

The Rayner Complex in MacRobertson Land, east Antarctica

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Declaration

I declare that this thesis is less than 100,000 words in length, exclusive of tables, figures, references and appendices, and that the work contained in this thesis has not been submitted for a higher degree at any other university or institution.

*Crevasse near Rumdoodle Peak, Masson
Range, January 1993*



Abstract

The Rayner Complex in MacRobertson Land, east Antarctica, is a Proterozoic mobile belt composed of granulite facies gneisses that were metamorphosed at mid-crustal levels during the *c.* 1000-900 Ma Rayner Structural Episode. At Cape Bruce on the Mawson Coast, gneisses are present that represent a metamorphosed supercrustal sequence of probable Proterozoic age. Felsic and pelitic sedimentary rocks were buried at mid-crustal levels prior to the emplacement of charnockite before *c.* 1000 Ma. These lithologies were metamorphosed, migmatized and deformed into a gneissic fabric by D₁-M₁ during the early stages of the Rayner Structural Episode. Granitoids and a 995±7 Ma biotite-rich orthogneiss were intruded after D₁. Peak metamorphic conditions of 6-7 kbar and ≥750°C during M₁ were followed by cooling at near-constant pressures, which produced hydrous M₂ mineral assemblages. M₂ occurred before or during intense D₂ deformation, that produced a strong foliation in the gneisses. At 944±10 Ma, felsic dykes and pegmatites were intruded that may have derived from partial melting during D₁ and D₂. Lithologies were subsequently deformed into upright north-trending folds by D₃, which represented the waning stage of continuous D₁ to D₃ deformation. Felsic pegmatites were emplaced after D₃, and were deformed into upright east-trending folds during D₄. The growth of M₃ minerals at *c.* 910Ma occurred after D₃ and before or during the D₄ event. M₃ occurred at lower temperatures than and similar pressures to M₁ and M₂. Narrow east-trending D₅ shear zones, containing mylonite and pseudotachylite, formed after D₄. These shear zones contain M₄ mineral assemblages that grew at approximately 6.5 kbar and 550-800°C. Whereas D₁ to D₄ occurred during the *c.* 1000-900 Ma Rayner Structural Episode, D₅ may have occurred at a much later time.

On the basis of geochronological data from Cape Bruce and other localities, terranes with distinct ages of formation can be recognised in Kemp and MacRobertson Land. Archaean gneisses are exposed on the Mawson Coast west of the Stillwell Hills. In MacRobertson Land, the protoliths of exposed gneisses were formed within the period 1800-1000 Ma. At Cape Bruce, isotopic data provide evidence for unexposed Archaean crust in the region, whereas isotopic data from the northern Prince Charles Mountains and the Framnes Mountains yield inherited ages of 2200-1800 Ma. If the isotopic data represent inherited crustal material from the lower crust, the Rayner Complex in MacRobertson Land was underlain by mostly Paleoproterozoic crust during the Rayner Structural Episode. By contrast, the Cape Bruce area was underlain in part by Archaean crust.

During the peak of metamorphism in the Rayner Structural Episode, pressures and temperatures were sufficiently high to cause ionic diffusion on a cm scale. At Rumdoodle Peak in the Framnes Mountains, a xenolith of metasedimentary granulite is enclosed in Mawson Charnockite. The xenolith contains a disrupted layer of black spinel-orthopyroxene-phlogopite gneiss, that was truncated by quartz-rich pegmatite. A reaction band that developed between these units is composed of a sequence of mineral zones that contain spinel, orthopyroxene, sapphirine, cordierite and plagioclase. The sequence of mineral zones approximately matches that predicted by a model of closed system diffusion metasomatism, involving the exchange of Si for Fe and Mg. The sequence differs from that predicted by the model in the presence of micro-scale disequilibrium textures, that include 'double coronas' composed of cordierite surrounding sapphirine and sapphirine surrounding spinel. The growth of the reaction band was controlled by diffusion along intergranular pathways, with local equilibrium maintained adjacent to grain boundaries. The presence of corona textures was a result of slow reaction rates, due to limited diffusive exchange of Si and Al across mineral grains.

Similarities between the structural histories of various localities in MacRobertson Land indicate the operation of regional scale processes during the Rayner Structural Episode. Tectonism involved the compression of Proterozoic crust against Archaean continental crust in Enderby Land and the Indian peninsula. The orogeny occurred in an oblique-collisional setting. The early stages of the Rayner Structural Episode involved high-strain deformation associated with a large component of east-west compression, which produced east-over-west thrust structures in the Stillwell Hills and upright north-trending structures in the Framnes Mountains. Charnockitic magmas were emplaced in the MacRobertson during or after this period of compression. In the late stages of the Rayner Structural Episode, there was a regionally significant transition from a contraction-dominated regime to a transcurrent tectonic regime. The latter produced lateral movement along sub-vertical east-trending high-strain zones, and north-south compression in domains of lower strain. The transition was a result of either: 1) progressive deformation during east-west or southeast-northwest convergence at an oblique angle to the continental margin, or 2) a change in the direction of tectonic convergence, from east-west to north-south. Similar transitions have been commonly observed in oblique-collisional orogens. East-trending semi-brittle shear zones, that had a predominantly south-over-north sense of movement, were generated by north-south compression at some time after the Rayner Structural Episode.

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“Oh the powers of the understanding, which dwells at the same moment with equal intensity upon an insect and a world! That considers a universe - worlds upon worlds - until, attempting infinity, its wings failing like those of Icarus, it drowns in the vasty deep of its own contemplative nothingness. Then the chastened soul seeks her accustomed channels and pensively attempts objects less calculated to confound her”...

- *Thomas Hawkins, ichthyosaur collector and madman, 1834.*

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This thesis is dedicated to the memory of my grandfather, Joe Bennett, from whom I inherited my need to know.

Introduction

Introduction

The Rayner Complex is a Proterozoic mobile belt that extends along the coast of east Antarctica between latitudes 45°E and 80°E. The belt is composed of upper amphibolite to granulite facies gneisses that were deformed and metamorphosed during the *c.* 1000 Ma Rayner Structural Episode of Sandiford & Wilson (1984). Numerous geological and geochronological studies in Enderby Land, Kemp Land and MacRobertson Land have refined our knowledge of the origin and evolution of the mobile belt, and of the associations of the complex with adjacent Archaean cratons in east Antarctica and India. Recent work has radically improved our understanding of the timing and extent of orogenic processes in the region. In particular, fresh perspectives have been obtained from the detailed investigation of small localities, in areas such as the Mawson Coast (*e.g.* White & Clarke, 1993), the northern Prince Charles Mountains (*e.g.* Kinny *et al.*, 1993; Hand *et al.*, 1994a & b), and Prydz Bay (*e.g.* Zhao *et al.*, 1995; Hensen & Zhou, 1995; Carson *et al.*, 1996; Fitzsimons 1996). In this study, geological investigations of the Cape Bruce area on the Mawson Coast, MacRobertson Land, are described and compared with other studies, in order to define the evolution of the Rayner Complex in MacRobertson Land.

Aims of this study

The present study was conducted to improve the understanding of the Rayner Complex in MacRobertson Land. In order to do this, a detailed geological investigation was made of the Cape Bruce locality, which lies on the Mawson Coast in MacRobertson Land, east Antarctica. New geological and geochronological information derived by this investigation was compared to work done in other parts

of the Rayner Complex, in order to define the timing and extent of the geological processes that formed the Rayner Complex. In particular, the regional significance of the *c.* 1000 Ma Rayner Structural Episode was addressed. The recognition of large scale tectonic processes in the development of the Rayner Complex in MacRobertson Land has enabled the author to tentatively propose a model for the Rayner Structural Episode, involving orogenesis in an oblique-collisional setting.

Thesis outline

An investigation was conducted by the author into the geology of Cape Bruce, an area of rock exposure on the Mawson Coast in MacRobertson Land, east Antarctica. Detailed field mapping and sampling was conducted during the 1992 and 1993 Australian National Antarctic Research Expeditions. Based on this field work, detailed studies were conducted on the petrology, geochronology and structural geology of the rocks of Cape Bruce. These studies enable the author to present an in-depth geological history of Cape Bruce, as well as to propose new ideas on the geological evolution of MacRobertson Land and Kemp Land.

To understand the geological processes that operated in the Cape Bruce area in particular and in the Rayner Complex in general, a review is required of work already published on the subject. In particular, geochronological data needs to be compiled, to provide reliable constraints on the timing of geological events in the Rayner Complex. **Chapter 1** provides an introduction to the geology of the Rayner Complex. The results of previous work on the rocks of the Mawson Coast are presented. A review is also provided of the large body of geochronological work published on the Rayner Complex.

