagioclase geobarometer (Hoisch, 1990) and plagioclase composition of An31. B) Garnet - sillimanite - quartz - plagioclase geobarometer (Kozikol, 1989) and plagioclase composition of An31. C) Garnet - biotite - quartz - plagioclase geobarometer (Hoisch, 1990) and plagioclase composition of An100. D) Garnet - sillimanite - quartz - plagioclase geobarometer (Kozikol, 1989) and plagioclase composition of An100. Garnet - biotite - quartz - plagioclase geobarometer (Hoisch, 1990) and plagioclase composition of An31.
Table 3.6. Summary of calculated peak and final metamorphic conditions.

<table>
<thead>
<tr>
<th>Metamorphism</th>
<th>Thermometer</th>
<th>Barometer</th>
<th>Plagioclase</th>
<th>Temperature (°C)</th>
<th>Pressure (kbar)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Peak</td>
<td>Gnt - bio</td>
<td>Gnt - plag - qz - bio</td>
<td>An₃₁</td>
<td>650 - 700</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>Gnt - bio</td>
<td>Gnt - plag - sill - qz</td>
<td>An₃₁</td>
<td>~700</td>
<td>10</td>
</tr>
<tr>
<td></td>
<td>Gnt - bio</td>
<td>Gnt - plag - qz - bio</td>
<td>An₃₁₀₀</td>
<td>=650</td>
<td>6</td>
</tr>
<tr>
<td>Final</td>
<td>Gnt - bio</td>
<td>Gnt - plag - qz - bio</td>
<td>An₃₁</td>
<td>500 - 600</td>
<td>1 - 3</td>
</tr>
</tbody>
</table>

In accordance with criteria 4 of Frost and Tracy (1991), the application of this technique can be tested by comparison between calculated values, mineral textures and previously determined P - T conditions. Calculations of peak metamorphic conditions that include plagioclase = An₃₁, result in pressure estimates that are inconsistent with the presence of sillimanite. More realistic estimates are obtained by constraining metamorphic conditions to the sillimanite stability field and using geobarometer calculations that include plagioclase = An₃₁₀₀ to give minimum pressure. The final metamorphic conditions at which biotite and garnet attained equilibrium are near the sillimanite - andalusite boundary, or, within the andalusite field. Estimates near the sillimanite - boundary are more consistent with the absence of andalusite, however, the absence of andalusite at lower pressures and temperatures probably reflects the bulk rock composition (Table 3.2) which is too K-rich for Al₂SiO₅ stability, except above the quartz - muscovite breakdown reaction.

Estimated peak and final pressure and temperature conditions for rocks at Menninnie Dam are illustrated as fields in pressure - temperature space (Fig. 3.9) with inclusion of the geobarometer error limits estimated by Kohn and Spear, (1991a, b), and the geothermometer error limits estimated by Ferry and Spear (1978). Peak metamorphic conditions calculated for Menninnie Dam rocks lie to the left of the quartz - muscovite breakdown curve estimated by Spear and Cheney (1989; Fig. 3.9). The occurrence of sillimanite and K-feldspar is interpreted to indicate that the activity of water was less than 1.0 (Spear and Cheney 1989). Relatively dry conditions are consistent with calculated pressure and temperature conditions which lie to the right of the minimum melting curve for common granite (wet solidus; after Turner, 1980).

Peak pressure and temperature conditions calculated for Menninnie Dam (5 - 7 kbar and 600 - 700°C) are lower than those estimated for KM-D₁₂ in the southern Eyre Peninsula (> 8 kbar and 800 - 900°C; Mortimer et al., 1980; Bradley, 1980; Fig. 3.9) but comparable to KM-D₁₂ estimated for the Cowell Cleve area of the central Eyre Peninsula (7 kbar and 700°C; Parker, 1978; Fig. 3.9). Parker (1993a) described KM-D₃ as a retrograde metamorphic event (KM-D₃ = 1.5 - 2 kbars and 450 - 550°C), with pressure and temperature conditions that peaked below the andalusite - sillimanite - kyanite triple point (Fig. 3.9).
Figure 3.9  Metamorphic grade of the Kimban Orogeny events on the eastern Eyre Peninsula (modified from Parker, 1993a). Higher metamorphic conditions during KM-D1-2 are recorded for the southern Eyre Peninsula compared to the central Eyre Peninsula. Peak and retrograde metamorphic conditions at Menninnie Dam overlap those estimated for the central Eyre Peninsula. Al$_2$SiO$_5$ polymorph curves after Holdaway, (1971), lower stability limit for biotite after Winkler, (1976), lower stability limit for amphibine after Wügler, (1976), stability field for bulk Fe(FeO + MgO) ratio of 0.6 (cordierite + garnet) after Currie, (1971). Muscovite - quartz - aluminium silicate - K-feldspar curve after Spear and Cheney (1989). Minimum melting point of common granite (wet solidus) after Turner (1980).
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Calculated pressure and temperature conditions for the final metamorphic event at Menninnie Dam (1 - 3 kbars and 450 - 650°C) are similar to those estimated by Parker (1988) for KM-D3. The Menninnie Dam host lithologies have undergone a similar metamorphic history to that of the Central Eyre Peninsula.

3.4 Granite dykes

The western, central and eastern suites at Menninnie Dam have been intruded by numerous granite dykes (MD granite dykes). MD granite dykes are more common in the western suite schists (= 40%) than in either the central or eastern suites (= 5%). Beeson (1990b) noted the occurrence of granitoids at Menninnie Dam but did not resolve whether they were intrusions or "metamorphosed tuffaceous units". Scattered exposures of granite of the Lincoln Complex occur = 5 km to the east of the Menninnie Dam prospect (Blissett et al., 1988). This granite was interpreted by Parker (1993a) to be a late, syntectonic granite, associated with KM-D3 (Moody Suite correlative, Fig. 3.10).

The aim of this section is to:

- Describe the morphology and mineralogy of the MD granite dykes.
- Compare the major and trace element geochemistry of the MD granite dykes with those of other granite bodies from the Eyre Peninsula.

3.4.1 MD granite dyke description

The width of MD granite dykes vary in drill intersection from 10 cm to in excess of 60 m (Figs. 3.1, 3.2; Appendix 3.1). MD granite dykes within the western suite schists strike north - south and have been traced through drilling over distances ~ 2 km whereas those in the central and eastern suites cannot be easily correlated between sections spaced 200 m apart. MD granite dykes in the central suite locally crosscut iron formations, typically at angles of 5 - 30°, which is considered to be evidence for an intrusive origin.

MD granite dykes in the eastern, central and western suites have similar textures and mineralogy. They do not have chilled margins and comprise uniform, medium-grained, leucogranites with less than 5% mafic minerals. The granite dykes are composed of 20 - 35% quartz (0.5 - 3 mm), 30 - 65% orthoclase and microcline (0.5 - 5 mm), 5 - 20% albite (0.1 - 2 mm), 1 - 7% muscovite (< 2 mm), < 2% biotite (< 0.5 mm), < 2% chlorite, < 2% calcite, < 3% sericite, and trace amounts of zircon, apatite, hematite, magnetite, leucoxene, pyrite, chalcopyrite and sphalerite.

Two types of quartz have been recognised. Quartz is present as 20 - 200 μm rounded inclusions within orthoclase and, rarely, albite. Quartz also occurs as discrete grains or clusters up to several mm in diameter interstitial to feldspar crystals. Poikilitic orthoclase contains inclusions of quartz, albite and biotite.
Fig. 3.10 Distribution of Lincoln Complex granitoids and mylonite zones on the eastern Eyre Peninsula (modified after Parker, 1993b).
Orthoclase also occurs as non-poikilitic euhedral crystals that locally have graphic intergrowths with quartz. Most of the orthoclase is microperthitic and some orthoclase has partially reverted to microcline. Muscovite and biotite occur interstitial to large quartz and feldspar crystals. Apatite occurs as small sub-rounded crystals within quartz and feldspar, and as fine euhedral needles at the boundaries of other minerals. Rare, sub-rounded, zoned zircon crystals were probably derived from an earlier source. Chlorite, calcite and sericite are alteration products of biotite, albite and K-feldspar respectively. Every MD granite dyke examined contains evidence of varying degrees of deformation. Deformation textures are described in Section 3.5.

3.4.2 Granite dyke geochemistry

MD granite dykes were analysed for major and selected trace element compositions to enable geochemical classification, and to attempt correlations with granites occurring elsewhere in the Eyre Peninsula (eg. the Lincoln Complex; 1845 - 1710 Ma., Parker, 1993a; St. Peters Suite; 1630 - 1620 Ma.; Flint et al., 1990; Hiltaba Suite; 1589 Ma., Creaser and Cooper, 1993). Approximately one kilogram portions of those MD granite dykes that had undergone minimal hydrothermal alteration were selected for geochemical analyses. Major and trace elements were analysed using a Phillips PW1480 X-ray fluorescence (XRF) spectrometer at the Geology Department, University of Tasmania. The granites were analysed for, SiO₂, TiO₂, Al₂O₃, Fe₂O₃, MnO, MgO, CaO, Na₂O, K₂O, P₂O₅, loss on ignition (L.O.I.), Y, Rb, Cu, Pb, Zn, Ni, U, Th, Nb, Zr, Sr, Ba, Sc, V. Total iron is expressed as Fe₂O₃. Preparation methods and XRF settings are presented in Appendix 3.2.

Analytical results

MD granite dyke analyses, analyses of granitoids and associated mafic intrusions of the Lincoln Complex, granitoids and granodiorite dykes of the St Peter Suite (Parker, 1993a), and granitoids of the Hiltaba Suite (Flint, 1993b), are summarised in Table 3.7 and detailed in Appendix 3.4.

MD granite dykes plot in the granite field defined by Le Maitre (1989; Fig. 3.11). Their chemistry, mineralogy and textures are consistent with crystallisation from a melt rather than pegmatite, or as products of metamorphic reactions. The granite dyke compositions plot as a cluster on Al₂O₃ vs SiO₂, TiO₂ vs SiO₂, TiO₂/Nb vs SiO₂ and Y vs TiO₂ variation diagrams (Fig. 3.12a - f), interpreted to indicate that the MD granite dykes were co-genetic, and probably coeval. MD granite dykes can be differentiated from granitoids of the Hiltaba Suite on P₂O₅ vs SiO₂ and TiO₂ vs SiO₂ variation diagrams (Fig. 3.12e, f). At similar SiO₂ levels, MD granite dykes are enriched in P₂O₅ and depleted in TiO₂ relative to the granitoids of the Hiltaba Suite. Insufficient analyses of granitoids of the St. Peters Suite are available to test for differences to MD granite dykes.
<table>
<thead>
<tr>
<th>Granite dykes</th>
<th>St Peters</th>
<th>Lincoln</th>
<th>Hiltaba</th>
<th>Hiltaba</th>
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<tr>
<td>Composition range</td>
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<td>Granite</td>
<td>Granitite</td>
<td>Granite</td>
</tr>
<tr>
<td>SiO₂ (wt.%)</td>
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<td>74.68</td>
<td>74.68</td>
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<td>≥TiO₂ (wt.%)</td>
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<td>0.34</td>
<td>0.20</td>
</tr>
<tr>
<td>Al₂O₃ (wt.%)</td>
<td>13.33 - 16.38</td>
<td>14.25</td>
<td>13.34</td>
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<tr>
<td>Fe₂O₃ (wt.%)</td>
<td>0.67 - 4.28</td>
<td>1.64</td>
<td>2.29</td>
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<tr>
<td>MgO (wt.%)</td>
<td>0.01 - 0.09</td>
<td>0.05</td>
<td>0.07</td>
<td>0.03</td>
</tr>
<tr>
<td>CaO (wt.%)</td>
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<tr>
<td>Na₂O (wt.%)</td>
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<td>0.91</td>
<td>1.61</td>
<td>0.74</td>
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<tr>
<td>K₂O (wt.%)</td>
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<td>3.88</td>
<td>3.42</td>
</tr>
<tr>
<td>P₂O₅ (wt.%)</td>
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<td>4.13</td>
<td>5.33</td>
</tr>
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<td>-</td>
<td>-</td>
<td>0.10</td>
</tr>
<tr>
<td>H₂O² (wt.%)</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>CO₂ (wt.%)</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
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<tr>
<td>L.O.I. (wt.%)</td>
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<td>0.99</td>
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<td>-</td>
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<tr>
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<td>201.38</td>
<td>179</td>
<td>420</td>
</tr>
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<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Sr (ppm)</td>
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<td>89.06</td>
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<td>50</td>
</tr>
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<td>11</td>
<td>36</td>
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<td>8</td>
</tr>
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<td>20</td>
</tr>
<tr>
<td>W (ppm)</td>
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<td>-</td>
<td>-</td>
<td>10</td>
</tr>
<tr>
<td>Y (ppm)</td>
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<td>15.50</td>
<td>31</td>
<td>34</td>
</tr>
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<td>Zn (ppm)</td>
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<td>58</td>
<td>80</td>
</tr>
<tr>
<td>Zr (ppm)</td>
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<td>190</td>
<td>170</td>
</tr>
<tr>
<td>Number of analyses</td>
<td>15</td>
<td>15</td>
<td>4</td>
<td>2</td>
</tr>
</tbody>
</table>

Table 3.7: Range and average analyses of the granite dykes from Menninnie Dam, selected analyses of granites of the Hiltaba Suite (Flint, 1993b) and Lincoln Complex (Moody Suite; Parker, 1993a) that occur close to the Menninnie Dam prospect, and a granite of the St. Peters Suite (Parker, 1993a).
Figure 3.11 Plutonic igneous rock classification schema of Le Maitre (1989). Assuming albite is altered calcic plagioclase, data for the granite dykes at Menninnie Dam plot in the granite field.
Figures 3.12A-G. Variation diagrams for Granites of the Eyre Peninsula and the granite dykes from Menninnie Dam. MD granite dykes plot as a cluster, consistent with a co-genetic and possibly coeval relationship. The MD granite dykes can be differentiated from other granite suites of the Gawler Craton by their higher P2O5 contents and lower TiO2 contents at equivalent SiO2 contents. Data for the MD granite dykes plot in the field for peraluminous granitoids (Shand, 1927).
Granitoids of the Lincoln Complex have been subdivided into:

- Donington Granitoid Suite associated with KM-D1 (≈ 1843 - 1818 Ma., Mortimer et al., 1986; Mortimer et al., 1988; Webb et al., 1986).
- Colbert Suite and Middle Camp Granite associated with KM-D2 (≈ 1757 - 1738 Ma., Mortimer et al., 1986; Fanning, 1987).

The MD granite dykes and granitoids of the Moody Suite both have high P2O5 contents (Fig. 3.12e). The MD granite dyke compositions have a positive correlation between P2O5 and SiO2. The correlation between P2O5 and SiO2 does not appear to be due to alteration and when loss on ignition (L.O.I.) is excluded and the granite dyke compositions recalculated, the positive correlation between P2O5 and SiO2 remains. The granite dykes are strongly peraluminous (Fig. 3.12g), with an average aluminium saturation index of 1.50 (The aluminium saturation index = Al2O3 / (CaO + Na2O + K2O); Shand 1927). The correlation between P2O5 and SiO2 is interpreted to have been caused by the magma not having achieved apatite saturation prior to emplacement as dykes. Apatite solubility in a melt increases linearly with the aluminium saturation index, although for many L.C.T. granites, feldspar is the principal reservoir for phosphorous (London, 1995; L.C.T. granites are defined as granites with Li, Cs and Ta as their signature elements, Cerny, 1991 a, b).

3.5 Deformation

The aim of this section is to:

- Describe the deformation textures of the MD granite dykes and metasediments at Menninnie Dam.
- Compare the effects of deformation at Menninnie Dam with the effects of regional tectonic events.

The textural classification of fault rocks adopted in this study follows that of Sibson (1977; Fig. 3.13). The definition of mylonite is modified from that proposed by Sibson (1977) to included the criteria of White et al. (1980), who defined mylonite as the rock produced in a ductile shear zone or fault in which the rock matrix was deformed by crystal-plastic processes. In addition, Lister and Snoke (1984) considered the presence of strong crystallographic orientations (e.g. quartz c-axis fabrics) to be important in the definition of mylonites. While cataclasis may occur within mylonites with subsidiary brittle deformation of harder mineral phases such as feldspar (Lister and Snoke, 1984), the dominant deformation process in mylonites is by crystal-plastic processes.
Figure 3.13 Textural classification of fault rocks (modified after Sibson, 1977).
Deformation textures in each of the rock suites will be described in turn, starting with the western suite. It was not possible to conduct a structural analysis of lithologies at the Menninnie Dam prospect due to the absence of orientated drill core or a consistent 'marker fabric' such as cleavage which could be used as a reference plane for structural studies.

3.5.1 Deformation of the western suite
The western suite was recognised as being problematic by Higgins et al. (1990) and Beeson (1990b), in that it did not easily correlate with the Hutchison Group stratigraphy as proposed by Parker et al. (1985; Fig 2.3). Drilling carried out after 1990, has confirmed the north-south trend, and steep easterly dip of the western suite - central suite contact. This contrasts with the steep westerly dip and NNE strike of the central and eastern suites. The discordant boundary between the western and the central suites is therefore interpreted to be a major north-south trending shear zone.

The western suite is cut by a number of sub-parallel faults which collectively form a deformation zone several hundred metres wide. Deformation styles form a continuum from brittle, to intense ductile deformation, which has resulted in substantial changes to rock textures and, to a lesser extent, mineralogy. In most of the deformed schist, the schistose protolith is easily identifiable and the texture is typical of a protocataclasite (Fig. 3.4a). Increased deformation was associated with retrograde alteration of biotite and feldspar to sericite and chlorite, which has produced dark green, or yellow green schists (Fig. 3.14a).

Intense deformation is typically confined to zones up to a few tens of meters wide in drill intersection. Deformation is heterogeneous within these zones, with less deformed clasts in a fine grained, highly deformed matrix. Matrix material, which comprises = 10 - 15% of deformed zones, consists of fine grained quartz and rare feldspar fragments, and abundant chlorite and sericite. The matrix formed mostly by brittle processes, resulting in grain size reduction of quartz and, to lesser extent, feldspar. The resultant texture is that of a cataclasite (Fig. 3.14a).

Apart from local zones of cataclastic deformation, most of the schists have protocataclastic textures. By contrast, granite dykes in the western suite have a wider range of deformation textures, from weakly deformed granite to mylonite. The interior of the granite dykes is typically weakly deformed, with more intense deformation concentrated in a narrow band within dykes, near the contact with the enclosing schist (typically less than 1m wide). It appears that strain has been partitioned into granite dyke margins, rather than the enclosing schist (Fig. 3.14b). Intense deformation is invariably associated with retrograde alteration of feldspar and biotite to sericite ± chlorite ± pyrite, resulting in a characteristic light yellow green colour.
Fig. 3.14a. Drill hole PD0021, 331.4 m. Deformed western suite schist with a cataclastic texture. Biotite and feldspar have been altered to sericite and chlorite, resulting in a yellow-green or dark green colour.

Fig. 3.14b. Drill hole PD0027, 179.8 m. Deformed margin of a granite dyke intruded into the shear zone bounding the central and western suites. Deformation and alteration are most intense at the outer margin of the dyke (RHS). In more deformed zones, feldspar is altered to yellow-green sericite. Towards the interior of granite dyke (LHS), deformation is less intense.

Figs. 3.14c–e. are from drill hole PD0027 at = 179 m. They illustrate the progressive textural and mineralogical changes that have occurred over a 2 m interval near the margin of a granite dyke. Fig. 3.14c is from towards the interior of the granite dyke and is least deformed, and Fig. 3.14e is near the granite dyke margin and is most deformed.

Fig. 3.14c. Transmitted light photomicrograph (x nicols). = 2 m from the granite dyke margin. Less deformed granite has a sub-gneissic foliation. Quartz crystals in contact with each other (Q) have serrated grain contacts typical of grain margin migration.

Fig. 3.14d. Transmitted light photomicrograph (x nicols). = 20 cm from the granite dyke margin. More deformed granite has a higher proportion of matrix and ductile processes are dominant. Most of the larger porphyroclasts are comprised of feldspar (F) of < 8 mm diameter, although quartz + feldspar + muscovite clasts, and rare quartz are locally present. The matrix is dominated by quartz ribbons (Q), with minor feldspar and muscovite (M) fragments, and interstitial fine grained sericite and calcite (C). Most tabular-elongate quartz grains and quartz ribbons have a consistent crystallographic orientation, typical of dynamic recrystallisation. Pressure shadows of fine grained quartz and mica are common in the lee of porphyroclasts (P). Quartz in pressure shadows has recrystallised with polygonal grain boundaries and no consistent crystallographic orientation.

Fig. 3.14e. Transmitted light photomicrograph (x nicols). = 5 cm from the granite dyke margin. The most intensely deformed granite is fine grained and contains pervasive sericite alteration and trace amounts of finely disseminated pyrite. The frequency and size of porphyroclasts has diminished, and most are commonly ovate to augen and comprised of feldspar (F). Fine grained matrix comprises 80% of the rock. Quartz ribbons make up an estimated 5% of the matrix and the size of sub-grains within ribbons is typically less than 200 μm (Q). The remainder of the matrix has an average grain size of < 50 μm and the abundance of sericite appears to be inversely proportional to the abundance of feldspar porphyroclasts and is typical of 'reaction softening' (ref. Section 3.8.3)
Progressive textural changes in the granite dyke can be compared due to the textural homogeneity of the granite protolith. Figures 3.14c-e illustrate changes in the intensity of deformation over a 1m interval, starting with weakly deformed granite (Fig. 3.14c) and progressing to more intensely deformed granite at the dyke margin (Fig. 3.14e).

### 3.5.2 Deformation of the central suite

Deformation is much less pronounced in rocks of the central suite compared to the western suite, and appears to be restricted to narrow zones that become less common away from the western-central suite boundary. Massive diopside zones are locally brecciated over intervals of up to 30 m wide. The degree of brecciation is highly variable, ranging from closed framework jigsaw fit (< 10% matrix), to open framework with more dispersed clasts and ≤ 40% matrix. The matrix consists of finer clasts of diopside with minor secondary carbonate and/or chlorite. The overall texture varies from crush breccia to protocataclasite.

A 17 m wide zone of diopside-quartz-dolomite in drill hole PD0017 contains both ductile and brittle deformation textures. The zone has a central core of banded quartz-diopside rock flanked by massive diopside, which grades outwards into diopside-marble rock. Throughout the 17 m wide zone, diopside clasts have a jigsaw fit, clast-supported breccia texture typical of those produced by brittle deformation processes (Ramsay and Huber, 1987). By contrast, quartz bands in the quartz-diopside core are intensely deformed, and have ultramylonite textures (Fig. 3.15a, b) due to competency contrasts between the two minerals.

Marbles of the central suite were probably also deformed, however, mylonitic deformation textures cannot be determined unequivocally due to recrystallisation (Fig. 3.15c, d).

All of the granite dykes in the central suite are deformed. The intensity of deformation varies from serrated quartz grain boundaries and thin zones of crush microbreccia to rare mylonites.

### 3.5.3 Eastern suite

Deformation in eastern suite schists is much less pervasive than in western suite schists. Deformation zones are typically less than a few metres wide and are concentrated within the schists, rather than at granite dyke margins. Brittle deformation textures are generally more common than the ductile deformation textures shown in Figure 3.15e.

### 3.6 Pegmatite dykes

Pegmatites occur as dykes up to 1m wide in close spatial association with granite dykes in rocks of the western, central, and eastern suites. Pegmatite dykes consist of hematite-dusted K-feldspar (usually microcline), quartz and, locally, minor tourmaline.
Fig. 3.15a. Drill hole PD0017, 286 m. Transmitted light photomicrograph (x nicols). Quartz bands in quartz-diopside rock, consist of quartz ribbons, and ribbon fragments (typically 0.5 - 1 cm long) in a fine grained mosaic of 20 - 100 μm polygonal quartz. The margins of quartz ribbons have a mortar texture of recrystallised, polygonal quartz, and most ribbons have been partially recrystallised to fine grained polygonal quartz.

Fig. 3.15b. Transmitted light photomicrograph. Rotated view of Fig. 3.15a with insertion of a gypsum plate. Fine grained polygonal quartz has a consistent crystallographic orientation that is commonly discordant to the crystallographic orientation of ribbon quartz. All of the quartz has recrystallised by ductile processes and this quartz band has an ultramylonitic texture.

Fig. 3.15c. Drill hole PD0010, 229.9 m, central suite. Finely laminated textures with relict 'porphyroclasts' apparent in hand specimen (P), are interpreted to have resulted from deformation of some 'marbles'.

Fig. 3.15d. Transmitted light photomicrograph of the sample shown in Fig. 3.15c. Fine grained dolomite (clear) and trace amounts of fine grained graphite (dark). Deformation textures interpreted from the hand specimen are not apparent in thin section possibly due to recrystallisation.

Fig. 3.15e. Drill hole PD0031, 241.5 m. Transmitted light photomicrograph. A 2 - 3 m wide deformed zone in the eastern suite schist has a 0.5 m wide core of mylonitic deformation, flanked by cataclasites and protocataclasites. The mylonite has well developed quartz ribbons (Q) wrapped around small feldspar porphyroclasts (F), a classical mylonitic texture.
Whereas all of the granite dykes are medium grained and deformed, the pegmatite dykes are coarse grained and either undeformed or weakly deformed. On the basis of the contrasting deformation styles, pegmatite dykes are considered to be younger than the granite dykes (Fig. 3.16a-d). Whereas granite dykes do not have alteration halos, larger pegmatite dykes typically have an extensive alteration halo which can extend for several metres into the enclosing country rock (described in Section 3.7.2).

3.7 Metasomatism

Naumann (1826) described metasomatism as the process by which one mineral is replaced by another of different chemical composition due to reactions caused by the introduction of material from external sources. Lindgren (1912) described metasomatic rocks as those whose chemical composition has been substantially changed by metasomatic alteration of the original constituents.

Some of the central suite calc-silicates and marbles have a distinct pinkish colouration that is interpreted to have resulted from metasomatic alteration. Marble that contains thin bands and disseminations of metapelitic material and/or thin iron formation bands has been more affected by metasomatism than other lithologies (Fig. 3.16e).

Higgins et al. (1990) described the metasomatized marbles of the central suite as being "vaguely banded marble", that contained small grains or aggregates of quartz, feldspar, diopside, amphibole and rare garnet, talc, forsterite and epidote. "Large porphyroblasts (up to 3 cm) and irregular layers and blebs of microcline may comprise 10% of the mineralised host rock". Higgins et al. (1990) interpreted this to be related to a tuffaceous component within the calcitic marble. Beeson (1990b) used whole rock geochemistry to calculated the average modal mineral composition of the metasomatised marbles as 65% calcite, 20% K-feldspar and 15% quartz. Beeson (1990b) considered that a primary volcanic component or authigenic K-feldspar contributed to modal K-feldspar. The positive correlation between K and other rock-forming elements such as Ti and variations in the Al : K ratios between nearby prospects led Beeson (1990b) to favour an authigenic origin.

In this study, interpretation of the origin of K-feldspar in the marbles of rock type 5 has centred around careful examination of the mineralogy and mineral textures. The whole rock geochemical database used by Beeson (1990b) was kindly made available by Dr. Beeson (then; Chief Geologist, Acacia Resources Ltd.). An attempt was made to utilise the whole rock geochemical data is Gresens (1967) type mass balance calculations, however, due to the nature of the sampling, intervals of similar rock types could not be unequivocally assigned to 'altered' and 'unaltered' groups. Mass balance calculations were therefore abandoned.
Fig. 3.16a. Drill hole PD0032, 294.8 m. Light, yellow-green, sericite altered, protomylonitic granite clasts (G) in an undeformed feldspar pegmatite (P). The clasts have alteration rinds of pink K-feldspar (K).

Fig. 3.16b. Transmitted light photomicrograph of the margin of an altered granite dyke clast illustrated in Fig. 3.16a. The contact between the altered margin of the granite clast and the unaltered interior is marked by a diffuse line (dashed line). Sericite that formed during deformation and alteration of the granite dyke is preserved in the interior of the clast (S). At the margin of the clast, sericite has been altered to K-feldspar (K). The relict texture of sericite is preserved by K-feldspar alteration.

Fig. 3.16c. Drill hole PD0007, 261.3 m. A 5-6 cm wide granite dyke hosted by metasomatically altered calcitic marble. The granite dyke has been brecciated, and hematite-dusted K-feldspar has formed the breccia matrix.

Fig. 3.16d. Drill hole PD0016, 281.4 m. Transmitted light photomicrograph (x nicols). A brecciated granite dyke adjacent to a pegmatite vein. Metasomatic microcline (M) occurs as overgrowths on granite breccia clasts (G).

Fig. 3.16e. Drill hole PD0020, 273.7 m. Metasomatic alteration of central suite marble has resulted in mineralogical changes and a pinkish colouration due to finely disseminated hematite. Bands of silicate minerals (iron formation or schist) and segmented bands (SB), are intensely altered around their margins whereas the interior of bands and segments is typically less altered.

Fig. 3.16f. Drill hole PD0016, 310.5 m. Transmitted light photomicrograph (x nicols). Metasomatic alteration of thin iron-rich band hosted by pinkish, calcitic marble has resulted in alteration of iron silicate minerals to microcline (M) and phyllosilicates (P). Remnant magnetite (MG) in the iron-rich band has retained a metamorphic orientation. Metasomatic epidote (E) and garnet (G) form a halo around the iron-rich band and passes into metasomatically altered calcite (C).
3.7.1 Mineralogy

The mineralogy of metasomatised marbles and silicate rocks has been determined by a combination of standard optical techniques and X-ray diffraction (XRD). XRD samples were analysed both prior to, and after, treatment with HCl. All samples were analysed as randomly orientated powders using the standard pressed powder technique (Brindley and Brown, 1980). XRD results and a description of the equipment and techniques used are presented in Appendix 3.5. When there was some uncertainty as to the nature of the feldspar, discrimination was made on the basis of staining using the techniques of Norman (1974), or by electron microprobe analysis.

Metasomatised marbles are medium grained, calcitic marbles, with between 1-40% disseminated silicate minerals or thin bands of silicate minerals. The pinkish colouration is caused by the presence of disseminated hematite and hematite-dusted microcline, or, in some cases, hematite-dusted albite. Other mineral species, in order of decreasing abundance, include: brown to olive green andradite garnet, pale brown grossular garnet, epidote, chlorite, sericite/mica, wollastonite, leucoxene, apatite and sphene. Trace amounts of cuspidine (Ca₄Si₂O₇(F,OH)₂), vesuvianite and scapolite have been observed by commercial petrographers (Cowan, 1990) but have not been confirmed in this study. Siderite ± calcite occurs in conjunction with metasomatic silicate minerals.

3.7.2 Metasomatic mineral textures

Metasomatic calcite and siderite are texturally different to metamorphic marbles. Most of the metamorphic calcite crystals are 1-3 mm in diameter, and commonly have deformation twins and sutured grain contacts. By contrast, metasomatic carbonates are commonly calcitic and locally sideritic, are finer grained (1-10 μm), do not have stress twins, and have irregular diffuse grain contacts. The most intense metasomatism of metamorphic marbles occurs adjacent to metapelitic bands and iron formation bands (Fig. 3.16). Metasomatic alteration of iron formations has resulted in partial, to complete replacement of Fe and Mg-silicates (hastingsite, grunerite, biotite and diopside) by fine grained sericite, chlorite and leucoxene (Fig. 3.17a). Magnetite has been partially replaced by hematite. Metasomatic sericite and chlorite were subsequently partially replaced by hematite-dusted microcline (Fig. 3.17b). In some instances, chlorite (after metamorphic biotite) and magnetite were overgrown by, and preserved as remnants within microcline (Fig. 3.17c). The remnant metamorphic fabric of the precursor ferro-magnesian minerals has been partially preserved as lined clots of leucoxene or sphene enclosed within metasomatic minerals.
Fig. 3.17a. Drill hole PD0001, 127.45 m. Hand specimen of altered thin iron-rich bands (IF) in metasomatically altered calcitic marble (M) adjacent to a thick iron formation. The dark iron-rich bands consist of biotite, hastingsite, quartz and minor disseminated magnetite. Bands are surrounded by halos of sericite, hematite-dusted microcline, minor leucoxene, remnant quartz and magnetite, and outer bands of olivine green garnet and apple green epidote (CS). Light green remnant diopside at the margins of iron-rich bands has been partially altered to fine grained chlorite + calcite.

Fig. 3.17b. Drill hole PD0001, 127.43 m. Transmitted light photomicrograph (x nicols) of the central portion of an iron-rich band. The left hand side of the diagram shows the core of the band (C), where the biotite and amphibole have been completely replaced by a fine grained sericite, quartz, microcline, chlorite and leucoxene. Relict grains of magnetite (MG) and trace amounts of chlorite-altered biotite are preserved, although most has been altered to sericite or microcline. Towards the margin of the iron-rich band (right hand side of the diagram), hematite-dusted microcline (M) has replaced metamorphic minerals. Some martitised magnetite (MG) has not been replaced and occurs as inclusions in microcline. All evidence of the original metamorphic fabric has been destroyed.

Fig. 3.17c. Drill hole PD0016, 278.4 m. Transmitted light photomicrograph (x nicols) of an altered iron-rich band. Remnant metamorphic biotite has been completely altered to dark green chlorite (C) and preserved as inclusions within metasomatic microcline (M). Disseminated magnetite (MG) and rare quartz have also been preserved within metasomatic microcline.

Fig. 3.17d. Drill hole PD0017, 549.9 m. Transmitted light photomicrograph (x nicols) of the outer portion of an altered iron formation band. The alteration assemblage includes pale brown, growth zoned grossular garnet (G). The central core of some garnets contains partially altered diopside relics (D) which have locally been overgrown by opaque andraditic garnet and growth zoned grossular garnet. Calcite (C) has infilled between grossular garnet crystals.

Fig. 3.17e. Drill hole PD0001, 291.3 m. Transmitted light photomicrograph (x nicols) of rounded microcline grains (M) produced through metasomatic alteration of disseminated biotite flakes in calcitic marble. Sphene (S) is a common accessory mineral.
Fig. 3.18a - e. Drill hole PD0026. 395.8 m - 420.8 m. A series of photographs across the alteration halo surrounding a pegmatite dyke hosted by massive diopside. Figure 3.18a is from near the margin of the alteration halo. Figure 3.18c is from near the centre of the dyke and Figure 3.18d - e are from the alteration halo on the opposite side of the pegmatite dyke.

Fig. 3.18a. 402.8 m. Massive diopsidic host rock (D) with minor calcite (C) and chlorite (CH) alteration.

Fig. 3.18b. 399.8 m. Near the margin of the pegmatite dyke. Brecciated diopside (LHS), has been altered to whitish calcite, dark green chlorite, epidote ± garnet and trace amounts of pyrite. The margin of the dyke (RHS), is indistinct due to partial to pervasive alteration of calcite, chlorite, epidote ± garnet to hematite-dusted K-feldspar.

Fig. 3.18c. 399 m. Core of the pegmatite dyke. Hematite-dusted K-feldspar and quartz have completely replaced diopside. Remnants of chlorite + calcite + epidote ± garnet alteration after diopside are preserved within feldspar + quartz alteration.

Fig. 3.18d. 395.8 m. The outer alteration zone (RHS) comprises dark green chlorite + calcite + epidote ± garnet alteration of breccia textured diopside.

Fig. 3.18e. 395.7 m. Transmitted light photomicrograph (x nicols) of Large epidote laths (E) with interstitial feldspar (F), from the outer alteration halo of a pegmatite vein hosted by massive diopside.

Figs. 3.18f - g. Drill hole PD0013, 129 - 134 m. Two diagrams of the alteration halo surrounding a pegmatite vein in the south of the prospect.

Fig. 3.18f. Diopside (D) has been brecciated and partially altered to epidote ± garnet (E).

Fig. 3.18g. Towards the centre of the pegmatite vein, hematite-dusted feldspar (red) has overprinted epidote + chlorite + calcite ± garnet alteration (green). The sulphide vein (S) cuts across the alteration and is interpreted to have post-dated metasomatic alteration.
Epidote and garnet form an outer halo to microcline (Fig. 3.17a). Epidote is present as irregular shaped crystals and locally, growth-zoned prisms. Some garnets have an irregular core of olive-green andradite overgrown by pale brown, growth-zoned grossular (Fig. 3.17d). Fine grained Fe-rich carbonate and finely dispersed hematite occurs as matrix fill. Diopside has been partially altered to a fine grained mix of carbonate ± talc ± chlorite.

Metasomatic alteration has also affected banded or disseminated aluminous minerals and dispersed biotite has been partially, to completely, replaced by small (10-100μm diameter), rounded 'grains' of hematite-dusted microcline and trace amounts of leucoxene or sphene (Fig. 3.17e). Although metasomatism has resulted in pervasive alteration of most impure marble units, it is most intense in proximity to pegmatite dykes, where the alteration assemblage locally includes wollastonite. Altered impure marble units are bound to the east by schist (eastern suite schist) and to the west by thick iron formations (Fig. 3.1). Marble and calc-silicate lithologies to the west of the iron formations have not been extensively altered by metasomatism.

Although not extensively altered, massive diopside is locally altered to feldspar + epidote + chlorite + calcite. A criteria for alteration of diopside appears to have been a pre-existing fracture network and proximity to pegmatite dykes. Pegmatite dykes in diopside have a zoned alteration pattern (Figs. 3.18a - g) with (i) a core of pinkish hematite-dusted feldspar, which grades outwards to the diopside host through zones of; (ii) hematite-dusted feldspar ± epidote ± garnet; (iii) epidote ± garnet + chlorite + calcite ± traces of disseminated pyrite; (iv) chlorite + calcite ± disseminated hematite with remnant diopside; (v) jig-saw fit brecciated massive diopside; (vi) massive diopside. In the absence of a biotite precursor, metasomatic feldspars in altered diopsidic host rocks are sodic rather than potassic (Fig. 3.18e).

3.8 Discussion
In this section, the Palaeoproterozoic stratigraphy at Menninnie Dam is compared with similar units that occur elsewhere on the Eyre Peninsula for the purposes of regional correlations. The timing of deformation, granite and pegmatite dyke emplacement, and metasomatism is discussed with reference to regional tectonic and intrusive events.

3.8.1 Origin of the central suite calc-silicates
Two hypotheses can be proposed for the origin of calc-silicate minerals in the central suite rocks.

- Prograde and retrograde metamorphism of a dolomite - calcite - quartz and iron oxide rich protolith.
- Metasomatism of a metamorphosed dolomite and / or limestone protolith.
Whereas isochemical regional metamorphism requires all the elements to be present prior to metamorphism, metasomatism can result in addition of Mg and SiO₂ into a carbonate rock with elemental and mineral zonation occurring over a large area around the heat or fluid source (e.g., Bingham Canyon; Atkinson and Einaudi, 1978). No such zonation has been observed at Menninnie Dam. The lack of zonation is not considered to be due to the limited area of the Menninnie Dam prospect, as marble and calc-silicate rocks with similar mineralogy and textures are common elsewhere on the Eyre Peninsula in units correlated with the Katunga Dolomite (Parker, 1993b). The mineralogy and textures of the central suite at Menninnie Dam are therefore concluded to be a product of regional metamorphism of a dolomite, calcite, quartz and iron oxide rich protolith.

The mineralogy of the central suite rocks are consistent with the metamorphic grade determined for the eastern suite schist (Section 3.3). Forsterite is absent from the central suite rocks whereas tremolite is common. The appearance of forsterite as a stable phase coincides with the disappearance of tremolite during prograde metamorphism of calc-silicate rocks (Trommsdorff, 1966, 1972; Slaughter et al., 1975; Metz, 1976). The presence of talc in central suite rocks is inconsistent with calculated peak metamorphic conditions and talc most likely formed during retrograde metamorphism through the break-down and hydration of calc-silicate minerals and carbonates (Turner, 1980). Disseminated magnetite occurs in calcitic marbles, dolomitic marbles, mixed calc-silicate-marble rocks and in massive diopside. Magnetite occurs within diopside crystals in massive diopside units and is commonly concentrated along the contact between diopside and calcite, or tremolite and calcite, in marble-calc-silicate lithologies (Fig. 3.4d). If iron oxide was evenly dispersed throughout the carbonate protolith, magnetite concentration could have occurred through consumption of carbonate during formation of calc-silicate minerals.

3.8.2 Regional correlations and implications for the Menninnie Dam stratigraphy.

Lithologies at Menninnie Dam can be correlated with the Middleback Subgroup of the Hutchison Group (Fig. 3.19). The marbles, calc-silicate lithologies, graphitic schists and iron formations of the central suite at Menninnie Dam are interpreted to correlate with the Katunga Dolomite and Lower Middleback Jaspilite respectively, as proposed by Parker (1993b). Marbles and calc-silicate lithologies similar to those of the central suite at Menninnie Dam crop out west of Cowell and at Sleaford Bay, and are interpreted to be equivalent to the Katunga Dolomite or Lower Middleback Jaspilite (Parker, 1993b). Iron formations at Menninnie Dam are similar to those of the Upper and Lower Middleback Jaspilites of the Middleback Subgroup. The graphitic schists at Menninnie Dam are similar to graphitic schists described elsewhere on the Eyre Peninsula.
Figure 3.19 Correlation between the Menninnie Dam outcrop pattern (schematic interpretation, not to scale) and the regional stratigraphy of the Hutchison Group (modified from Parker et al., 1988 and Drexel et al., 1993). Red lines connect correlated lithologies. The heavy dashed line is the faulted contact between the central and western suites at Menninnie Dam.
Graphite is a minor component of some of the iron formations in the Middleback Ranges (Parker, 1993b) and a significant component of Middleback Subgroup schists in the Uley and Koppio regions of the southern Eyre Peninsula (Parker, 1993b).

The gradational contact between the central suite marble and calc-silicate lithologies and the eastern suite schist, along with the interpreted facing direction of metamorphosed graded beds, implies that the eastern suite is younger than the central suite. The eastern suite schist is interpreted to correlate with the Cook Gap Schist which is stratigraphically younger than the Katunga Dolomite and Lower Middleback Jaspite (Fig. 2.3). The mineralogy, textures, occurrence of amphibolite layers, calc-silicate units, migmatites, and garnet-rich layers in the eastern suite schist are consistent with Parkers (1993b) description of the Cook Gap as "dominantly semipelitic, quartz-veined garnet-mica schist and gneiss with minor calc-silicate gneiss, magnetite-bearing gneiss, and discordant amphibolites ... the schists and gneisses are layered, comprising bands either rich or poor in mica (biotite or muscovite), garnet, feldspar, sillimanite, and quartzo-feldspathic or pegmatitic segregations". Garnetiferous bands were interpreted by Parker (1993b) to result from sedimentary layering. Migmatites are restricted to zones of higher metamorphic grade and discordant amphibolites occur throughout the lower part of the Cook Gap Schist (Parker, 1993b).

The Yadnarie Schist, interpreted to be younger than the Cook Gap Schist, is mineralogically and texturally similar to the Cook Gap Schist (Parker, 1993b), however, calc-silicate lithologies, garnetiferous bands and migmatites have not been described from the Yadnarie Schist and amphibolites are rare or absent (Parker, 1993b). Correlation of the eastern suite schist with the Cook Gap Schist is consistent with interpretation of the Menninnie Dam stratigraphy by Parker (1993b).

The contact between the western and central suites is discordant relative to stratigraphic layering in the central and eastern suites and is interpreted to be a sheared contact. In the Gawler Craton, KM-D3 is associated with the formation of broad, northerly-trending mylonite zones, such as the Kalinjala Mylonite (Parker, 1978; Fig. 3.10). The sense of movement of the mylonites has been interpreted to include a major component of dextral shear (Parker, 1993a). Parker (1993a) mapped the trace of a major north-northwesterly trending mylonite zone near Refuge Rocks. If this mylonite zone was projected northwards along strike for a further 50 km, it would pass close to, or through the Menninnie Dam prospect (Fig. 3.10).

Recognition of the western suite / central suite contact as a major shear zone assists interpretation and correlation of the local stratigraphy at Menninnie Dam. The western suite schist is mineralogically and texturally similar to the eastern suite schist and the Cook Gap Schist.
Fig. 3.20. Cartoon. The NNE trending central suite rocks (brick pattern) may have been truncated by a dextral shear bounding the central and western suites resulting in juxtaposition of the Cook Gap Schist (stippled pattern) against the lower part of the Katunga Dolomite.
Garnetiferous bands, migmatites and rare amphibolites in the western suite schist are similar to those described in the Cook Gap Schist by Parker (1993b). The western suite at Menninnie Dam is interpreted to be a structural repetition of the eastern suite that was emplaced during dextral movement along the bounding fault zone, and is therefore also correlated with the Cook Gap Schist (Fig. 3.20). Structural repetition is interpreted to have resulted in the juxtaposition of younger Cook Gap Schist against the lower units of the Katunga Dolomite (Fig. 3.2).

Higgins et al. (1990) noted the thickness of the Katunga Dolomite (central suite) at Menninnie Dam is greater than the same unit in the Middleback Ranges. Local deformation zones recognised within the central suite may have resulted in structural thickening of this unit, although this cannot be proved.

### 3.8.3 Timing of deformation and MD granite dyke emplacement

The common occurrence of deformed MD granite dykes in zones of intense deformation is consistent with a link between deformation and MD granite dyke emplacement. Factors which contribute to constraining the timing of dyke emplacement are:

- The granite dykes all have similar chemistry, and are interpreted to be co-genetic and coeval.
- Units containing abundant silicate minerals, such as iron formations and schist layers, that are enclosed in marble are commonly surrounded by a halo of calc-silicate minerals. This halo is interpreted to have formed at near peak-metamorphic conditions, and probably formed by element diffusion between silicate layers and marble. In contrast, MD granite dykes in contact with marbles have no such halo. Therefore, either the MD granite dykes did not react with the marbles, or, they were emplaced after peak-metamorphic conditions (KM-D1,2), i.e. sometime after 1745 Ma. (Parker, 1993a).
- MD granite dykes were probably intruded while the host rocks were at elevated temperatures. MD granite dykes less than a few tens of centimetres wide are not likely to have propagated any distance unless they maintained low viscosity through elevated temperatures. The absence of chilled margins to larger MD granite dykes is also consistent with hot host rocks. It therefore seems likely that the MD granite dykes were intruded prior to, or during, KM-D3 (1745 - 1710 Ma., Parker 1993a), or within the thermal aureole of a granite pluton.
- Scattered outcrop of granite of the Lincoln Complex occurs ~ 5 km to the east of the Menninnie Dam prospect and was considered by Parker (1993a) to have been emplaced during KM-D3 (Moody Suite correlative; Fig. 3.10).
- The MD granite dykes have similar geochemical compositions to granitoids of the Moody Suite. The Moody Suite and the MD granite dykes are elevated in P₂O₅ at...
similar SiO₂ values (Section 3.4.2). The age of the Moody Suite is ≈ 1745 - 1710 Ma. (Parker, 1993a; Fanning, 1984).

- MD granite dykes are tectonically deformed, and have textural and geochemical attributes which distinguish them from granitoids of the Hiltaba Suite (1589 ± 2 Ma., Creaser and Cooper, 1993) which are not tectonically deformed. Therefore, MD granite dykes must have been intruded prior to 1589 Ma., and are likely to have been intruded prior to or during the previous deformation event (KM-D3).

The style of MD granite dyke deformation can be used to help constrain the timing of granite dyke emplacement. Ductile deformation was more common at the margins of MD granite dykes, whereas deformation in the enclosing schist is characterised by cataclastic deformation. Strain may intuitively have been expected to partition into the phyllosilicate-rich schist, however, at Menninnie Dam, it has partitioned into the margins of granite dykes. Two mechanisms could account for this:

1) Strain partitioning into MD granite dyke margins may have resulted from ductility contrasts, where the granite magma was more ductile than the enclosing schist. This situation may have arisen if hot, low viscosity granite magma was intruded into an active deformation zone in which the host rocks were cooler and less plastic.

2) Interaction of aqueous fluids with the margin of a hot MD granite dyke may have resulted in hydrothermal alteration of the MD granite dyke margin with destruction of feldspar and formation of fine grained phyllosilicates. The presence of abundant phyllosilicates may have enhanced strain partitioning. This process is similar to 'reaction softening', described by White et al. (1980).

The first mechanism requires that the MD granite dykes were intruded into an active deformation zone. If this was so, then MD granite dykes might be expected to show a lineation fabric due to both magmatic flow and solid state flow (Paterson et al., 1989). However, magmatic flow may be masked by rapid cooling and/or solid state flow in high strain environments (Miller and Paterson, 1994).

The second mechanism also requires MD granite dyke emplacement to have occurred either prior to, or during deformation. The fluid required for hydrothermal reactions and formation of chlorite and sericite at the MD granite dyke margin may have been derived from fluid entering the shear system. Beach (1980) discussed retrograde metamorphic mineral formation associated with shearing in the Lewisian Complex, NW Scotland. Hydration reactions converted feldspar to muscovite, amphibole to chlorite, and biotite to chlorite. The amount of fluid involved in such reactions can be considerable, although the source of the fluid was not known (Beach, 1980).
MD granite dykes that have intruded marbles and calc-silicate lithologies of the central suite do not show the same degree of deformation and alteration at their margins, nor are the host rocks pervasively altered to a retrograde mineral assemblage. It appears that most of the fluid flow which resulted in alteration of the MD granite dykes was focused into a zone of maximum deformation within the western suite.

The common association of hydrothermal alteration and more intense deformation at the margins of MD granite dykes in the western suite shear zone is interpreted as evidence for reaction softening as a mechanism of strain partitioning. However, strain partitioning due to ductility contrasts cannot be discounted and may have occurred prior to reaction softening.

When factors interpreted to constrain the timing of MD granite dyke are considered in conjunction, the MD granite dykes are most likely Moody Suite correlates that were emplaced during deformation associated with KM-D3, and probably related to the exposed granite = 5 km to the east of the Menninnie Dam prospect (Moody Suite correlative, Parker, 1993a). This constrains the timing of MD granite dyke emplacement to = 1745 - 1710 Ma. (Parker, 1993a; Fanning, 1984). The pegmatite dykes (Section 3.6) are spatially associated with MD granite dykes, however the undeformed nature of pegmatite dykes is consistent with post-kinematic emplacement. Weakly deformed pegmatites intruded into the Kalinjala Mylonite Zone were dated at = 1710 Ma. and were considered to have been emplaced during the latter phases of KM-D3 (Fanning, 1984). A similar age of = 1710 Ma. could be inferred for the pegmatite dykes at Menninnie Dam, however, on the basis of data presented in this study, the timing of the pegmatite dyke emplacement cannot be unequivocally determined.

3.8.4 Timing of metasomatism

None of the silicate minerals produced through metasomatic alteration have textures that are compatible with high grade metamorphism:

- Triple junctions are absent.
- There is no tendency towards metamorphic refining of minerals; for example, microcline is poikolitic (Section 3.7.1).
- Epidote, talc and chlorite are incompatible with the peak metamorphic conditions calculated for the Menninnie Dam rocks.
- Carbonate and talc are fine grained, and do not appear to have been recrystallised.
- Garnet and epidote contain well-preserved complex growth zoning.

Metasomatic alteration is spatially and genetically related to pegmatite dykes, if pegmatite dykes were associated with the latter stages of MD granite dyke emplacement, then metasomatism would also have occurred around 1710 Ma. The timing of pegmatite dykes and metasomatism is addressed again in Section 4.4.12 following presentation of data on the Mesoproterozoic stratigraphy at Menninnie Dam.
Impure marbles bound to the west by iron formations and to the east by schist (eastern suite) were pervasively altered during metasomatism. The iron formations and the schist may have acted as aquatards, impeding the passage of hydrothermal fluids.

3.9 Summary

The Palaeoproterozoic stratigraphy at the Menninnie Dam prospect has been subdivided into three suites, a western suite, a central suite and an eastern suite. Marbles, diopsidic calc-silicates, iron formations and minor graphitic schist bands of the central suite are correlated with the Katunga Dolomite and Lower Middleback Jaspilite of the Middleback Subgroup. The eastern suite biotite-quartz-feldspar-muscovite-sillimanite-garnet schist correlates with the Cook Gap Schist of the Middleback Subgroup. Deposition of the host rocks can therefore be constrained to $\sim 1964 - 1845$ Ma.

Thermobarometric calculations indicate peak conditions corresponding to upper amphibolite facies followed by a retrograde greenschist event. Calculated P-T conditions correspond with those determined elsewhere on the Eyre Peninsula for peak-metamorphic (KM-D1,2) and retrograde metamorphic (KM-D3) phases of the Kimban Orogeny.

Regionally, the Warrow Quartzite underlies Katunga Dolomite. However, in contrast to earlier reports (Higgins and Hellsten, 1986; Higgins et al., 1990), the Warrow Quartzite is not known to occur in the immediate vicinity of Menninnie Dam. The contact between the central and western suites is interpreted to be a large protocataclastic to mylonitic shear zone. Assuming a dextral sense of movement, the western suite schist is probably a faulted repetition of the eastern suite. The shear zone was probably initiated during KM-D3, synchronous with the formation of large, N-S trending, mylonite zones elsewhere on the Eyre Peninsula. Granite dykes are interpreted to have intruded the western suite protocataclastic to mylonitic schists during the deformation event with partitioning of ductile strain into the margin of the granite dykes. Granite dykes are tentatively considered to be Moody Suite correlates and have an inferred age of 1740 - 1710 Ma. Pegmatitic dykes are spatially associated with granite dykes but have textures consistent with post-kinematic emplacement. Pegmatitic dykes may have been emplaced during the latter stages of KM-D3 ($\sim 1710$ Ma.; Fanning, 1984), however the timing of emplacement is uncertain. Fluids associated with pegmatitic dykes resulted in pervasive metasomatism of central suite marbles, particularly those that contained a minor pelitic component.
4.1 Introduction
The Mesoproterozoic geology at Menninnie Dam is described in this chapter. Excluding a veneer of Quaternary cover, all rocks younger than Palaeoproterozoic at Menninnie Dam are interpreted to be of Mesoproterozoic age (1600 Ma. - 1000 Ma.). At Menninnie Dam, Mesoproterozoic rocks overlie Palaeoproterozoic rocks, however, unlike Palaeo-Proterozoic lithologies, they have not been metamorphosed, tectonically deformed or metasomatised. Mesoproterozoic rocks have been deeply weathered to depths commonly in excess of 100m and much of the weathered profile was percussion drilled, therefore, drill core of Mesoproterozoic rocks is less complete than that for Palaeoproterozoic rocks. All descriptions and interpretations are based on limited drill core and a few percussion holes.

The orientation of some Mesoproterozoic rock units at Menninnie Dam are poorly constrained due to their limited lateral extent and the wide spacing of drill holes (typically spaced 200m apart, Fig. 3.1). Where possible, the orientation of Mesoproterozoic rock units has been interpreted from their intersection in two or more drill holes, and on two or more sections. Summary drill hole logs and cross sections are presented in Appendix 3.1.

The aims of this chapter are to:
- Describe the regional geological setting.
- Describe the Mesoproterozoic rock units at Menninnie Dam.
- Compare the geochemistry of intrusive rocks to the geochemistry of the Gawler Range Volcanics (GRV) and co-magmatic Hiltaba Suite.
- Interpret the processes which resulted in formation of Mesoproterozoic rock units.

4.2 Geological setting
Mesoproterozoic rock units in the vicinity of the Menninnie Dam prospect comprise volcaniclastic lithologies, lavas, and dykes of the GRV. A simplified stratigraphy and outcrop map of the GRV in the southern Gawler Range area are illustrated in Figures 4.1 and 4.2 respectively. Two kilometres to the west of the Menninnie Dam prospect are low hills comprising Bittali Rhyolite. Scattered, low-lying, Bittali Rhyolite also crops out to the east and south of the Menninnie Dam prospect (Fig. 4.2). The Nonning Rhyodacite crops out ten kilometres to the northeast of the Menninnie Dam prospect and forms low lying, dissected foothills of the Gawler Ranges.
The Eucarro Dacite crops out two kilometres to the north of the Menninnie Dam prospect as low-lying dissected foothills of the Gawler Ranges.
Stratigraphy of the Southern Gawler Ranges area

Yardea Dacite: Massive porphyritic dacite and rhyodacite. Phenocrysts of plagioclase (5 cm across), smaller K-feldspar and clinopyroxene set in a fine-granular felsic matrix.


Yannabie Rhyodacite: Fine grained to porphyritic rhyodacite with phenocrysts of plagioclase, K-feldspar and some spherical quartz in tuffaceous, vesicular matrix. Grades into rhyolite with more abundant quartz phenocrysts.

Eucarro Dacite: Massive porphyritic dacite grading to rhyodacite with scattered plagioclase (~5 mm across), spherical quartz, and minor K-feldspar, and clinopyroxene phenocrysts set in a finely-granular or granophytic matrix of K-feldspar and quartz.

Nonning Rhyolite: Massive porphyritic rhyolite to dacite. Phenocrysts of plagioclase (~1 cm across), smaller K-feldspar crystals, clinopyroxene and round quartz are set in a reddish-brown, granophytic or microgranular felsic matrix.

Bittali Rhyolite: Variable assemblage of rhyolitic to rhyodacitic lavas, intrusive fender dykes and welded ash flows with flow banding, fiamme and pumiceous streaks. Texture ranges from fine-grained to porphyritic with phenocrysts of K-feldspar, plagioclase and round quartz.

Waganny Dacite: Porphyritic dacite to rhyodacite with phenocrysts of plagioclase and minor K-feldspar up to 2 mm across set in a aphanitic or microgranular matrix.

PA01 - (rare basal member of succession). Unnamed porphyritic rhyolite.

Proterozoic or unknown basement

Fig. 4.1  Schematic, simplified stratigraphy of the Gawler Range Volcanics in the southern Gawler Ranges area (not to scale; stratigraphy and descriptions modified after Blissett et al., 1988), with subdivision into the developmental and mature phases (after Stewart, 1992).
Figure 4.2 Location of the Menninnie Dam grid and simplified map of the Gawler Range Volcanics units in the southern area of the Gawler Ranges (modified after Elsset et al., 1988), with subdivision into the developmental and mature phases after Stewart (1992).
4.3 Prospect geology

Mesoproterozoic rock units encountered during drilling at Menninnie Dam are described starting with units interpreted to be the oldest followed by progressively younger units. These include:

- Rhyolite intrusions
- Discordant polymictic breccia units
- Fiamme-bearing discordant polymictic breccia units
- Rhyolite-bearing discordant polymictic breccia units
- Layered polymictic breccia units
- Rhyolite-bearing layered polymictic breccia units
- Coherent rhyolite
- Muddy sandstone and pumice breccia units
- Menninnie Dam ignimbrite

An interpretation of rhyolite-bearing discordant polymictic breccia units is presented immediately following the description, however all other units are interpreted in a later discussion section.

4.3.1 Rhyolite intrusions

Brick-red, quartz- and feldspar-phryric rhyolite intrusions were intersected in four drill holes in the south of the Menninnie Dam prospect (Table 4.1 and Fig. 4.3a). The lower contacts of two rhyolite intrusions (drill holes PD0030 and PD0031) were not intersected by drilling, and their extent is not known.

Table 4.1 Drill hole intersections of rhyolite intrusions. Northing and easting coordinates are for drill collars. 'From' and 'To' are down hole depths. The orientation of rhyolite intrusions is not known, therefore the width is equal to that intersected in drill holes.

<table>
<thead>
<tr>
<th>Drill hole</th>
<th>Northing</th>
<th>Easting</th>
<th>From(m)</th>
<th>To(m)</th>
<th>Width (m)</th>
<th>Type*</th>
</tr>
</thead>
<tbody>
<tr>
<td>PD0019</td>
<td>838</td>
<td>10408</td>
<td>283.5</td>
<td>296.5</td>
<td>13.0</td>
<td>?</td>
</tr>
<tr>
<td>PD0027</td>
<td>9293</td>
<td>10025</td>
<td>76</td>
<td>82</td>
<td>6</td>
<td>?</td>
</tr>
<tr>
<td>PD0030</td>
<td>7764</td>
<td>10250</td>
<td>373.9</td>
<td>388.25</td>
<td>14.35</td>
<td>W.A.</td>
</tr>
<tr>
<td>PD0030</td>
<td>7764</td>
<td>10250</td>
<td>395.00</td>
<td>417.00</td>
<td>22.0</td>
<td>S.A.</td>
</tr>
<tr>
<td>PD0031</td>
<td>8023</td>
<td>10300</td>
<td>644.2</td>
<td>&gt;644.8 (EOH)</td>
<td>&gt;0.6</td>
<td>W.A.</td>
</tr>
</tbody>
</table>

* W.A. = weakly hydrothermally altered; S.A. = strongly hydrothermally altered; ? = intensity of hydrothermal alteration unknown as masked by intense weathering. EOH = end of hole.

Rhyolite intrusions in drill holes PD0027 and PD0019 are intensely weathered and will not be considered further. With so few drill hole intersections, it is not possible to determine the dimensions or shape of rhyolite intrusions although lenticular or pipe-like shapes are considered likely (Fig. 3.1 and Appendix 3.1).
Fig. 4.3a. Drill hole PD0030, 471 m. Hard specimen of weakly altered, brick-red quartz- and feldspar-phyric rhyolite intrusion. A single 1.5 x 1 cm xenolith consisting of altered silicate minerals and pyrite occurs in the centre of the figure (X).

Fig. 4.3b. Drill hole PD0030, 471 m. Transmitted light photomicrograph (x-nicols) of the groundmass of a rhyolite intrusion. The groundmass consists of a mosaic of fine-grained quartz and feldspar with a granophyric texture.

Fig. 4.3c. Drill hole PD0030, 471 m. Photomicrograph (reflected light) of round and embayed pyrite from the 1.5 x 1 cm xenolith illustrated in Figure 4.3a.

Fig. 4.3d. Drill hole PD0030, 410 m. Hand specimen of strongly altered, light yellow-green, quartz- and feldspar-phyric rhyolite intrusion. Relict feldspar and ferro-magnesian phenocrysts have been pervasively altered to dark, fine grained sericite, quartz, calcite and chlorite. The groundmass is an alteration assemblage of fine grained, felted sericite, quartz, calcite and chlorite.

Fig. 4.3e. Drill hole PD0030, 410 m. Reflected light photomicrograph of irregularly shaped crystals of chalcopyrite (CP) have been enclosed by galena (G) in the strongly altered rhyolite intrusion.
Description

All rhyolite intrusions are non-vesicular and have similar primary mineralogy and textures. Their mineralogy and mineral proportions have been visually estimated as: subhedral, round and embayed quartz phenocrysts, typically 0.5 - 4 mm across (15%); euhedral albite phenocrysts, typically 0.1 - 2 mm across (6%); euhedral K-feldspar, typically 0.1 - 4 mm across (10%), in a fine-grained groundmass. Altered remnants of ferro-magnesian phenocrysts comprise less than 1%. Rhyolite intrusions can be subdivided into two groups based on the intensity of hydrothermal alteration:

1) Weakly altered rhyolite intrusions
2) Strongly altered rhyolite intrusions

Weakly altered rhyolite intrusions

Weakly altered rhyolite intrusions are brick-red. The groundmass has a granophyric texture and consists of a fine grained mosaic of quartz and feldspar (Fig. 4.3b), a texture commonly attributed to primary crystallisation of relatively slowly cooled lavas, welded ignimbrites, and shallow intrusions (Lofgren 1971b).

Ferro-magnesian minerals have been altered to fine grained quartz, calcite, hematite and chlorite. Rare trace amounts of chalcopyrite are locally associated with relict ferro-magnesian phenocrysts. Albite phenocrysts have been partially, to pervasively, altered to sericite, quartz, minor calcite, trace amounts of chlorite and a dusting of hematite. K-feldspar phenocrysts are mostly unaltered, although locally, some have been partially altered to sericite, calcite, and quartz with traces of chlorite, and a dusting of hematite. Groundmass feldspar has been partially altered to sericite. The groundmass has a pervasive dusting of hematite which imparts a characteristic brick red colour to weakly altered rhyolite intrusions.

Trace amounts of pyrite (< 0.5 mm across) are sparsely, but evenly, dispersed throughout the groundmass. The shape of pyrite grains is irregular, to blebby, with rare subhedral grains. A single 1.5 x 1 cm clast comprising pyrite and altered silicate minerals occurs at 471 m in drill hole PD0030. Pyrite 'grains' within the clast are up to 3 mm across and have blebby and embayed shapes, similar to dispersed pyrite grains. The clast has distinct margins consistent with inclusion of a sulphide-rich xenolith (Figs. 4.3a, c).

The size of phenocrysts in the rhyolite intrusion intersected near the end of drill hole PD0030 has a zonal distribution. Within seven metres of the intrusion margin, quartz and feldspar phenocrysts are 0.5 - 2 mm across. Greater than seven metres from the margin of the intrusion, quartz and feldspar phenocrysts are 1 - 4 mm across. Hutchison Group calc-silicate lithologies at the upper contact of the rhyolite intrusion are fractured over an interval of 1.3 m. 'Fingers' of rhyolite, typically a few centimetres thick, have intruded fractures in calc-silicate lithologies.
Chapter 4 Local geology: The Mesoproterozoic

Strongly altered intrusion(s)

The rhyolite intrusion at 395.00 - 417.00 m in drill hole PD0030 is light yellow-green due to pervasive hydrothermal alteration (Fig. 4.3d). Quartz and relict felspar phenocrysts are of similar size and abundance ($\approx 31\%$) to those in the outer seven metres of weakly altered rhyolite intrusions. Relict feldspar phenocrysts have been pseudomorphed by fine grained sericite, quartz, calcite and chlorite. The former presence of ferro-magnesian phenocrysts cannot be determined due to the intensity of alteration. The groundmass has been altered to fine grained mosaic of sericite, quartz, calcite and chlorite. Some groundmass quartz micropoikilitically encloses fine grained sericite ± calcite ± chlorite and quartz pseudomorphs of feldspar microlites, similar to the weakly altered rhyolite intrusions.

Trace amounts of fine grained (< 0.2 mm) pyrite, sphalerite, galena and chalcopyrite are locally associated with relict feldspar and/or ferro-magnesian phenocrysts (Fig. 4.3e). Sphalerite, galena and chalcopyrite have highly irregular grain shapes. Most pyrite also has irregular grain shapes, with rare subhedral and cubic crystals, in contrast to the round and embayed shape of pyrite in the weakly altered rhyolite intrusion.

The upper margin of the intrusion is in contact with eastern suite schist which shows little evidence of disruption or alteration. The lower margin is in contact with brecciated and altered Hutchison Group marble.

4.3.2 Polymictic breccia units

Localised polymictic breccia units have discordant contacts with Hutchison Group and MD granite dyke lithologies, and also occur as layered units overlying the Hutchison Group (discordant and layered polymictic breccias are abbreviated to 'DP breccia' and 'LP breccia' throughout the remainder of the text). DP and LP breccias consist of unsorted to poorly sorted, matrix supported, polymictic breccia (Fig. 4.4a). Clast are commonly randomly oriented, and range from angular to rounded. There is no correlation between clast lithology and the degree of rounding and angular and sub-rounded clasts of the same size and lithology are locally adjacent. A complete range of clast sizes is present. The largest clasts observed are $\approx 8$ cm across and most clasts are generally $\leq 5$ cm across. There is no lower clast size limit and clearly discernible clasts grade down to sub-millimetre size and form the matrix to larger clasts. For the purpose of this study, clasts $< 2$ mm diameter are considered to form the matrix. In order of decreasing abundance, clast lithologies include marbles and calc-silicate lithologies, granite dyke fragments, pegmatite dyke fragments, western and/or eastern suite schist, iron formation, graphitic schist, rhyolite intrusion fragments, rhyolite fiamme, and rare sulphide clasts.

Although DP and LP breccia units are composed of similar rock types and have textural similarities, sufficient differences are present to justify separate consideration.
Fig. 4.4a. Drill hole PD0010, 173.6 m. Hand specimen of a typical poorly sorted, matrix supported, discordant polymictic breccia (DP breccia). Angular to sub-rounded clasts of marble (M), pyrite (PY), sulphide-bearing marble (SM) and granite dyke fragments (G) are randomly orientated. The matrix (< 2 mm) consists of fragments of similar lithologies.

Fig. 4.4b. Drill hole PD0025. **Upper**: 544.6 m, **Middle**: 543.0 m, **Lower**: 539.2 m. Hand specimens sampled at different localities across a DP breccia unit that is graded from one margin to next. The upper sample contains coarser, more angular lithic clasts (L) and rhyolite clasts (R) with a distinct orientation. The middle sample is finer grained with minor rounded lithic clasts (L) and a moderately distinct fabric. The lower sample is the finest grained, all lithic clasts are well rounded and the breccia has a distinct fabric.