Chapter 2 presents the results of new field mapping in the Cape Bruce area. A geological history is defined that consists of a series of deformational and intrusive events, that were accompanied by metamorphism at granulite facies conditions.

Correlations by previous authors between the geological histories of localities on the Mawson Coast are re-assessed in light of the new evidence from Cape Bruce.

The development of metamorphic mineral assemblages in the Cape Bruce gneisses is addressed in **Chapter 3**. Four distinct episodes of granulite facies metamorphism are defined, based on the detailed petrographic study of thin sections. Geothermobarometry on key assemblages is described to provide an estimate of P-T conditions during the peak of metamorphism. A diagram is constructed, showing the stability of minerals in P-T space for Fe^{3+} and Ti-rich pelitic rocks. The petrogenesis of mineral assemblages during the four stages of metamorphism is discussed.

Chapter 4 describes the results of SHRIMP geochronology on six samples from Cape Bruce. A new statistical method, the mixture modelling technique of Sambridge & Compston (1994), is applied to four of the samples. From this modelling, the ages of two periods of metamorphic activity were resolved. Ages of emplacement for orthogneiss samples provide constraints on the timing and duration of the Rayner Structural Episode.

Chapter 5 discusses the formation of reaction bands in a sapphirine-bearing granulite from Rumdoodle Peak in the Framnes Mountains. The reaction bands are attributed to diffusion metasomatism under near-anhydrous granulite facies conditions. The relative chemical potentials of minerals in the reaction bands are used to develop a model for the generation of reaction textures by diffusive transfer.

The evolution of the Rayner Complex in MacRobertson Land is addressed in **Chapter 6**. Comparisons are made between geological data from Cape Bruce and other localities. The geochronology of various localities are correlated, and these correlations are used to infer the presence of age-defined terranes in the Rayner Complex. Structural histories are also correlated, in order to produce a regional framework for the Rayner Structural Episode. From this framework, tectonic models

are proposed that involved the oblique collision of the Rayner Complex with Archaean continental crust in Enderby Land and east India.

All mineral abbreviations are as *per* Kretz (1983). Other abbreviations are explained where used in the text.

Originality of thesis and published components

Unless otherwise stated, all research and interpretations presented in this thesis are the work of the author. No part of the work in this thesis has been submitted for any other degree. Some material from this thesis has been published or submitted for publication in the following references:

Dunkley, D.J., Clarke, G.L., Nelson, D. and White, R.W., 1995. Temporal resolution of complex tectono-metamorphic episodes: a case Proterozoic orogen at Cape Bruce, MacRobertson Land, east Antarctica (abstract). In: J. Müller (Editor), VII International Symposium on Antarctic Earth Sciences Abstracts, p. 116.

Clarke, G.L., Sun, S.S., White, R.W. and Dunkley, D.J., 1995. Review and correlation of Grenville age belts and associated older terrains in Australia and Antarctica (abstract). In: J. Müller (Editor), VII International Symposium on Antarctic Earth Sciences Abstracts, p. 83.

Dunkley, D.J., Clarke, G.L., and Harley, S.L., 1996. Diffusion metasomatism in aluminous sapphirine-bearing granulite from Rumdoodle Peak, Framnes Mountains, east Antarctica (abstract). In: I.S. Buick and I. Cartwright (Editors), Evolution of Metamorphic Belts. Geological Society of Australia Abstracts 42: 15.

Dunkley, D.J., Clarke, G.L., and Harley, S.L., submitted. Diffusion metasomatism in silica-undersaturated sapphirine-bearing granulite from Rumdoodle Peak, Framnes Mountains, east Antarctica. Submitted to Contributions to Mineralogy and Petrology, October 1997.

Chapter 1

Regional Geology

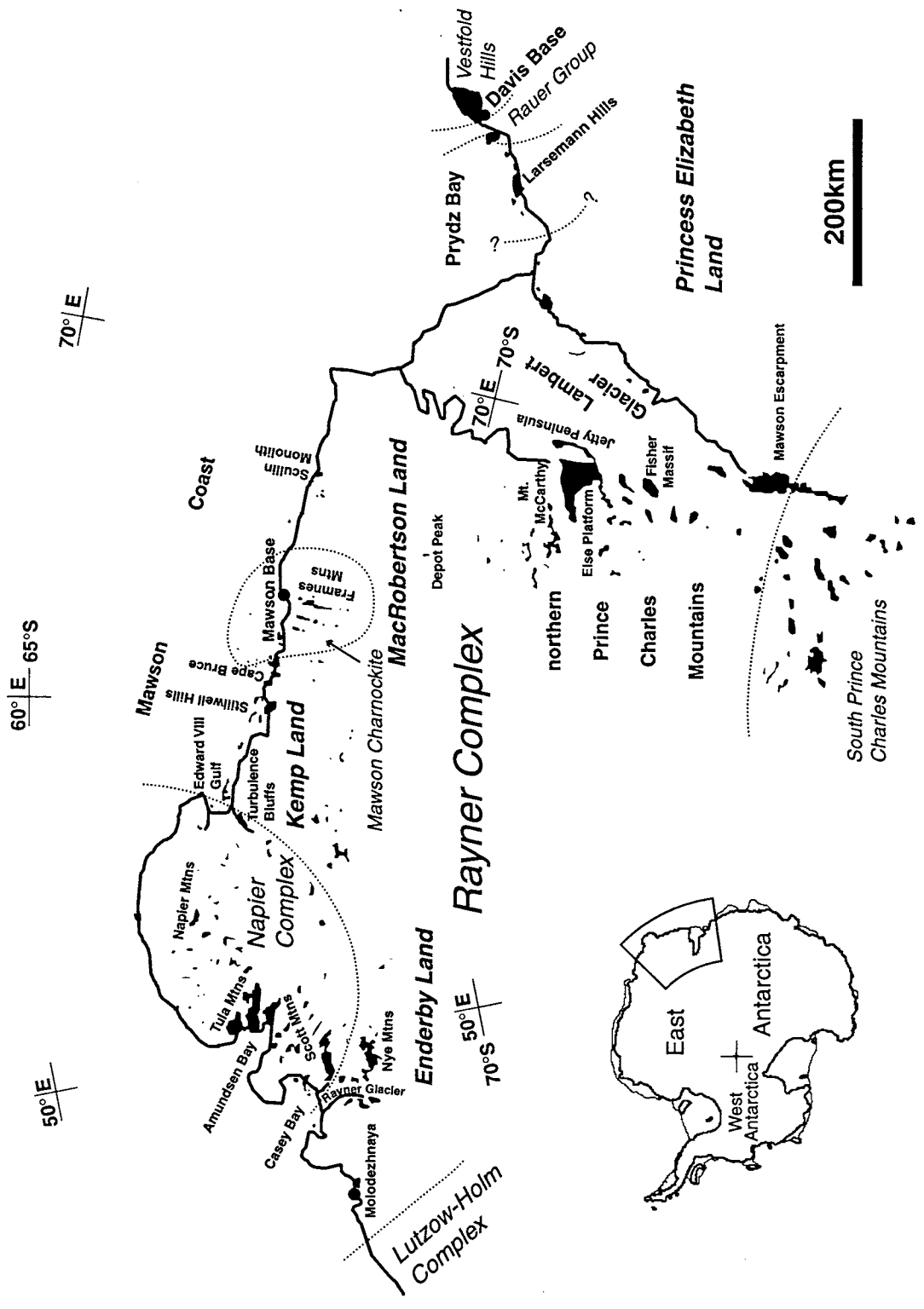
1. Regional Geology

1.1 The Rayner Complex

The East Antarctic continent consists of a large Precambrian shield, composed of numerous Archaean cratons separated by Proterozoic to early Cambrian mobile belts (Tingey, 1991). Geological exposures between latitudes 45°E and 80°E along the East Antarctic coast are considered to form a single mobile belt, known as the *Rayner Complex* (Sheraton *et al.*, 1980; Clarke, 1988; Harley & Hensen, 1990; Tingey, 1991). The Rayner Complex was defined by Kamenev (1972), who used the term to distinguish granulites in Enderby Land that contained biotite and/or hornblende from anhydrous granulites of higher metamorphic grade. The latter granulites were defined as belonging to the Napier Complex (Fig. 1.1). Gneisses of similar metamorphic grade to Rayner Complex exposures in Enderby Land were recognised in Kemp Land (Sheraton *et al.*, 1980) and MacRobertson Land (Tingey, 1982; Clarke, 1987). In this study, the term 'Rayner Complex' is used to define all exposures of this mobile belt in Enderby Land, Kemp Land and MacRobertson Land, consistent with the usage of Clarke (1988).