Fig. 4.4c. Drill hole PD0014, 207.7 m. Hand specimen near the centre of a graded DP breccia unit. Larger, rounded lithic clasts are more common near the centre of this DP breccia unit. Late calcite veins (C) have cross-cut the DP breccia.

Fig. 4.4d. Drill hole PD0031, 196 m. Hand specimen of a fiamme-bearing DP breccia unit. The DP breccia consists of small lithic clasts (L), rhyolite clasts (R) and rhyolite fiamme (F). Rhyolite clasts and rhyolite fiamme have a common orientation. The angle between the drill core axis and the orientation of fiamme is consistent with either a sub-horizontal or sub-vertical orientation.

Fig. 4.4e. Drill hole PD0031, 196 m. Transmitted light photomicrograph of sericite altered rhyolite fiamme (F) in a fiamme-bearing DP breccia unit. Fiamme commonly contain round and embayed quartz crystals (Q).

Fig. 4.4f. Drill hole PD0031, 196 m. Transmitted light photomicrograph of the groundmass of a rhyolite clast in a fiamme-bearing DP breccia unit. The groundmass consists of quartz that micropoikilitically encloses feldspar microlites.
DP breccia units

DP breccia units appear to be restricted to marble and calc-silicate lithologies of the central suite, and marble - calc-silicate lithologies of the eastern suite (Table 4.2), however this may be a function of the restricted location of drill holes.

Table 4.2 Drill hole intersections of DP breccia units. Northing and easting are drill collar coordinates. All depths are down hole depths. The orientation of DP breccia units is not known, therefore, the width is equal to that intersected in drill holes.

<table>
<thead>
<tr>
<th>Drill hole</th>
<th>Northing</th>
<th>Easting</th>
<th>Interval (m)</th>
<th>Width (m)</th>
<th>Fault zone interval (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PD0003</td>
<td>10000</td>
<td>11235</td>
<td>183.10 - 183.2</td>
<td>0.10</td>
<td>182.93 - 187.20</td>
</tr>
<tr>
<td>PD0007</td>
<td>10400</td>
<td>10791</td>
<td>105.20 - 107.65</td>
<td>2.45</td>
<td></td>
</tr>
<tr>
<td>PD0007</td>
<td>10400</td>
<td>10791</td>
<td>215.50 - 215.55</td>
<td>0.25</td>
<td>214.2 - 220.50</td>
</tr>
<tr>
<td>PD0008</td>
<td>9200</td>
<td>10314</td>
<td>223.65 - 223.85</td>
<td>0.20</td>
<td>223.33 - 224.13</td>
</tr>
<tr>
<td>PD0009</td>
<td>9400</td>
<td>10350</td>
<td>231.00 - 231.40</td>
<td>0.40</td>
<td></td>
</tr>
<tr>
<td>PD0009</td>
<td>9400</td>
<td>10350</td>
<td>236.85 - 236.95</td>
<td>0.10</td>
<td></td>
</tr>
<tr>
<td>PD0010</td>
<td>9000</td>
<td>10300</td>
<td>153.60 - 153.66</td>
<td>0.06</td>
<td></td>
</tr>
<tr>
<td>PD0010</td>
<td>9000</td>
<td>10300</td>
<td>158.30 - 160.90</td>
<td>2.60</td>
<td></td>
</tr>
<tr>
<td>PD0010</td>
<td>9000</td>
<td>10300</td>
<td>167.80 - 179.18</td>
<td>11.38</td>
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</tr>
<tr>
<td>PD0011</td>
<td>9200</td>
<td>10210</td>
<td>422.70 - 422.60</td>
<td>0.90</td>
<td>421.20 - 423.90</td>
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<tr>
<td>PD0014</td>
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<td>11140</td>
<td>179.78 - 179.88</td>
<td>0.10</td>
<td>179.66 - 180.00</td>
</tr>
<tr>
<td>PD0014</td>
<td>11200</td>
<td>11140</td>
<td>196.10 - 196.14</td>
<td>0.04</td>
<td></td>
</tr>
<tr>
<td>PD0014</td>
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<td>11140</td>
<td>207.47 - 207.56</td>
<td>0.39</td>
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</tr>
<tr>
<td>PD0016</td>
<td>9392</td>
<td>10663</td>
<td>102.10 - 132.51</td>
<td>30.81</td>
<td></td>
</tr>
<tr>
<td>PD0021</td>
<td>7564</td>
<td>10073</td>
<td>379.24 - 379.96</td>
<td>0.72</td>
<td></td>
</tr>
<tr>
<td>PD0024</td>
<td>8961</td>
<td>10125</td>
<td>322.70 - 325.37</td>
<td>2.67</td>
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</tr>
<tr>
<td>PD0024</td>
<td>8961</td>
<td>10125</td>
<td>385.25 - 385.85</td>
<td>0.60</td>
<td>384.50 - 385.85</td>
</tr>
<tr>
<td>PD0025</td>
<td>7957</td>
<td>10051</td>
<td>536.73 - 545.30</td>
<td>8.57</td>
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<tr>
<td>PD0026</td>
<td>9060</td>
<td>10104</td>
<td>299.80 - 302.75</td>
<td>2.95</td>
<td>296.13 - 303.74</td>
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<tr>
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<td>9293</td>
<td>10025</td>
<td>460.33 - 461.53</td>
<td>1.20</td>
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</tr>
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<td>462.70 - 462.90</td>
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</tr>
<tr>
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</tr>
<tr>
<td>PD0027</td>
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<td>473.15 - 477.17</td>
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<td>473.15 - 477.47</td>
</tr>
<tr>
<td>PD0028</td>
<td>8746</td>
<td>10461</td>
<td>507.23 - 509.60</td>
<td>2.37</td>
<td>507.23 - 518.4</td>
</tr>
<tr>
<td>PD0028</td>
<td>8746</td>
<td>10461</td>
<td>513.91 - 516.00</td>
<td>2.09</td>
<td>507.23 - 518.4</td>
</tr>
<tr>
<td>PD0029</td>
<td>8776</td>
<td>10329</td>
<td>274.02 - 275.20</td>
<td>1.18</td>
<td>274.00 - 280.10</td>
</tr>
<tr>
<td>PD0030</td>
<td>7764</td>
<td>10250</td>
<td>373.90 - 388.25</td>
<td>14.35</td>
<td></td>
</tr>
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<td>PD0031</td>
<td>8023</td>
<td>10300</td>
<td>195.5 - 197.35</td>
<td>1.85</td>
<td></td>
</tr>
<tr>
<td>PD0031</td>
<td>8023</td>
<td>10300</td>
<td>422.50 - 424.80</td>
<td>2.30</td>
<td>418.10 - 424.80</td>
</tr>
</tbody>
</table>

*DP breccia in drill hole PD0016 is intensely weathered and its thickness is uncertain.

DP breccia units vary greatly in drill intersection width from 0.04m to 14.35 m (and possibly 30.81 m) and occur from = 100m to > 500m depth. The orientation and shape of DP breccia units is not known and they cannot be interpolated between adjacent drill holes, or adjacent sections. In general, DP breccia units are more common between 8500 mN and 9500 mN and their drill intersection width is greater to the south of 9500 mN compared to those north of 9500 mN (Fig. 4.5a). The drill intersection width of DP breccias does not show any systematic change with depth (Fig. 4.5b). Contacts between DP treccia units and the host rocks are sharp and curvilinear and fracturing of the host rock adjacent to DP breccia units is common.
Fig. 4.5A - B. Diagrams of the width of DP breccias vs northing (A), and down hole depth vs width of DP breccia (B). DP breccias are more common between 8500 mN and 9500 mN and commonly wider in the south of the Menninnie Dam prospect. DP breccias occur over a range of depths with no systematic relationship between depth and the width of DP breccia bodies.
Many of the clast lithologies present in DP breccias are not present in the adjacent wall rock. Where DP breccias are hosted by a thick sequence of a single rock type, clasts of the same lithology as the host rock are more abundant. Different drill intersections of DP breccia commonly have different proportions of clasts to matrix. DP breccias with ≥ 40% matrix commonly contain smaller, rounded clasts, and DP breccia with a < 40% matrix commonly contain larger, angular to rounded clasts. Clasts in some DP breccia units are laterally graded. Three different types of clast grading have been recognised:

- 'Double coarse margin grading': Smaller, rounded clasts with a higher proportion of matrix occur near the centre of the DP breccia unit and larger, more angular clasts with a diminished matrix content occur near the breccia margins.
- 'Single coarse margin grading': Gradational variation in the size and proportion of clasts to matrix from one margin to the next (Fig. 4.4b).
- 'Coarse centre grading': Small rounded clasts and a higher proportion of matrix occur near the margins of the DP breccia unit, and coarser well rounded clasts with a smaller proportion of matrix near the centre of the DP breccia unit (Figs. 4.4c).

The fine-grained portion of graded DP breccias commonly has a faint fabric with the long axis of clasts oriented parallel to the margin of the DP breccia unit (Fig. 4.4b).

**Fiamme-bearing DP breccia units**

DP breccias in drill holes PD0021 and PD0031 at 379.24 - 379.96m and 195.5 - 197.35m respectively contain rhyolite fiamme* and rhyolite clasts (Fig. 4.4d). Both units have a poorly defined layered fabric orientated = 30 - 60° to the drill core axis (the drill hole is orientated = 60° from horizontal). Layers are typically 0.5 - 5 cm wide and comprise layers of poorly sorted sand to mud sized matrix, and less commonly, layers of matrix supported clasts. Contacts between layers are diffuse but planar. Fiamme-bearing DP breccia units consist of 5 - 15% rounded lithic clasts, typically 2 - 10 mm across, 1 - 5% sub-rounded rhyolite clasts, typically 10 - 50 mm across, and 5 - 15% of light green, sericite altered rhyolite fiamme, typically 2 - 20 mm long, set in a fine grained altered matrix. The orientation of fiamme is consistently parallel to layering. Fiamme have ragged terminations and locally include round and embayed quartz crystals and crystal fragments (< 2 mm across) similar to those in the strongly and weakly altered rhyolite intrusions (Fig. 4.4e). The groundmass of fiamme comprises a fine grained mosaic of quartz, sericite, chlorite and calcite. The groundmass of rhyolite clasts commonly comprises quartz that micropoikilitically encloses partially altered feldspar microlites (Fig. 4.4f). Round and

* Fiamme is derived from the Italian 'fiamma' = flame, and relates to the ragged to wispy flame like shape of clasts. In this study, the term fiamme is used as purely descriptive term without genetic connotations or any implication of hot emplacement or welding. This is in recognition of near identical shapes derived from the sintering together of hot, juvenile pyroclasts and from burial-compaction flattening of cool diagenetically altered pumice lapilli as well as shapes which resemble clasts but are in-fact a product of hydrothermal alteration (Allen, 1988).
embayed quartz crystals and crystal fragments are also scattered throughout the matrix of the fiamme-bearing DP breccia.

**Alteration of DP breccias**
The matrix of DP breccias is commonly pervasively hydrothermally altered and some clasts have an outer altered rind, typically < 3 mm wide. Alteration minerals in order of decreasing abundance are: chlorite, calcite, sericite, minor disseminated pyrite, rare quartz and galena, and trace amounts of fine grained hematite. Marble clasts are more commonly hydrothermally altered than other clast lithologies which are generally weakly altered, or unaltered.

**Association of DP breccias and faults**
Nearly half of the DP breccia units are hosted by zones of fractured host rock or fault breccias (Table 4.2; and Fig. 4.6). Fault breccias are monomict and the clast lithology is identical to the surrounding host rock. Fault breccia clasts are angular and have either a jigsaw-fit texture or a more disrupted texture. Tabular clasts are rare or absent. The contact between DP breccia and fault breccia is typically sharp and curviplanar.

DP breccias not hosted by fault breccias also have sharp host rock contacts, although the host rock is commonly fractured. Tabular host rock clasts at the contact with DP breccias have not been observed, however this may be a function of the small sample size provided by drill core. The contact between DP breccia and fractured host rock is usually inclined to the drill hole axis at such an angle that the contact must be either subvertical or subhorizontal. A subvertical orientation is considered to be more likely as this conforms with the orientation of major faults, such as the faulted contact between the western and central suites (Fig. 3.1).

**Rhyolite-bearing DP breccias**
DP breccias in the southern end of the prospect are spatially associated with rhyolite intrusions. Three of the DP breccia units either partially merge with a rhyolite intrusion or contain clasts derived from rhyolite intrusions (Table 4.3).

Table 4.3 Drill hole intersections of DP breccias that either merge with a rhyolite intrusion or contain clasts of rhyolite intrusion. Northing and easting are drill collar coordinates. ‘From’ and ‘To’ are down hole depths. The orientation of DP breccia units is not known, therefore, the width is equal to that intersected in drill holes.

<table>
<thead>
<tr>
<th>Drill hole</th>
<th>Northing</th>
<th>Easting</th>
<th>From (m)</th>
<th>To (m)</th>
<th>Width (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PD0025</td>
<td>7957</td>
<td>10051</td>
<td>536.73</td>
<td>545.3</td>
<td>8.57</td>
</tr>
<tr>
<td>PD0030</td>
<td>7764</td>
<td>10250</td>
<td>373.9</td>
<td>388.25</td>
<td>14.35</td>
</tr>
<tr>
<td>PD0031</td>
<td>8023</td>
<td>10300</td>
<td>422.5</td>
<td>424.8</td>
<td>2.30</td>
</tr>
</tbody>
</table>

**Descriptions**
The three DP breccia intersections listed in Table 4.3 have different textures and are described separately.
Fig. 4.6. Drill hole PD0027, 473 m (upper) - 479 m (lower). Drill core of a DP breccia unit hosted by a fault zone in dolomitic marble. The upper (UP) and lower (LOW) contacts between the DP breccia unit and brecciated marble are sharp. The fault breccia (FB) is monomineral and consists of partially altered dolomite marble clasts. The angle between the host rock contact and the marble fabric is consistent with a sub-vertical orientation.
Drill hole PD0030
The DP breccia at 373.9 - 388.25m in drill hole PD0030 is partially merged with a rhyolite intrusion (Fig. 4.7a). The centre of rhyolite intrusion contains minor clots of DP breccia and ~5% dispersed DP breccia clasts. The proportion of DP breccia clots and dispersed clasts increases in towards the margin of the rhyolite intrusion, and eventually grades into DP breccia devoid of rhyolite intrusion.

Near the outer margin of the rhyolite intrusion, rhyolite has formed a complex interpenetration into the DP breccia, partially surrounding portions of DP breccia and larger clasts (Fig. 4.7a). Unbroken, penetrating 'fingers' of rhyolite intrusion can be traced back to more coherent rhyolite intrusion. In the DP breccia-dominated outer zones, the rhyolite has formed isolated clots and globules with delicate, irregular, feathery margins (Fig. 4.7b). The feathery margins of rhyolite clots and globules contrast with the angular to sub-rounded shape of lithic clasts in the DP breccia. Quartz phenocrysts within rhyolite clots and globules are commonly intact, however, minor brecciated jigsaw-fit quartz phenocrysts locally occur.

The rhyolite intrusion has a devitrified groundmass of quartz which micropoikilitically encloses feldspar microlites. Feldspar microlites and feldspar phenocrysts have been partially altered to sericite ± calcite. Alteration of the DP breccia is more intense than that of the rhyolite intrusion, and rhyolite clots and globules. DP breccia included in the rhyolite intrusion is partially to pervasively altered to chlorite, sericite and calcite. DP breccia external to the rhyolite intrusion has been pervasively hydrothermally altered to chlorite, calcite, sericite and traces of disseminated pyrite.

Drill hole PD0031
DP breccia at 422.5 - 424.8m in drill hole PD0031 contains abundant, dispersed, randomly oriented, blocky rhyolite clasts that have similar mineralogy and textures to coherent rhyolite intrusions (Fig. 4.7c). Some rhyolite clasts have delicate feathery margins whereas others have more curviplanar margins. The delicate feathery margins of rhyolite clasts contrasts with the ubiquitous angular and sub-rounded shape of lithic clasts. Feldspar phenocrysts, and the groundmass of rhyolite clasts have been pervasively altered to light green sericite, quartz, calcite, disseminated pyrite ± chlorite (Fig. 4.7d). Alteration has destroyed former groundmass textures. The DP breccia matrix is similarly altered to dark green chlorite, calcite, sericite, quartz and disseminated pyrite.
Fig. 4.7a. Drill hole PD0030, 389.8 m (lower right) - 379 m (upper left). Drill core of rhyolite intrusion that has merged with DP breccia. DP breccia clasts and domains of DP breccia (DP) have been incorporated in the rhyolite intrusion.

Fig. 4.7b. Drill hole PD0030, 385.52 m. Hand specimen of peperite. Pinkish quartz- and feldspar-phyric rhyolite intrusion (R) has formed a network of interconnected, irregularly shaped penetrations into DP breccia as well as isolated 'globules' of rhyolite with delicate margins.

Fig. 4.7c. Drill hole PD0031, 422.5 m. Hand specimen of peperite with yellow-green sericite, calcite, chlorite altered quartz- and feldspar-phyric rhyolite clasts (R). Most rhyolite clasts have curviplanar margins, however some have delicate, irregular margins which contrast with the angular to rounded shape of lithic clasts (eg. 'L').

Fig. 4.7d. Drill hole PD0031, 422.5 m. Reflected light photomicrograph of altered peperite with rounded to embayed quartz crystals (Q) in an altered matrix which includes trace amounts of galena (G) and pyrite (P).

Fig. 4.7e. Drill hole PD0025, 544.5 m. Hand specimen of peperitic quartz- and feldspar-phyric rhyolite clasts in a DP breccia. The rhyolite clasts (R) have delicate, fluidal shapes which contrast with the angular to rounded shape of lithic clasts (L). The rhyolite clasts and the lithic clasts have a consistent preferred orientation.
Drill hole PD0025

In contrast to the blocky rhyolite clasts described from drill hole PD0031, dispersed rhyolite clasts within DP breccia in drill hole PD0025 have fluidal shapes (Fig. 4.7e). Fluidally-shaped rhyolite clasts have delicate, elongated shapes, and an internal finely laminated fabric parallel to the elongation direction of the clast (flow banding?). Lithic clasts are also weakly aligned parallel to the orientation of fluidal rhyolite clasts, however the fluidal shape of rhyolite clasts contrast with the more angular to rounded shape of lithic clasts. The groundmass of fluidal rhyolite clasts and the matrix of DP breccia has been partially altered to calcite, sericite, chlorite and trace amounts of disseminated pyrite. Alteration is more intense in the outer = 0.5 mm of fluidal rhyolite clasts.

Interpretation of rhyolite-bearing DP breccias

Rhyolite clasts have similar textures to the weakly and strongly altered rhyolite intrusions, and in one instance (373.9 - 388.25 m in drill hole PD0030), rhyolite clasts are connected to rhyolite intrusion that has merged with DP breccia. The contrasting shapes of rhyolite and lithic clasts is interpreted to indicate formation by different mechanisms. Inclusion of disaggregated and altered DP breccia into less altered rhyolite intrusion (Fig. 4.7a, b) is interpreted to indicate that the DP breccia was unconsolidated and wet at the time of rhyolite intrusion.

The rhyolite-bearing DP breccias have similarities to descriptions of peperite (Fisher, 1960; Williams and Mc Birney, 1979; Hanson and Schweickert, 1982; Hanson and Wilson, 1993). Busby-Spera and White (1987) described two textural types of peperite:

- Blocky peperite
- Globular (fluidal) peperite

The formation of one type over another was considered to be strongly influenced by the nature of the host sediment. Coarser-grained sediments were considered to give rise to blocky peperite whereas globular (fluidal) peperite was considered to have formed in fine-grained sediment where an insulating vapour film formed between the magma and the sediment.

The rhyolite-bearing DP breccia at 422.5 - 424.8m in drill hole PD0031 is interpreted as blocky peperite, whereas that at 536.73 - 545.3m in drill hole PD0025 has a more fluidal shape than the globular peperite described by Busby-Spera and White (1987). Kokelaar (1982) and Brooks (1995) discussed the intrusion of magma into water-saturated, unconsolidated sediments, and suggested that as the water was heated, it expanded and locally vaporised, resulting in turbulent mixing of magma clasts and sediment. If rhyolite clasts at 536.73 - 545.3m in drill hole PD0025 were able to maintain a low viscosity, then fluidal shapes of magma clasts may have resulted through turbulent mixing with the sediment (cf. Branney and Suthren, 1988; Brooks, 1995). Heat retention could have been maintained.
through an insulating vapour film if clasts were surrounded by sufficient fine grained DP breccia matrix (Busby-Spera and White, 1987), or, if sufficient space was generated by expansion of the mixture, clasts could have been mostly surrounded and insulated by steam. The consistent orientation of fluidal rhyolite clasts may have resulted from fluidisation and streaming of the vapour-sediment mix. Fluidisation of DP breccias is discussed in Section 4.4.2.

LP breccia units

Intersections of LP breccias are restricted to the south of the Menninnie Dam prospect (Table 4.4).

Table 4.4 Drill hole intersections of LP breccia. Northing and easting are drill collar coordinates. 'From' and 'To' are down hole depths. The thickness of LP breccia is calculated assuming they are horizontal (drill holes are generally angled at between 50-70° from horizontal).

<table>
<thead>
<tr>
<th>Drill hole</th>
<th>Northing</th>
<th>Easting</th>
<th>From (m)</th>
<th>To (m)</th>
<th>Thickness (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PD0019</td>
<td>8380</td>
<td>10048</td>
<td>273.50</td>
<td>296.50</td>
<td>21.40</td>
</tr>
<tr>
<td>PD0021</td>
<td>7564</td>
<td>10073</td>
<td>324.80</td>
<td>328.79</td>
<td>3.27</td>
</tr>
<tr>
<td>PD0025</td>
<td>7957</td>
<td>10051</td>
<td>201.42</td>
<td>226.57</td>
<td>13.80</td>
</tr>
<tr>
<td>PD0030</td>
<td>7764</td>
<td>10250</td>
<td>201.42</td>
<td>208.03</td>
<td>5.40</td>
</tr>
<tr>
<td>PD0032</td>
<td>7496</td>
<td>10250</td>
<td>279.37</td>
<td>282.20</td>
<td>2.83</td>
</tr>
</tbody>
</table>

Some LP breccias can be correlated between drill holes and adjacent sections (Fig. 4.8). The contact between LP breccias and the underlying Hutchison Group is sharp and unconformable (Fig. 4.9a). Hutchison Group and granite dykes dip steeply to the west and steeply to the east respectively, whereas the orientation of LP breccia units is sub-horizontal (Fig. 4.8). Hutchison Group, metasomatised lithologies, and/or granite dykes directly beneath the LP breccia are extensively fractured with a jigsaw-fit breccia texture (Fig. 4.9a) and the intensity of fracturing diminishes with depth over several metres.

LP breccia clasts are typically angular to sub-rounded. Rounded clasts are uncommon in LP breccias, which contrasts to the rounded shape of many smaller DP breccia clasts. Angular and sub-rounded clasts of the same size and lithology are locally adjacent. As for DP breccias, the matrix of LP breccias is considered to be all clasts < 2 mm and consists of polymictic clasts from < 2 mm diameter to microscopic. The matrix of LP breccias has been partially altered to fine grained carbonate, sericite, minor chlorite and traces of pyrite. Galena has not been observed as an alteration mineral in LP breccias.
Fig. 4.8 Section 8000 mN. Schematic geological interpretation. LP breccia units overlie Hutchinson Group lithologies and have a sub-horizontal orientation (Maximum forward and back extrapolation ± 100m)
Key to interpretative geology plan and 8000 mN cross section

**Quaternary**
- Surficial, transported clays and sand. Minor ferricrete and silcrete.

**Mesoproterozoic**
- Lithic rich Menninnie Dam ignimbrite.
- Layered polymictic breccia (LP breccia)
- Discordant polymictic breccia (DP breccia)
- Rhyolitic intrusions and coherent rhyolite.

**Palaeoproterozoic**
- Granite dykes.
- Central suite. Calc-silicate facies, and silicate facies iron formations.
- Central suite, dolomitic and calcitic marbles. Minor disseminated and chaotically banded magnetite.

Horizontal projection of drill holes
- Prefix 'PD' = diamond drill hole with percussion precollar
- Prefix 'PE' = reverse circulation percussion drill hole
- Prefix 'RP' = rotary air blast percussion drill hole
Fig. 4.9a. Drill hole PD0025, 221.7 m (upper left) - 230.5 m (lower right). Contact (C) between a granite (G) dyke and LP breccia (LP). The granite dyke is strongly altered and fractured / brecciated with a jigsaw-fit texture. The contact between LP breccia and the granite dyke is sharp. The proportion of lithic clasts to matrix is higher near the base of the LP breccia and lithic clasts are commonly larger.

Fig. 4.9b. Drill hole PD0032, 282.0 m. Hand specimen of altered LP breccia. The matrix and some clasts have been pervasively altered to a quartz - carbonate - chlorite - calc-silicate + hematite assemblage.

Fig. 4.9c. Drill hole PD0032, 280.4 m. Transmitted light photomicrograph of quartz - carbonate - chlorite - calc-silicate + hematite altered LP breccia. Sphalerite (S) and galena (G)-bearing sulphide clasts in the LP breccia have undergone partial dissolution and have corroded margins. Sulphides also occur in trace amounts disseminated throughout the matrix (Small dark flecks).

Fig. 4.9d. Drill hole PD0025, 221.9 m (lower) - 209 m (upper). Drill core of LP breccia with the base of each layer marked by blue bars. Layers are defined by the abundance of coarse lithic clasts near the base and finer clasts near the top. The upper layer is altered to clays with minor relict lithic clasts of < 1 cm diameter. The contact between LP breccia and the overlying MD ignimbrite is also marked.

Fig. 4.9e. Drill hole PD0025, 226.1 m (LHS) and 218.8 m (RHS). The sample on the LHS is from near the base of a LP breccia layer and has a high lithic clast to matrix ratio. Note the rounded sulphide clast (S). The sample on the RHS straddles the contact (indicated by arrow) between two LP breccia layers. The top of the underlying layer is finer grained and incorporates more matrix than the base of the overlying layer.
LP breccia in drill hole PD0032 is more intensely altered than other LP breccia intersections. The matrix, and some carbonate clasts, are pervasively altered to dark green chlorite, carbonate, epidote, sericite and tremolite with minor serpentine. Patches of hematite-dusted quartz locally overprint chlorite - carbonate - epidote - sericite - tremolite - serpentine alteration, particularly near the top of the LP breccia (Fig. 4.9b). Traces of galena, sphalerite, chalcopyrite and pyrite are scattered throughout the altered matrix (Fig. 4.9c). Alteration of marble clasts has resulted in a pattern of concentric rings of alteration minerals including epidote, tremolite, chlorite, calcite, locally overprinted by hematite-dusted quartz.

Layering
The LP breccia in drill hole PD0025 is layered (Fig. 4.9d, e). Layering is oriented approximately normal to the drill hole axis which corresponds to approximately 30° from horizontal. Layers range in thickness from 0.9 m to 8.5 m (Fig. 4.10). The tops and bases of layers are either irregular and diffuse, or sharp and near planar. Clasts of ≥5 cm diameter are more abundant in the basal 20 - 50 cm of layers, and are commonly clast supported with interstitial matrix. The interior of layers consists of matrix-supported LP breccia with ≥30% clasts. The size of clasts decreases from ≥0.2 - 4 cm to ≥0.2 - 3 cm across towards the top of each layer with a corresponding increase in the amount of matrix. The tops of layers comprise a 1 - 4 cm thick interval with ≥60% matrix in which clasts are typically 0.2 - 1 cm across (Fig. 4.9e).

The uppermost layer in drill hole PD0025 contains ≥5 - 10% clasts, all of which are 0.2 - 1 cm across (Fig. 4.9d). The matrix and many of the clasts have been pervasively altered to light-grey, puggy clays and quartz fragments. Layering is less well defined in other LP breccia intersections and correlation of layers between drill holes has not been attempted.

Rhyolite-bearing LP breccia
Overlying the Hutchison Group in drill hole PD0031 is an LP breccia that contains fiamme and clasts of crystal-bearing rhyolite referred to as 'rhyolite-bearing LP breccia'. The rhyolite-bearing LP breccia is overlain by non-vesiculated coherent rhyolite whose upper contact was not cored (Table 4.5).

<table>
<thead>
<tr>
<th>Drill hole</th>
<th>Northing</th>
<th>Easting</th>
<th>From (m)</th>
<th>To (m)</th>
<th>Width (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coherent rhyolite</td>
<td>PD0031</td>
<td>8023</td>
<td>10300</td>
<td>171</td>
<td>&lt;146</td>
</tr>
<tr>
<td>Rhyolite-bearing LP breccia</td>
<td>PD0031</td>
<td>8023</td>
<td>10300</td>
<td>180.7</td>
<td>171</td>
</tr>
</tbody>
</table>
Fig. 4.10 Graphic logs of LP breccia, muddy sandstone, pumice breccia, rhyolite-bearing LP breccia, and coherent rhyolite in the five most southerly drill holes. PD0032 is the most southerly drill hole and PD0031 is the most northerly drill hole. Dotted lines connect units between drill holes.
There is a gradational variation in the abundance, size, and texture of rhyolite fiamme and clasts from the base to the top of the rhyolite-bearing LP breccia unit and a corresponding change in the nature of the matrix from lithic dominated near the base, to volcanic dominated near the top. (Fig. 4.11a and Table 4.6). The proportion and size of rhyolite fiamme and clasts increases from the base upwards from ~30% at 1.8m from the base to >95% above 5.39m (Figs. 4.11b-f, 4.12a-f). Rhyolite fiamme and clasts throughout the unit are consistently oriented sub-parallel to the basal contact with the Hutchison Group.

Nearer the base of the unit, rhyolite fiamme are typically <5 cm long, <1 cm wide and have ragged terminations. Fiamme contain crystals and crystal fragments of ~2% quartz (0.2-2 mm), ~2% of 0.1-3 mm K-feldspar and ~1% of 0.1-1.5 mm relict plagioclase. Intact quartz crystals are commonly round and embayed. Fragmented crystals are less common and where present, have a jigsaw-fit texture. The groundmass of fiamme comprises a fine grained mosaic of alteration minerals including quartz, calcite, sericite and chlorite.

Towards the top of the unit, fiamme become progressively larger and are typically longer than can be determined from drill core (>8 cm). Those that are longer than the drill core are referred to as ‘rhyolite clasts’. Rhyolite clasts contain a higher proportion of crystals than fiamme and crystal fragments are rare (ie. ~4% quartz (0.2-2 mm), ~4% of 0.1-3 mm K-feldspar and ~3% of 0.1-1.5 mm relict plagioclase). Above 5.39m from the base, the groundmass of fiamme and rhyolite clasts comprises a fine grained mosaic quartz that micropoikilitically encloses sericite-calcite altered feldspar microlites. The matrix between clasts is predominantly fine grained volcanic fragments with a minor fine grained lithic content and comprises patches of quartz that micropoikilitically encloses sericite-calcite altered feldspar microlites. Due to the presence of a micropoikilitic texture in fiamme, rhyolite clasts and the matrix, the upper and lower margins of rhyolite clasts are not well defined but can be differentiated from matrix by their higher crystal content, and the occurrence of lithic fragments and crystal fragments in the matrix. Micropoikilitic texture is consistent with high temperature devitrification from a glassy precursor (Lofgren 1971a, b, 1974; Anderson, 1969).

An 8 cm wide composite clast of volcanic and lithic breccia is present in the centre of the example illustrated in Figure 4.11d. The upper and lower margins of the composite clast are irregular and feathery to cuspatate. The composite clast comprises ~30% attenuated fiamme (<2 cm long) and ~30% lithic clasts (0.2-3 cm across), in a fine grained groundmass. Fiamme in the clast have a consistent orientation parallel to the to the general fabric of the unit, and are commonly plastically deformed around lithic clasts. Fiamme contain micropoikilitic and rare spherulitic devitrification textures (Fig. 4.11e) consistent with welding and devitrification from a glassy precursor (Lofgren 1971a, b, 1974; Anderson, 1969).
<table>
<thead>
<tr>
<th>Height above Hutchison Group (m)</th>
<th>Proportion of rhyolite clasts</th>
<th>Size of rhyolite clasts</th>
<th>Matrix</th>
<th>Shape of rhyolite clasts</th>
<th>Juvenile rhyolite clast devitrification textures</th>
<th>Breccia fabric</th>
<th>Alteration</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.80</td>
<td>(\approx 30%)</td>
<td>2 mm to 13 cm</td>
<td>Typically 2 to 2 cm long and locally up to 4 cm across</td>
<td>Lithic fragments, minor altered rhyolite fragments and quartz crystal fragments</td>
<td>Clasts of 2 mm to 2 cm have fiamme shapes with foliation margins. Clasts &gt; 2 cm are equant. Flammen with feathery ends weakly cuspate upper and lower margins.</td>
<td>None apparent</td>
<td>Rhyolite clasts and the breccia matrix partially altered to fine grained quartz, calcite, chlorite, sericite, and trace amounts of disseminated hematite. Rhyolite clasts and the breccia matrix are partially altered to fine grained quartz, calcite, chlorite, and sericite.</td>
</tr>
<tr>
<td>5.01</td>
<td>(\approx 60%)</td>
<td>Typically &lt; 1.5 cm across</td>
<td>2 mm to 5 cm</td>
<td>Lithic fragments, minor altered rhyolite fragments and quartz crystal fragments</td>
<td>Flammen with feathery ends weakly cuspate upper and lower margins.</td>
<td>None apparent</td>
<td>Planar fabric defined by the orientation of fiamme.</td>
</tr>
<tr>
<td>5.39</td>
<td>(\approx 90%)</td>
<td>Typically &lt; 1 cm and rarely &gt; 1.5 cm across</td>
<td>40% are &gt; 8 cm long (diameter of the drill core) and 1 - 2 cm wide</td>
<td>Altered rhyolite fragments, minor lithic fragments, quartz and feldspar crystal fragments</td>
<td>Rhyolite clasts have patches with a micropoikilitic texture.</td>
<td>Rhyolite clasts have a foliation fabric parallel to layering.</td>
<td></td>
</tr>
<tr>
<td>6.05</td>
<td>(\approx 95%)</td>
<td>Typically &lt; 2 mm (matrix) and rarely up to 1 cm across</td>
<td>60% are &gt; 8 cm long and 1 - 7 cm wide</td>
<td>Altered rhyolite fragments, quartz and feldspar crystal fragments and minor lithic fragments (1-2%)</td>
<td>Anisometric shapes</td>
<td>Rhyolite clasts and the matrix to clasts have a micropoikilitic texture. Clast boundaries are indistinct.</td>
<td>Weak planar fabric due to orientation of rhyolite clasts and trails of crystal within rhyolite clasts. K-feldspar crystals are less altered.</td>
</tr>
<tr>
<td>9.17</td>
<td>Coherent rhyolite contact</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Micropoikilitic textures</td>
<td>Matrix partially altered to calcite and sericite. K-feldspar crystals are unaltered.</td>
</tr>
</tbody>
</table>
Fig. 4.1a. Drill hole PD0031, 181.6 m (lower right) - 168.8 m (upper left). Drill core of altered carbonate facies iron formation (IF) overlain by rhyolite-bearing LP breccia at 180.7 m (LP) which is overlain by coherent rhyolite at 171.0 m (R). The contacts are marked by arrows.

Fig. 4.1b. Drill hole PD0031, 178.73 m. Hand specimen from near the base of the rhyolite-bearing LP breccia unit. Lithic clasts are (L) mostly angular to sub-rounded. Rhyolite clasts (R) are brick red and comprise smaller fiamme clasts and larger, more equidimensional quartz-feldspar phyric clasts. Both fiamme shaped clasts and larger rhyolite clasts have delicate feathery margins in contrast to the shape of lithic clasts.

Fig. 4.1c. Drill hole PD0031, 177.3 m. Transmitted light photomicrograph of rhyolite-bearing LP breccia from 3.4 m from the base of this unit. Fiamme have been altered to a fine grained mosaic of quartz, sericite, chlorite and carbonate. Round and embayed quartz crystals (Q) and are common throughout fiamme.

Fig. 4.1d. Drill hole PD0031, 175.5 m. Hand specimen of rhyolite-bearing LP breccia unit. Rhyolite clasts have fiamme shapes (F). The centre of the figure consists of an eight centimetre wide composite clast made up of lithic clasts and welded pyroclastics. The upper and lower margins of the composite clast are cuspat with embayments and small fine protrusions into LP breccia.

Fig. 4.1e. Drill hole PD0031, 175.5 m. Transmitted light photomicrograph of the composite clast illustrated in Figure 4.1d. The groundmass comprises fiamme (F) and relict shards that have a pronounced eutaxitic texture and have commonly been plastically deformed around lithic clasts (L).

Fig. 4.1f. Drill hole PD0031, 175.5 m. Transmitted light photomicrograph of the composite clast illustrated in Figures 4.1e. Relict shards (S) have a pronounced eutaxitic texture and have been plastically deformed around lithic clasts (L). Larger shards commonly have ragged terminations of former bubble walls (T).
The groundmass comprises attenuated and plastically deformed relict shards with a well developed compactional foliation parallel to fiamme (Fig. 4.11f). Lithic clasts are angular to rounded, poorly sorted, and consist of Hutchison Group lithologies and granite dyke fragments, similar to those in LP breccia.

At 9.17 m from the base, the rhyolite-bearing LP breccia has a sharp contact with overlying massive, coherent rhyolite (Fig. 4.12c, f).

4.3.3 Coherent rhyolite
Quartz and feldspar phenocrysts in the coherent rhyolite are mostly intact with rare fragmented phenocrysts. The coherent rhyolite has a weak fabric accentuated by trails of quartz and feldspar phenocrysts (Fig. 4.12f). The orientation of the fabric is parallel to that developed in the underlying rhyolite-bearing LP breccia. The upper surface of coherent rhyolite is not preserved in drill core, and the measured thickness of 22.9 m is a minimum thickness.

Coherent rhyolite was not intersected in drill holes 200 - 300m to the north, south and west. There is no drill hole control to the east. The rhyolite-bearing LP breccia unit and the coherent rhyolite occupy the same stratigraphic position as muddy sandstone, and pumice breccia units intersected in drill holes to the south (Fig. 4.10).

4.3.4 Muddy sandstone and pumice breccia units
Overlying the LP breccia is a thin sequence of interbedded, poorly sorted, lithic and muddy sandstone units and pumice breccia units referred to as 'muddy sandstone' and 'pumice breccia' respectively. Muddy sandstone and pumice breccia units were intersected in 5 drill holes in the southern end of the prospect (Table 4.7).

| Table 4.7 Drill hole intersections of muddy sandstone and pumice breccia. All depths are down hole depths. The true width is calculated assuming muddy sandstone and pumice breccia units are horizontal. |
|-----------------|---------------|-----------|--------|
| Muddy sandstone | PD0021        | 324.37    | 324.41 | 0.03   |
| Pumice breccia  | PD0021        | 324.41    | 324.59 | 0.18   |
| Muddy sandstone | PD0021        | 324.59    | 324.80 | 0.18   |
| Muddy sandstone | PD0030        | 201.36    | 201.38 | 0.02   |
| Pumice breccia  | PD0030        | 201.38    | 201.42 | 0.04   |
| Muddy sandstone | PD0032        | 278.50    | 278.53 | 0.03   |
| Pumice breccia  | PD0032        | 278.53    | 279.08 | 0.55   |
| Muddy sandstone | PD0032        | 279.08    | 279.37 | 0.25   |

Muddy sandstone units
Muddy sandstone units consist of laminated, to medium bedded, poorly sorted, non-graded beds that range in thickness from 2 cm to 29 cm (Fig. 4.13 a, b).
Fig. 4.13a. Drill hole PD0032, 279.15 m. Hand specimen of a thinly bedded to thickly laminated muddy sandstone unit.

Fig. 4.13b. Drill hole PD0032, 279.15 m. Transmitted light photomicrograph (x nicols) of poorly sorted muddy sandstone. This unit contains quartz fragments derived from the underlying Hutchison Group or granite dykes (Q1), and round and embayed quartz crystals (Q2).

Fig. 4.13c. Drill hole PD0032, 278.6 m. Hand specimen of pumice breccia unit. Most of this unit is comprised of rhyolite fiamme (F).

Fig. 4.13d. Drill hole PD0032, 279.15 m. Transmitted light photomicrograph of the pumice breccia unit with rhyolite fiamme (F) in a fine-grained finely-laminated matrix. Fiamme are commonly bent around lithic fragments (L), and fragments of round and embayed quartz (Q).

Fig. 4.13e. Drill hole PD0032, 279.15 m. Transmitted light photomicrograph (x nicols) of a rhyolite fiamme in the pumice breccia unit. The interior of fiamme consists of quartz that locally micropoikilitically encloses feldspar microlites and resembles a micropoikilitic texture.
The tops and bases of beds are sharp, and planar to irregular. Clasts of 2 mm - 2 cm constitute = 33% of the unit and consist of lithic fragments (= 19%), fiamme (= 9%), round and embayed quartz crystals and crystal fragments (1%), feldspar crystals and crystal fragments (< 1%). Fiamme are elongated parallel to bedding and contain < 3% round and embayed quartz crystals and crystal fragments of < 1.5 mm diameter, and sericite pseudomorphs of feldspar crystals, in a fine-grained felted intergrowth of quartz, sericite, chlorite and trace amounts of hematite. Quartz crystals and crystal fragments have straight extinction. Lithic fragments consist of quartz with undulose extinction, schist fragments and partially altered diopside and marble. The matrix to muddy sandstone units (= 70%) comprises fine grained felted quartz, carbonate, sericite, chlorite and traces of hematite.

**Pumice breccia units**
Pumice breccia units consist of thickly bedded, to very thinly-bedded, poorly sorted, non-graded beds (Fig. 4.13c) which comprise juvenile volcanogenic clasts (= 50%), lithic clasts (1 - 5%) and matrix = 45%. Pumice breccia units range in thickness from 4 cm to 55 cm (Table 4.7). The tops and bases of pumice breccia beds are sharp, planar and conformable. Volcanogenic clasts have fiamme shapes (0.2 - 5 cm long) and comprise rare embayed to round quartz crystals and crystal fragments of < 2 mm diameter, rare K-feldspar crystals and crystal fragments of < 1 mm diameter in a fine-grained felted groundmass of quartz, sericite, calcite, chlorite and locally, K-feldspar (Fig. 4.13d). The groundmass of some fiamme contain zones of quartz which micropoikilitically encloses fine grained feldspar microlites, and rare spherulitic quartz and K-feldspar intergrowths (Fig. 4.13e), consistent with devitrification from a glassy precursor (Lofgren 1971a, b, 1974; Anderson, 1969). Lithic clasts consist of quartz fragments with undulose extinction and minor granite fragments and partially to pervasively altered grains (marble or calc-silicate ?).

**4.3.5 Geochemistry of rhyolite intrusions, peperites and coherent rhyolite.**
The rhyolite intrusions, the rhyolite component of peperite, and rhyolite clasts in the various LP and DP breccias have a common crystal / phenocryst population of K-feldspar, plagioclase, and embayed to round quartz in similar proportions. Direct physical links cannot be examined from drill core, however, comparison of their geochemistry is a means of testing the hypothesis that they were derived from the same batch of magma. In this section, the geochemistry of some of these units is examined and compared to published analyses of units of the GRV and granitoids of the Hiltaba Suite.

**Methods**
The samples listed in Table 4.8 were jaw crushed to less than 5 mm, and split to obtain a few hundred grams which was then ground in a chrome-steel ring mill. The resultant powder was made into pills and beads for X-ray fluorescence analyses (XRF) using the procedures of Norrish and Hutton (1969). Pills and beads were analysed using a Phillips PW1480
XRF spectrometer at the Geology Department, University of Tasmania. XRF analytical conditions are detailed in Appendix 3.2.

Table 4.8 Samples used for geochemical analysis.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Drill hole</th>
<th>Depth</th>
</tr>
</thead>
<tbody>
<tr>
<td>Strongly altered rhyolite intrusion</td>
<td>PD0031</td>
<td>644m</td>
</tr>
<tr>
<td>Strongly altered rhyolite intrusion</td>
<td>PD0030</td>
<td>403m</td>
</tr>
<tr>
<td>Weakly altered rhyolite intrusion</td>
<td>PD0030</td>
<td>471m</td>
</tr>
<tr>
<td>Coherent rhyolite</td>
<td>PD0031</td>
<td>157m</td>
</tr>
<tr>
<td>Rhyolite component of peperite</td>
<td>PD0030</td>
<td>385.6</td>
</tr>
</tbody>
</table>

Results

Detailed analytical results and are presented in Appendix 3.4 along with analyses of granitoids of the Hiltaba Suite (Flint, 1993b), and units of the GRV (Blissett et al., 1993, Stewart, 1992, and Jagodzinski, 1985). The degree of alkali ion mobility and exchange during cooling, devitrification and hydration of glassy rocks has been described as negligible (Lipman, 1965; Lipman et al., 1969) to variable and complex behaviour (Scott, 1971). Stewart (1992) considered the degree of alkali ion mobility in the Eucarro and Yarda Dacites of the GRV to be minimal and suggested that ion mobility may be lower in coherent lavas and densely welded ignimbrites due to lower porosity, resulting in restricted fluid movement. Assuming negligible alkali ion exchange occurred during cooling and devitrification of the weakly altered rhyolite intrusion, its composition plots in the field for rhyolite defined by Le Bas et al. (1986) on a Na2O + K2O versus SiO2 variation diagram (Fig. 4.14a).

Elements interpreted to have been immobile under most hydrothermal conditions (Rollinson, 1993) have been used to determine the composition of altered lithologies. Their compositions lie near the boundary of the rhyolite and trachyandesite fields defined by Winchester and Floyd, (1977; Fig. 4.14b). The concordance between alkali element and immobile trace element variation diagrams is consistent with interpretation as rhyolite.

Compositions of the rhyolite intrusions, coherent rhyolite and the rhyolite component of peperite plot as a cluster on various immobile element variation diagrams, interpreted to indicate derivation from the same batch of magma (Fig. 4.15a - f). Compositions of granitoids of the Hiltaba Suite, and units of the mature and developmental phases of the GRV plot as overlapping fields on a number of immobile element variation diagrams (Fig. 4.15a - f). Compositions of the rhyolite intrusions, igneous component of peperite and coherent rhyolite lie outside the fields for analyses of the mature and developmental phases of the GRV and close to, or within, the field for analyses of granitoids of the Hiltaba Suite.
The Mesoproterozoic rhyolite intrusion

![Diagram A](image1)

**Fig. 4.14a** Total alkali vs silica (TAS) variation diagram for classification of igneous rocks with least altered sample of rhyolite intrusion plotted (after Le Bas et al. 1986).

**Fig. 4.14b** Immobile trace element ratio variation diagram for the compositions of the strongly and weakly altered rhyolite intrusions, coherent rhyolite and the igneous component of peperite (after Winchester and Floyd, 1977).
Figure 4.15 Immobile element variation diagrams with composition fields for Hiltaba Suite granite (Flint, 1993b), developmental, and mature phases of the GRV (Blissett et al., 1993; Stewart, 1992; Jagodzinski, 1985). The rhyolite intrusions, coherent rhyolite and igneous component of peperite from Menninnie Dam have compositions that lie close to, or within, the composition field for Hiltaba Suite granite and are distinct from the mature and developmental phases of the GRV.
Fig. 4.16a. Drill hole PD0021, 266.1 m (vert.). Basal contact of the MD ignimbrite with a muddy sandstone unit. The muddy sandstone unit (M) is 3 cm thick and overlies a pumice breccia unit (P). The contact between the MD ignimbrite (MD) and the muddy sandstone is sharp and conformable. Fiamme in the pumice breccia are parallel to lithologic contacts. Fiamme within the lower 1 cm of the MD ignimbrite are parallel to the basal contact whereas those a few centimetres higher are at an angle to the contact.

Fig. 4.16b. Drill hole PD0021, 132.54 m (vert.). A thin rind (R) = 8 mm wide comprised of fine relict ash and minor fine grained lithics has coated the margin of a brecciated and altered 16m diameter granite clast (G). The rind differs from the enclosing MD ignimbrite (MD) which contains coarser pyroclastic and lithic clasts.
Fig 4.17 3D Wire mesh model of the basal surface of the Menninnie Dam ignimbrite viewed from the south-east and looking to the north west. The greatest thickness of ignimbrite occurs in the south of the prospect close to the faulted contact between western and central suites (the trace of the fault is shown as a dashed line). The fault zone is interpreted to have been a valley at the time of ignimbrite deposition.
Drill hole PD0021
MD ignimbrite intersection

0 - 123.2 m from surface; Percussion pre-collar

141.3 m Medium red-brown welded lithic-rich ignimbrite. Shards and fiamme define a pronounced eutaxitic texture. Fiamme have granophytic, spherulitic and micropoikilitic devitrification textures (Fig. 4.18b, c).

133.7 m Light yellow-brown welded lithic-rich ignimbrite. Shards and fiamme define a eutaxitic texture. Fiamme have both spherulitic and micropoikilitic devitrification textures with partial alteration of K-feldspar to sericite.

114.5 m Light yellow-green sericite altered welded lithic rich ignimbrite. Shards and fiamme define a strong eutaxitic texture. Fiamme have spherulitic, micropoikilitic and minor axiolitic devitrification textures with partial alteration of K-feldspar to sericite (Fig. 4.18d).

87.1 m Light brown welded lithic-rich ignimbrite. Shards and fiamme define a strong eutaxitic texture. Fiamme and some larger shards have spherulitic and micropoikilitic devitrification textures with partial alteration of K-feldspar to sericite.

70.1 m Light brown welded lithic rich ignimbrite. Shards and fiamme define a strong eutaxitic texture. Fiamme and some larger shards have spherulitic or micropoikilitic devitrification textures. Minor alteration of K-feldspar to sericite (Fig. 4.18e, f).

39.1 m Light yellow-brown welded lithic rich ignimbrite. Individual shards and fiamme have been partially altered to sericite ± chlorite alteration (Fig. 4.18g).

0 m (Base of the MD ignimbrite). Light yellow-brown welded lithic-rich ignimbrite. Relict shards and fiamme are preserved and define a strong eutaxitic texture (Fig. 4.18h and Fig. 4.16a).

Fig. 4.18a  Summary graphic log of textures and alteration mineralogy in the Menninnie Dam ignimbrite in drill hole PD0021.
Fig. 4.18b. Drill hole PD0021, 141.3 m from the base of the MD ignimbrite. Transmitted light photomicrograph of welded lithic-rich MD ignimbrite. Shards and fiamme define a pronounced eutaxitic texture. Many fiamme (F) are curved around lithic fragments (L).

Fig. 4.18c. Drill hole PD0021, 141.3 m from the base of the MD ignimbrite. Transmitted light photomicrograph of welded lithic-rich MD ignimbrite. Intensely flattened shards (S) are commonly bent around small lithic fragments (L).

Fig. 4.18d. Drill hole PD0021, 114.5 m from the base of the MD ignimbrite. Transmitted light photomicrograph (x nicols). The interior of a fiamme contains spherulites (SP) that have partially recrystallised (R). The groundmass has a layered fabric due to the orientation of compacted shards (S).

Fig. 4.18e. Drill hole PD0021, 70.1 m from the base of the MD ignimbrite. Hand specimen. Granite clasts (G) are the most common clast lithology throughout the MD ignimbrite. Fiamme (F) are clearly visible in hand specimen and define a eutaxitic texture.

Fig. 4.18f. Drill hole PD0021, 70.1 m from the base of the MD ignimbrite. Transmitted light photomicrograph (x nicols). Part of the interior of a fiamme. The fiamme has inner margins with a spherulitic texture (S) and a core of quartz and feldspar (C).

Fig. 4.18g. Drill hole PD0021, 39.1 m from the base of the MD ignimbrite. Transmitted light photomicrograph (x nicols). Quartz and sericite altered fiamme (F) contained with a fine grained matrix.

Fig. 4.18h. Drill hole PD0021, Transmitted light photomicrograph of MD ignimbrite from within 10 cm of the basal contact. Relict shards (S) have a pronounced eutaxitic texture and are commonly curved around lithic fragments (L).
There is no apparent change in the length of fiamme throughout the drill core, although those higher up in the MD ignimbrite are commonly thinner and more attenuated (Figs. 4.18b, d, e, f, g, h). The orientation of fiamme varies locally over distances as small as 1 cm (Fig. 4.16a), however, the general orientation of fiamme is = 25 - 45° to the drill core axis, which translates to either a sub-vertical or a sub-horizontal orientation (drill holes are inclined at = 60° from horizontal). Given that the base of the MD ignimbrite is sub-horizontal (Fig. 4.17), and there has been little deformation since the late Palaeoproterozoic (Section 2.3.1), the orientation of fiamme is interpreted to indicate a sub-horizontal flattening orientation. Local perturbations in the orientation of fiamme commonly occur in proximity to lithic clasts with fiamme curving around the margins of lithic clasts.

**Lithic clasts**

Lithic clasts comprise Hutchison Group lithologies, including marble, calc-silicate lithologies, schist, iron formation, graphitic schist, as well as granite fragments, pegmatite fragments and fragments of metasomatized lithologies. Lithic clasts range in size from 2 mm to 19.85 m. Those smaller than 2 mm are considered to be part of the matrix. The shape of clasts varies from angular to sub-rounded with no correlation between the shape of clasts, size or lithology (as well as can be determined from drill core). The concentration of lithic clasts is not homogenous throughout the MD ignimbrite.

The lithic content of the MD ignimbrite in drill holes PD0021, PD0025 and PD0032 has been estimated with subdivision of clasts into two groups:

- Clasts greater than 10 cm
- Clasts between 2 mm and 10 cm

The position and size of lithic clasts greater than 10 cm have been recorded. The proportion of clasts less than 10 cm have been visually estimated at various depths by comparison with the 'grain percentage' diagrams of Terry and Chilingar (1955). The accuracy of visual estimates was periodically checked by sectioning the core length-ways, overlaying 1 mm square graph paper onto the flat surface of the core and tracing all clasts greater than 2 mm. Visual estimates were within 5% of the measured value. Estimated percentages of clasts < 10 cm and measurements of clasts > 10 cm are illustrated in Fig. 4.19 and detailed in Appendix 4.1.

The proportion of clasts 2 mm - 10 cm across ranges from 10 - 80% and systematically increases with depth to = 234 m (drill holes PD0032 and PD0021; Fig. 4.19). Near 234m in drill hole PD0021 the clast size is diminished and again shows a systematic increase towards the base of the MD ignimbrite at 266 m. Clasts > 5m are more common above 214 m (Fig. 4.19). Interpretation of the variation in abundance and size of lithics is discussed in Section 4.4.8.
Drill hole PD0021

Drill hole PD0032

Drill hole PD0025

Fig. 4.19  Diagrams illustrating the percentage of lithic clasts < 100 mm across and lithic clast > 100 mm across throughout the MD ignimbrite for the three drill holes with significant intersections of MD ignimbrite. Lithic clasts < 100 mm diameter have a systematic increase in abundance with depth to ~ 234 m. Lithic clasts > 5 m diameter are more common in zones that also have a higher percentage of lithic clasts < 100 mm diameter. The position of the lithic-rich unit is also illustrated.
Thin, lithic-rich zones are present in the two most southerly drill holes (Table 4.10 and Fig. 4.19).

Table 4.10 Lithic-rich zones in the MD ignimbrite. Northing and easting are drill collar coordinates. 'Depth' is the vertical depth below surface (the land surface is horizontal). Thickness is the vertical thickness.

<table>
<thead>
<tr>
<th>Drill hole</th>
<th>Northing</th>
<th>Easting</th>
<th>Depth (m)</th>
<th>Thickness (m)</th>
<th>Estimated % lithic clasts</th>
</tr>
</thead>
<tbody>
<tr>
<td>PD0021</td>
<td>764</td>
<td>10073</td>
<td>165.39</td>
<td>0.85</td>
<td>80</td>
</tr>
<tr>
<td>PD0032</td>
<td>7405</td>
<td>10035</td>
<td>165.85</td>
<td>0.45</td>
<td>80</td>
</tr>
</tbody>
</table>

Both lithic-rich zones are at similar depths (the land surface is horizontal), have similar lithic clast contents, range of lithic clast types and sizes. The matrix to lithic-rich zones comprises fine lithic fragments devoid of a juvenile volcanic component. The bases of lithic-rich zones are irregular and diffuse over ~ 20 cm and incorporate a juvenile volcanic component. The tops of lithic rich units are sharp, irregular and parallel to the orientation of fiamme in the overlying MD ignimbrite. A lithic-rich zone was not intersected at the same depth in drill hole PD0025, 390 m to the north. The corresponding interval in drill hole PD0030, 200m to the north was percussion drilled and the data cannot be evaluated as percussion samples are 2 m composites.

**Relationship between lithic clast lithology, abundance and size**

Granite clasts are mineralogically and texturally indistinguishable from the deformed granite dykes that locally intruded Hutchison Group and interpreted to be ~ 1745 - 1710 Ma. (Section 3.8.3). Of the 64 clasts > 10 cm across in drill holes PD0021, PD0025 and PD0032, 73% are composed entirely of granite and 92% are composed entirely of granite, or, are composite clasts which included granite (Table 4.11). Granite clasts are also the most common lithology for clasts of 2 mm - 10 cm (Fig. 4.18e).

Table 4.11 Lithology of clasts > 10 cm across in drill holes PD0021, PD0025 and PD0032.

<table>
<thead>
<tr>
<th>Clast lithology</th>
<th>Number of clasts</th>
</tr>
</thead>
<tbody>
<tr>
<td>Granite</td>
<td>47</td>
</tr>
<tr>
<td>Granite and schist</td>
<td>10</td>
</tr>
<tr>
<td>Granite and marble</td>
<td>2</td>
</tr>
<tr>
<td>Schist</td>
<td>4</td>
</tr>
<tr>
<td>Marble</td>
<td>1</td>
</tr>
</tbody>
</table>

Some granite clasts > 10 cm across are moderately fractured and some contain zones of tight jigsaw-fit breccia. Fractures and breccia zones do not extend beyond clast margins. Clasts > 2m commonly have a thin rind separating the clast from the enclosing ignimbrite (Fig. 4.16b). The rind varies from 3 mm to 2 cm in width and consists of greyish tuff (< 2 mm), and minor lithic fragments (< 2 mm). Rinds commonly have a weak lineation fabric parallel to the clast margin and commonly discordant to layering of fiamme outside the rind.
MD ignimbrite matrix
Matrix is defined as all components < 2 mm. The matrix comprises various proportions of lithic fragments and relict glassy shards (excluding the two lithic rich zones, Table 4.10). Flattened, and plastically deformed relict glassy shards are present throughout the MD ignimbrite, including within a few millimetres of the basal contact. More attenuated relict, glassy-shards common occur at higher levels in the MD ignimbrite (Figs. 4.18b, c).

Fragments of round and embayed quartz crystals (with straight extinction) are present within a few metres of the base of the MD ignimbrite. These are identical to those in the underlying muddy sandstone and/or pumice breccia. Throughout the remainder of the MD ignimbrite, fragments of round and embayed quartz crystals are absent. The apparent absence of a juvenile crystal population in the MD ignimbrite matrix concurs with the absence of a crystal population within fiamme.

Evidence for hot emplacement of the MD ignimbrite
The MD ignimbrite has textures which are indicative of hot emplacement and welding (Fig. 4.18b-i). Fiamme are strongly attenuated and commonly plastically deformed around lithic clasts. The interior of fiamme commonly comprise spherulitic, or axiolitic textures, or quartz which micropoikilitically encloses feldspar microlites. Cuspate shards are plastically deformed around small lithic fragments. The presence of spherulites and micropoikilitic quartz and feldspar are interpreted to indicate high temperature devitrification from a glassy precursor of flattened, welded, pumice lapilli (Lofgren, 1971 a and b, Lofgren, 1974, Wright, 1915, Ross and Smith, 1961, Anderson, 1969).

MD ignimbrite in the basal 34 m of drill hole PD0032
The basal 34 m of drill hole PD0032 differs from other intersections of MD ignimbrite in that it is non-welded and selectively, to pervasively altered. The intensity of alteration increases with depth and the alteration minerals show a corresponding systematic change. A summary of the textures and alteration mineralogy of MD ignimbrite in drill hole PD0032 is illustrated in Figures 4.20a-i.

Description (ref. Figs. 4.20a-i)
The basal 34 m of drill hole PD0032 contains a lithic clast content of 54% which is slightly higher than the average lithic clasts content determined for the remainder of the MD ignimbrite (47%). Above ~58.5 m from the base of drill hole PD0032, the MD ignimbrite is welded with a strong eutaxitic texture. Between ~58.5 - 33.5 m from the base, the MD ignimbrite consists of light greenish-brown, chlorite altered, lithic-rich, partly-welded ignimbrite with a moderate eutaxitic texture, and common micropoikilitic and spherulitic devitrification textures within fiamme.
Drill hole PD0032
MD ignimbrite intersection

0 - 99.7 m from ground level; Percussion pre-collar

143.48 m. Light brown, lithic-rich welded ignimbrite. Platey and bubble wall shards are weakly attenuated and some are plastically deformed. Some fiamme have spherulitic and micropoikilitic devitrification textures. Secondary K-feldspar has been altered to sericite and fine grained quartz (Fig. 4.20b).

97.48 m. Light greenish-brown, lithic-rich welded ignimbrite with a strong eutaxitic texture. Shards and fiamme are commonly plastically deformed. Some fiamme have spherulitic and micropoikilitic devitrification textures (Fig. 4.20c).

84.48 m. Light brown, lithic-rich welded ignimbrite with a strong eutaxitic texture. Shards and fiamme are commonly plastically deformed around lithic clasts. Fiamme have micropoikilitic and minor zones of granophytic devitrification textures. Secondary K-feldspar has been altered to sericite, quartz and traces of chlorite. The lithic-rich unit is at 77.5 m from the base of the MD ignimbrite.

51.48 m. Light greenish-brown, chlorite altered, lithic-rich partly welded ignimbrite with a moderate eutaxitic texture. Platey and bubble-wall shards are strongly aligned and many have been plastically deformed. Fiamme vary from weakly to strongly attenuated. Strongly attenuated fiamme have micropoikilitic and spherulitic devitrification textures. Secondary K-feldspar has been altered to sericite, chlorite and calcite (Fig. 4.20d).

31.48 m. Dark green chlorite - carbonate - epidote - sericite - tremolite - serpentine - hematite altered non-welded lithic-rich ignimbrite. Relict pumice lapilli are pervasively altered and do not have a preferred orientation. Alteration of marble clasts has resulted in a pattern of concentric rings of alteration minerals including epidote, tremolite, chlorite, calcite; Fig. 4.20e.

20.48 m. Dark green chlorite - sericite - epidote - tremolite - calcite - hematite altered non-welded lithic-rich ignimbrite. Rare relict shard-like textures have a random orientation. Lithic clasts (other than quartz and feldspar) commonly have indistinct outlines and are pervasively altered. The presence of relict pumice lapilli cannot be determined due to the intensity of alteration. Traces of galena sphalerite and pyrite are scattered throughout the matrix; Fig. 4.20f, g.

17.48 m. Mottled, dark reddish-green chlorite - sericite - epidote - tremolite - quartz - calcite - hematite altered non-welded lithic-rich ignimbrite. Hematite-dusted quartz overprints earlier chlorite - sericite - epidote - tremolite. Relict shard and pumice lapilli are preserved. Lithic clasts other than quartz and feldspar commonly have indistinct outlines and are similarly altered (Fig. 4.20f, g).

4.48 m. Medium red-brown quartz - chlorite - hematite - sericite - epidote - tremolite - calcite altered, non-welded lithic rich ignimbrite. Relict shard-like textures are present and define a weak fabric. Chlorite - sericite - carbonate altered relict pumice lapilli are weakly flattened (Fig. 4.20h, i). The base of the MD ignimbrite is at 243.48 m below ground level.

Underlying units.

Fig. 4.20a Summary graphic log of textures and alteration mineralogy in the Menninnie Dam ignimbrite in drill hole PD0032.
Fig. 4.20b. Drill hole PD0032, 143.48 m from the base of the MD ignimbrite. Hand specimen. Fiamme are common, but not pronounced in hand specimen, and define a weak eutaxitic texture.

Fig. 4.20c. Drill hole PD0032, 97.48 m from the base of the MD ignimbrite. Transmitted light photomicrograph of fiamme (F) and relict shards (S) that are commonly curved around lithic clasts and fragments (L) and define a eutaxitic texture.

Fig. 4.20d. Drill hole PD0032, 51.48 m from the base of the MD ignimbrite. Hand specimen. Light greenish-brown, sericite, chlorite, calcite altered, lithic-rich partly welded ignimbrite with a moderate eutaxitic texture. Fiamme (F) are altered to yellow-green sericite, and minor chlorite and calcite.

Fig. 4.20e. Drill hole PD0032, Transmitted light photomicrograph. 31.48 m from the base of the MD ignimbrite. Fiamme (F) in the MD ignimbrite have been altered to sericite, chlorite and calcite.

Fig. 4.20f. Drill hole PD0032, 17.48 m from the base of the MD ignimbrite. Hand specimen. Mottled, dark reddish-green altered non-welded lithic-rich ignimbrite. Pumice lapilli (P) and some shards (S) are preserved as yellow-green sericite - chlorite - calcite altered relics. Sericite - chlorite - calcite alteration of the MD ignimbrite matrix has mostly been overprinted by dark-green chlorite - sericite - epidote - tremolite - quartz - calcite - hematite alteration. Some marble clasts contain pyrite (PY) and galena (G) as alteration minerals.

Fig. 4.20g. Drill hole PD0032, 17.48 m from the base of the MD ignimbrite. Transmitted light photomicrograph. Pumice lapilli (within the dashed line) are commonly weakly flattened and have been altered to sericite, chlorite and calcite. Pumice lapilli are enclosed in a darker green, chlorite - sericite - epidote - tremolite - quartz - calcite - hematite altered groundmass.

Fig. 4.20h. Drill hole PD0032, 4.48 m from the base of the MD ignimbrite. Hand specimen. Medium red-brown non-welded lithic rich ignimbrite. Early sericite - chlorite - calcite alteration has been preserved in relict pumice lapilli (PL) whereas the MD ignimbrite matrix has been overprinted by dark green epidote - tremolite - chlorite - sericite - carbonate alteration which has been later partially overprinted by red-brown quartz - hematite alteration.

Fig. 4.20i. Drill hole PD0032, 4.48 m from the base of the MD ignimbrite. Transmitted light photomicrograph. Relict shard-like textures (S) are preserved and have been altered to quartz. Some relict pumice lapilli (PL) have been preserved as sericite - chlorite - calcite altered pseudomorphs.
Between $33.5 - 18.5$ m from the base, the MD ignimbrite consists of dark green chlorite-carbonate-epidote-sericite-tremolite-serpentine-hematite altered non-welded lithic-rich ignimbrite. Fiamme are absent and relict pumice lapilli are uncompacted and pervasively altered. Alteration of marble clasts has resulted in a pattern of concentric rings of alteration minerals including epidote, tremolite, chlorite and calcite. Trace amounts of galena, sphalerite and pyrite are scattered throughout the altered matrix. Between $18.5 - 10.5$ m from the base, the MD ignimbrite is non-welded, dark reddish-green, and has been pervasively altered to chlorite, sericite, epidote, tremolite, quartz, calcite, serpentine and hematite. Hematite-dusted quartz has overprinted earlier chlorite-sericite-epidote-tremolite-serpentine alteration. Lithic clasts other than quartz and feldspar, commonly have indistinct margins and are similarly altered. Altered relict pumice lapilli (and shards?) are preserved within hematite-dusted quartz. The basal $10.5$ m of the MD ignimbrite comprises medium red-brown, non-welded lithic rich ignimbrite with more intense hematite-dusted quartz alteration.

4.4 Discussion
In this section an interpretation of the units described in Sections 4.3.1 - 4.3.4 is presented and a genetic model for their emplacement is proposed. This is followed by an interpretation of the MD ignimbrite and the volcanic architecture at the time of its emplacement.