Although the timing of granulite facies metamorphism in the Rayner Complex was originally considered to be Archaean (Kamenev, 1972), numerous geochronological studies have established that the major metamorphic and deformation events occurred in most localities at *c.* 1000 Ma (*e.g.* Black *et al.*, 1987; Tingey, 1991; Kinny *et al.*, 1997). Archaean cratonic blocks abut the Rayner Complex in Enderby Land (the Napier Complex; Sheraton *et al.*, 1987) and MacRobertson Land (the southern Prince Charles Mountains; Tingey, 1991). Recent geochronological work has shown that in areas on the western and eastern extremities of the Rayner Complex, *i.e.* near Molodezhnaya Base and Prydz Bay (Fig. 1.1), gneisses were formed during a *c.* 500

Fig. 1.1. Locality map of the Rayner Complex, as defined by the author. Dotted lines mark the approximate boundaries of major geological blocks. Inset shows the location of the Rayner Complex in East Antarctica.



Ma metamorphic event (e.g. Shiraishi *et al.*, 1994; Zhao *et al.*, 1992). The geochronology of the Rayner Complex is reviewed in Section 1.3.

The c. 1000 Ma event deformed and metamorphosed a mixture of igneous and sedimentary lithologies. Lithological relationships have been mapped in Enderby Land (Sheraton *et al.*, 1980), on the Mawson Coast (Trail, 1970; Clarke, 1988; White & Clarke, 1993), and in the northern Prince Charles Mountains (Fitzsimons & Thost, 1992; Hand *et al.*, 1994b). In most localities, felsic orthogneiss predominates, and is associated with lesser amounts of intermediate to mafic pyroxene orthogneiss and rare ultramafic orthogneiss (Sheraton *et al.*, 1980; Clarke, 1988; Fitzsimons & Thost, 1992). Layered sequences of paragneiss are also present in all areas, and form a significant part of the outcrop area at Cape Bruce (Fig.1.1), the Colbeck Archipelago (White & Clarke, 1993), the Casey Range (Clarke *et al.*, 1989) and the Jetty Peninsula (Hand *et al.*, 1994b). Paragneiss includes psammitic and pelitic lithologies, with common but volumetrically minor occurrences of calc-silicate granulite.

The c. 1000 Ma episode that formed granulites in the Rayner Complex was termed the *Rayner Structural Episode* by Sandiford & Wilson (1984). This term was also applied to metamorphism and deformation on the Mawson Coast by Clarke (1988) and White & Clarke (1993). This study re-defines the Rayner Structural Episode as a c. 1000 Ma tectonic event that affected all gneisses in the Rayner Complex. The Rayner Structural Episode in all localities is associated with high-strain plastic deformation, syn-tectonic igneous activity, high-grade metamorphism and the migmatization of pre-existing lithologies.

1.2 Geological work on the Mawson Coast

Geological mapping was conducted along the Mawson Coast from MacRobertson Land to Kemp Land and reported in Crohn (1959), McCarthy & Trail (1964), Trail

et al. (1967) and Trail (1970). Gneisses were subdivided by Trail *et al.* (1967) into four major units (Fig. 1.2). The Stillwell Gneiss groups together paragneiss and felsic to mafic orthogneiss that outcrop west of the Scoble Glacier (Fig. 1.2). The Colbeck Gneiss dominates outcrops from Campbell Head to Chapman Ridge (Fig. 1.2), and comprises metasedimentary and granitic gneisses. The Framnes Mountains (Fig. 1.2) is the type locality of the Painted Gneiss, which comprises a variety of metapelites, felsic gneisses, calc-silicates and mafic to felsic orthogneiss. Painted Gneiss outcrops as isolated xenoliths or roof pendants within the late Proterozoic Mawson Charnockite, which also intruded the Colbeck Gneiss. The Mawson Charnockite is an extensive batholith of syn-tectonic orthogneiss, that outcrops throughout the Framnes Mountains and as far west as Chapman Ridge (Fig. 1.2).

Detailed geological mapping has been conducted in the Stillwell Hills (Clarke, 1987; Clarke, 1988), the Framnes Mountains (Clarke, 1988) and the Colbeck Archipelago (White & Clarke, 1993). The sequence of geological events in these areas are summarised in Table 1.1. In the Stillwell Hills, Clarke (1988) subdivided the Colbeck Gneiss into layers of quartzofeldspathic, pelitic and mafic gneiss. The Stillwell Gneiss was identified as being composed of charnockitic to mafic orthogneiss. These lithologies contained a gneissic 'compositional layering' (*sic* Clarke, 1988), that was truncated by layers of basic gneiss, which were correlated by Clarke (1988) with 1190 ± 200 Ma tholeiitic dykes in the Napier Complex (Sheraton *et al.*, 1980). These lithologies were folded and metamorphosed during the Rayner Structural Episode (D₁-D₃, Clarke, 1988).

Simple geological maps were produced by Clarke (1988) for Painted Hill and Rumdoodle Peak in the Framnes Mountains (Fig. 1.2), and for islands in the vicinity of Mawson base (Clarke, 1988; Clarke & Norman, 1993). Painted Hill (Fig. 1.2) consists of felsic, pelitic and calc-silicate gneisses, with a gneissic compositional layering that was deformed during the Rayner Structural Episode (D₁ & D₃, Clarke 1988). Gneissic layering in paragneiss was truncated by the Mawson Charnockite

Fig. 1.2. Simplified geological map of the Mawson Coast, based on mapping by Trail *et al.* (1967).

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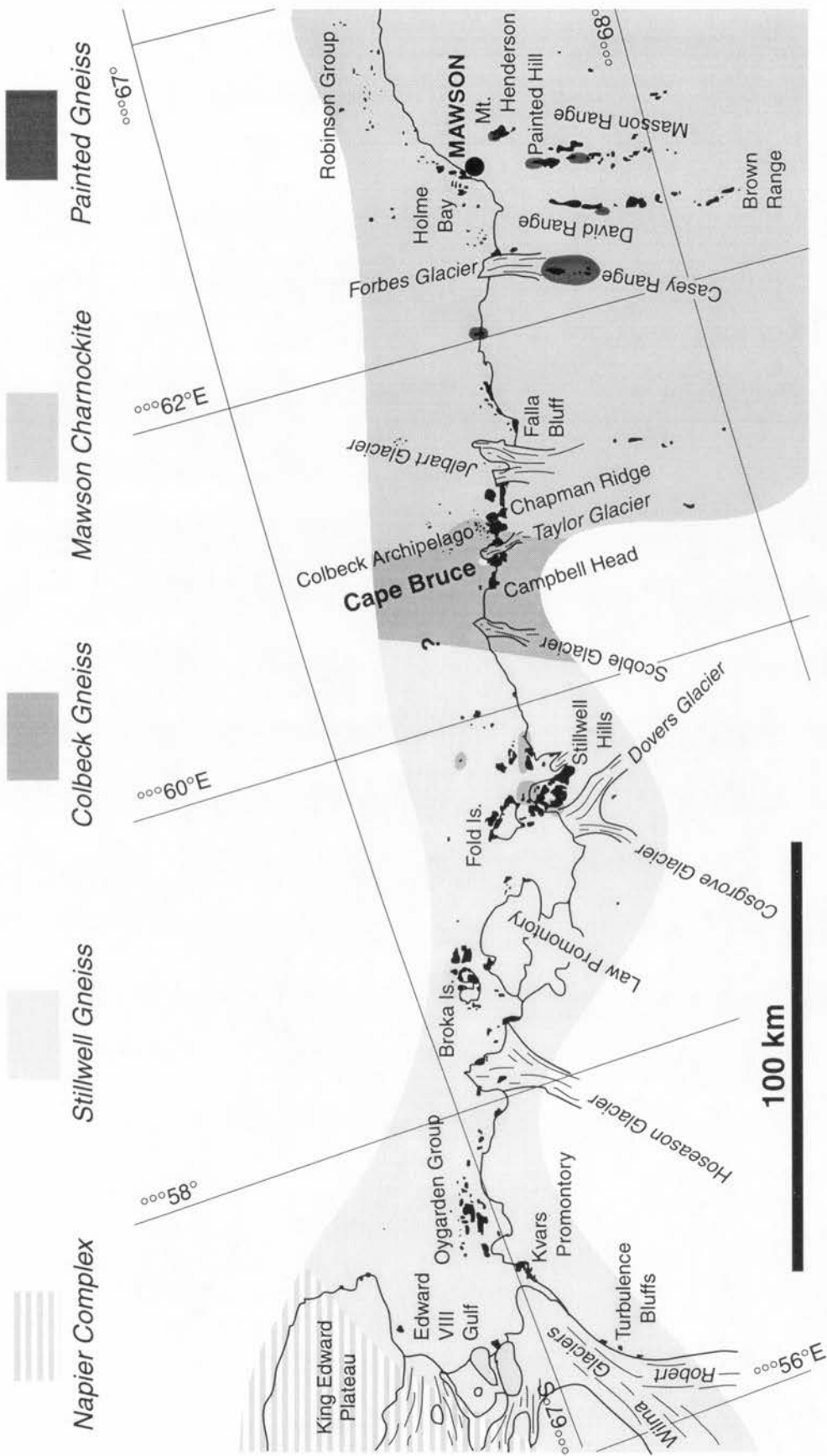