4.4.1 Rhyolite intrusions elsewhere in the Gawler Range Volcanic Province
The geometry of the rhyolite intrusions at Menninnie Dam is uncertain, and they may be either sills, dykes or plugs of restricted extent. The rhyolite intrusions are texturally similar to rhyolite dykes described elsewhere in the Gawler Range Volcanic Province. Giles (1977) described brick-red, alkali-feldspar, quartz, and plagioclase-phyric, rhyolitic to dacitic dykes up to $80$ m wide (Moonamby Dyke Suite) in the Glyde Hill Volcanic Complex in the north-west of the GRV Province. Dykes were considered to have been co-magmatic with the GRV and emplacement of granitoids of the Hiltaba Suite. Watmuff (1973a and b) described similar dykes from the Glenloth Goldfield area in the far north-west of the GRV Province. Rhyolite dykes were described as containing small phenocrysts of euhedral, sericitised K-feldspar, and spherical quartz in a devitrified, fine-grained, reddish-brown matrix with a micropoikilitic texture. Similar dykes occur in the Tarcoola area (Wiltabbie Volcanics; Daly, 1985), and north-west of Corunna (Turner, 1975), and in the Kokatha area, (Blissett et al., 1993). With the exception of Wiltabbie Volcanics, most dykes were interpreted to have intruded units of the GRV and to have been co-magmatic with the Hiltaba Suite (Flint, 1993b). On the basis of their immobile element geochemistry (Section 4.3.5), rhyolite intrusions at Menninnie Dam are also interpreted to be genetically linked to the Hiltaba Suite which constrains their age to $\sim 1589$ Ma. (Creaser and Cooper, 1993).
4.4.2 Interpretation of DP breccias.

DP breccias are hosted by and include clasts of Hutchison Group lithologies, deformed granite dykes, pegmatite dykes and metasomatised lithologies. Hutchison Group lithologies have been subjected to upper amphibolite facies metamorphism whereas DP breccias are neither deformed nor metamorphosed and therefore, cannot have been primary sedimentary breccias.

DP breccias are tentatively interpreted to be pipe-like bodies with a sub-vertical orientation ('breccia pipes') because:

- DP breccias have limited lateral extent and significant vertical extent.
- Many DP breccias occupy fault zones. Fault zones were interpreted to be sub-vertical (Section 3.5.1), therefore, DP breccias are also interpreted to be sub-vertical.

Reference to breccia pipes interpreted to be related to collapse, or associated with volcanic or igneous activity is common in the literature (e.g. Wenrich, 1985; Sillitoe, 1985; Hazlitt and Thompson, 1990). Baker et al. (1986) reviewed the processes proposed for formation of hydrothermal breccia pipes and suggested that the processes that gave rise to breccia pipes can be interpreted from a variety of breccia textures considered in combination. The major features of hydrothermal breccia pipes with emphasis on the characteristics relating to their recognition are summarised in Table 4.12 (modified after Baker et al., 1986).

Inverted cone or ellipsoidal-cylindrical shapes are common for breccias formed by different processes (Table 4.12). In the absence of exposure, or more detailed drilling, it cannot be determined as to whether DP breccia bodies flare towards the surface or not (inverted cone shape). McCallum (1985) conducted scaled experiments on breccia pipe formation. Smaller, less flared pipes were developed in less permeable hosts whereas larger diameter pipes formed in more permeable hosts. The competent and impermeable nature of the host rocks at Menninnie Dam may account for the apparent absence of an inverted cone shape of DP breccia pipes. DP breccias south of 8200 mN are spatially associated with rhyolite intrusions. Formation of peperite at the contact between rhyolite intrusions and DP breccia requires DP breccias to have been wet and unconsolidated at the time of rhyolite intrusion. It is therefore considered likely that DP breccias immediately preceded rhyolite intrusions and could have been genetically related to rhyolite intrusions.

DP breccias have many similarities to the maar / diatreme breccias described by Baker et al. (1986; Table 4.12). The sharp host rock contacts, polymict nature of DP breccias, and the angular to rounded shape of clasts distinguishes them from fault related breccia pipes. The matrix supported nature of DP breccias is similar to that for hydrothermal and maar / diatreme breccias but differs from the commonly clast supported nature of hypabyssal breccias.
Table 4.12 Summary of the major features of hydrothermal breccia pipes with emphasis on those characteristics relating to their recognition (modified after Baker et al., 1986).

<table>
<thead>
<tr>
<th>Hypabyssal Maar / Diatreme</th>
<th>Hydrothermal eruption</th>
<th>Fault related</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>General form</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Steeply dipping cylindrical</td>
<td>Inverted cones, mostly circular, becoming cylindrical at depth.</td>
<td>Pipes of irregular form, ovoid, cylindrical, linear to irregular, generally small scale.</td>
</tr>
<tr>
<td><strong>General environment</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Intrusive related</td>
<td>Volcanic related</td>
<td></td>
</tr>
<tr>
<td><strong>Wallrock contact features</strong></td>
<td>Sharp contacts. Margins commonly contain large, fractured blocks of host rocks.</td>
<td>Sharp complex finger-like contacts.</td>
</tr>
<tr>
<td>Typically sharp, with sub-parallel shingled mapping. Upper sections encased by shatter zones.</td>
<td></td>
<td>Irregular contacts, poorly defined. Strong relation to adjacent fault zones.</td>
</tr>
<tr>
<td><strong>Internal fragment characteristics</strong></td>
<td>Sections of the pipe show wide variation in clast type, size and angularity. Tendency for rounding. Pronounced in interior (granite) and smaller clasts. Metasediments etc.- less well rounded. Usually altered.</td>
<td>Fragments mixed, of near surface wall rocks. Often rounded. Pervasively altered.</td>
</tr>
<tr>
<td><strong>Matrix</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Hydrothermal components</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Replacement of fragments and matrix. Usually completely infilled. Vug of similar mineralogy to the alteration. Mineralisation infilled commonly in vugs and replaced matrix often peripheral.</td>
<td>Replacement of fragments and matrix, especially in upper levels. Mineralisation in breccia matrix and overprinting veins and fissures.</td>
<td>Pervasive mineralisation mostly as replacement of matrix and clasts. Also in overprinting veins.</td>
</tr>
<tr>
<td><strong>Examples</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Formation depth</strong></td>
<td>0.5 - 2.0 km, possibly deeper</td>
<td>500m to surface</td>
</tr>
</tbody>
</table>

NB: Some of the above types incorporate several breccia styles. All incorporate fault block shuffling / 'shatter breccia' style phenomena. Some areas of hypabyssal, maar / diatreme, and to a lesser extent eruption breccias contain areas of collapse. Fragments here are large tabular, clast supported with vug / infill components. This summary does not include collapse of solution styles prevalent in carbonate / Mississippi Valley type regimes.
The depth extent of DP breccias (> 545 m, Table 4.2) is more than the maximum depth attributed to hydrothermal breccias, but similar to that for hypabyssal and maar / diatreme breccias. Sillitoe (1985) reviewed breccia pipes formed in volcano-plutonic arcs. The "phreatic breccias" described by Sillitoe (1985) are interpreted to be similar to the hydrothermal breccias of Baker et al. (1986). Sillitoe (1985) described the dominant feature that separated phreatic breccias from most magmatic hydrothermal breccias was the widespread occurrence of quartz as a cement and replacing the matrix and breccia fragments, as well as the common occurrence of argillic alteration in magmatic hydrothermal breccias. Quartz is not a pervasive alteration mineral or cement in DP breccias at Menninnie Dam and argillic alteration is absent.

Cas and Wright (1987) described diatreme breccias as a "pipe-like volcanic conduit filled with volcanioclastic debris". While DP breccias have many similarities to maar / diatreme breccias, they do not comply with the definition of Cas and Wright (1987) in that they are not obviously volcanic conduits and are not filled with volcanic debris. Other authors (eg. Wolfe, 1986) considered the definition of diatreme to be too restrictive and suggested it should be broadened to include structures such as kimberlite pipes and intrusive breccias.

Sillitoe (1985) applied the term 'porphyry-type breccias' to a group of pipe and dyke-like breccia bodies that are commonly associated with base metal mineralisation (eg. the breccias in the Leadville District, Colorado). Porphyry-type breccias were interpreted to be associated with a particular intrusive phase that commonly post-dated ore deposition. Hazlitt and Thompson (1990) described heterolithic breccia bodies from the Leadville District that locally crosscut manto and chimney* style Pb - Zn - Ag mineralisation. Breccias bodies locally exceed 600 m vertical extent and are localised by high angle faults. Breccias were interpreted to have been associated with rhyolite porphyry dykes and plugs and "to have formed much like diatremes by fluidisation of rock along hydrothermal fluid conduits as the ore-forming magmatic hydrothermal system collapsed and meteoric water encroached" (Hazlitt and Thompson, 1990). Of the four breccia stages identified by Hazlitt and Thompson (1990), stages 1 and 2 have many similarities to DP breccias at Menninnie Dam. Stage 1 breccias are polymict and incorporate angular and rounded clasts in a finer matrix of similar material. Stage 2 breccias have a distinct banding of fragments and matrix and commonly include wispy fragments of "flow banded material", considered to resemble a local flow banded porphyry unit. Breccia pipes and pebble dykes associated with Pb - Zn - Ag manto ore bodies have been observed in a number of other districts including; Tintic and Park City, Utah (Hazlitt and Thompson, 1990), the Potossi Mine, Santa Eulalia District, Chihuahua, Mexico (Miguel et al., 1986) and Butte, Montana (Meyer et al., 1968). On the

* manto' is Spanish for blanket and refers to the sub-horizontal shape of carbonate replacement style Pb - Zn - Ag mineralisation. Chimneys are sub-vertical equivalents.
basis of this discussion, a mechanism is proposed for the formation of DP breccias at Menninnie Dam and presented as part of a genetic model in Section 4.4.7.

**Fluidisation**

The term 'fluidisation' rather than 'gas or fluid streaming' (Wolfe, 1986) is used in this study in keeping with common usage. The polymict nature and lateral grading across some DP breccia bodies (Section 4.3.2) is interpreted to have resulted from fluidisation processes. Burnham (1985) attributed breccia formation in sub-volcanic environments to second boiling and subsequent decompression of the host rock, and calculated breccias could form by these processes at subsurface depths in excess of 1 km. Fracture-generated decompression of the host rocks was postulated to result in increased effervescence of the magma at depth (Burnham, 1985), possibly resulting in fluidisation of material in the lower part of the system and rafting of wall-rock fragments hundreds of metres upwards.

Alternatively, DP breccia formation and fluidisation may have resulted from rapid magmatic heating of ground water and conversion to a gaseous phase. Analytical and numerical solutions to ground water flow and pore pressure increases associated with rapid intrusion of magma into wet rock by Delaney (1982), were interpreted to indicate that greater fluid pressures were generated by rapid intrusion of magma into relatively impermeable rocks at depths of less than 1 km. Hydraulic fracturing of brittle rocks, or rapid migration of heated fluids along pre-existing fractures, faults or lithologic contacts open to the surface, could result in flashing of the fluid to steam above the boiling point - depth curve. Rapid expansion caused by conversion of the fluid to steam could result in brecciation of the host rock and possibly fluidisation of breccia clasts similar to that proposed by Burnham (1985).

Experimental simulation of fluidisation processes lead McCallum (1985) to attribute fluidisation by a gaseous fluid as a possible method of clast transport resulting in homogenous mixing of breccia clasts. McCallum (1985) attributed lateral grading of clasts within modelled breccia pipes to the duration of fluidisation. In short lived pipes, coarser clasts were concentrated near the centre of the pipe and finer clasts nearer the margins due to elutriation of finer particles (equivalent to 'coarse centre grading'; Section 4.3.2). Where fluidisation persisted for longer periods, the reverse grading pattern was found with coarser particles concentrated near the margins of pipes and finer, more rounded, clasts near the centre of the pipe in zones of higher gas velocity attributed to increased particle - particle interaction and clast milling (equivalent to 'double coarse margin grading'; Section 4.3.2). The lateral grading observed in DP breccias at Menninnie Dam is interpreted to have resulted from processes similar to those observed by McCallum (1985).
4.4.3 Interpretation of LP breccias

LP breccias do not fulfil any of the criteria of Holland and Zbinden (1985) for the identification of Precambrian paleosols and there is no evidence of a subaqueous environment having been present during deposition, therefore, LP breccias are not considered to be products of weathering or sub-aqueous deposition.

The poor sorting and weak grading of individual LP breccia layers are common features of subaerial debris flow deposits (Schmincke, 1967; Ballance, 1984; Smith 1986). However, LP breccias have features that are not consistent with a debris flow origin. Pb - Zn - Ag sulphide mineralisation is interpreted to have been present only at depths of > 80 m below the palaeosurface at the time of LP breccia deposition and cannot have been incorporated as clasts in a debris flow (ref. Sections 5.6.7; 5.6.8; and 8.6).

Incorporation of lithic clasts of identical lithology to the underlying Hutchison Group, and Pb - Zn - Ag sulphide clasts, can be interpreted to indicate the LP breccias were locally derived. The local terrain at the time of LP breccia deposition has been established as gently sloping (Fig. 4.17). The rounded shape of many LP breccia clasts is typical of abrasion (Blatt et al., 1980) which is unlikely to have occurred during plug or laminar flow of a debris flow for short distances over gently sloping terrain (Johnson, 1970; Enos, 1977).

Most debris flows incorporate a significant clay component derived from weathering or hydrothermal alteration of the precursor lithologies (Vaught et al., 1981; Scott et al, 1995). LP breccias do not appear to contain a clay component, and lithic clasts do not show any evidence of having been weathered prior to incorporation in LP breccia.

LP breccia bodies have many features in common with less fluidised DP breccias. They have the same clast compositions (including clasts of Pb - Zn - Ag sulphides), a similar clast : matrix ratio, similar matrix composition, and similar textures. Differences lie in the contrasting styles of clast grading, geometry of the units, and alteration mineralogy. Two of the more substantial thicknesses of LP breccia occur in drill holes that also intersected DP breccia bodies of > 8.5m thickness. McCallum (1985) observed that laboratory models of breccia pipes which broke through to the surface formed a bedded cone of ejecta immediately surrounding the pipe, which is similar to surficial deposits associated with maars (Lorenz, 1986; Cas and Wright, 1987; Self et al., 1980).

Although LP breccias are not interpreted to be associated with maar volcanoes or tuff rings, the eruptive mechanism involving interaction of magma and ground water (Kienle et al., 1980; Lorenz, 1986; White, 1991), may have been similar to that which resulted in the formation of the DP and LP breccias at Menninnie Dam. Cas and Wright (1987) noted that the first fragments ejected from maars and tuff rings comprise coarse angular fragments of
country rock. Lorenz (1973) described the ejecta from the Boos Maar, Eifel District, Germany, as constituting 60 - 80% country rock debris. Self et al. (1980) described a coarse grained, lithic fall deposit near the basal Ukinrek maar deposit. Kienle et al. (1980) cite two examples of maar deposits in which non-juvenile clasts were the only material ejected.

LP breccias are interpreted to have resulted from DP breccias stopping their way through to the surface and erupting onto the surface forming proximal breccia deposits with similarities to the basal lithic rich deposits associated with maars and tuff rings (Lorenz, 1973; Self et al., 1980; Kienle et al., 1980). The layering developed in LP breccia may have resulted from sequential eruptions or fluctuations in the intensity of eruptions of DP breccia, similar to deposits observed to have formed during eruptions at Taal volcano, Philippines (Wolfe, 1986). Layering has also been reported from a hydrothermal eruption breccia deposit, but is not common (Hedenquist and Henley, 1985a). The more intense alteration of DP breccias compared to LP breccias is interpreted to have resulted from a sub-surface hydrothermal fluid which did not interact with surficial LP breccias.

4.4.4 Interpretation of the rhyolite-bearing LP breccia
The rhyolite fiamme in the rhyolite-bearing LP breccia are interpreted to have been juvenile vesiculated clasts. The common occurrence of crystal fragments, evidence of high temperature devitrification textures, consistent compaction fabric parallel to the basal contact of the unit, curvature of fiamme around lithic fragments and absence of evidence for a subaqueous depositional environment are interpreted to indicate a subaerial pyroclastic origin for rhyolite clasts. Formation of micropoikilitic textures in rhyolite clasts and fiamme as well as the matrix between clasts in the upper part of the unit is consistent with welding followed by devitrification (Lofgren 1971a, b, 1974; Anderson, 1969). The progressive increase in the size of fiamme and rhyolite clasts is consistent with a decrease in the degree of vesiculation of the parental melt (Fisher and Schmincke, 1984; Cas and Wright, 1987).

The apparent reverse grading of pyroclasts, poor sorting and absence of impact structures are not consistent with a conventional fallout origin (Cas and Wright, 1987). A pyroclastic flow origin is also considered unlikely. If the rhyolite-bearing LP breccia had resulted from a pyroclastic flow, it might be expected to have been laterally extensive and intersected in surrounding drill holes. It could be argued that the deposit generated by a pyroclastic flow was eroded and the rhyolite-bearing LP breccia is an isolated remnant. This is considered unlikely as it would require preferential erosion of a pyroclastic flow deposit with a welded upper surface and preservation of less consolidated (or most likely unconsolidated) LP breccia in drill holes PD0021, PD0030 and PD0032.
Feathery margins of the welded, composite clast at 175.4m in drill hole PD0031 (Fig. 4.11d), are interpreted to indicate the clast was re-fragmented and re-sedimented while still hot, plastic, and partially vesiculated. The composite clast has similarities to descriptions of spatter in felsic and intermediate pyroclastic deposits (Turbeville, 1992; Mellors and Sparks, 1991; Perrotta and Scarpati, 1994.). Most spatter deposits are considered to be vent proximal (Cas and Wright, 1987, Turbeville, 1992; Stevenson et al., 1993) or proximal facies of pyroclastic flow deposits (Turbeville, 1992, Mellors and Sparks, 1991; Perrotta and Scarpati, 1994).

On the basis of this discussion, a mechanism is proposed for the formation of rhyolite-bearing LP breccia and presented as part of a genetic model in Section 4.4.7.

4.4.5 Interpretation of fiamme-bearing DP breccias.
The fiamme-bearing DP breccia in drill hole PD0031 is = 12.8 m (vert.) below the base of the rhyolite-bearing LP breccia in the same drill hole. The ragged terminations of fiamme are interpreted to have resulted from flattening of vesiculated volcanic clasts rather than shredding of magma. The fiamme-bearing DP breccia is interpreted to be the near surface equivalent of the overlying rhyolite-bearing LP breccia. The fiamme-bearing DP breccia in drill hole PD0021 is 43.7m (vert.) below an intersection of LP breccia in the same drill hole, although in this instance, rhyolite clasts are not present in the LP breccia. The absence of rhyolite pyroclasts in the LP breccia may relate to the greater depth at which fiamme-bearing DP breccia occurs in this drill hole and solidification of the magma prior to eruption.

4.4.6 Interpretation of muddy sandstone and pumice breccia
Rhyolite clasts in the rhyolite-bearing LP breccia, muddy sandstone unit and the pumice breccia all contain crystals or crystal fragments of feldspar, and round and embayed quartz. Rhyolite-bearing LP breccia occupies the same stratigraphic position as muddy sandstone and pumice breccia units in drill holes a few hundred metres to the south. These units are correlated, and the muddy sandstone and pumice breccia units are interpreted to be a primary distal facies equivalent of the rhyolite-bearing LP breccia.

The emplacement mechanism of pumice breccia and muddy sandstone units cannot be easily determined from drill core. They may have been deposited by fall out, pyroclastic flow (pumice breccia) or re-sedimentation of ash, pumice lapilli, and lithic fragments. Pyroclastic deposits immediately preceding rhyolite lava flows are a common phenomenon (Newhall and Melson, 1983; Heiken and Wohletz, 1987).

The local presence of micropoikilitic textures in some fiamme in the pumice breccia unit are interpreted to indicate high temperature devitrification of a glassy precursor (Lofgren 1971a, b, 1974, Anderson, 1969). If the pumice breccia unit was derived from a pyroclastic flow, it would be expected to have dissipated heat rapidly due to the small thickness (≤ 55 cm) and
unlikely to have resulted in micropoikilitic devitrification textures. If the pumice breccia was not partially welded, then the devitrification textures may have resulted from secondary welding due to transfer of heat from the considerable thickness of overlying MD ignimbrite. Many fiamme show extreme attenuation which is unlikely to have resulted from cold compactional process alone. The processes and effects of secondary welding have been described by Smith (1960), Christiansen and Lipman (1966), Schmincke (1967), McPhie and Hunns (1995).

4.4.7 Genetic model for the formation of polymictic breccias, coherent rhyolite, muddy sandstone and pumice breccia units

A genetic model is proposed for the rhyolite intrusions, peperite, DP breccias, fiamme-bearing DP breccias, LP breccias, rhyolite-bearing LP breccia, overlying coherent rhyolite, muddy sandstone units and pumice breccia units.

The rhyolite intrusions, rhyolite component of peperite and coherent rhyolite have mineralogical, textual and geochemical similarities and were probably derived from the same batch of magma. (possibly Hiltaba Suite?). DP breccias are interpreted to have formed near vertical pipe like bodies that resulted from interaction of rhyolite intrusions and fault hosted ground waters (Fig. 4.21a-b) similar to the mechanism proposed by Delaney (1982). Rapid expansion due to flashing of the ground water to steam may have brecciated the host rocks and resulted in fluidisation of the resultant breccia with formation of lateral grading (McCallum, 1985). DP breccia pipe formation post-dated early formed Pb-Zn-Ag mineralisation and locally included mineralised clasts (Fig. 4.21b). DP breccia pipes may have been a path of least resistance to the ascending rhyolite intrusion resulting in formation of peperite at the contact of wet DP breccia and the rhyolite intrusion.

Some DP breccia pipes are interpreted to have erupted onto the palaeo-surface resulting in proximal deposition of poorly sorted LP breccias. Repeated eruptions or changes in the intensity of eruptions resulted in crude layering of LP breccia deposits. Muddy sandstone units occupy the same stratigraphic position as LP breccia deposits and are interpreted to distal facies equivalents (Fig. 4.21c).

Some rhyolite intrusions continued to intrude along DP breccia pipes and as they neared the surface decompression of dissolved gas resulted in explosive fragmentation, formation of fiamme-bearing DP breccias followed by eruption of the breccia and rhyolite pyroclasts through earlier accumulations of LP breccia. The resultant deposit is interpreted to have been vent proximal rhyolite-bearing LP breccia (Fig. 4.21d). Pumice breccia is interpreted to be the distal facies equivalent of rhyolite-bearing LP breccia.
G.L. /I
Rhyolite intrusion along fault zone

Interaction of hot magma and groundwater in the fault, or second boiling of magma and formation of DP breccia. Some mineralised clasts included in DP breccia

DP breccia breaches surface and deposits LP breccia as vent proximal, poorly sorted layers. Muddy sandstone unit is deposited distal to the vent

Rhyolite-bearing LP breccia

Vesiculated rhyolite pyroclasts in the rhyolite-bearing LP breccia are overlain by less vesicular coherent rhyolite lava

MD ignimbrite

Units are overlain by the MD ignimbrite

Fig. 4.21 Cartoon (not to scale) of a genetic model for the formation of DP breccia (B) and peperite (D), deposition of the LP breccia (C), muddy sandstone (C), formation of the rhyolite-bearing LP breccia (D), and coherent rhyolite (E).
As the upper vesiculated portion of the rhyolite intrusion was consumed, pyroclasts became less vesiculated forming agglutinated proximal spatter which eventually gave way to coherent rhyolite lava and formation of a restricted flow or dome (Fig. 4.21e). The rhyolite flow or dome, LP breccias and pumice breccias were later overlain by MD ignimbrite (Fig. 4.21f).

Although not interpreted to be a maar volcano, the proposed genetic model is consistent with formation of the DP breccia and deposition of the LP breccia by similar processes to those involved in the formation of maar volcanoes. Spatter, pyroclastic deposits, and coherent lava have been recognised from a number of maar deposits (e.g., Hopi Buttes, Arizona, White, 1991; Tower Hill, Victoria, Cas and Wright, 1987; Ukinrek, Alaska, Self et al.). These are all of basaltic composition, however, similar deposits have been described from vent proximal settings of felsic volcanics (e.g., Cas and Wright, 1987; Turbeville, 1992; Stevenson et al., 1993). The sequence of probable vent breccias (LP breccia and muddy sandstone units), pyroclastics (rhyolite-bearing LP breccia, and pumice breccia units) and rhyolite lava (coherent rhyolite) has similarities to the cycle of silicic dome emplacement described by Heiken and Wohletz (1987).

4.4.8 Interpretation of the MD ignimbrite

In this section, the thickness of the MD ignimbrite, source of lithic clasts realtionship between lithic clast content and welding and distribution of lithic clasts in the MD ignimbrite are discussed and used to interpret the environment of deposition in Section 4.4.9.

The absence of sediments of unequivocally marine or lacustrine origin at the base of the MD ignimbrite and the occurrence of welding to the base of the MD ignimbrite (Fig. 4.18i) are interpreted to be evidence for subaerial emplacement onto a dry landscape, which is consistent with interpretation of the palaeo-geographic setting for the GRV (Blissett et al., 1993).

The abundance of fiamme, relict glass shards, and evidence for hot emplacement are consistent with a pyroclastic origin for the MD ignimbrite. The absence of complete cooling breaks or other significant interruptions during deposition are interpreted to indicate a single cooling unit with a gradation from the base upwards in the degree of welding (Fig. 4.18a-i). The occurrence of unsorted lithic clasts throughout > 260 m of MD ignimbrite and absence of bedding, are not consistent with a fall-out origin. The evidence for dense welding is interpreted to indicate the MD ignimbrite was not deposited by a debris flow mechanism, and the most likely mode of emplacement for this unit is interpreted to be by pyroclastic flow.
Lipman and Christiansen (1964), noted a similar variation in the degree of welding in the basal section of the Yucca Mountain Tuff which they considered to be a thick ignimbrite unit. Sparks et al. (1978) explained the variable welding of ignimbrites using the column collapse model. Eruptions with low gas content, and low gas velocity, were interpreted to result in low collapse heights and little heat loss during collapse, which was interpreted to favour the formation of high emplacement temperatures and densely welded ignimbrites (Sparks et al., 1978). This condition was interpreted to have resulted in less expanded flows which lost only minor amounts of their vitric ash, as appears to be the case for many densely-welded intracaldera ignimbrites (Cas and Wright, 1987).

The original thickness of the MD ignimbrite is not known as it forms the current weathering surface, and a minimum thickness of > 260 m is interpreted from drill core (Table 4.9). Single ignimbrite thicknesses of > 500 m are commonly interpreted as ponded intra-caldera fill sequences (eg. Lipman, 1984; Sawyer and Lipman, 1983; Howells et al., 1991; Branney et al., 1992; Shawe and Snyder, 1988; Lipman et al., 1993). Single units of outflow facies ignimbrites rarely reach thickness comparable to the MD ignimbrite in all but the largest ignimbrite eruptions (eg. Fish Canyon Tuff up to 200m thick; Self and Wright, 1983). Thick outflow facies are likely to have been vent proximal, or ponded in depressions such as valleys or earlier calderas.

**Origin of lithic clasts in the MD ignimbrite**
Lithic clasts in the MD ignimbrite may have either been derived from the surface through gravity in-sliding of country rock along a caldera margin, incorporation of surficial lithics into a pyroclastic flow, or, from the sub-surface as either ballistic particles or entrained in a pyroclastic eruption column. Attributes of coarse lithic breccia associated with pyroclastic flows are summarised in Table 4.13.

<table>
<thead>
<tr>
<th>Surface derivation</th>
<th>Summary of coarse lithic breccias associated with pyroclastic flow deposits.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mesobreccia (Lipman, 1976)</td>
<td>Occur in intracaldera fill sequences and form tabular sheets of a few metres to a few 10's of metres thick which are typically thickest near the caldera margin and thin towards the centre. They comprise lithic clasts of &lt; 1 m and a matrix of lithic-rich micro-breccia of the same composition. Clasts do not have a preferred orientation and units have many similarities to debris flows.</td>
</tr>
<tr>
<td>Megabreccia (Lipman, 1976)</td>
<td>Occur in intracaldera fill sequences, form more bulbous deposits than mesobreccias and are commonly thickest at caldera margins. They comprise lithic clasts of ≥ 1 m diameter and single clasts may approach 1 km in length. Minor pyroclastic material forms the matrix to clasts. Clasts are of similar composition to caldera walls and are interpreted to have resulted from syn-eruptive gravity insliding of over-steepened caldera margins.</td>
</tr>
</tbody>
</table>
Table 4.13 cont.

**Sub-surface derivation**

Proximal, coarse lithic breccia, co-ignimbrite lag breccia, and proximal layer 2bL breccia

Lithic segregation zones within ignimbrites (layer 2bL, of Sparks et al., 1973) may be graded, lack internal stratification, are commonly matrix supported, but rarely exceed 1 m in thickness (Walker, 1985). Ground breccias (Druitt and Sparks, 1982) were interpreted to have segregated out from the strongly fluidised head of pyroclastic flows. They may be graded, lack internal stratification, are clast supported and fines depleted, and seldom more than 1 m thick. Lag breccias are characterized by their coarseness, fines depletion, presence of internal stratification, and are commonly between 1 - 20 m thick (Walker, 1985).

Co-ignimbrite lag-fall breccias

Co-ignimbrite lag-fall breccias formed at, or near, the site of continuous eruption column collapse through accumulation of clasts that were too large, or too heavy for the column to support. Pumice and fine pyroclasts were interpreted to have been transported away from the vent by flows (Wright and Walker, 1977, 1981; Druitt and Sparks, 1982).

Eruptive megabreccia

Shawe and Snyder (1988) described lithic breccias associated with the caldera fill sequence from the Manhattan and Mount Jefferson Calderas, Nye County, Nevada. They described breccia units of >200 m thickness with a lithic clast content between 1 - 40%, and a matrix of poorly consolidated juvenile ash to lapilli size pyroclasts. Lithic clasts of all sizes (microscopic to >200 m) and shapes, with varying degrees of rounding occur side by side in breccia units. Lithic clasts were interpreted to be derived from the vent or conduit, and entrained in the eruption column. Shawe and Snyder (1988) suggested that most (but not all) clasts were in thermal equilibrium with the tuff matrix. Some larger clasts interpreted to be from cooler parts of the vent and in thermal disequilibrium with the tuff have an indurated rind of a few centimetre thick composed of fine-grained, chilled glass shards.

Lithic clasts in the MD ignimbrite

The matrix of the thin lithic-rich breccia lens at 165 m in drill holes PD0021 and PD0032 does not incorporate any juvenile pyroclasts and is similar to descriptions of surface derived mesobreccia units (Lipman 1976). Interpretation as mesobreccia implies a break in accumulation of the MD ignimbrite although cooling breaks are not evident. Mesobreccia emplacement is catastrophic and geologically instantaneous and may have occurred during a short break in deposition of the MD ignimbrite which was not sufficient to have resulted in significant cooling.

The distribution of lithic clasts throughout the remainder of the MD ignimbrite differs from coarse, proximal, lithic breccias described by Walker (1985), in that clasts are much coarser (up to 20 m diameter), the lithic clasts lack obvious stratification, lithic clasts are completely enclosed in a matrix of juvenile pyroclasts, and the lithic-bearing part of the MD ignimbrite (all of it) is considerably thicker. Lithic breccias in the MD ignimbrite are dissimilar to the lithic-rich portions of outflow facies ignimbrites (Ground layers, layer 2bL; Walker, 1985) but have some similarities to descriptions of surface-derived mega and meso breccias associated with caldera collapse (Lipman, 1976; Lipman et al., 1993; Fiske and Tobisch, 1994; Thompson, 1985; Branney and Kokelaar, 1992; Kokelaar et al., 1994; Howells et al., 1991). Lithic breccias in the MD ignimbrite also have similarities to descriptions of the sub-surface derived, proximal vent breccias described by Shawe and Snyder (1988) although differences lie in the welded and non-welded nature respectively of the juvenile volcanic component.
Lithologies similar to the lithic clasts in the MD ignimbrite are interpreted to have been present in both the sub-surface and at the palaeosurface immediately prior to deposition of the MD ignimbrite. Clasts of granitic composition are the most common clast type (Table 4.11) and granite dykes are more abundant in the fault zone separating the central and western suites (Fig. 3.2). Two possibilities are considered:

- Lithologies of the western suite formed a topographic high at the time of emplacement of the MD ignimbrite.
- Vent excavation and flaring occurred within a zone of abundant granite dykes. The fault zone may have been a favourable site for establishment of a vent.

Localised areas containing up to 30% lithic clasts between 5 cm to 50 m across, supported in a matrix of Eucarro or Yardea Dacites occur in the southwestern and southern Gawler Range Volcanic Province (Garner, 1996). Granite clasts with metamorphic textures are the most common lithology, and many show effects of partial melting attributed to a sub-surface explosive origin (Garner, 1996). The sub-surface/surface derivation of lithic clasts in the MD ignimbrite cannot easily be determined. An average lithic content of 47% is determined for the MD ignimbrite (ie. 30% ≤ 10 cm + 17% > 10 cm). Incorporation of cool lithic clasts (surface derived) or hotter lithic clasts (sub-surface) may have resulted in differences in the extent of compaction and welding of the MD ignimbrite. For example, the lithic clast content of the basal 34m of drill hole PD0032 is estimated as 53% and this is the only part of the MD ignimbrite that was not welded.

Eichelberger and Koch (1978) demonstrated that the time required for 10% cool lithic fragments of ≤ 20 cm diameter to reach thermal equilibrium with a 100 m thick ignimbrite erupted at 750°C was around 1 hour which was insignificant compared to the time required to achieve welding (= 1 year). Incorporation of cooler lithics was interpreted to have the same effect as a lower emplacement temperature of the ignimbrite.

Eichelberger and Koch (1979) used the welding model of Riehle (1973) which is based on the assumption that viscous strain occurs. Bierwirth (1982) considered this to be invalid for porous volcanic ash. Riehle et al (1995) re-evaluated earlier modelling, concluded the assumption of viscous strain was valid, and compiled a computer program for modelling of cooling, degassing and compaction of rhyolitic ignimbrites. Parameters necessary for modelling, such as the emplacement temperature, water content, thickness, permeability, and initial porosity of the MD ignimbrite are not known and approximations would invalidate interpretation of sub-surface versus surface derivation of the lithic clasts. A further complication is that welding may have been initiated during ignimbrite deposition and prior to attainment of thermal equilibrium with larger lithic clasts (ie rheomorphism; Branney and
Kokelaar, 1994). On the basis of this discussion, a surface or sub-surface origin for the lithic clasts in the MD ignimbrite cannot currently be determined.

4.4.9 Genetic model for the MD ignimbrite
The thickness of the MD ignimbrite is consistent with either an intracaldera ignimbrite or a large, ponded outflow facies. Lithic clasts within the MD ignimbrite are consistent with a an intracaldera facies ignimbrite irrespective of whether lithic clasts were derived from the surface or the sub-surface. The volcanic stratigraphy at Menninnie Dam of rhyolitic lava ('coherent rhyolite') followed by eruption of the MD ignimbrite is similar to the pre-caldera and main caldera forming events of the 'general caldera cycle' described by Lipman (1984). Based on the above discussion, the MD ignimbrite is interpreted to be the eroded remnant of a ponded, intracaldera ignimbrite.

4.4.10 Distribution of lithic clasts in the MD ignimbrite
Lithic clasts > 10 cm comprise two intervals, each of which has a systematic increase in the abundance and size of clasts with depth (Fig. 4.19). If lithic clasts are interpreted as caldera collapse breccia, then the two zones can be interpreted as distinct phases of caldera margin collapse that occurred during essentially uninterrupted pyroclastic deposition (the MD ignimbrite is a single cooling unit). If the lithic breccias were derived from the subsurface, they can be interpreted as distinct phases of vent excavation during otherwise uninterrupted pyroclastic deposition. Alternatively, a combination of the surface and sub-surface derivation may have contributed to the 'two phases' of lithic clasts.

4.4.11 Timing and duration of Mesoproterozoic intrusive and volcanic events at Menninnie Dam
Emplacement of the rhyolite intrusions, peperite, coherent rhyolite, DP and LP breccias, muddy sandstone and pumice breccia are interpreted to have occurred over a short space of time and were essentially coeval. The LP breccias, muddy sandstone, and possibly the pumice breccia, are interpreted to have been deposited as unconsolidated sediments. Their preservation probably resulted from deposition of the MD ignimbrite soon afterwards. It can be concluded that all of the Mesoproterozoic units preserved at Menninnie Dam were emplaced in a geologically short space of time. The interpretation of a genetic, and therefore temporal, association between the rhyolite intrusions at Menninnie Dam and granitoids of the Hiltaba Suite can be extended to encompass all Mesoproterozoic units at Menninnie Dam and an age of \( \approx 1589 \text{ Ma.} \) is inferred (Creaser and Cooper, 1993).

4.4.12 Timing of metasomatism
In Section 3.8.4, it was suggested that the timing of pegmatite dyke emplacement and metasomatic alteration of marble and calc-silicate lithologies at Menninnie Dam probably occurred during the final phase of KM-D3 \(( \approx 1710 \text{ Ma.}, \) Fanning, 1984). Metasomatised
lithologies occur locally directly beneath LP breccia (Section 4.3.2) and have been incorporated as clasts in DP and LP breccia units and the MD ignimbrite. The LP breccias are interpreted to have been deposited onto the palaeosurface at \( \approx 1589 \text{ Ma.} \) and it is unlikely that metasomatic alteration occurred at the palaeosurface, therefore, an age for metasomatism of \( \approx 1710 \text{ Ma.} \) is considered to be consistent with available data.

4.5 Summary

At Menninnie Dam, rocks younger than 1600 Ma. consist of intrusive and volcanic rocks and breccias interpreted to be the remnants of a deeply eroded cauldron associated with eruption of the GRV and intrusion of the co-magmatic Hiltaba Suite. Rock units include:

- Quartz- and feldspar-phyric rhyolite intrusions ('rhyolite intrusions'), locally with peperitic margins into DP breccia.
- Discordant polymictic breccia dykes ('DP breccia') hosted by Hutchison Group lithologies.
- Layered polymictic breccia ('LP breccia'), deposited on top of Hutchison Group lithologies.
- Rhyolite-bearing LP breccia interpreted to be a mix of DP breccia, LP breccia, and subaerially deposited rhyolite pyroclastic deposits.
- Muddy sandstone, interpreted to be a distal facies correlative of LP breccia.
- Pumice breccia, interpreted to be a distal facies correlative of rhyolite-bearing LP breccia.
- Coherent rhyolite lava that overlies the rhyolite-bearing LP breccia.
- Lithic-rich Menninnie Dam ignimbrite ('MD ignimbrite') interpreted to be the eroded remnant of an intracaldera ignimbrite.

Rhyolite intrusions have textural and geochemical similarities to granitoids of the Hiltaba Suite and porphyritic dykes that occur elsewhere on the northern Eyre Peninsula. Interaction of the rhyolite intrusions with ground water resulted in subsurface explosive fragmentation of the host rocks and formation of near vertical, pipe-like bodies of DP breccia. Vaporisation of ground water adjacent to the rhyolite intrusions resulted in fluidisation, clast mixing, and upwards transport of DP breccia clasts. Locally, the rhyolite intrusion continued to migrate upwards, intruding into wet, unconsolidated DP breccia forming peperitic contacts. Some of the DP breccia pipes vented on to the palaeosurface forming proximal deposits of LP breccia and distal deposits of muddy sandstone. The rhyolite intrusion locally erupted through the same vent, resulting in mixed DP breccia, LP breccia and rhyolite pyroclasts (rhyolite-bearing LP breccia). Pumice breccia units are interpreted to be distal facies correlates of rhyolite-bearing LP breccia. Following eruption of the upper vesiculated portion of the rhyolite intrusion, less vesiculated rhyolite lava was extruded and formed a small dome or restricted flow (coherent rhyolite).
Eruption of the lithic-rich Menninnie Dam ignimbrite mantled earlier deposited units to a depth in excess of 260 m. An intracaldera eruptive setting is interpreted for the MD ignimbrite and all of the intrusive and volcanic units at Menninnie Dam are interpreted to have been emplaced in a geologically short interval of time around 1589 Ma. Their present occurrence is consistent with a deeply eroded cauldron.
Chapter five
Sulphide mineralisation

5.1 Introduction and previous work

The aim of this chapter is to:
- describe the relationship between the sulphides and the host rocks,
- describe the mineralogy and textures of sulphide mineralisation,
- investigate paragenetic relationships of the sulphide and gangue minerals, and
- investigate the distribution of Pb, Zn, and other metals throughout the Menninnie Dam prospect.

Higgins and Hellsten (1986) described the sulphide mineralisation at Menninnie Dam as having formed clusters of 0.5m wide bands, comprising massive pyrite, sphalerite, galena and minor chalcopyrite. Higgins et al. (1990) briefly described the sulphide mineralogy and estimated an inferred resource to 300m depth, of 1.7 million tonnes containing 5% lead, 8% zinc and 100 grams per tonne silver. Sulphides were considered to show ample evidence for remobilisation and metamorphism including: exsolution of chalcopyrite along sphalerite cleavages and sub-grain boundaries, and curved cleavages of polygonal galena. Higgins and Hellsten (1986), Higgins et al. (1990) and Beeson (1990) considered the mineralisation to be syn-sedimentary having accumulated in a restricted basin. A 20 m wide zone of chert-dolomite - sulphide breccia near the geographic centre of the prospect was interpreted to have been a footwall stringer system to the mineralisation (Higgins et al., 1990).

The work by Higgins and Hellsten (1986), Higgins et al. (1990) and Beeson (1990), was based on interpretation of 6 percussion and 14 diamond drill holes. Since these publications, an additional 18 diamond drill holes have been drilled into the Menninnie Dam prospect, and a more comprehensive examination of the sulphide mineralisation can now be made.

Throughout this study, the sulphide mineralisation has been subdivided into three zones (Fig. 3.1):

**Northern zone.** The northern zone is defined as: north of, and including drill hole PD0015 (9391 mN).

**Central zone.** The central zone is defined as: north of drill hole PD0032 (7406 mN) and south of drill hole PD0015 (9391 mN).

**Southern zone.** The southern zone is defined as: south of, and including drill hole PD0032 (7406 mN).

The basis for these subdivision relates to differences in the style of mineralisation, gangue mineral assemblage, and sulphide textures. Sulphide mineralisation which is not
incorporated in the above subdivisions includes; trace amounts of disseminated pyrite associated with sericite - chlorite - quartz alteration of deformed granite dykes in the shear system bounding the western and central suites (Section 3.5.1), trace amounts of disseminated pyrite in weakly altered rhyolite intrusions (Section 4.3.1) and trace amounts of pyrite associated with metasomatic alteration (Section 3.7.1). These occurrences are considered to be unrelated to Pb - Zn - Ag sulphide mineralisation at Menninnie Dam and will not be considered further in this chapter. The sulphide and gangue mineralogy in the northern and central zones, and the southern zone are described separately.

5.2 Mineralisation in the northern and central zones

The major occurrence of Pb - Zn - Ag sulphide mineralisation at Menninnie Dam is hosted by marble, calc-silicate lithologies, and mixed marble - calc-silicate lithologies of the central suite, and by similar units in the eastern suite. In the northern zone, four bands (0.1 - 2m wide) of galena - sphalerite - pyrite - chalcopyrite mineralisation hosted by metasomatised marble were intersected in drill holes PD0006 and PD0016 (Fig. 3.1; Appendix 3.1). Sulphide bands were not intersected at the same stratigraphic position in drill hole PD0012 which lies midway between drill holes PD0006 and PD0016. Most of the Pb - Zn - Ag mineralisation occurs in the central zone hosted by units of the central suite. South of 7600 mN, the central suite thins and pinches out against the fault bounding the western and central suites, and minor amounts of sulphide mineralisation are hosted by marble and calc-silicate lithologies of the eastern suite.

Pb - Zn - Ag sulphides appears to occupy a 'window' between 80 - 400m depth (vertical), however, this may be a function of the extent of drilling. The deepest mineralisation encountered is a 6 cm wide vein at 643m (vert.) in drill hole PD0015. The upper limit of the 'window' marks the average extent of weathering which ranges from 8m in drill holes PD0032 and PD0021 to 344m in drill hole PD0019. The base of weathering is often marked by a 1 - 20m thick zone of secondary Pb - Zn - Ag enrichment. Sulphides other than pyrite are absent in this zone and Zn, Pb and Ag are probably associated with iron and manganese oxides or adsorbed onto clay minerals. This zone of secondary Pb - Zn - Ag enrichment was percussion drilled and samples are not suitable for the purposes of this study.

Minor Pb - Zn - Ag sulphide mineralisation locally overprints cataclasites in the western suite (Fig. 5.1a) and contributes to the matrix of brecciated granite dykes in both the western and central suites. DP breccias contain minor amounts of Pb - Zn - Ag sulphide, either as discrete clasts of galena - sphalerite - pyrite - chalcopyrite mineralisation, or as pyrite, minor galena and traces of sphalerite and chalcopyrite associated with alteration of both lithic clasts and the matrix (Section 4.3.2). Rare clasts of galena - sphalerite - pyrite - chalcopyrite mineralisation occur in LP breccias. The matrix of LP breccias has locally been altered to an assemblage that includes trace amounts of pyrite (Section 4.3.2).
Fig. 5.1a. Drill hole PD0026, 197.3 m (central zone). Pyrite 1, sphalerite 2, galena 1 and quartz 1 have overprinted cataclasite in the western suite shear zone. Breccia fragments are comprised of metamorphic quartz (Q) veins.

Fig. 5.1b. Drill hole PD0013, 131.8 m (northern zone). Hand specimen of a 5 - 6 cm wide sulphide vein in massive diopside. The attitude of the vein is \( \approx 90^\circ \) to metamorphic layering developed in the diopside host rock.

Fig. 5.1c. Drill hole PD0016, 273.4 m (northern zone). Hand specimen of a sulphide vein in metasomatically altered marble. The vein has a 2 cm wide alteration halo with bands of pink hematite-dusted metasomatic microcline (M) surrounded by light yellow - green sericite 1 (SE). Pyrite 1 (P) is more abundant in the centre of the vein and galena 1 (G) is more abundant nearer the vein margin. Sphalerite 1 and 2 (S) have formed overgrowths on pyrite 1. Sphalerite 1 and 2 are more abundant in a zone between pyrite 1 and galena 1. The core of the vein has been infilled with Ca-Mn-Mg-Fe carbonate 1 (CA). Larger growth zoned sphalerite crystals have locally attained 9mm across (9MM).

Fig. 5.1d. Drill hole PD0008, 326.0 m (central zone). Breccia developed in stockwork mineralisation hosted by banded calc-silicate - marble - quartz rock. Marble and calc-silicate bands have been replaced by sulphide and gangue minerals. Metamorphic quartz (Q) bands are unaltered and form breccia fragments.

Fig. 5.1e. Drill hole PD0006 (northern zone). Hand specimens. upper 257.62 m - 258.03m, lower 258.4 - 258.68 m. Upper (UH) and lower (DH) contacts of a sulphide vein (S) hosted by metasomatic marble (M). Dotted lines are drawn parallel to the metamorphic fabric of the host marble. The vein has a 2 - 3 cm alteration halo (AH) of fine-grained, altered metasomatic marble. Mineralising fluids have permeated a fracture (F) in metasomatic marble near the up hole contact of the sulphide vein.
Strongly altered rhyolite intrusions and peperite contain trace amounts of disseminated pyrite, sphalerite, galena and chalcopyrite (Section 4.3.1).

Approximately one square metre of weathered, silicified marble subcrop (or calc-silicate lithology?) with 1 - 3% finely disseminated pyrite and trace amounts of galena and sphalerite was uncovered during shallow excavations of a drill sump (<1m deep) at 9000 mN, 10250 mE which is close to the faulted contact between the western and central suites.

5.2.1 Relationship between sulphides and host rocks
Mineralisation is not confined to any particular lithological horizon, nor does it parallel stratigraphic layering. Mineralisation is not directly associated with the major north-south fault separating the central and western suites, however, there is a relationship between mineralisation and deformation which will be discussed in more detail on Section 5.6.3. It is not possible to define the shape, or lateral extent of mineralised zones as drill intersections cannot be meaningfully connected between adjacent drill holes and this may, in part, result from the wide spacing between drill holes (typically 200 m). The approximate extent of known mineralisation is indicated on Figure 3.1.

Sulphides have formed veins (Figs. 5.1b, c), anastomosing stockworks and breccias (Fig 5.1d) which crosscut marble, calc-silicate lithologies and mixed marble - calc-silicate lithologies. Vein mineralisation is more common north of 9300 mN and south of 8200 mN whereas stockwork, breccia and vein mineralisation occurs between 8200 mN and 9300 mN.

The width of veins varies between 1 cm to 2m. The orientation of individual sulphide veins cannot be accurately established due to the unorientated nature of drill core, however it is likely they have a wide variety of orientations as the angle between vein margins and the long axis of the drill core varies from parallel to 90°. The contact between veins and the host rock is sharp and defines an irregularly shaped horizon with an alteration halo that rarely extends beyond a few centimetres into the host rock (Fig. 5.1e). Sulphide and gangue minerals within veins are mostly intimately intergrown and veins do not have crustiform banding or cockscomb textures. Some veins hosted by massive diopsidic units have poorly developed banding that comprises irregularly shaped, millimetre to centimetre wide lenses of sulphide and minor gangue minerals, alternating with similarly shaped lenses in which gangue minerals are more abundant than sulphide minerals (Fig. 5.1b). Some marble hosted veins also have a poorly developed mineralogical banding (Fig. 5.1c).

Stockwork and breccia zones extend vertically for tens of metres and are commonly better developed in host rocks of calc-silicate lithology (mostly diopside, or diopside - quartz) and mixed marble - calc-silicate lithology compared to marble (Fig. 5.1d). The host rock
surrounding stockwork zones has a well developed fracture pattern, and locally, a breccia texture. The network of veins in stockwork zones are commonly poorly defined as the host rock in between veins has undergone selectively-pervasive alteration with preferential alteration of marble followed by calc-silicate lithologies, whereas quartz bands have remained unaltered (Fig. 5.1d).

Alteration of the host rock between stockwork veins has resulted in breccia textures (Fig. 5.1d). Breccias are poorly sorted, non-graded, and matrix supported. The proportion of clasts to matrix is variable and ranges between 10 and 80%. Some clasts have a weak preferred orientation, commonly orientated 45 - 70° to the drill core axis, which translates to either a sub-vertical or sub-horizontal orientation (drill holes are commonly inclined 60° from horizontal). Breccia clasts range from angular to sub-rounded and consist of metamorphic quartz, and minor marble and calc-silicate lithologies identical to that of the surrounding unaltered host rock (Fig. 5.1d). Clasts of marble and calc-silicate lithologies have diffuse corroded margins that 'blend' into the matrix and are more likely to be sub-rounded, whereas metamorphic quartz clasts have sharp margins and are invariably angular. The breccia matrix comprises abundant, small remnant clasts of highly corroded marble and calc-silicate lithology (< 2 mm across), angular quartz fragments, and sulphide and gangue minerals. Unlike vein mineralisation, the matrix to stockwork mineralisation hosted by calc-silicate lithologies commonly has a significant proportion of cryptocrystalline quartz which locally comprises up to 95% of the rock.

5.2.2 Sulphide and gangue mineralogy
Sulphides consist of a simple assemblage of pyrite, sphalerite, galena and chalcopyrite. Pyrrhotite has been described as a minor phase by Higgins et al. (1990) but has not been identified from this study following examination of 167 polished thin sections covering all diamond drill holes. No sulphate minerals have been identified throughout the Menninnie Dam prospect. Gangue minerals include; dolomite, phlogopite, talc, chlorite, adularia, sericite, quartz, rhodonite, fluorite, Ca-Mn-Mg-Fe carbonate, hematite and calcite.

5.2.3 Sulphide and gangue mineral textures and their paragenesis
Sulphides hosted by marbles have different textures to those hosted by calc-silicate lithologies, therefore, mineralisation hosted by different rock types is described separately. Descriptions of sulphide and gangue minerals follows a paragenetic sequence, starting at the first formed mineral. Minerals of the same composition that appear to have formed at different times during the mineralising event are assigned a number following the mineral name with '1' representing the first formed phase and 2, 3 ... representing progressively later phases. A summary of the sulphide and gangue mineral paragenesis applicable to the northern and central zones is illustrated in Figure 5.2
Fig. 5.2. Paragenesis of sulphide and gangue minerals in the northern and central zones. Thicker lines represent more abundant minerals, and dotted lines represent trace amounts of minerals of uncertain paragenesis. The grey bars have no significance other than to assist the reader.
Sulphides hosted by marbles

Dolomite
Within 1 - 2 cm of the margin of sulphide veins hosted by calcitic marble, the marble has been altered along crystal margins, and to a lesser extent along twin planes and replaced by finely crystalline dolomite rhombs, typically 10 - 50 μm across (Fig. 5.3a). The size of dolomite rhombs is increased closer to the centre of the sulphide vein, and locally attains 0.5 mm across. Replacement by finely crystalline dolomite has not occurred in dolomitic marbles.

Pyrite 1
Pyrite 1 consists of disseminated subhedral to euhedral crystals, typically 1 - 20 μm across and larger aggregates. In calcitic marble host rocks, deposition of finely crystalline pyrite 1 followed, and commonly replaced finely crystalline dolomite (Figs. 5.3b, c). In dolomitic host rocks, finely crystalline pyrite 1 replaced the host dolomitic marble along crystal margins. Pyrite 1 locally forms small atoll structures with cores of later sulphide minerals (Fig. 5.3d), many of which have relict rhombic shapes inherited from replacement of finely crystalline dolomite. Replacement of dolomite by pyrite 1 is illustrated in Figures 5.3e, f where replacement has been halted by replacement of dolomite by finely crystalline quartz 1.

Small pyrite 1 crystals have locally coalesced forming randomly oriented, irregularly shaped aggregates and larger, euhedral pyrite crystals. Contacts between coalesced crystals of pyrite 1 are rounded to irregular with no evidence of polygonal crystal contacts or 120° triple junctions. Some larger pyrite 1 aggregates and euhedra have been fractured or have a jigsaw-fit breccia textures with a matrix of paragenetically later sulphide minerals (Fig. 5.3h). Pyrite 1 crystals hosted by metasomatically altered marble commonly either overgrow and / or contain inclusions of metasomatic minerals such as andraditic garnet, epidote and microcline (Fig. 5.3g).

Chlorite 1
Trace amounts of light green chlorite 1 locally occurs as overgrowths on pyrite 1. Chlorite 1 has only been observed near the contact between sulphide minerals and the host rocks and is not present within the central part of veins or stockwork breccias.

Adularia and hematite
Adularia has been identified by its rhombic shape, common finely crystalline inclusions of hematite and other unidentified minerals (typically < 1.0 μm), and by electron microprobe analysis (Table 5.1 and Appendix 5.1). Adularia typically forms rhombic crystals of 0.1 - 1 mm across, and rarely exceeds 1 - 2% of mineralised zones.
Fig. 5.3a. Drill hole PD0017, 319.5 m (central zone). Photomicrograph, crossed nicols. Calcitic marble (M) in the outer margin of a 2 cm wide alteration halo surrounding a sulphide vein has been altered along grain boundaries to fine-grained dolomite (D).

Figs. 5.3b, c. Drill hole PD0017, 319.5 m (central zone). Photomicrographs; (5.3b) transmitted light, (5.3c) reflected light. Fine-grained pyrite 1 (P) has replaced fine-grained dolomite and surrounds corroded marble crystals (M).

Fig. 5.3d. Drill hole PD0008, 311.4 m (central zone). Photomicrograph, reflected light. Pyrite 1 (P) has locally formed small atoll structures with cores of later sulphides including galena (G). Some of the larger atolls have a relict rhombic shape and have pseudomorphed dolomite.

Figs. 5.3e, f. Drill hole PD0019, 298.2 m (central zone). Photomicrographs, (5.3e) transmitted light, (5.3f) reflected light. Relict dolomite has been partially replaced along cleavage planes by pyrite 1, and resulted in formation of pyrite atoll structures (A). Not all of the fine pyrite grains are at the surface, and visible in Fig. 5.3f. Some scattered pyrite 1 grains have retained the relict rhombic shape of dolomite (A). Quartz 1 (Q) has replaced remaining dolomite and overgrown pyrite 1.

Fig. 5.3g. Drill hole PD0015, 731 m. Photomicrograph, combination of reflected and transmitted light. Pyrite 1 (P) has overgrown andraditic garnet (G) and hematite-dusted microcline (M). Andradite garnet and microcline formed by metasomatic alteration of marble and calc-silicate lithologies.

Fig. 5.3h. Drill hole PD0011, 452.2 m (central zone). Photomicrograph, reflected light. Fractures in brecciated pyrite 1 (P) have been infilled with sphalerite 2 (S), galena 1 (G) and Ca-Mn-Mg-Fe carbonate 1 (C). Galena 1 has a straight cleavage pattern and has not been deformed.
Contact relationships between chlorite 1 and adularia have not been observed and adularia locally occurs as overgrowths on pyrite 1 or as overgrowths on rounded microcline 'grains' formed by metasomatic alteration of disseminated biotite (Fig. 5.4a, b and Section 3.7.2). Although trace amounts of finely disseminated hematite occur throughout adularia, hematite alteration of pyrite 1 has not been observed.

Table 5.1 Average of 21 analyses of adularia from Menninnie Dam. An adularia analysis from Switzerland is listed for comparison. All oxide values are wt.%.

<table>
<thead>
<tr>
<th></th>
<th>SiO₂</th>
<th>Al₂O₃</th>
<th>K₂O</th>
<th>Na₂O</th>
<th>CaO</th>
<th>Fe₂O₃</th>
<th>SrO</th>
<th>BaO</th>
</tr>
</thead>
<tbody>
<tr>
<td>*1 Menninnie Dam</td>
<td>64.18</td>
<td>18.64</td>
<td>15.19</td>
<td>0.25</td>
<td>0.01</td>
<td>0.04</td>
<td>0.20</td>
<td>0.10</td>
</tr>
<tr>
<td>*2 Menninnie Dam</td>
<td>64.28</td>
<td>19.19</td>
<td>15.30</td>
<td>0.92</td>
<td>0.11</td>
<td>0.09</td>
<td>0.11</td>
<td></td>
</tr>
</tbody>
</table>

Sericite 1

Adularia has been partially altered to sericite 1 and in many instances, sericite 1 forms an outer rind on adularia crystals (Fig. 5.4c). Sericite 1 comprises < 1% of mineralised zones and is locally more abundant (1 - 5%) in the 0.5 - 3 cm wide alteration halo that surrounds veins hosted by metasomatically altered marble (Fig. 5.1c). Sericite flakes are typically less than 4 μm long and can easily be distinguished from muscovite flakes present prior to the mineralising event that occur as 20 - 100 μm long platelets overgrown by all sulphide minerals, including pyrite 1 (Figs. 5.4d, e).

Sphalerite 1 and chalcopyrite 1

Three phases of sphalerite are present in the northern and central zones at Menninnie Dam referred to as sphalerite 1 - 3. Sphalerite 1 commonly occurs as a dark, corroded, growth zoned core to later sphalerite 2 (Fig. 5.5a). The dark colour of sphalerite 1 relates to a high proportion of small chalcopyrite 1 and minor pyrite 1 inclusions (typically < 5 μm across) in light yellow - brown sphalerite. Inclusions of chalcopyrite 1 define growth zones whereas the small inclusions of pyrite 1 are randomly distributed throughout sphalerite 1. Sphalerite 1 is rarely in contact with earlier minerals but has locally mantled sericite 1 rims on adularia or in the absence of sericite 1, sphalerite 1 has mantled corroded crystals of adularia (Fig. 5.5a). Sphalerite 1 constitutes less than 5% of the sphalerite present at Menninnie Dam.

Sphalerite 2 and chalcopyrite 2

Sphalerite 2 forms pale to light yellow - brown isolated, and more commonly, coalesced growth zoned crystals that typically range in size from 50 μm to 2 mm across (Fig. 5.5a, b). Crystals of up to 9 mm across are locally present with growth zoning clearly visible in hand specimen (Fig. 5.1c). Growth zoning is well developed in most sphalerite 2 crystals and defined by slight colour variations and finely crystalline inclusions chalcopyrite 2 (typically < 5 μm; Fig. 5.5b). Inclusions of chalcopyrite 2 also occur along cleavage planes in sphalerite 2 and as randomly scattered crystals (Fig. 5.5c).
Figs. 5.4a, b. Drill hole PD0016, 262.77 m (northern zone). Photomicrograph; (5.4a) transmitted light, crossed nicols, (5.4b) reflected light. Subrounded 'grains' of metasomatic microcline (M) have been overgrown by adularia (A). Microcline grains in contact with pyrite 1 (P) do not have an intervening overgrowth of adularia. Adularia and pyrite 1 are enclosed within sphalerite 2 (S). Adularia in contact with sphalerite 2 has corroded margins.

Fig. 5.4c. Drill hole PD0016, 262.77 m (northern zone). Photomicrograph transmitted light, crossed nicols. Subrounded 'grains' of metasomatic microcline (M) have been overgrown by adularia (A) and enclosed within sphalerite 2 (S). The margins of adularia have been partially altered to sericite 1 (SE) prior to being overgrown by sphalerite 2.

Figs. 5.4d, e. Drill hole PD0016, 262.77 m (northern zone). Photomicrograph; (5.4d) transmitted light, crossed nicols, (5.4e) reflected light. Metamorphic muscovite (M) locally occurs as inclusions in pyrite 1 (P), sphalerite 2 (S) and galena 1 (G). Metamorphic muscovite has a consistent orientation which is not present in sulphide minerals.
Fig. 5.5a. Drill hole PD0016, 262.77 m (northern zone). Photomicrograph, transmitted light. Adularia rims (A) formed on metasomatic microcline (M) have been corroded and overgrown by sphalerite 1 (S1) which contains abundant inclusions of opaque, fine-grained chalcopyrite 1. Sphalerite 1 forms corroded cores to light yellow-brown sphalerite 2 (S2). Sphalerite 2 forms coalescing growth zoned crystals defined by inclusions of chalcopyrite 2 (C2).

Fig. 5.5b. Drill hole PD0006, 258 m, (northern zone). Photomicrograph, transmitted light. Growth zones in sphalerite 2 (S2) are defined by inclusions of fine-grained chalcopyrite 2 (opaque) which commonly have a gradational distribution across growth zones, and a higher proportion of inclusions occurs near the outer margin of each growth zone (arrow).

Fig. 5.5c. Drill hole PD0011, 452.2 m, (central zone). Photomicrograph, transmitted light. Small opaque inclusions of chalcopyrite 2 are scattered throughout sphalerite 2 (S2) as well as concentrated along cleavage planes. A small flake of hematite (H) has been included in sphalerite 2.

Fig. 5.5d. Drill hole PD0016, 262.77 m, (northern zone). Reflected light photomicrograph of an etched thin section. Sphalerite 2 commonly has broad growth twins (GT) with little evidence of deformation. Grain boundaries between coalesced grains of sphalerite 2 are commonly rounded to irregular (arrow).

Fig. 5.5e. Drill hole PD0011, 452.2 m, (central zone). Reflected light photomicrograph of an etched thin section. Sphalerite 2 (S2) overgrown by quartz 1 (Q). Sphalerite 2 near sphalerite 3 commonly has curved stress twins (ST). Broad growth twins are rare or absent.

Fig. 5.5f. Drill hole PD0029, 258.9 m, (central zone). Reflected light photomicrograph. Sphalerite 2 (S2) has been overgrown by chalcopyrite 3 (C3) which is overgrown by galena 1 (G). Chalcopyrite 3 results from expulsion of chalcopyrite 2 inclusions during dynamic recrystallisation of sphalerite 2 to form sphalerite 3. Chalcopyrite 3 and galena 1 have infilled fractures in sphalerite 2. Galena 1 is not obviously deformed and has straight cleavage planes.
The abundance of chalcopyrite 2 inclusions is gradational across individual growth zones (Fig. 5.5b). The inner margin of a growth zone typically has < 1% of small (< 5 μm), scattered, rounded blebs of chalcopyrite 2. Towards the outer margin of a growth zone, the proportion of chalcopyrite 2 inclusions is increased to = 20% resulting in a near opaque band typically < 20 μm wide. The outer most margin of a growth zone is defined by a sudden drop in the abundance of chalcopyrite 2 inclusions to < 1% which marks the start of the next growth zone.

Sphalerite 2 typically occurs without a core of earlier minerals but has locally overgrown sphalerite 1, sericite 1 or adularia (Figs. 5.4c, 5.4a). Some sphalerite 2 contains trace amounts of pyrite 1 inclusions (typically 5 - 70 μm across) and rare hematite flakes (typically < 20 μm across; Fig. 5.5c). In metasomatised marbles sphalerite 2 locally contains small inclusions of epidote, garnet, microcline, and remnants of carbonate - quartz - chlorite altered diopside. Sphalerite 2 constitutes = 94% of the sphalerite present at Menninnie Dam.

**Etching**

Sphalerite 2 was etched to assess the nature of the contact between coalescing sphalerite crystals and possible strain effects within sphalerite crystals. The etch consisted of a H₂SO₄ - KMnO₄ solution prepared as follows:

- Dissolve 1.0 grams of KMnO₄ in 40 ml of distilled water.
- Add 40 ml of 25% H₂SO₄.

Polished thin sections were immersed in the etch solution for between 35 - 45 seconds, then flushed with distilled water.

Coalescing sphalerite 2 crystals have sub-rounded to irregular margins, and polygonal crystal contacts and triple junctions are absent (Fig. 5.5d). Ramdohr (1969), Clark and Kelly (1973) and Cox (1987) described growth twins in etched, undeformed sphalerite and also fabrics that resulted from deformation of sphalerite. Growth twins formed broad, parallel sided, or stepped and bluntly terminated bands (Cox, 1987). Deformational twinning was formed by dislocation glide and can be distinguished from growth twins on the basis of their fine, tapering and discontinuous habit (Cox, 1987). Sphalerite 2 has well developed broad growth twins (Fig. 5.5d).

**Sphalerite 3**

Sphalerite 3 occurs in zones of sulphide mineralisation that contain brecciated crystals of pyrite 1. Sphalerite 3 consists of coalesced, polygonal crystals (typically 10 - 30 μm across) of homogenous, light brown sphalerite, devoid of micron size chalcopyrite inclusions. Sphalerite 3 is intimately intergrown with the margin of sphalerite 2 and locally crosscuts sphalerite 2 growth zones. Etched sphalerite 2 adjacent to sphalerite 3 has well developed
deformational twins (Fig. 5.5e). The texture of sphalerite 3 is similar to a mortar texture that results from dynamic recrystallisation (White, 1976; Brown et al., 1980; Cox, 1987).

**Chalcopyrite 3**
Chalcopyrite 3 has only been observed in the presence of dynamically recrystallised sphalerite 3. Chalcopyrite 3 commonly mantles sphalerite 3 and has in-filled fractures developed in sphalerite 2 (Fig. 5.5f). Where chalcopyrite 3 has infilled fractures in sphalerite 2, the margins of sphalerite 2 fragments are nearly opaque due to abundant micron size inclusions of chalcopyrite 3.

**Quartz 1**
The outer margins of sphalerite 2 locally incorporate small inclusions of euhedral quartz 1, but more commonly quartz 1 rims sphalerite 2 and forms euhedral growth zoned crystals, and crystal aggregates (Fig. 5.6a, b). Quartz 1 contains abundant small fluid inclusions (typically < 1 μm and rarely up to several μm diameter) and a dusting of hematite 1 which collectively, define growth zones. Quartz 1 locally infills fractures developed in sphalerite 2 and mantles adjacent dynamically recrystallised sphalerite 3 and chalcopyrite 3. Some quartz 1 adjacent to sphalerite 3 and chalcopyrite 3 has undulose extinction and is also deformed.

**Ca-Mn-Mg-Fe carbonate 1**
Ca-Mn-Mg-Fe carbonate 1 forms overgrowths on quartz 1 (Figs. 5.6a, b), sphalerite 2, and locally, sphalerite 3. Ca-Mn-Mg-Fe carbonate 1 forms both poorly crystalline, pale brownish aggregates, and euhedral rhombic crystals, and crystal aggregates, locally with a radiating fan texture, attributed to open space growth (Ineson, 1989).

**Galena 1**
Galena 1 forms overgrowths on Ca-Mn-Mg-Fe carbonate 1 and paragenetically earlier minerals. Galena 1 forms irregularly shaped crystals whose margins are dictated by the shape of earlier minerals (Figs. 5.3h, 5.4e, 5.5f). Deposition of galena 1 has not resulted in corrosion of dolomite or sphalerite 2 and 3, however, it has corroded euhedral pyrite 1 along growth zones or cleavage planes (Fig. 5.7). Most galena 1 has straight cleavage planes, however, curved cleavage planes are locally present.