Table 1.1. Summary of the geological histories of the Colbeck Archipelago, Stillwell Hills and the Framnes Mountains. Events at the same level in each column have been correlated by previous authors; correlations between the Colbeck Archipelago and the Stillwell Hills are by White & Clarke (1993), and correlations between the Stillwell Hills and the Framnes Mountains are by Clarke (1988).

	Colbeck Archipelago	Stillwell Hills	Framnes Mtns & Mawson Base
	White & Clarke, 1993	Clarke, 1988	Clarke 1988
	Metasediments, mafic orthogneiss, charnockitic orthogneiss	Metasediments, granitoids, c.2700Ma pre-syn-Dc charnockite	Metasediments, mafic orthogneiss & dykes, granitoids
	D₁ S1 dominant gneissosity, partial melting. Intrafolial F1	D_c Sc compositional layering, partial melting. Intrafolial Fc	D_c Sc compositional layering, partial melting. Intrafolial Fc
	Biotite-pyroxene orthogneiss, granitic gneiss, layered gneiss	1190±200 Ma mafic dykes	
R A Y N E R	D₂ S2 dominant foliation, partial melting. Macro-mesoscopic isoclinal F2 with steep E dipping axial planes in areas of low D3 strain. Steep NE-E plunging F2 axes. L2//F2 axes	D₁ S1 dominant foliation // Sc. N-NE trending recumbent isoclinal F1. W over E shear sense. Sub-horizontal W trending L1, normal to parallel with F1 axes	D₁ S1 dominant foliation, partial melting. Mostly moderate-steep E dip S1. Tight-isoclinal E plunging F1, L1//F1 axes.
	c. 980-950Ma Mawson Charnockite, pegmatites	940±80 Ma pegmatites	c. 980-950Ma Mawson Charnockite
	D₃ Macro-mesoscopic open-isoclinal F3 with steep S dipping axial planar S3. F3 axes plunge steeply E. F3 possibly coaxial with F2. Controls NE trend of S2/S1 in high D3-strain zones	D₂ Low strain. Vertical W trending S2. Open-isoclinal upright F2 with horizontal W trending axes and stretching direction	
S T R U C T U R A L		940±80 Ma pegmatites	
	D₄ ?Weak vertical N-trending S4 in Mawson charnockite	D₃ Steep E dipping S3. Open-isoclinal upright F3 with sub-horizontal N trending axes. E plunging L3	D₃ S3 dominant foliation in Mawson Charnockite. Vertical to steep E dipping S3. Minor open-isoclinal F3 with sub-horizontal N-trending axes
		≥ 770Ma? pegmatites	
E P I S O D E	D₅ MPS shear zones with steep N-NW dip and S over N movement. N plunging L5.	D₄ MPS shear zones with S dip and S over N movement. SW-SE plunging L4	D₄ MPS shear zones with moderate to steep SW-SE dips and S over N movement. S-SW plunging L4

between Painted Hill and Rumdoodle Peak. The Mawson Charnockite contains a foliation that has been correlated with D₃ structures in the Painted Gneiss (Table 1.1; Clarke, 1988). All lithologies in the Framnes Mountains and on islands around Mawson Base were transected by discrete shear zones that contain ultramylonite and pseudotachylite (D₄, Clarke, 1988; Clarke & Norman, 1993). Similar shear zones were observed in the Stillwell Hills (Clarke, 1988).

The gneisses of the Colbeck Archipelago and Chapman Ridge, that were previously mapped as Colbeck Gneiss by Trail *et al.* (1967), form a complex association of paragneisses and pre- to syn-tectonic orthogneisses (White & Clarke, 1993). The oldest lithologies comprise the *charnockitic orthogneiss* (italics after White & Clarke, 1993), and the interlayered paragneisses and mafic gneisses of the *proclamation Gneiss*. These units were deformed and metamorphosed to granulite facies during D₁, which White & Clarke (1993) correlated with the compositional layering observed by Clarke (1988) in the Stillwell Hills and the Framnes Mountains (Table 1.1). Gneissic fabrics that were produced by D₁ were truncated by the intrusion of the *biotite±pyroxene, granitic and layered gneisses* (White & Clarke, 1993). Mafic dykes were also emplaced after D₁, and were correlated by White & Clarke (1993) with the 1190±200 Ma tholeiitic dykes in the Napier Complex (Sheraton *et al.*, 1980). Post-D₁ orthogneisses were considered by White & Clarke (1993) to have intruded just before or during high-strain D₂ deformation, in the early stages of the Rayner Structural Episode (Table 1.1). D₂ was followed by the intrusion of the Mawson Charnockite and numerous associated felsic dykes and pegmatites at Chapman Ridge. F₃ open folding occurred after the intrusion of the Mawson Charnockite, and was correlated by White & Clarke (1993) with the D₂ event of Clarke (1988) in the Stillwell Hills. All lithologies were subsequently transected by ultramylonite-pseudotachylite shear zones (D₅, White & Clarke, 1993).

1.3 Geochronology of the Rayner Complex: a review

A large number of studies have provided temporal constraints on the geological evolution of the Rayner Complex. Comparisons between geochronological data from different localities can be used to determine the regional significance of tectonic, structural and metamorphic events. In order to make such comparisons, the author has compiled the following review of the available geochronological data. A summary of localities that have been dated in Enderby, Kemp, MacRobertson and Princess Elizabeth Land is presented in the map in Fig. 1.3. Geochronological data are described below. The review is broken up into sections according to age. Data in each section is presented in order of location, from west to east across the Rayner Complex.

Archaean crustal ages

The Proterozoic Rayner Complex is abutted at three points by Archaean terranes (Fig. 1.3). In Enderby Land, the Napier Complex is composed of a series of igneous bodies and supercrustal sequences that evolved over the period 3950 - 2450 Ma (Sheraton *et al.*, 1987, Black *et al.*, 1991). These lithologies experienced multiple episodes of granulite facies metamorphism, with the earliest episode having occurred at 2837 ± 15 Ma (U-Pb in zircon, Harley & Black, 1997), under anhydrous conditions with pressures of up to 10kbar. Rehydration of lithologies during metamorphism around the western and eastern boundaries of the Napier Complex has been correlated with the formation of granulites within the Rayner Complex (Sheraton *et al.*, 1987).