**Sericite 2, chlorite 2 and rhodonite**
Some mineralised zones contain trace amounts of a second phase of sericite (sericite 2) which occurs in association with trace amounts of chlorite and rhodonite rimming galena 1. Some sericite flakes are up to 10 μm long and are more correctly termed muscovite, however to avoid confusion it is referred to as sericite 2.

*The composition of Ca-Mn-Mg-Fe carbonates were determined using the electron microprobe (ref. Section 5.5.2).*
Figs. 5.6a, b. Drill hole PD0011, 452.2 m, (central zone). Photomicrograph; (5.6a) transmitted light, (5.6b) transmitted light, crossed nicols. Sphalerite 2 (S2) has been overgrown by growth zoned quartz 1 (Q) with later, interstitial Ca-Mn-Mg-Fe carbonate 1 (C). Ca-Mn-Mg-Fe carbonate 1 has been partially altered to chlorite 2 (CH).

Fig. 5.7. Drill hole PD0024, 401.3 m, (central zone). Photomicrograph, reflected light. Pyrite 1 (P) overgrown by sphalerite 2 (S2) and later galena 1 (G). Galena 1 has corroded and replaced pyrite 1 crystals along cleavage planes (arrow).

Figs. 5.8a, b. Drill hole PD0017, 234.15 m, (central zone). Photomicrograph; (5.8a) reflected light, (5.8b) transmitted light, crossed nicols. The outer margin of galena 1 (G) locally contains small laths of sericite 2 (SE) giving the appearance of 'spines'. Sericite 2 has been overgrown by pyrite 2 (P2) which is overgrown by Ca-Mn-Mg-Fe carbonate 2 (C2).

Fig. 5.8c. Drill hole PD0011, 452.2 m, (central zone). Photomicrograph, combination of reflected and transmitted light. Pyrite 1 (P) has been overgrown by Ca-Mn-Mg-Fe carbonate 1 (C) which has later been altered and locally pseudomorphed by chlorite 2 (CH) and overgrown by fluorite (F).

Fig. 5.8d. Drill hole PD0017, 367 m, (central zone). Photomicrograph, transmitted light. Sphalerite 2 (S2) and quartz 1 (Q) with interstitial rhodonite (R).
Sericite 2 forms finely crystalline aggregates surrounding earlier formed minerals and has locally formed radiating elongated laths around the margin of galena 1, giving the appearance of 'spines' (Figs. 5.8a, b). Trace amounts of finely crystalline chlorite 2 are intergrown with sericite 2 and also occur as pseudomorphs after euhedral Ca-Mn-Mg-Fe carbonate 1 (Fig. 5.8c). Trace amounts of rhodonite occur in some sulphide veins in drill hole PD0017 as scattered, poorly crystalline, irregularly shaped straw coloured aggregates of up to 50 μm across (Fig. 5.8d). Due to the trace amounts of chlorite 2, sericite 2 and rhodonite, a paragenetic sequence for the three minerals cannot be established, therefore, they are tentatively considered coeval and paragenetically later than galena 1.

Pyrite 2
Pyrite 2 forms overgrowths on sericite 2, galena 1, and locally sphalerite 2. Pyrite 2 forms small (typically 20 - 50 μm across), subhedral to euhedral crystals and larger crystal aggregates around the margins of earlier minerals (Fig. 5.9). Undeformed pyrite 2 has locally overgrown brecciated pyrite 1.

Fluorite and Ca-Mn-Mg-Fe carbonate 2
Clear, to purplish fluorite occurs in minor to trace amounts interstitial to galena 1 and pyrite 2 crystals (Fig. 5.10). The shape of fluorite crystals was dictated by earlier formed minerals. Fluorite has been overgrown by Ca-Mn-Mg-Fe carbonate 2 or in the absence of fluorite, Ca-Mn-Mg-Fe carbonate 2 has overgrown earlier minerals. Ca-Mn-Mg-Fe carbonate 2 forms clear, to pale brown, poorly crystalline aggregates as well as isolated euhedral crystals, and crystal aggregates.

Quartz 2
Rare veinlets of quartz 2 cut across all earlier formed minerals. Veinlets are typically a few millimetres long and < 0.5 mm wide. Quartz 2 veinlets appear to be restricted to zones of sulphide mineralisation and have not been observed in the adjacent host rocks.

Late calcite veins
White calcite veins referred to as 'late calcite veins' have crosscut all earlier minerals and are equally abundant in mineralisation, the host rocks, and DP breccia units. Late calcite veins are typically a few millimetres wide and locally exceed 1 cm. Their length is not known as they extend beyond the confines of drill core. Their relationship to sulphide mineralisation (if any) is uncertain.
Fig. 5.9. Drill hole PD0017, 351.7 m, (central zone). Reflected light photomicrograph. Pyrite 1 (P1) has been overgrown by sphalerite 2 (S2). Both pyrite 1 and sphalerite 2 have been brecciated and overgrown by non-brecciated pyrite 2 (P2).

Fig. 5.10. Drill hole PD0016, 305.6 m, (northern zone). Transmitted light photomicrograph. Pyrite 1 (P) with interstitial sphalerite 2 (S2) have been overgrown by clear to purple fluorite (F). A 'grain' of microcline (M) adjacent to a crystal of pyrite 1 has been overgrown by adularia (A) which is also overgrown by fluorite.
**Sulphides hosted by quartz - diopside and massive diopside rock types**

Only the diopside component of quartz - diopside rock and massive diopside has been replaced by sulphide mineralisation. Metamorphic quartz is commonly brecciated in stockwork zones, but remains unaltered and is not host to sulphide mineralisation. Sulphides and gangue minerals hosted by quartz - diopside and massive diopside rock types have a similar paragenesis to sulphides and gangue minerals hosted by marbles, however, there are pronounced textural differences and minor mineralogical differences. Description of the sulphide and gangue mineralogy and textures will follow a paragenetic sequence starting with the first formed minerals.

**Dolomite, talc 1, phlogopite and chlorite 1**

Alteration and mineralisation developed in diopside units has preferentially occurred along fractures and in breccia zones. Mineralising fluids have preferentially permeated along fractures resulting in a series of lobate alteration ‘fronts’ (Fig. 5.1a). ‘Fronts’ comprise discontinuous laminations of finely crystalline sulphide and gangue minerals (Fig. 5.1b). On a microscopic scale, alteration has occurred along diopside cleavage planes resulting in a finely crystalline, felted mix of dolomite, talc 1, phlogopite and chlorite 1 (Fig. 5.1c). The matrix to brecciated diopside and the margins of diopside breccia clasts have been similarly altered. Alteration has resulted in chemical rounding and a reduction in the size of diopside clasts.

**Pyrite 1, sphalerite 1, sphalerite 2, chalcopyrite 1 and chalcopyrite 2**

Pyrite 1 forms lobate bands and isolated euhedral to subhedral crystals of 2 - 100 μm across, and locally, small atoll structures and rare framboids (Fig. 5.1c). Adularia and sericite 1 are uncommon alteration minerals in diopside-bearing host rocks and pyrite 1 is commonly overgrown by sphalerite 1 or in the absence of sphalerite 1, pyrite 1 is overgrown by sphalerite 2. Sphalerite 1 and 2 have formed isolated anhedral to subhedral crystals of 40 - 240 μm diameter and appear to have nucleated on small crystals of pyrite 1 (Fig. 5.1c). In addition to pyrite 1 inclusions, sphalerite 1 and 2 commonly contain micron sized inclusions of chalcopyrite 1 and 2 respectively. The fine banding of sulphide and gangue minerals produced through alteration of diopside are typically spaced 600 microns apart (Fig. 5.1b). Some pyrite 1 crystals in bands further away from the diopsidic host rock are coarser (5 - 70 μm diameter) compared to those in bands closer to the diopsidic host rock (5 - 10 μm diameter). Sphalerite 1 and 2 crystals are similarly larger further away from the unaltered diopsidic host rock.

**Quartz 1**

Quartz 1 is significantly more abundant in mineralisation hosted by diopside-bearing lithologies compared to mineralisation hosted by marble.
Fig. 5.11a. Drill hole PD0008, 289.83 m, (central zone). Hand specimen from near the contact between mineralisation and diopsidic host rock. The diopside has been altered along fractures (F) resulting in lobate 'fronts' of sulphide and gangue minerals. The direction of front migration is towards the top of the diagram.

Fig. 5.11b. Drill hole PD0017, 377 m, (central zone). Transmitted light photomicrograph of alteration 'fronts' similar to those in Figure 5.11a with bands of fine-grained pyrite 1 (opaque) and dolomite, chlorite ± phlogopite ± talc (mottled light brown). Less altered diopside is towards the top of the diagram. Larger crystals of pyrite 1 are more common near the rear of each front.

Fig. 5.11c. Drill hole PD0017, 377 m, (central zone). Transmitted light photomicrograph. Diopsidic host rock (D) has been preferentially altered along grain boundaries and cleavage planes to fine grained dolomite ± chlorite ± phlogopite ± talc. Fronts that contain fine-grained pyrite 1 (small black flecks) are further from the diopsidic host rock.

Fig. 5.11d. Drill hole PD0017, 377 m, (central zone). Reflected light photomicrograph. Small pyrite 1 framboinds (P) in cryptocrystalline quartz 1 (Q) that have resulted from alteration of diopsidic host rock.

Fig. 5.11e. Drill hole PD0017, 377 m, (central zone). Transmitted light photomicrograph from near the contact between mineralisation and diopsidic host rock. Small sphalerite 1 crystals (S1) have been enclosed in fine grained dolomite ± chlorite ± phlogopite ± talc gangue (G). Sphalerite 1 has nucleated on pyrite 1 (P1) crystals which form dark cores to sphalerite 1 crystals.

Fig. 5.11f. Drill hole PD0019, 317.7 m, (central zone). Photomicrograph, transmitted light, crossed nicols. Cryptocrystalline quartz 1 (almost all of Figure 5.11f) is common in mineralised diopsidic units. Minor amounts of sulphides dispersed throughout cryptocrystalline quartz 1 has resulted in a fine-grained dark grey silicic rock.
Quartz 1 commonly forms an interlocking mosaic of fine euhedral and rare anhedral crystals that appear to have nucleated on pyrite 1 sphalerite 1 and sphalerite 2 crystals and have replaced dolomite, talc 1 and chloride 1 (Fig. 5.11f). Pyrite 1, sphalerite 1 and sphalerite 2 crystals that have been overgrown by quartz 1 are generally smaller compared to those that have not been overgrown by quartz 1. Quartz 1 contains diffuse hematite-dusted patches, however, the presence of hematite does not appear to have had any affect on pyrite 1. Quartz 1 was commonly the last mineral to have formed in diopside host rocks, resulting in a medium to dark grey, finely crystalline, siliceous rock with disseminations of finely crystalline pyrite 1 and trace amounts of sphalerite 1 and 2. The abundance of quartz 1 appears to have inhibited the formation of paragenetically later minerals such as Ca-Mn-Mg-Fe calcite 1 and 2, galena 1, sericite 2, chloride 2, pyrite 2 and fluorite and these minerals locally occur as infill minerals in millimetre sized veins that crosscut the finely crystalline mosaic of interlocking quartz 1 crystals.

Sulphides hosted by marble - calc-silicate lithologies
Alteration of mixed marble and calc-silicate lithologies has resulted in a combination of textures described for marble hosted mineralisation and diopside - quartz rock or massive diopside hosted mineralisation.

Mineralisation hosted by DP and LP breccias
Both DP and LP breccias contain discrete sulphide clasts and composite sulphide - host rock clasts whose sulphide mineralogy consists of pyrite 1, sphalerite 1 and 2, chalcopyrite 1 and 2, and galena 1 (Fig. 4.9e and 5.12a). The matrix of most DP breccias and some marble and diopside clasts in DP breccias have been altered to various proportions of finely crystalline galena 1, pyrite 2, Ca-Mn-Mg-Fe carbonate 2, chloride 2, sericite 2, and rarely, trace amounts of sphalerite 2. The matrix of some LP breccias is altered to an assemblage which contains disseminated pyrite of unknown paragenesis. The sulphide-bearing clasts in both LP and DP breccias can be subdivided into three types on the basis of their sulphide and gangue mineralogy and textures. These are referred to as clast types 1, 2 and 3, and described below.

Clast type 1
Clast type 1 consists of various combinations and proportions of pyrite 1, adularia, sericite 1, sphalerite 1 and 2, chalcopyrite 1 and 2, quartz 1, Ca-Mn-Mg-Fe carbonate 1 galena 1 ± marble (Fig. 5.12a). Clasts typically range from 2 mm to 5 cm across and have a range of shapes from angular to rounded. The paragenesis and textures of the sulphides and gangue minerals are identical to those of mineralisation hosted by marble except that sulphide and gangue minerals paragenetically later than galena 1 are absent.
Fig. 5.12a. Drill hole PD0031, 178.8 m, (central zone). Hand specimen of LP breccia. LP breccia consists of lithic clasts (LC), rhyolite pyroclasts (R) and rare sulphide clasts (S) in a matrix of lithic fragments and fine pyroclastics. Sulphide clasts have similar mineralogy and textures to sulphide mineralisation hosted by the underlying Hutchison Group.

Fig. 5.12b. Drill hole PD0008, 179.92 m, (central zone). Hand specimen of a deformed and brecciated granite dyke with undeformed sulphide (S) and gangue mineral (GA) overgrowths on breccia clasts (G). Sulphide and gangue minerals have similar mineralogy and textures to those hosted by marble and calc-silicate lithologies.

Fig. 5.12c. Drill hole PD0031, 241.5 m, (central zone). Transmitted light photomicrograph. A vein comprised of undeformed, growth zoned sphalerite 2 (S2) with abundant fine-grained inclusions of chalcopyrite 2 (opaque), has cross-cut a mylonite (M) formed in eastern suite schist.

Fig. 5.13. Drill hole PD0032, 331.5 m, (southern zone). Hand specimen of brecciated granite dyke (G) that has been pervasively altered to fine-grained, dark green chlorite, carbonate, quartz and light yellow green sericite (LAO). This style of pervasive, texturally destructive alteration is typical of the 'late alteration overprint'.

Fig. 5.14. Drill hole PD0032, 437.5 m, (southern zone). Hand specimen of a sulphide vein hosted by dolomitic marble (M). The sulphide vein comprises sphalerite 2 and 4 (S), chalcopyrite 4 (CP4), pyrite 1 (P1) and trace amounts of galena 1 or 2 (G) and hematite 2. The dark green alteration halo (AH) comprises, chlorite 3, Ca-Mn-Mg-Fe carbonate 3, and minor calc-silicate minerals.
Chapter 5  Sulphide mineralisation

Clast type 2
Clast type 2 consists of finely crystalline, disseminations of pyrite 1 in a finely crystalline matrix of quartz 1. Trace amounts of sphalerite 1 and 2, chalcopyrite 1 and 2, Ca-Mn-Mg-Fe carbonate 1 and galena 1 are present in some clasts. Clasts typically range from 1 mm to 5 cm across and have a range of shapes from angular to rounded. Sulphide and gangue minerals paragenetically later than galena 1 have not been observed. Partially altered remnants of diopside are present in some clasts. The sulphide and gangue paragenesis and textures are similar to mineralisation hosted by massive diopside and diopside bands.

Clast type 3
Clast type 3 consists of angular fragments of pyrite 1 aggregates and typically range from 1 mm to 1 cm across. The size range of clasts type 3 is similar to pyrite 1 aggregates hosted by marble units.

Mineralisation hosted by granite dykes, graphitic schist, eastern and western suite schists.
Minor amounts of sulphide mineralisation are locally hosted by granite dykes, graphitic schists (Fig. 3.14g), and eastern and western suite schists. Mineralisation hosted by granite dykes invariably occurs in the matrix of brecciated granite dykes (Fig. 5.12b). Mineralisation hosted by graphitic schist occurs as thin veinlets typically 0.5 - 2 mm wide which crosscut the schistose fabric (Fig. 3.5e). Mineralisation hosted by eastern and western suite schists is commonly associated with fault zones but crosscuts the fault zone fabric and locally crosscuts mylonitic fabrics (Figs. 3.15e, 5.12c). Sulphide and some gangue minerals have primary mineral zoning and have not been deformed (Fig. 5.12c). The mineralogy, textures and paragenesis of sulphide and gangue minerals hosted by brecciated granite dykes, graphitic schist, and eastern and western suite schists are similar to those of marble hosted mineralisation.

5.3 Mineralisation in the southern zone
Interpretation of mineralisation and alteration in the southern zone is based on a single diamond drill hole (drill hole PD0032; Fig. 3.1) and a few percussion drill holes to the south of drill hole PD0032. Summary geological logs for drill hole PD0032 and the southern percussion drill holes are illustrated in Appendix 3.1.

Distribution of sulphides
Most sulphide mineralisation in the southern zone is hosted by marbles and calc-silicate lithologies of the eastern suite, interpreted to be part of a transition zone between the marbles and calc-silicate lithologies of the central, and schist of the eastern suite (Section 3.2.3). Minor amounts of disseminated sulphides are hosted by brecciated granite dykes and altered eastern suite schist. LP breccia and the lower 34m of the MD ignimbrite (Section 4.3.2)
contain trace amounts of disseminated sulphides. Marble clasts scattered throughout the remainder of the MD ignimbrite are locally altered, and some contain trace amounts of sulphide mineralisation. Rare sulphide-bearing veins crosscut the eutaxitic texture of the MD ignimbrite. Due to the absence of outcrop and the limited number of drill holes, it is not possible to further define the shape or lateral extent of mineralisation in the southern zone.

**Late alteration overprint**

A significant difference between rock units and mineralisation in the southern zone compared to those in the northern and central zones, is the occurrence of pervasive chlorite - carbonate - quartz - calc-silicate - hematite alteration that has affected both the host rocks and mineralisation. Pervasive chlorite - carbonate - quartz - calc-silicate - hematite alteration is referred to as the 'late alteration overprint' throughout the remainder of this thesis. The late alteration overprint has resulted in significant mineralogical and textural modifications of all rock units with both selective and pervasive alteration and destruction of metamorphic textures. Marble and calc-silicate lithologies have been altered to chlorite, calcite, epidote, tremolite ± garnet ± talc ± serpentine ± graphite ± rhodochrosite ± hematite. Schist units have been partially altered to quartz, chlorite, calcite, epidote, tremolite and sericite with partial destruction of metamorphic textures. Feldspar in granite dykes has been selectively altered to sericite, chlorite, calcite, quartz and minor epidote. Locally, granite dykes have been brecciated and pervasively altered to chlorite, calcite, sericite, quartz and minor epidote with total destruction of primary textures (Fig. 5.13).

**5.3.1 Relationship between sulphides and host rocks**

The style of mineralisation in the percussion holes cannot easily be determined due to the nature of the drilling process and interpretation of the relationship between mineralisation and the host rocks is based solely on diamond drill hole PD0032.

Mineralisation in the southern zone occurs as veins and less commonly as disseminations throughout the host rock (Fig. 5.14). Veins vary from 1 - 10 cm wide and are generally less common, and thinner, than veins in the northern and central zones. Veins are commonly associated with a zone of host rock alteration several centimetres to several metres wide zone which is significantly more extensive than the alteration halo surrounding veins in the northern and central zones (Fig. 5.14). Unlike mineralisation in the northern and central zones, the relationship between mineralisation and fractured or brecciated host rock is not obvious. This may have resulted from destruction of host rock textures by the late alteration overprint.
5.3.2 Sulphide and gangue mineralogy and textures
The sulphide and gangue mineralogy and textures are described in paragenetic order starting with the first formed mineral. Minerals in the southern zone that have a similar label to those in the northern and central zones are paragenetically equivalent.

e.g. Sphalerite $\gamma_{\text{south}} = \text{Sphalerite } \gamma_{\text{north and central}}$

A summary of the sulphide and gangue mineral paragenesis applicable to the southern zone is illustrated in Figure 5.15.

**Pyrite 1 and sphalerite 2**
Pyrite 1 was the earliest formed mineral, present as euhedral to subhedral crystals and aggregates of 5 µm - 1 mm diameter. Sphalerite 1 has not been observed in the southern zone and 5 - 100 µm crystals of pyrite 1 are commonly overgrown by sphalerite 2. Sphalerite 2 forms corroded and embayed crystals and aggregates of 0.1 - 3 mm diameter and rare, isolated, rounded to subbedral crystals (Fig. 5.16a). Euhedral crystal margins are locally preserved where sphalerite 2 is in contact with pyrite 1, quartz 1, Ca-Mn-Mg-Fe carbonate 1, or galena 1 (Figs. 5.16c, d, 5.17).

Sphalerite 2 commonly comprises two, and locally, three sectors, referred to as the 'inner sector', the 'mid sector' and the 'outer sector' (Fig. 5.16a). The inner sector of sphalerite 2 is light to medium yellow - brown with well developed growth zoning. Trace amounts of hematite 1 locally occur as < 20 µm diameter inclusions scattered throughout the inner sector of sphalerite 2. The inner sector is surrounded by the mid sector which consists of light yellow - brown sphalerite 2 typically devoid of growth zoning, or locally with a remnant, partially corroded, outermost growth zone (Figs. 5.16a, c). The contact between the darker coloured inner sector and the lighter coloured mid sector is irregular and diffuse and cuts across growth zoning developed in the inner sector. The width of the mid sector is variable and ranges from 50 microns to almost 1 mm. The mid sector (or in its absence, the inner sector) is mantled by an outer sector of pale brown sphalerite 2 devoid of growth zoning. The contact between the mid sector and the outer sector is either abrupt, occurring at the single outer growth zone of the mid sector (Figs. 5.16a, c), or transgresses the outer growth zone and is diffuse.

Growth zoning in the inner sector of sphalerite 2 is defined by micron sized inclusions of chalcopyrite 2. As with sphalerite 2 in the northern and central zones, individual growth zones often show a gradation in the intensity of micron sized inclusions of chalcopyrite 2. The inner part of a growth zone has < 1% scattered small (≤ 5 µm) rounded blebs of chalcopyrite 2. As the outer margin of the growth zone is approached, the proportion of chalcopyrite 2 inclusions is increased to ≥ 20%.
Fig. 5.15 Paragenesis of sulphide and gangue minerals in the southern zone. Thicker lines represent more abundant minerals, and dotted lines represent trace amounts of minerals of uncertain paragenesis. The grey bars have no significance other than to assist the reader.
Fig. 5.16a, b. Drill hole PD0032, 245.7 m, (southern zone). Photomicrograph, (5.16a) transmitted light (5.16b) reflected light. Light yellow - brown sphalerite 2 with inner (I), mid (M) and outer (O) sectors. The outer margin of the outer sector has a dark rind of abundant fine grained chalcopyrite 4 (CP4). The dark rind is also present at the boundary between sphalerite 2 grains in mutual contact and along fractures in sphalerite 2. Sphalerite 2 in contact with galena 1 (G) and Ca-Mn-Mg-Fe carbonate 2 (C2) has euhedral crystal margins whereas sphalerite 2 in contact with Ca-Mn-Mg-Fe carbonate 3 and chlorite 3 (C3) has corroded crystal margins.

Figs. 5.16c, d. Drill hole PD0032, 245.7 m, (southern zone). Photomicrograph, transmitted light, (5.16c) reflected light (5.16d). Mid (M) and outer (O) sectors of pale yellow - brown sphalerite 2 with dark chalcopyrite rich outer margins. Sphalerite 2 has euhedral margins in contact with Ca-Mn-Mg-Fe carbonate 2 (C2) and corroded margins in contact with tremolite (T), Ca-Mn-Mg-Fe carbonate 3 (C3) and trace amounts of graphite (opaque flecks). Corroded sphalerite 2 (S2) crystals have been mantled by a thin rim of galena 2 (G2).

Fig. 5.16e. Drill hole PD0032, 245.7m, (southern zone). Reflected light photomicrograph of an etched thin section. Sphalerite 2 grains in mutual contact have round to irregular shapes (arrow) and broad growth twins (GT).
The outer-most margin of a growth zones is defined by a sudden decrease in the proportion of chalcopyrite 2 inclusions to < 1% which also marks the inner-most boundary of the next growth zone.

The mid sector of sphalerite 2 is either devoid of chalcopyrite 2 inclusions or contains scattered inclusions of chalcopyrite 2 and a concentration of chalcopyrite 2 inclusions along a single outer-most growth zone. Chalcopyrite 2 inclusions in the mid sector are commonly smaller than those in the inner sector. The outer margin of the outer sector is corroded and is commonly packed with fine chalcopyrite 2 inclusions of < 1 μm diameter that are smaller than those in either the mid sector or the inner sector. The abundance of fine chalcopyrite 2 inclusions along the outer margin of the outer sector forms a dark, opaque rind of = 40 μm wide on most sphalerite 2 crystals. (Figs. 5.16a, c). The boundaries of sphalerite 2 crystals in mutual contact and sphalerite 2 adjacent to fractures have a similar dark rind of sub-micron chalcopyrite 2 inclusions (Fig. 5.16a).

Broad growth twins and irregular crystal boundaries are apparent in etched sphalerite 2 (Fig. 5.16e). Deformation twins have not been observed in sphalerite 2 in the southern zone.

Quartz 1
The outer margin of sphalerite 2 locally contains small euhedral crystals of quartz 1. More commonly, quartz 1 forms euhedral growth zoned crystals and crystal aggregates that overgrow sphalerite 2. Growth zoning is defined by abundant fluid inclusions and mineral inclusions typically < 1 μm across, and rarely up to several microns.

Ca-Mn-Mg-Fe carbonate 1
Ca-Mn-Mg-Fe carbonate 1 has overgrown quartz 1, sphalerite 2, or pyrite 1 (Fig. 5.16c, d). Trace amounts of small hematite 1 flakes (< 5 μm diameter) are locally scattered throughout Ca-Mn-Mg-Fe carbonate 1.

Galena 1 and Ca-Mn-Mg-Fe carbonate 2
Galena 1 has overgrown Ca-Mn-Mg-Fe carbonate 1, quartz 1 and sphalerite 2 (Fig. 5.17) and forms coarse crystal aggregates with irregular corroded margins. In the northern and central zones, galena 1 locally corroded pyrite 1, however, the nature of the contact between pyrite 1 and galena 1 has not been established in the southern zone. Galena 1 has locally been overgrown by Ca-Mn-Mg-Fe carbonate 2.

Up to this point, descriptions of the mineralogy and mineral paragenesis of the southern zone are similar to those of the northern and central zones, however, minerals paragenetically later than Ca-Mn-Mg-Fe carbonate 2 in the southern zone are significantly different from those of the northern and central zones.
Fig. 5.17. Drill hole PD0032, 245.7 m, (southern zone). Reflected light photomicrograph. Sphalerite 2 (S2) contains small inclusions of chalcopyrite 2 (yellow). Sphalerite 2 adjacent to galena 1 (G1) has euhedral crystal margins. Galena 1 has a straight cleavage pattern. Galena 2 (G2) has resulted from remobilisation of galena 1 and forms rims on corroded crystals of sphalerite 2.

Fig. 5.18. Drill hole PD0032, 245.7 m, (southern zone). Transmitted light photomicrograph. Sphalerite 2 (S2) has been overgrown by Ca-Mn-Mg-Fe carbonate 1 (C1). Ca-Mn-Mg-Fe carbonate 1 has been partially replaced by a Ca-Mn-Mg-Fe carbonate 3 and chlorite 3 (C3). The contact between Ca-Mn-Mg-Fe carbonate 1, and Ca-Mn-Mg-Fe carbonate 3 and chlorite 3 is an irregular front (arrow).

Figs. 5.19a, b. Drill hole PD0032, 383.25 m, (southern zone). Photomicrograph, transmitted light (5.19a), reflected light (5.19b). Corroded sphalerite 2 crystals (S2) have been cross-cut by a thin vein of sphalerite 4 (S4), chalcopyrite 4 (CP4), galena 2 (G2), and Ca-Mn-Mg-Fe carbonate 3 (C3). Sphalerite 4 is devoid of chalcopyrite inclusions.

Figs. 5.20a, b. Drill hole PD0032, 392.95 m, (southern zone). Photomicrograph, transmitted light (5.20a), reflected light (5.20b). Sphalerite 2 (S2) contains abundant small inclusions of chalcopyrite 2 (CP2; yellow flecks in Fig. 5.20b). Sphalerite 4 (S4) adjacent to sphalerite 2 is devoid of chalcopyrite 2 inclusions and contains small inclusions of remnant galena 1 (G1; white flecks in Fig. 5.20b). A rind of hematite 2 (H2) has partially encompassed sphalerite 2 and 3.
Pervasive chlorite-carbonate-quartz-calc-silicate-hematite alteration (late alteration overprint) of the host rocks and mineralisation has occurred irrespective of the nature of the host rocks. The paragenetic sequence of late alteration overprint minerals is not always clear and the following descriptions follow a paragenetic sequence as closely as could be determined (Fig. 5.15).

**Chlorite 3 and Ca-Mn-Mg-Fe carbonate 3**
In mineralised zones hosted by marble, pervasive, finely crystalline, chlorite 3 and Ca-Mn-Mg-Fe carbonate 3 alteration has corroded and replaced earlier Ca-Mn-Mg-Fe carbonate 1 and 2 (Fig. 5.16a, 5.18).

**Galena 2, sphalerite 4, chalcopyrite 4 and pyrite 3**
Galena 1, sphalerite 2, chalcopyrite 2 and pyrite 1 have undergone partial dissolution, and have been locally re-precipitated, or more commonly re-precipitated elsewhere as galena 2, sphalerite 4, chalcopyrite 4 and pyrite 3 respectively. Dissolution of sulphides is evidenced by rare veinlets of sphalerite 4, chalcopyrite 4, galena 2 and pyrite 3 that crosscut sphalerite 2 (Fig. 5.19a, b). Dissolution of sphalerite 2 resulted in corrosion and embayment of sphalerite 2 crystal margins and leaching of iron and micron sized inclusions of chalcopyrite 2. This resulted in the lighter colour, and the diminished content and smaller size of chalcopyrite 2 inclusions in the mid and outer sectors of sphalerite 2 (Fig. 5.16a, c).

Partial dissolution of galena 1 has resulted in highly corroded crystal shapes and the outer = 50 μm of galena 1 crystals is devoid of the characteristic, well defined, cleavage pattern, of galena 1. Galena 2 also forms 10 - 50 μm wide overgrowths on corroded sphalerite 2 crystals immediately adjacent to the dark rind of sub-micron chalcopyrite 2 inclusions (Fig. 5.17). Galena 1 adjacent to sphalerite 2 has locally been mostly replaced by pale yellow brown sphalerite 4. Small irregular to elongated, rounded 'globs' of remnant galena 1 (10 - 60 μm long) remain randomly scattered throughout sphalerite 4 (Fig. 5.20a, b).

**Tremolite, talc 2 and graphite**
Tremolite, talc 2, and locally, trace amounts of graphite, have overprinted chlorite 3 and partially replaced Ca-Mn-Mg-Fe carbonate 3 (Fig. 5.16c, d). Tremolite has formed randomly orientated, scattered laths (typically 80 μm, and locally up to 5 mm long) and radiating aggregates. Talc 2 is present as a finely crystalline mosaic interstitial to tremolite laths. Graphite has only been observed in a single polished thin section where it forms small randomly orientated disseminated flakes (< 5 μm length) dispersed between tremolite laths. Remnant Ca-Mn-Mg-Fe carbonate 3 has a mottled pinkish colour.
Quartz 3
Euhedral growth zoned quartz 3 has partially replaced Ca-Mn-Mg-Fe carbonate 3 and locally, overprinted tremolite (Fig. 5.21). Quartz 3 locally forms a mosaic of finely crystalline, irregularly shaped, interlocking crystals with poorly defined crystal margins. Chlorite 3 does not appear to have been replaced by quartz 3 and is locally present as inclusions in quartz 3.

Epidote and garnet
Epidote appears to have formed in association with tremolite and has locally replaced quartz 3, Ca-Mn-Mg-Fe carbonate 3 and chlorite 3 (Fig. 5.22). Epidote is present as isolated, randomly orientated, light yellow - green and colourless prismatic crystals, up to 2 mm in length as well as larger crystal aggregates. Locally, grossular garnet is present as scattered, pale coloured, euhedral crystals, typically 100 - 300 μm across and appears to be paragenetically later than epidote (Fig. 5.22).

Serpentine and rhodochrosite
Minor amounts of serpentine locally occur in association with epidote and garnet. In zones that contain epidote and garnet, remnant Ca-Mn-Mg-Fe carbonate 3 has been converted to rhodochrosite (Fig. 5.22).

Chlorite 4 and hematite 2
Chlorite 4 and hematite 2 are the latest minerals in the paragenetic sequence. Hematite 2 forms irregular shaped clusters up to 500 μm diameter, that infills fractures in sphalerite and epidote as well as overgrowths on garnet (Fig. 5.23). Hematite also occurs as small needle shaped laths within chlorite 4.

MD ignimbrite hosted mineralisation
The basal 34m (vert.) of the MD ignimbrite in drill hole PD0032 and the underlying LP breccia have been pervasively altered to a mineral assemblage similar to that described for the southern zone. Alteration of the MD ignimbrite has been described in Section 4.3.6 and is summarised below. This description applies equally to the underlying LP breccia. Minor sulphide mineralisation also occurs in thin veins at higher levels in the MD ignimbrite.

Basal 34m of the MD ignimbrite in drill hole PD0032
Marble clasts in the LP breccia and overlying MD ignimbrite have been altered to tremolite ± epidote ± Ca-Mn-Mg-Fe carbonate 3 ± trace amounts of pyrite 3, chalcopyrite 4, sphalerite 4 and galena 2. Lithic clasts have been partially altered to sericite, chlorite and Ca-Mn-Mg-Fe carbonate 3. The groundmass has been pervasively altered to finely crystalline, felted, sericite, chlorite, quartz, serpentine, Ca-Mn-Mg-Fe carbonate 3 and trace amounts of pyrite 3, chalcopyrite 4, sphalerite 4, galena 2, and hematite 2 (Figs. 4.20a - i).
Fig. 5.21. Drill hole PD0032, 383.25 m, (southern zone). Transmitted light photomicrograph. Crystals of sphalerite 2 (S2) have been corroded, and overgrown by Ca-Mn-Mg-Fe carbonate 3 (C3), chlorite 3 (CH3) and tremolite (T). Chlorite 3 and tremolite have been overgrown by quartz 3 (Q3). Quartz 3 has partially replaced tremolite but not chlorite 3. Ca-Mn-Mg-Fe carbonate 3 has a pale brownish colour with a pinkish hue.