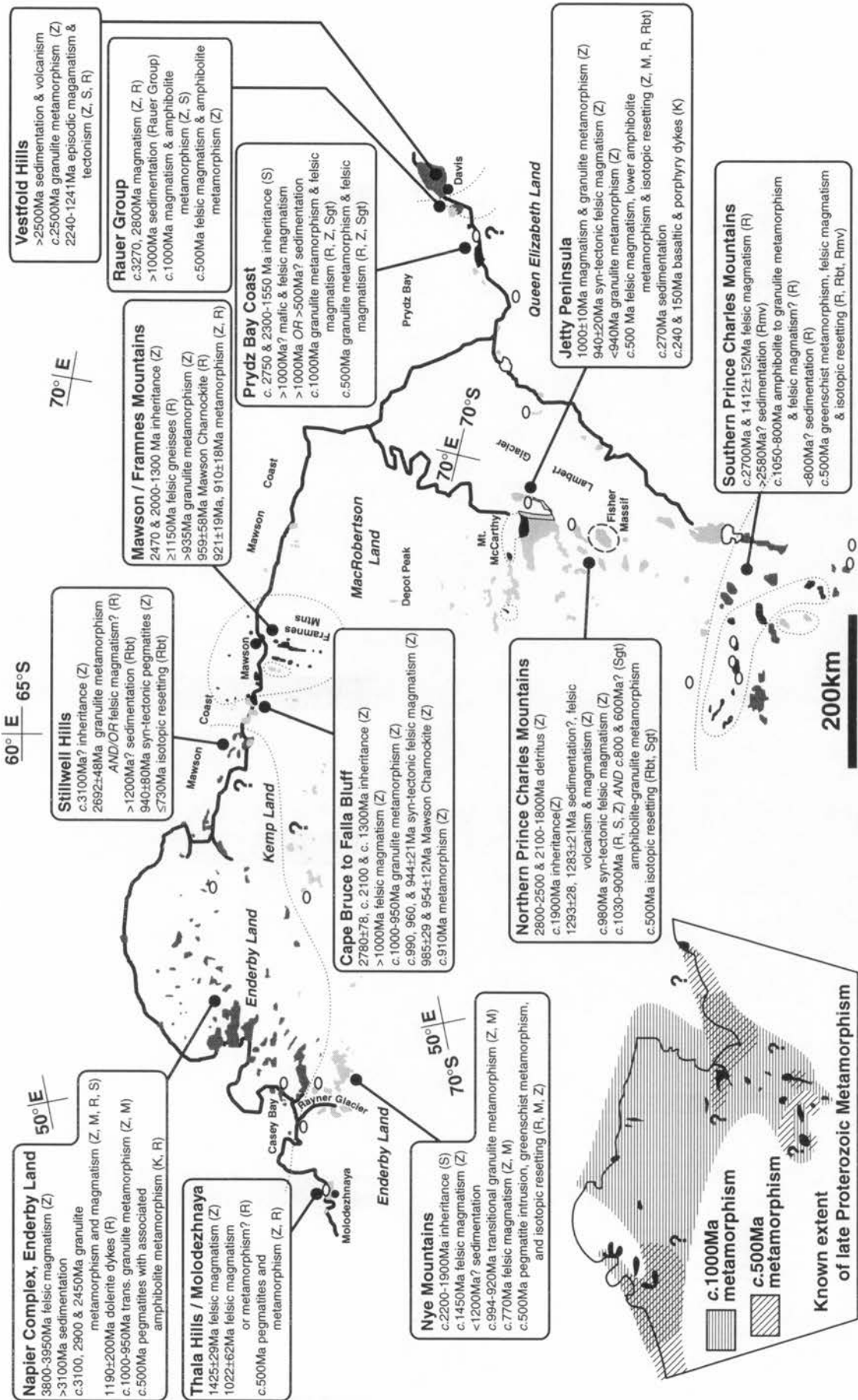
The second Archaean terrane occurs in the southern Prince Charles Mountains, where *c.* 2700 Ma granitoids were cut by Mesoproterozoic granitoids and subsequently metamorphosed during the Neoproterozoic (Tingey, 1991). The third site of Archaean exposures lies at the eastern extremity of the Rayner Complex (Fig. 1.3). The Vestfold Hills preserve granulite facies gneisses, which include *c.* 2532 - 2478 Ma

Fig. 1.3. Geochronological map of the Rayner Complex. Age constraints on geological events are covered in more detail in the text. Abbreviations in brackets indicate the isotopic system on which each estimate is based: Z = U-Pb in zircon; M = U-Pb in monazite; R = Rb-Sr whole rock; Rbt = Rb-Sr in biotite; Rmv = Rb-Sr in muscovite; S = Sm-Nd whole rock; Sgt = Sm-Nd garnet-whole rock isochrons.

Inset shows the known extent of metamorphism that occurred in the region during the *c.* 1000Ma and *c.* 500Ma events.

Shading of outcrops indicates the age of formation of unmetamorphosed lithologies and the protoliths of metamorphosed lithologies.

-  Archaean igneous and sedimentary lithologies & protoliths (4000-2500 Ma)
-  Mesoproterozoic igneous and sedimentary lithologies & protoliths ($\approx$ 1450-1150 Ma)
-  Neoproterozoic igneous and sedimentary lithologies & protoliths ($\approx$ 1000-600 Ma)
-  Late Neoproterozoic to early Cambrian felsic pegmatites and granitoids ($\approx$ 600-450 Ma)
-  Permian sedimentary lithologies (*c.* 270 Ma)



syn-tectonic intrusions (dated using U-Pb in zircon; Black *et al.*, 1991). Archaean orthogneisses with ages from *c.* 3270 to *c.* 2800 Ma occur in the adjacent Rauer Group (Kinny *et al.*, 1993). The orthogneisses were metamorphosed and intruded by felsic and mafic bodies during the Proterozoic (Sheraton *et al.*, 1984; Harley, 1987; Kinny *et al.*, 1993; Lanyon *et al.*, 1993). Differences in geochemical and age data between the Archaean gneisses in the Rauer Group and those in the Vestfold Hills have been interpreted by Snape & Harley (1996) as evidence for the presence of two unrelated terranes, that amalgamated during the Proterozoic.

Evidence for the presence of Archaean crust in the Rayner Complex is scarce. In the Stillwell Hills (Fig. 1.3), a Rb-Sr whole rock age of 2692 ± 48 Ma was derived from charnockitic orthogneiss by Clarke (1987). Clarke interpreted this age as that of the granulite facies recrystallization of the orthogneiss. Zircon separates from a felsic orthogneiss in the same area yielded a discordant population of U-Pb ages, with intercepts of 3081 ± 170 Ma and 820 ± 480 Ma (Grew *et al.*, 1988). Since Grew *et al.* (1988) consider this zircon to be xenocrystic to the orthogneiss, the lower intercept is interpreted as the age of crystallization of the orthogneiss, and the upper intercept as deriving from Archaean material incorporated into the intrusion.

Various gneisses in the Rayner Complex have provided Archaean ages from xenocrystic zircon. At Mawson Base, cores in zircon grains from a felsic orthogneiss yield U-Pb ages between 2470 and 1013 Ma (Young & Black, 1991). The *c.* 2470 Ma core analysis is significantly older than other zircon cores, and may have been inherited from source rocks that melted to form the orthogneiss. Detrital zircons in a felsic paragneiss from the northern Prince Charles Mountains range in U-Pb age from 2800 - 2500 and 2100 - 1800 Ma (Kinny *et al.*, 1997). Zircons with discordant ages in a felsic paragneiss from the Nye Mountains had an upper intercept of *c.* 2200 Ma (Black *et al.* 1987). These ages indicate the reworking of late Archaean and Paleoproterozoic crust in the formation of Rayner Complex.

Paleoproterozoic data

The formation of continental crust in the Paleoproterozoic is indicated by Sm-Nd (T_{CHUR}) model ages of 2000-1760 Ma for three granitic orthogneisses from the Nye Mountains and Thala Hills (Black *et al.*, 1987). These gneisses have U-Pb zircon ages of *c.* 1450 Ma, which are interpreted as the crystallization ages of granitic magmas derived from the melting of Paleoproterozoic crust (Black *et al.*, 1987). A similar Sm-Nd (T_{CHUR}) model age of 2020 Ma was derived for a sample of Mawson Charnockite from Mawson Base (Black *et al.*, 1987). Inherited zircon in orthogneiss samples from the northern Prince Charles Mountains have U-Pb ages of *c.* 1900 Ma (Kinny *et al.*, 1997). However, no samples of unequivocally Paleoproterozoic age have been found in the Rayner Complex.

Mesoproterozoic data

c. 1450 Ma ages derived from Enderby Land provide an age of intrusion for much of the granitic orthogneiss present there (Black *et al.*, 1987). Lithologies of this age are not found elsewhere in the Rayner Complex. A series of tholeiitic dykes that intrude the Napier Complex have been dated at 1190 ± 200 Ma (Amundsen Type 1 dykes of Sheraton & Black, 1981). In the Rayner Complex, deformed and metamorphosed dykes and mafic orthogneisses in Kemp Land, that were intruded into the protoliths of paragneiss lithologies, have been correlated with the Amundsen dykes (Sheraton & Black, 1981; Clarke, 1988). A poorly defined Rb-Sr whole-rock age of 1256 ± 17 Ma was obtained for layered paragneiss samples from the Stillwell Hills by Clarke (1987). The Rb-Sr data were inferred by Clarke (1987) to represent partial resetting during *c.* 1000 Ma metamorphism. In the Framnes Mountains, Rb-Sr whole rock ages of 1153 ± 47 and 1254 ± 13 Ma were derived from felsic Painted Gneiss (Young & Black, 1991, recalculated from unpublished data by Paul Arriens). It is not recorded if the Painted Gneiss samples were of orthogneiss or paragneiss.

At Fisher Massif in the northern Prince Charles Mountains (Fig. 1.3), metavolcanics and felsic orthogneisses have been dated from U-Pb in zircon at *c.* 1300 Ma (Beliatsky

et al., 1994; Kinny *et al.*, 1997). This provides an age of both magmatic activity and surface exposure for this area. In the Rauer Group, Prydz Bay, a series of mafic dykes that intruded Archaean basement rocks were correlated by Sims *et al.* (1994) with dyke swarms in the Vestfold Hills that were emplaced from *c.* 2240 to 1240 Ma (Lanyon *et al.*, 1993). The deposition of sediments in the Rauer Group was inferred by Sims *et al.* (1994) to have occurred between *c.* 1240 and 1000 Ma. However, Snape & Harley (1996) have suggested that the Archaean *and* Proterozoic histories of the Rauer Group and the Vestfold Hills are unrelated. Other authors (Hensen & Zhou, 1995; Zhao *et al.*, 1992) have argued that sedimentation in the Prydz Bay area occurred during the Neoproterozoic (see below).

Early Neoproterozoic data

The concept that the Rayner Complex represents a single extensive mobile belt is underpinned by the correlation of geochronological data into a single episode of tectonism. A period of tectonic activity, termed the Rayner Structural Episode by Sandiford & Wilson (1984), involved the metamorphism and deformation of Archaean and Proterozoic crust, with syn-tectonic magmatism, at *c.* 1000 Ma. The extent and importance of this episode is indicated by the large number of early Neoproterozoic ages obtained for high-grade gneisses from Enderby Land, the Mawson Coast, the Prince Charles Mountains and the Prydz Bay region (Fig. 1.3).

In western Enderby Land, Archaean and Proterozoic lithologies were strongly deformed and metamorphosed under hydrous transitional granulite facies conditions. In the Casey Bay region of the Napier Complex (Fig. 1.3), monazites from a paragneiss yielded a U-Pb age of 1087 ± 29 Ma (Black *et al.*, 1987). In the Rayner Complex, granitic orthogneiss from the Thala Hills gave a Rb-Sr whole rock age of 1022 ± 62 Ma (Grew *et al.*, 1978, recalculated by Sheraton *et al.*, 1987). In the Nye Mountains, metamorphic zircons from samples of *c.* 1450 Ma felsic orthogneiss and paragneiss yielded U-Pb ages in the range 920-994 Ma (Black *et al.*, 1987).

Along the Mawson Coast, ages are available for intrusive rocks that were emplaced during the Rayner Structural Episode. An extensive intrusion of Mawson Charnockite, which outcrops along the Mawson Coast from Mawson Station to the Colbeck Archipelago, and in the Framnes Mountains, was foliated and folded during the late stages of the Rayner Structural Episode (D₃ of Clarke, 1988; D₄ of this study). Samples of Mawson Charnockite taken from the Mawson Coast were dated with U-Pb in zircon at 985 ± 29 Ma and 954 ± 12 Ma (Young & Black, 1991). U-Pb zircon dating of Mawson Charnockite in the Framnes Mountains yielded a minimum age of emplacement of 923 Ma (Black *et al.*, 1987). Rb-Sr whole-rock dating of the charnockite gave similar or slightly younger ages of 886 ± 48 , 959 ± 58 and 910 ± 18 Ma (Black *et al.*, 1987; Young & Black, 1991, recalculated from unpublished data by Paul Arriens). In the Stillwell Hills region, syn-tectonic felsic pegmatites were dated by U-Pb zircon at 940 ± 80 Ma (Grew *et al.*, 1988).

Metamorphic zircon in an enclave of layered orthogneiss in the Mawson Charnockite at Mawson station was dated at 921 ± 19 Ma (Young & Black, 1991). This age is similar to the younger Rb-Sr whole rock ages obtained from the Mawson Charnockite. Young & Black (1991) interpret the Rb-Sr ages as a product of isotopic resetting during metamorphism.

There are several plutons of charnockite in the northern Prince Charles Mountains, which were emplaced during the Rayner Structural Episode. The charnockites intruded granulite to amphibolite facies gneisses which possessed strong gneissosities (S₁ and S₃ of Fitzsimons & Thost, 1992). They were subsequently deformed by high-strain shear zones (D₆ of Fitzsimons & Thost, 1992). Several samples of charnockitic and granitic orthogneiss from the area yielded U-Pb zircon ages within the range 951-1001 Ma (Kinny *et al.*, 1997). Garnet-bearing leucogneiss, probably derived from the partial melting of local lithologies (Fitzsimons & Thost, 1992), truncates a charnockite pluton at Mt. McCarthy, and has been dated at 990 ± 30 Ma (U-Pb zircon; Kinny *et al.*, 1997). At the Jetty Peninsula, U-Pb in zircon separates provided

ages of 1000 ± 10 Ma for a pre-D₁ pyroxene orthogneiss and 940 ± 20 Ma for a pre-D₆ leucogneiss (Manton *et al.*, 1992). These data indicate that D₁-D₄ tectonism, magmatic activity and partial melting occurred during a relatively brief episode at *c.* 1000 Ma (Kinny *et al.*, 1997). This is supported by Rb-Sr whole-rock ages for several gneisses, which lie in the range 1030-870 Ma (Tingey, 1991). However, the age of D₅-D₆ tectonism has not been constrained. On the basis of structural similarities between the earlier and later tectonic events, Hand *et al.* (1994a) argue that both stages are part of the Rayner Structural Episode. Similarly, Kinny *et al.* (1997) argue that the absence of post-1000 Ma age data in the samples indicates that D₅-D₆ occurred shortly after D₁-D₄.

Most of the high-grade gneisses of the Prydz Bay region have been previously interpreted as being produced by metamorphism and magmatism during the Rayner Structural Episode (*e.g.* Harley, 1987; Dirks *et al.*, 1994; Harley & Fitzsimons, 1991). The timing of metamorphism was based on only one Rb-Sr whole-rock age of 1050 ± 110 Ma from the Rauer Group (Tingey 1981). Kinny *et al.* (1993) used U-Pb in zircon to date gneisses from the Rauer Group. Felsic orthogneisses, and leucosome in a mafic orthogneiss, yielded ages between 1080 and 954 Ma, which were interpreted by Kinny *et al.* (1993) as ages of emplacement and granulite facies metamorphism. However, other studies (*e.g.* Hensen & Zhou 1995, Zhao *et al.*, 1995) have suggested that metamorphism occurred at *c.* 500 Ma instead (see below).

Mid-late Neoproterozoic data

Evidence for a widespread episode of metamorphism and magmatic activity in the interval 600-450 Ma has been found in most parts of the Rayner Complex and the adjacent Archaean terranes (Fig. 1.3). Closure temperatures for Rb-Sr in biotite were last exceeded at *c.* 500 Ma for samples from Enderby Land, Kemp Land, the Prince Charles Mountains and the Prydz Bay region (Collerson & Sheraton, 1986; Black *et al.*, 1987; Tingey, 1991). In Enderby Land and Kemp Land, *c.* 500 Ma felsic pegmatites and shear zones associated with localised amphibolite facies

metamorphism are found on both sides of the boundary between the Rayner and Napier Complexes (Black *et al.*, 1987; Sheraton *et al.*, 1987). Monazite from a biotite-bearing granite from the Nye Mountains yielded a discordant U-Pb age population with intercepts at 766 ± 15 Ma and 528 ± 51 Ma, which were interpreted by Black *et al.* (1987) as crystallization and resetting ages, respectively.

A Rb-Sr biotite age of 718 ± 10 Ma was obtained by Clarke (1987), from an undeformed pegmatite from the Stillwell Hills. This date provides a lower constraint on the age of tectonism and metamorphism in this area. No dates, however, have been obtained from the Mawson Coast or the Framnes Mountains that unequivocally represent geological events that occurred after *c.* 900 Ma. On the basis of Sm-Nd garnet-whole rock isochrons derived for paragneisses from the northern Prince Charles Mountains, Hensen *et al.* (1998) suggest that granulite facies metamorphism also occurred in the periods 825-809 Ma and 630-550 Ma.

In the southern Prince Charles Mountains, a calcareous sedimentary sequence unconformably overlies Archaean gneisses (Grikurov & Soloviev, 1974). A granitic clast from conglomerate in the sequence was dated at *c.* 800 Ma (Rb-Sr, Halpern & Grikurov, 1975) and the whole sequence was deformed and metamorphosed to greenschist facies, probably during a *c.* 500 Ma event (Tingey, 1991).