Fig. 5.22. Drill hole PD0032, 392.6 m, (southern zone). Transmitted light photomicrograph. Corroded remnants of sphalerite 2 (S2), galena 1 (and 2; GN) have been overgrown by later epidote (E), garnet (G) and rhodochrosite (R).

Fig. 5.23. Drill hole PD0032, 392.6 m, (southern zone). Photomicrograph, transmitted light. Sphalerite 2 (S2) and garnet (G) have been overgrown by hematite 2 (H2).

Fig. 5.24a. Drill hole PD0032, 120.65 m (105 m vert.), southern zone. Hand specimen of an altered marble clast within the MD ignimbrite. The marble clast has been altered to concentric bands with variable amounts of disseminated fine-grained pyrite calcite and sericite.

Fig. 5.24b. Drill hole PD0032, 170m (148 m vert.), southern zone. Photomicrograph, low level transmitted light. A quartz (Q'), sphalerite (S), galena (G), chlorite, carbonate (C) vein has cross-cut welded MD ignimbrite (MD). Euhedral quartz crystals have later quartz overgrowths (Q'). The MD ignimbrite has a well developed eutaxitic texture.
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The paragenesis of sericite, chlorite, quartz and some carbonate in the MD ignimbrite is uncertain as they may equally have been associated with either the late alteration overprint or with devitrification of the volcaniclastics.

A 30 cm marble clast occurs completely enclosed within altered MD ignimbrite at 245.7m (30.8m from the base, in drill hole PD0032. The marble clast has been crosscut by a sulphide-bearing vein that does not extend beyond the clast margin. The sulphide minerals have identical mineralogy, textures and paragenesis to southern zone, marble-hosted mineralisation (Section 5.2.3). Both the enclosing ignimbrite, marble clast, and mineralisation have been partially altered to chlorite 3, Ca-Mn-Mg-Fe carbonate 3, and quartz 3 with minor amounts of epidote, tremolite, talc, pyrite 3, sphalerite 4, chalcopyrite 4 and galena 2 and trace amounts of graphite.

Higher levels in the MD ignimbrite
Some marble clasts > 34m from the base of the MD ignimbrite have been altered to a pattern of concentric, alternating bands of finely crystalline pyrite 3, carbonate and sericite (Fig. 5.24a). The alteration bands approach a sub-spherical or ellipsoidal shape irrespective of the angularity of the original marble clast.

A 5 mm wide quartz 3 - calcite - sphalerite 4 - galena 2 - pyrite 3 - chlorite 3 vein crosscuts welded MD ignimbrite at 96.5m (vert.) from the base of the MD ignimbrite in drill hole PD0032 (Fig. 5.24b).

Sulphides in percussion drill holes
Percussion drill holes to the south of drill hole PD0032 intersected either MD ignimbrite, granite dyke or rhyolite intrusion (Appendix 3.1). MD ignimbrite from these drill holes has been pervasively altered to chlorite 3, quartz 3, carbonate + epidote + tremolite with traces of disseminated pyrite 3, chalcopyrite 4, sphalerite 4 and galena 2.

5.4 Chemical analysis of mineralised zones
Many ore bodies have a vertical and/or lateral zonation in their mineralogy and the abundance of sulphide and gangue minerals (eg. Gilman, Colorado, Beaty and Merchant, 1990; VHMS deposits, Large, 1992; porphyry copper deposits, Beane and Titley, 1981). Variations in mineral abundances have a corresponding variation in their constituent elements, therefore, examination of the whole rock chemistry can be used to examine mineralogical abundances. The aim of this section is to examine prospect scale mineral variations using the existing geochemical database.

With the exception of drill hole PD0018 and the upper part of drill hole PD0021, all drill holes have been assayed throughout their length for a range of elements including Pb, Zn,
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Cu, Ag, Mn, Fe and less commonly, Au, Hg, As, Sn, W and Ba. Samples for analysis were collected by grinding a strip along the length of non-mineralised core and collecting the powder for each interval of between 2 - 6 m. Core which contained visible sulphides was quartered. Samples were submitted to commercial laboratories for analysis using either atomic absorption spectrometry (AAS) or X-ray fluorescence (XRF). Summary geochemical logs for each of the 32 diamond drill holes are illustrated in Appendix 3.1. Analyses of mineralised core intervals, below the zone of weathering are presented in table format in Appendix 5.2.

Only assays from samples below the zone of weathering that have either Pb or Zn greater than 1000 ppm (0.10 wt.%) are used throughout this section. Assay values for the northern and central zone, and the southern zone are summarised in Tables 5.2 and 5.3 respectively.

Table 5.2 Summary of chemical analyses of mineralisation from the northern and southern zones. Only samples that have Pb or Zn > 0.10 wt.% have been included.

<table>
<thead>
<tr>
<th>Element</th>
<th>Number of analyses</th>
<th>Range</th>
<th>Median</th>
<th>Average</th>
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<tbody>
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<td>Pb (wt.%)</td>
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<td>0 - 44.60</td>
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<td>1.36</td>
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<td>Zn (wt.%)</td>
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<td>Cu (ppm)</td>
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<td>Ag (ppm)</td>
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<td>Mn (wt.%)</td>
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<td>Fe (wt.%)</td>
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<tr>
<td>Au (ppm)</td>
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<tr>
<td>Hg (ppm)</td>
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<td>As (ppm)</td>
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<td>Ba (ppm)</td>
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<td>10 - 793</td>
<td>142</td>
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Table 5.3 Summary of chemical analyses of mineralisation from the southern zone. Only samples that have Pb or Zn > 0.10 wt.% have been included. Hg, As, Sn, W, and Ba were not analysed.

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<th>Average</th>
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<td>Zn (wt.%)</td>
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<td>Cu (ppm)</td>
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<tr>
<td>Ag (ppm)</td>
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<td>11</td>
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<tr>
<td>Mn (wt.%)</td>
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<tr>
<td>Fe (wt.%)</td>
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<td>0.021</td>
<td>0.022</td>
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</table>

5.4.1 Spatial variations in mineralisation chemistry

Variations in metal abundances over the length of the prospect are illustrated in Figures 5.2Sa-g. Diagrams have a bell shaped distribution with the highest analyses in the vicinity of 9000 mN and tailing off to both the north and south. The absence of data in the vicinity of 8500 mN does not necessarily represent an absence of mineralisation, but rather the paucity of drilling and the extreme depth of weathering.
Fig. 5.25 Chemical analyses of mineralised zones (with Pb or Zn > 1000 ppm) vs northing. The boundary between the southern zone, and the northern and central zones is shown as a dotted line. Higher analyses of each element are centered around 9000 mN.
There is no systematic variation between metal values or metal ratios with depth below the zone of weathering. Insufficient analyses for As, W and Sn are available for investigation of spatial variances.

Copper to zinc ratios
Copper to zinc ratios (100 Cu / Cu + Zn) for the southern zone and the combined northern and central zones are illustrated in Figure 5.26. Copper ratios are commonly higher in the southern zone compared to the combined northern and central zones.

5.5 Mineral chemistry
Sulphide mineralisation at Menninnie Dam averages = 0.84 wt.% manganese with a maximum of 16 wt.%. Silver is present in concentrations up to 900 ppm and typically = 16 ppm. Analyses of up to 3240 ppm tungsten have been recorded from mineralisation in the northern and central zones. Up to 3% of rhodochrosite (MnCO₃) is present in the southern zone and may account for analyses with elevated Mn. The only obvious Mn-bearing mineral is the northern and central zones is trace amounts of rhodonite [(Mn,Ca,Fe)SiO₃] which cannot account for the high Mn analyses of mineralised zones. Similarly, no silver or tungsten-bearing minerals have been observed throughout the Menninnie Dam prospect. The aim of this section is to identify the host minerals to Mn, Ag and W using the electron microprobe.

5.5.1 Method
Variation diagrams with Ag, W, or Mn plotted against other elements were used to establish element correlations. Elements with a significant correlation were then related to a likely host mineral (eg. Pb and galena). The likely host mineral was analysed for either Ag, Mn or W using the electron microprobe in the Central Science Laboratory, University of Tasmania under the supervision of W. Jablonski. Analytical results are detailed in Appendices 5.3 - 5.5.

5.5.2 Results
Silver
Pb and Ag analyses from mineralisation in the northern and central zones have a positive correlation with a correlation coefficient of 0.50 (Fig. 5.27a). Pb and Ag analyses from mineralisation in the southern zone do not have an obvious correlation (Fig. 5.27b). Bismuth and silver were identified by electron microprobe analysis of galena 1 from the northern and central zones and galena 1 and 2 from the southern zone (Table 5.4 and Appendix 5.3). Silver and bismuth from analyses of galena 1 in the northern and central zones have a positive correlation (correlation coefficient of 0.77; Fig. 5.27c).
Fig. 5.26 Frequency diagrams of copper : zinc ratios for the northern and central zones, and for the southern zone.
Fig. 5.27  A) Drill hole analyses for lead and silver in the northern and central zones. B) Drill hole analyses for lead and silver in the southern zone. C) Electron microprobe analytical results for silver and bismuth in galena 1 and 2 from the northern and central zones, and the southern zone.
Insufficient data are presented to enable assessment of a correlation between Bi and Ag in galena 1 and 2 from the southern zone. The trace element chemistry of galena 1 and galena 2 from all three zones is similar and they cannot be distinguished on the basis of the data presented.

Table 5.4. Trace element compositions of galena 1 and 2

<table>
<thead>
<tr>
<th>Zone</th>
<th>Mineral paragenesis</th>
<th>Element</th>
<th>Detection limit (ppm)</th>
<th>Number of analyses</th>
<th>Range (ppm)</th>
<th>Median (ppm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northern and central</td>
<td>galena 1</td>
<td>Ag</td>
<td>62</td>
<td>13</td>
<td>46 - 639</td>
<td>238</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Bi</td>
<td>283</td>
<td>13</td>
<td>0 - 3253</td>
<td>1857</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Sb</td>
<td>300</td>
<td>13</td>
<td>below detct.</td>
<td></td>
</tr>
<tr>
<td>Southern</td>
<td>galena 1</td>
<td>Ag</td>
<td>62</td>
<td>7</td>
<td>0 - 203</td>
<td>91</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Bi</td>
<td>283</td>
<td>7</td>
<td>651 - 2136</td>
<td>1810</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Sb</td>
<td>300</td>
<td>7</td>
<td>below detct.</td>
<td></td>
</tr>
<tr>
<td>Southern</td>
<td>galena 2</td>
<td>Ag</td>
<td>62</td>
<td>5</td>
<td>75 - 214</td>
<td>120</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Bi</td>
<td>283</td>
<td>5</td>
<td>1331 - 2316</td>
<td>1674</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Sb</td>
<td>300</td>
<td>5</td>
<td>below detct.</td>
<td></td>
</tr>
</tbody>
</table>

Manganese

Variation diagrams involving Mn and other elements produce a scatter with no obvious correlations. Pyrite rarely contains significant Mn, however, sphalerite can accommodate up to a few percent of Mn (Deer et al. 1978). Electron microprobe analyses of sphalerite 2 from the northern, central and southern zones shows significant quantities of iron, manganese and cadmium. There is no appreciable difference in Cd and Mn contents of sphalerite 2 from the northern and central zone, and sphalerite 2 from the southern zone (Table 5.5 and Appendix 5.4).

Table 5.5. Mn, Cd and Fe contents of sphalerite 2 from the northern, central and southern zones.

<table>
<thead>
<tr>
<th>Zone</th>
<th>Element</th>
<th>Detection limit (ppm)</th>
<th>Number of analyses</th>
<th>Range (ppm)</th>
<th>Median (ppm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northern and central</td>
<td>Mn</td>
<td>202</td>
<td>21</td>
<td>410 - 6449</td>
<td>2990</td>
</tr>
<tr>
<td></td>
<td>Cd</td>
<td>290</td>
<td>24</td>
<td>1160 - 2456</td>
<td>1760</td>
</tr>
<tr>
<td></td>
<td>Fe</td>
<td>0.020 wt.%</td>
<td>29</td>
<td>1.14 - 4.32 wt.%</td>
<td>4.08 wt.%</td>
</tr>
<tr>
<td>Southern</td>
<td>Mn</td>
<td>202</td>
<td>4</td>
<td>2019 - 3703</td>
<td>2811</td>
</tr>
<tr>
<td></td>
<td>Cd</td>
<td>290</td>
<td>4</td>
<td>1769 - 2237</td>
<td>1953</td>
</tr>
</tbody>
</table>

A small number dolomite, Ca-Mn-Mg-Fe carbonate 2 and late calcite veins were analysed using the electron microprobe. Results are illustrated in Figure 5.28 and detailed in Appendix 5.5. Dolomites and late calcite veins are close to end member dolomite and calcite compositions respectively. Ca-Mn-Mg-Fe carbonate 2 has a range of compositions that lie between kutnahorite, calcite, ankerite and dolomite and constitutes a significant reservoir of manganese. The compositional variations in Ca-Mn-Mg-Fe carbonate 1, 2 and 3 have not systematically been investigated, and although such a study has merit, it is not consistent with the aims of this thesis.
Fig. 5.28 Electron microprobe analysis of dolomite, Ca-Mg-Mn carbonate 2 and the late vein calcite.
Fig. 5.29. Tungsten vs Zinc variation diagram of drill core analyses from the northern and central zones. Tungsten and zinc have a positive correlation with a correlation coefficient of 0.96.
Tungsten

W and Zn analyses from mineralisation in the northern and central zones have a positive correlation with a correlation coefficient of 0.96 (Fig. 5.29). No tungsten bearing minerals have been detected despite examination of 167 thin sections and ultraviolet light irradiation of both thin sections and core. Sphalerite and all other minerals from samples with high tungsten contents were analysed using the electron microprobe to produce back scattered element maps. No tungsten was detected.

5.6 Discussion

In this section, the relative timing of sulphide mineralisation and the host rocks, metamorphic events, rhyolite intrusions and the MD ignimbrite are interpreted from mineral textures. The association of mineralisation and brittle deformation, the spatial distribution of mineralisation, and attributes of the mineral chemistry are also discussed.

5.6.1 Sulphide textures

The host rocks to mineralisation have undergone upper amphibolite facies metamorphism (Section 3.3.3). Sulphide and gangue mineral textures in the northern, central and southern zones are not consistent with those expected in high grade metamorphic terrains. Inconsistencies include:

• Sulphide and gangue minerals form veins, stockwork and breccias and have clearly replaced the host marble and calc-silicate lithologies (Figs. 5.1b-e and 5.3a, 5.3a, b, 5.11a-c).

• Adjacent sulphide or gangue minerals have not formed polygonal contacts (most Figs. from 5.3a-5.24b).

• Growth zoning in sphalerite 1, sphalerite 2, and quartz 1 is preserved and unlikely to have survived upper amphibolite facies metamorphism (Figs. 5.5a, b, 5.6a, b, 5.16a, c).

• Sphalerite 2 contains broad growth twins which are unlikely to have survived upper amphibolite facies metamorphism (Figs. 5.5d and 5.16e).

• Some sulphide and gangue minerals contain abundant inclusions of other minerals (eg. inclusions of chalcopyrite 2 in sphalerite 2, Figs. 5.5a-c, 5.16a-c). High grade metamorphism commonly results in 'refining' of minerals and 'expulsion' of inclusions (Turner, 1980).

• Finely crystalline minerals such as dolomite, talc 1 and 2, phlogopite, sericite 1 and 2, quartz 1 and Ca-Mn-Mg-Fe carbonate 1, 2 and 3 and chlorite 1, 2 and 3 have not been recrystallised as might be expected at higher metamorphic grades (Figs. 5.3a, b, 5.11b, c).

• Adularia is a disordered form of K-feldspar more commonly associated with low temperature epithermal systems (Henley et al. 1984; Figs. 5.4a-c).
The gangue minerals talc 1 and 2, chlorite 1, 2 and 3, and epidote are not stable under the peak P - T conditions interpreted for the Menninnie Dam host rocks (Turner 1980 and Section 3.3.3).

Sulphides have post dated deformation and locally, veins containing growth zoned sphalerite 2 have crosscut mylonitic textures (Fig. 5.12c).

Sulphides have post dated emplacement of granite dykes which were interpreted to postdate upper amphibolite facies metamorphism (Section 3.8.3; Fig. 5.12b)

Sulphide minerals have locally overprinted minerals interpreted to have formed during metasomatism (Fig. 5.3g).

Based on the interpretation of sulphide and gangue mineral textures, mineralisation cannot have been deposited at the same time as the host rocks and clearly post-dated peak metamorphism (KM-D1-2), retrograde metamorphism (KM-D3), granite dyke emplacement, and metasomatism.

5.6.2 Paragenetic stages

Northern and central zones

The sulphide and gangue mineral paragenesis has been divided into four stages (Fig. 5.2):

- An early pyrite stage
- A sphalerite - galena stage
- A late pyrite stage
- A late vein stage.

The early pyrite stage is associated with the dissolution of marble or diopside and the initial precipitation of dolomite, chlorite 1 ± talc 1 ± phlogopite, pyrite 1, and adularia. The start of the sphalerite - galena stage is marked by partial alteration of adularia to finely crystalline sericite 1. Sericite 1 was followed by sphalerite 1 and chalcopyrite 1, sphalerite 2 and chalcopyrite 2, quartz 1, Ca-Mn-Mg-Fe carbonate 1 and galena 1. The late pyrite stage is marked by the re-appearance of sericite (sericite 2) as well as chlorite 2 ± rhodonite + pyrite 2 followed by fluorite and Ca-Mn-Mg-Fe carbonate 3 respectively. The late vein stage consists of rare crosscutting veins of quartz 2 and common calcite veins (late calcite veins). The relationship between mineralisation and the late calcite veins is uncertain.

Southern zone

The sulphide and gangue mineral paragenesis in the southern zone is subdivided into four stages (Fig. 5.15):

- An early pyrite stage
- A sphalerite - galena stage
- A chlorite - quartz - carbonate - calc-silicate stage
- A late hematite stage
The early pyrite, and sphalerite - galena stages have similar mineralogy, textures and paragenesis to the early pyrite, and sphalerite - galena stages of the northern and central zones. The chlorite - quartz - carbonate - calc-silicate stage and the late hematite stage have been collectively termed the 'late alteration overprint'.

The apparent absence of chlorite 1 and sericite 1 in the early pyrite and galena - sphalerite stages of the southern zone is interpreted to have resulted from overprinting of these minerals by the late alteration overprint. The apparent absence of adularia in the southern zone is interpreted to have resulted from either local non-deposition of adularia or alteration of adularia during the late alteration overprint. Feldspar does not appear to have been a stable phase throughout the late alteration overprint, and feldspar in granite dykes has been altered to felted, finely crystalline chlorite, sericite, calcite and quartz (Fig. 5.13).

5.6.3 Association of mineralisation with fractures and brittle faults
Mineralisation and gangue textures are not typical of open space vein-fill and are interpreted to have resulted from replacement of host rock marble or diopside. Textures are interpreted to be consistent with replacement of marble along crystal margins and twin planes (Figs. 5.3a, 5.3b, c) and replacement of diopside along fractures and cleavage planes (Figs. 5.11a - c). In many hydrothermal systems such as skarns, Mississippi Valley Type deposits and carbonate replacement Pb - Zn - Ag bodies, carbonate is a reactive rock type and commonly host to mineralisation. While marble is commonly hosts mineralisation at Menninnie Dam, a significant amount of mineralisation is hosted by diopsidic units.

Diopsidic units that are host to mineralisation are invariably fractured or brecciated, and mineralising fluids are interpreted to have preferentially permeated along fractured or brecciated zones (Figs. 5.11a). Brittle deformation of diopside is interpreted to have resulted from competency contrasts between diopside and adjacent marbles or quartz units during deformation associated with KM-D3 (Section 3.5.2), or following KM-D3.

Marbles do not have a well developed network of fractures or breccia zones similar to that of diopsidic units and this may be due to recrystallisation of marble units. Rarely, fracturing of marbles has been observed (Fig. 5.1e) and in such instances, fracturing is interpreted to have post-dated KM-D3. Despite the apparent absence of a fracture network in marble units, mineralisation hosted by marbles is interpreted to have been associated with faulting.

Some sulphide textures are consistent with post-depositional deformation and others are interpreted to have resulted from syn-depositional deformation. Pyrite 1 locally has a jigsaw-fit breccia texture and has been overgrown by galena 1 that has a straight cleavage pattern (Fig. 5.3h). The margins of sphalerite 2 have locally undergone dynamic recrystallisation resulting in formation of sphalerite 3. Growth twins in sphalerite 2 adjacent
to sphalerite 3 have been replaced by stress twins (Fig. 5.5e). Recrystallised sphalerite 3
does not contain chalcopyrite inclusions and is locally overgrown by chalcopyrite 3,
interpreted to have been expelled from sphalerite 2 during dynamic recrystallisation.
Chalcopyrite 3 is overgrown by galena 1 with a straight cleavage pattern (Fig. 5.5f).
Brecciated pyrite 1 and sphalerite 2 fragments are locally overgrown by undeformed pyrite 3
(Fig. 5.9). Fractures and brittle fault structures may have been important fluid conduits
during the mineralising event with permeability maintained through syn-mineralisation
faulting.

5.6.4 Inclusions of chalcopyrite in sphalerite
Mechanisms that result in small inclusions of chalcopyrite in sphalerite are contentious and a
variety of processes have been proposed including:
- Recrystallisation or exsolution of copper and iron-bearing sphalerite (Ramdohr 1969).
- Replacement of FeS in sphalerite by a hot copper-bearing solution ('chalcopyrite
disease'; Barton, 1978; Eldridge et al., 1983; Barton and Bethke, 1987; Eldridge et al.,
- Co-precipitation of sphalerite and chalcopyrite (Kojima, 1990; Bortnikov et al., 1991)
- Replacement of sphalerite by chalcopyrite with introduction of both copper and iron
(Bortnikov et al., 1991)
- Solid state diffusion (Bente and Doering, 1993).

Barton and Bethke (1987) described a number of different textures of chalcopyrite inclusions
in sphalerite. 'Watermelon', 'dusting' and 'bimodal' textures were attributed to replacement
of FeS by copper rich fluids. 'Bead chain' textures were attributed to exsolution and
'orchard textures' were tentatively attributed to co-precipitation. Chalcopyrite diseased
sphalerite was produced experimentally by Kojima and Sugaki (1987) and Eldridge et al.
(1988) by reacting hot copper-bearing solutions with natural sphalerite. Chalcopyrite was
demonstrated to have replaced FeS in sphalerite. Experiments by Kojima (1990)
demonstrated that some chalcopyrite disease textures could be produced by co-precipitation
of chalcopyrite and sphalerite by cooling of a solution supersaturated with respect to both
sphalerite and chalcopyrite.

Bortnikov et al. (1991) suggested that replacement of FeS by copper-rich fluids may have
occurred in some cases, but was not the sole cause of chalcopyrite disease. Either co-
precipitation of sphalerite and chalcopyrite, or replacement of ZnS by chalcopyrite were
interpreted to have occurred in a number of Russian deposits. Bortnikov et al. (1991)
suggested in the absence of volume changes, the replacement of ZnS and FeS in sphalerite
by copper rich solutions should be associated with evidence for zinc mobility. Bortnikov et
al. (1991) also suggested that in many deposits where chalcopyrite inclusions were attributed
to co-precipitation, there was little correlation between the iron content of sphalerite and the occurrence of chalcopyrite inclusions.

Solid state diffusion experiments by Bente and Doering (1993) produced many of the chalcopyrite - sphalerite textures seen in nature and they proposed the term "diffusion induced segregations" (DIS) to replace the term chalcopyrite disease. Their findings included (among others):

- Significant DIS reactions can occur at temperatures below 300°C.
- There is a minimum Fe content below which no DIS are produced.
- Following precipitation of an Fe-bearing sphalerite, an increase in the sulphur fugacity of the fluid is required for the formation of DIS.
- DIS textures depend on the Fe content of sphalerite, time, temperature, the Cu : Fe ratio in the diffusing source and the sulphur fugacity.
- During formation of DIS, the Fe content of sphalerite is not only used to form DIS-phases, but Fe is also removed from sphalerite resulting in depletion of Fe in sphalerite.

Inclusions of chalcopyrite in sphalerite in the northern and central zones
Chalcopyrite disease (or DIS) is ubiquitous in sphalerite 1 and 2 throughout the northern and central zones of the Menninnie Dam prospect and has similar textures to the orchard texture described by Barton and Bethke (1987; Figs. 5.5a - c), however there is no evidence for a later, hotter, Cu rich fluid having overprinted sphalerite 2. Apart from local dynamic recrystallisation around the margin of sphalerite 2, there is no evidence for zinc mobility after deposition of sphalerite 2. Dynamic recrystallisation of sphalerite 2 has resulted in local segregation of chalcopyrite 3 from sphalerite 3 (Fig. 5.5f). Sphalerite 2 with chalcopyrite disease (or DIS) is commonly completely enclosed within quartz 1 or galena 1 interpreted to indicate either co-precipitation of sphalerite 2 and chalcopyrite 2, or, introduction of chalcopyrite 2 immediately after deposition of sphalerite 2 and prior to deposition of quartz 1 and galena 1. On the basis of textural evidence, chalcopyrite 1 and 2 are interpreted to be part of the sphalerite - galena stage and not related to a later, hot copper-bearing fluid.

Inclusions of chalcopyrite 2 in sphalerite 1 in the southern zone at Menninnie Dam
The late alteration overprint is interpreted to have resulted in dissolution, and re-deposition of pre-existing sulphides, particularly sphalerite 2, galena 1 and chalcopyrite 2 (Figs. 5.16a, c, 5.19a, h, 5.17, 5.20a, b). The lighter colours of the mid and outer sectors of sphalerite 2 in the southern zone compared to sphalerite 2 in the northern and central zones is interpreted to have resulted from removal of iron by fluids associated with the late alteration overprint (Figs. 5.16a, c). Chalcopyrite 2 inclusions in the mid and outer sectors of sphalerite 2 in the southern zone are commonly absent or smaller than those in the inner sector (Figs. 5.16a, c). The distribution and size of chalcopyrite 2 inclusions in sphalerite 2 is interpreted to have
resulted from partial, or complete, dissolution of chalcopyrite 2 in the mid and outer sectors of sphalerite 2 and local re-deposition as chalcopyrite 4 at the margins of sphalerite crystals. Veins of sphalerite 4, chalcopyrite 4 and pyrite 3 that crosscut sphalerite 2, rinds of galena 1 on sphalerite 2, and replacement of galena 1 by sphalerite 4 are interpreted as evidence for sulphide dissolution and re-precipitation (Figs. 5.16a-d; 5.17; 5.19a, b; 5.20a, b).

5.6.5 Spatial distribution of metals
Analytical values of lead, zinc, manganese, iron, copper, silver and gold commonly have a maxima centred around 9000 mN and progressively lower values to the north and south. The bell shaped distribution of analytical values (Fig. 5.25) is interpreted to have resulted from a single mineralising system whose conduit zone was centred around 9000 mN. A single mineralising system is also interpreted to account for the spatial distribution of mineralisation styles with stockwork and vein style mineralisation between 8200 mN and 9200 mN, flanked by vein style mineralisation to the north and south.

The copper to zinc ratio of the southern zone is higher than that of the northern and central zones (Fig. 5.26) and is interpreted to have resulted from either introduction of minor amounts of copper during the late alteration overprint, or local dissolution and removal of zinc. In the absence of unequivocal evidence for the introduction of copper during the late alteration overprint, removal of zinc is considered more likely.

5.6.6 Mineral chemistry
Silver
Van Hook (1960) demonstrated that argentite (Ag2S) has a maximum solubility in PbS of 0.4 mole % at 700°C, which is probably close to the assumed maximum solubility at the binary eutectic temperature of 615°C in the PbS-AgS system. The solubility of AgS in galena is negligible below 455°C which is near the upper limit for most ore fluid temperatures (Van Hook, 1960). Assuming the mineralisation at Menninnie Dam formed from a fluid of < 455°C, then, if argentite was present in galena, it would be present as an exsolved phase. However, argentite has not been observed in galena from Menninnie Dam.

The solubility of AgS in galena increases dramatically in the presence of bismuth and antimony with coupled substitution of Ag(1+) + Bi(3+) for 2Pb(2+), or Ag(1+) + Sb(3+) for 2Pb(2+) (Van Hook, 1960). PbS forms a series of solid solutions with AgBiS2 (matildite) or AgSbS2 (miargyrite) above 215°C. At lower temperatures, matildite is commonly exsolved, although about 8 mole % of matildite may remain dissolved within the galena structure at room temperature (Van Hook, 1960; Craig, 1967). The apparent absence of specific silver bearing minerals at Menninnie Dam and the occurrence of silver and bismuth in galena is best explained by solid solution of matildite in galena.
Galena from the Menninnie Dam prospect contains much more bismuth than can be accommodated in matildite alone. Van Hook (1960) noted the composition range of galena was not confined to the PbS-AgBiS₂ join line, and considerable amounts of Bi₂S₃ (up to 9 mole % outside the PbS-AgBiS₂ join line) could be accommodated.

**Tungsten**

The presence of tungsten in mineralised zones is questionable. No tungsten bearing minerals have been detected and tungsten could not be located within other minerals. The nature of the sample preparation techniques used by commercial laboratories prior to analysis of tungsten are unknown. Tungsten values of > 3000 ppm are considered unlikely to have been derived from contamination if a tungsten carbide mill was used for sample preparation. Tungsten and zinc have a positive, linear relationship and a correlation coefficient of 0.96 (Fig. 5.29). ‘Tungsten-bearing samples’ were analysed at Analabs, Perth, W.A., by XRF using the 1st order L beta and K beta peaks for tungsten and zinc respectively (pers. comm. A. Ronald, Analabs, Perth, W.A.). In the presence of significant amounts of zinc, the zinc K beta and the tungsten L beta peaks cannot be resolved. It seems likely that the presence of tungsten may have been misinterpreted in the presence of abundant zinc.

### 5.6.7 Relative timing of DP breccias and rhyolite dykes

The sulphide and gangue paragenesis established for the northern and central zones can be used to interpret a relative timing for formation of DP breccias. Sulphide and gangue minerals in sulphide clasts are paragenetically earlier than the late pyrite stage of mineralisation. The presence of pyrite 2 and trace amounts of galena 1 and sphalerite 2 as part of the alteration assemblage that affected the matrix of DP breccia is interpreted to indicate DP breccias formed near the end of the sphalerite - galena stage of mineralisation. DP breccias are interpreted to be temporally and genetically related to emplacement of rhyolite intrusions (Section 4.4.7), therefore, the timing of rhyolite intrusions is also interpreted to have occurred near the end of the sphalerite - galena stage of the mineralising event.

### 5.6.8 Relative timing of mineralisation, the MD ignimbrite and the late alteration overprint.

The relative timing of the mineralising event, deposition of the MD ignimbrite and the late alteration overprint can be established. LP breccias are interpreted as surficial deposits that resulted from DP breccia bodies erupting onto the palaeosurface (Section 4.4.7) and therefore also formed near the end of the sphalerite - galena stage of mineralisation. Sulphide mineralisation in the southern zone is interpreted to have been overprinted by the late alteration overprint. LP breccia and the basal 34m of the MD ignimbrite in drill hole PD0032 have been altered by the late alteration overprint (Section 5.3.2) and sulphide-bearing veins crosscut welded MD ignimbrite (Fig. 5.24b). The pattern of concentric
alteration bands in marble clasts in the MD ignimbrite is interpreted to have resulted from in-situ alteration (Fig. 5.24a). Therefore, the late alteration overprint is interpreted to have post dated both the mineralising event and emplacement, welding and cooling of the MD ignimbrite. The time period between the mineralising event and the late alteration overprint cannot be established from geological associations alone.

5.7 Summary
Sulphides in the northern and central zones have replaced marbles and calc-silicate lithologies of the central suite. Sulphides in the southern zone have replaced marble and calc-silicate lithologies of the eastern suite. Sulphide and gangue minerals form veins and stockwork style mineralisation. Stockwork and vein style mineralisation is more common between 8200 mN and 9300 mN flanked by vein style mineralisation to the north and south. Mineralisation in marbles is interpreted to have been associated with both pre-mineralisation and syn-depositional brittle faulting. Calc-silicate lithologies have been preferentially replaced along fractures and breccia zones. Sulphide clasts occur scattered throughout DP breccias and LP breccias and minor sulphides occur throughout the matrix of DP breccias and to a lesser extent, the matrix of LP breccias.

Sulphide and gangue minerals in the northern and central zones consist of a simple assemblage of pyrite, quartz, sphalerite, galena and Ca-Mn-Mg-Fe carbonate with minor amounts of chalcopyrite and chlorite and trace amounts of adularia, sericite, fluorite, rhodonite, talc, phlogopite, dolomite, hematite and matildite. Chalcopyrite occurs as inclusions in sphalerite and is interpreted to have resulted from either co-precipitation or was included immediately after sphalerite deposition. Sulphide and gangue minerals have not been metamorphosed and most primary textures are preserved. The sulphide and gangue mineral paragenesis has been used to interpret the a syn-mineralisation timing for rhyolite intrusion, and DP and LP breccia formation.

Pb, Zn, Mn, Fe, Cu, Ag and Au analytical values have bell shaped distributions that have a maxima at ~9000 mN and decrease to both the north and south. Maximum assay values correspond with the zone of stockwork and vein style mineralisation. In the northern and central zones, manganese is incorporated in Ca-Mn-Mg-Fe carbonate 1 and 2 with minor amounts in sphalerite and rhodonite. In the southern zone manganese is incorporated in Ca-Mn-Mg-Fe carbonate 1 and 2 and rhodochrosite with minor amounts in sphalerite. Silver is present as matildite dissolved in galena. Tungsten analyses reported by commercial laboratories resulted from mis-interpretation of XRF analyses and significant tungsten is absent in the Menninnie Dam mineralisation.

The southern zone is the southern periphery of the mineralised system and has been overprinted by a later hydrothermal event (‘late alteration overprint’). The late alteration
overprint resulted in pervasive calcite - chlorite - quartz - carbonate - calc-silicate - hematite alteration of the Hutchison Group, granite dykes, mineralisation and the basal 34m of the MD ignimbrite. Pre-existing sulphide and gangue mineralisation was partially dissolved by the late alteration overprint and re-precipitated both locally and distally. The late alteration overprint postdates emplacement of the MD ignimbrite.