In the northern Prince Charles Mountains, felsic pegmatites and granitic stocks intruded into Proterozoic crust at about 500 Ma (Manton *et al.*, 1992; Tingey, 1991). In the Jetty Peninsula area, the intrusion of felsic dykes at *c.* 500 Ma (Manton *et al.*, 1992) was accompanied by partial melting of host lithologies, implying that the dykes intruded under upper amphibolite facies conditions (M_3 of Hand *et al.*, 1994a). Later shear zones, that produced minor retrogression of gneisses under lower amphibolite facies conditions, are attributed by Hand *et al.* (1994a) to the waning stages of the *c.* 500 Ma event. Brittle shear zones, with attendant greenschist facies metamorphism, are found throughout the northern Prince Charles Mountains, and have been

attributed to a *c.* 500 Ma event (Fitzsimons & Thost, 1992). The dating of garnet in granulite facies gneisses by Zhou & Hensen (1995) and Hensen *et al.* (1998) suggest that this event had a more pervasive effect on the northern Prince Charles Mountains, and at significantly higher grade conditions than previously recognised (Thost, 1996).

Archaean amphibolites and Neoproterozoic sediments in the southern Prince Charles Mountains were intruded by granitoid dykes, deformed and metamorphosed under greenschist facies conditions (Tingey, 1991). Dykes from the area have been dated at 551 ± 74 Ma (Rb-Sr, Tingey, 1991).

In the Prydz Bay region, the true extent of *c.* 500 Ma metamorphism is only beginning to be recognised. Along the coastline, mafic to felsic gneisses of indeterminate age form a basement to supercrustal units, which were metamorphosed at granulite facies and intruded by syn-tectonic granitoids (Stüwe *et al.*, 1989; Zhao *et al.*, 1992; Fitzsimons & Harley, 1991; Carson *et al.*, 1996). Kinny *et al.* (1993) obtained U-Pb zircon ages of around 1000 Ma for the syn-tectonic granitoids. Granitoids and pegmatites that intruded subsequently were dated using U-Pb in zircon, at 516 ± 7 Ma (Carson *et al.*, 1996), 500 ± 12 Ma (Kinny *et al.*, 1993) and 563–523 Ma (Zhao *et al.*, 1992). Most of the samples dated by Kinny *et al.* (1993) also yielded *c.* 500 Ma U-Pb analyses, as well as discordant data that lay along chords with lower intercepts at *c.* 500 Ma.

A number of interpretations have been made from these data. In the Larsemann Hills, Zhao *et al.* (1992, 1995) have suggested that sediments were deposited on a *c.* 1000 Ma granulite facies basement, and that these lithologies were intruded and metamorphosed to granulite facies at *c.* 500 Ma. This model is supported by Hensen & Zhou (1995), who obtained Sm-Nd garnet whole-rock isochrons with ages of 1000–898 Ma for ‘basement’ mafic granulite and 521–503 Ma for paragneiss and leucogneiss samples. Hensen & Zhou (1995) also suggest that the *c.* 1000 Ma data obtained by Kinny *et al.* (1993) represent zircon inherited from basement gneisses,

and that the host gneisses were intruded and metamorphosed at *c.* 500 Ma. An alternate interpretation is proffered by Carson *et al.* (1996), who suggest that the Rauer Group orthogneisses dated by Kinny *et al.* (1993) were intruded at *c.* 1000 Ma and metamorphosed at *c.* 500 Ma. In either case, granulite facies metamorphism and magmatic activity occurred at Prydz Bay and in parts of the northern Prince Charles Mountains at *c.* 500 Ma. Elsewhere in the Rayner Complex, a high degree of tectonic activity at *c.* 500 Ma has not been identified unambiguously.

1.4 A general history of the Rayner Complex

With the geochronological data in mind, a generalised history of the Rayner Complex can be described. This provides a starting point from which the complexities of local geology and geological events can be addressed. A detailed correlation of geological events across the Rayner Complex is discussed in the Synthesis (Chapter 6).

Gneisses in the Rayner Complex had a complex history before the Rayner Structural Episode. Protoliths consisted of a mixture of reworked Archaean to Proterozoic crust and juvenile Proterozoic crust (*e.g.* Black *et al.*, 1987; Tingey, 1991). Sedimentary and volcanic supracrustal protoliths were deposited at various stages during the Paleoproterozoic and Mesoproterozoic. Igneous activity, at surface and sub-surface levels, occurred at various times and localities during the Paleoproterozoic and Mesoproterozoic. There is considerable variation between the pre-Rayner Structural Episode histories of different localities in the Rayner Complex.

All of these lithologies were affected by the Rayner Structural Episode of Sandiford & Wilson (1984), which occurred between the mid-Mesoproterozoic and the mid-Neoproterozoic. The Rayner Structural Episode was associated with multiple stages of deformation, granulite to upper amphibolite facies metamorphism and syn-tectonic

intrusive activity (*e.g.* Sheraton *et al.*, 1987; Harley & Hensen, 1990). The largest intrusion emplaced during the Rayner Structural Episode was the Mawson Charnockite, which was intruded at the same time as charnockitic bodies in the northern Prince Charles Mountains. Ultramylonite-pseudotachylite shear zones were developed after plastic deformation in all parts of the Rayner Complex.

Chapter 2

Geology of the Cape Bruce area

2. Geology of the Cape Bruce area

Introduction

Cape Bruce is a 2 by 4km area of rock exposure on the Mawson Coast, immediately west of the emergent tongue of the Taylor Glacier (Fig. 1.2). Geological mapping was conducted in the area by the Australian Bureau of Mineral Resources in 1961. Trail (1970) divided the gneisses at Cape Bruce into two lithological types, the Colbeck Gneiss and the Stillwell Gneiss, both of which have a regional distribution (Fig.1.2).

In this study, new field work was conducted, which separated the gneisses into eight distinct lithologies. The gneisses record a complex geological history, marked by five deformation episodes and at least seven magmatic events. The Colbeck and Stillwell Gneisses, as defined by Trail (1970), both include lithologies that are genetically and temporally disparate. The *proclamation gneiss*, *charnockitic orthogneiss* and the *white granitic orthogneiss*, as defined below, were included in the Stillwell Gneiss by Trail (1970). The *metapelitic paragneiss*, *felsic paragneiss*, *biotite-pyroxene orthogneiss* and *red granitic orthogneiss* were included in the Colbeck Gneiss (*N.B.* All informal names for lithological units are italicised).

This chapter describes the lithologies and the structural relationships mapped in the present study. A geological history is proposed for the evolution of the gneisses at Cape Bruce, and this history is compared to the interpretations made by previous studies on other geological outcrops along the Mawson Coast.

2.1 Lithologies

Proclamation gneiss

Proclamation gneiss is distinguished as a group of felsic, metapelitic, calc-silicate and mafic gneiss units, that occur as a sequence of repeated layers 0.5-10m thick (Fig. 2.1, Plate 2.1a). The lithology was originally named by White & Clarke (1993), who correlated the gneisses around Proclamation Point (Fig. 2.1) with *proclamation gneiss* in the Colbeck Archipelago. *Proclamation gneiss* consists mostly of a single layer that is at least 100m thick, extending from Proclamation Point to low hills in the middle of the Cape Bruce area. A small outcrop of *proclamation gneiss* also occurs in the mid-eastern part of the Cape Bruce area, adjacent to the Taylor Glacier (Fig. 2.1). The gneiss is composed of high-grade minerals that define a strong S_1 gneissosity (Plate 2.1b). Most *proclamation gneiss* is made up of pale grey/violet felsic layers 1-5m thick, which are composed of quartz, K-feldspar, plagioclase and garnet. Most minerals, including garnet, occur as granoblastic grains less than 1mm across. Small amounts of biotite and magnetite are also present in the felsic gneiss units. In the south-most outcrops of *proclamation gneiss*, grey felsic gneiss grades into 1m thick layers of white quartzite.

About 20% of the outcrop of *proclamation gneiss* is composed of thin (1-80cm) layers of biotite-rich felsic and metapelitic gneiss. Both types possess an intense foliation defined by biotite grains. Biotite-rich felsic gneiss also contains abundant amounts of quartz, plagioclase, K-feldspar, and garnet porphyroblasts up to 1cm across. Metapelitic gneiss contains fine-grained quartz, cordierite, garnet, plagioclase and K-feldspar, and a pervasive S_2 foliation and lineation defined by sillimanite and biotite grains.

Massive leucosome is composed of random grains of quartz, plagioclase and K-feldspar that form layers up to 0.7m wide. Altogether, these form approximately 5-10% of the outcrop of *proclamation gneiss* (Plate 2.1b). Leucosome also occurs as

Fig. 2.1. Lithological map of the Cape Bruce locality. The markers N-S and W-E indicate the locations of the vertical cross sections in Fig. 2.4.

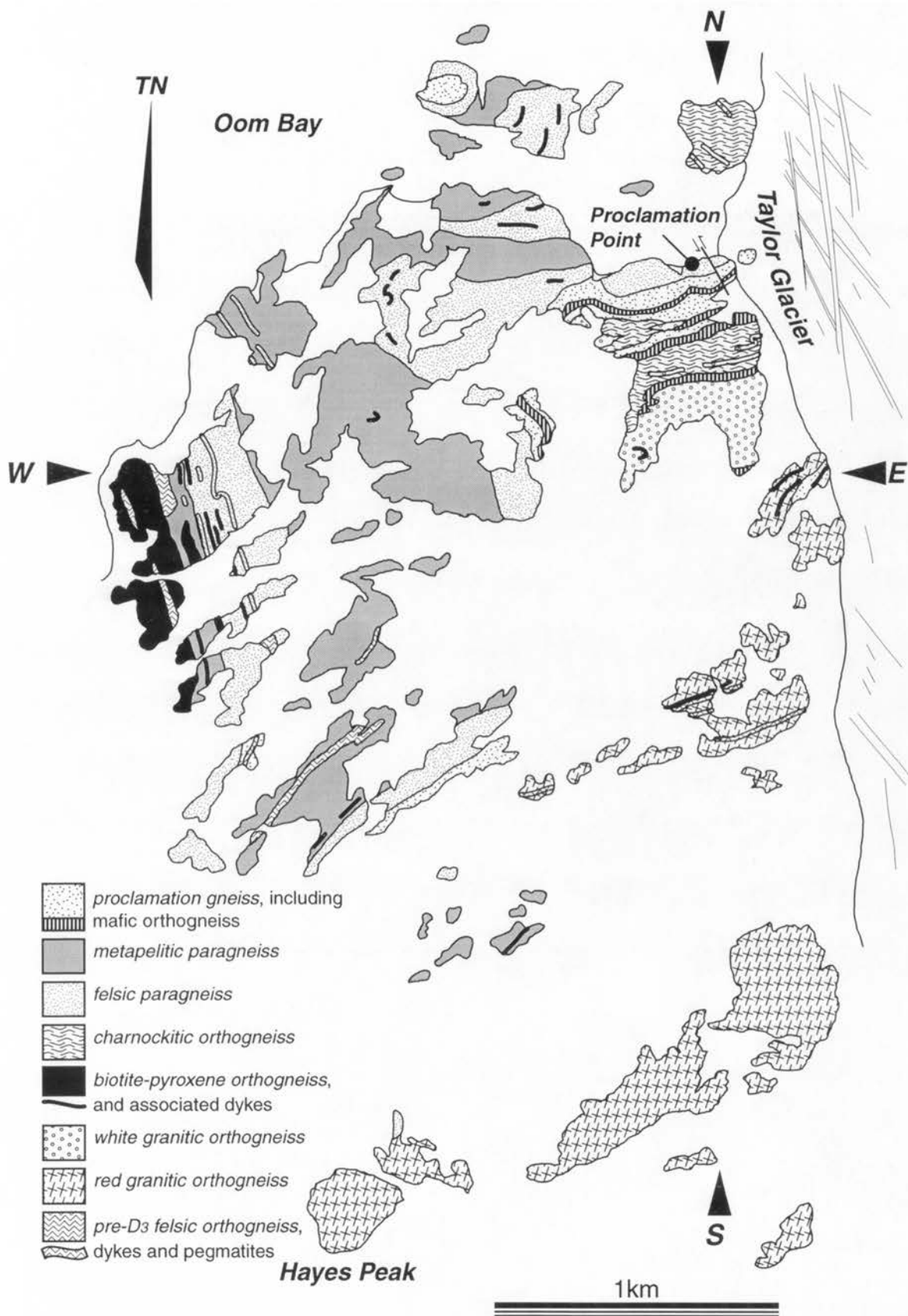


Plate 2.1

a) The Proclamation Point area, viewed from a ridge 300m to the south-south-east. The labelled hill in the right half of the view is about 35m high. The lithologies visible include *proclamation gneiss* (p) that includes dark-coloured layers of mafic orthogneiss (m); *charnockitic orthogneiss* (c), and *white granitic orthogneiss* (w). A person in the foreground is arrowed for scale.

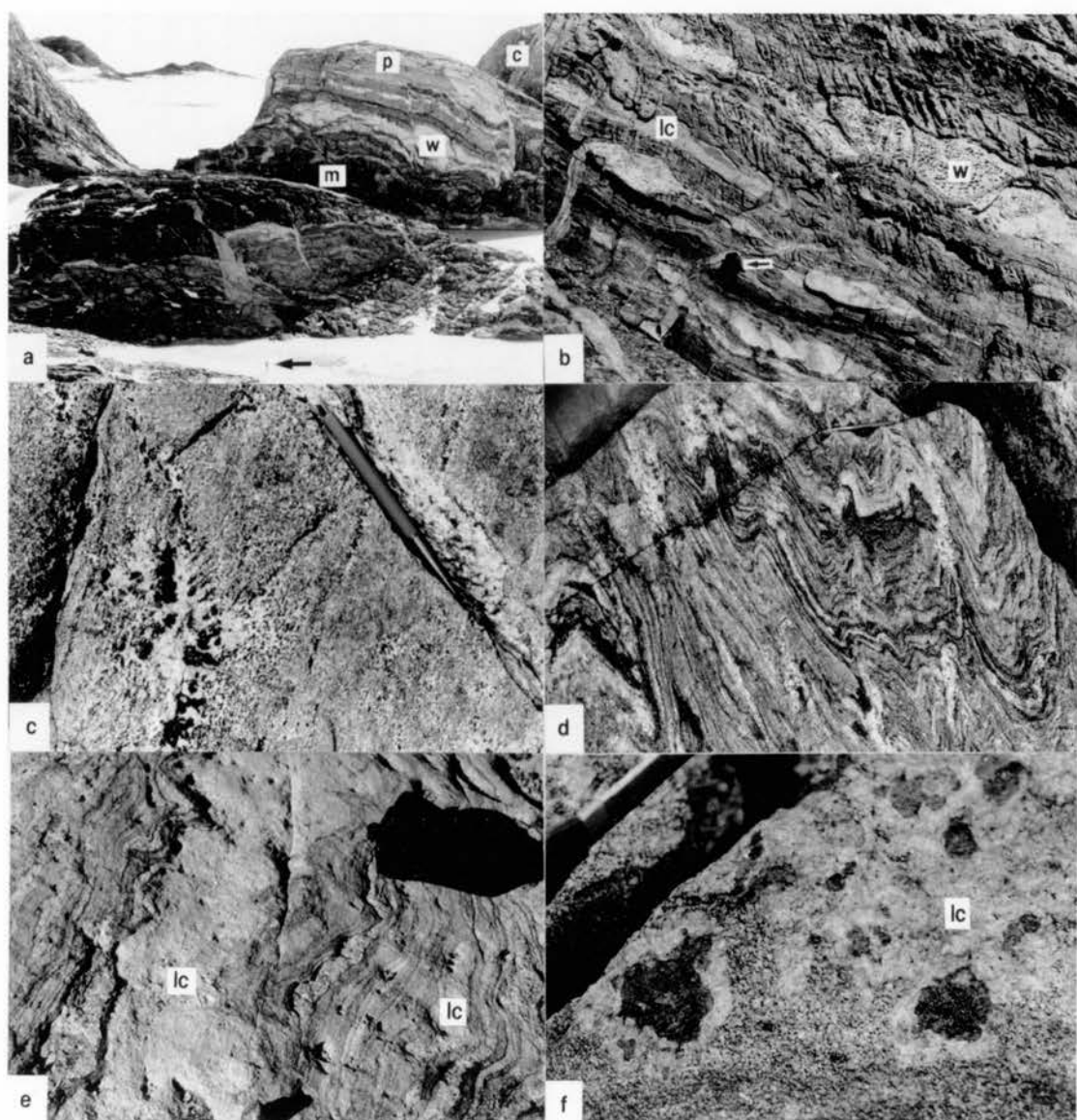
b) Hillside exposure of *proclamation gneiss*, that is also visible in the upper left-hand corner of Plate 2.1a. The *proclamation gneiss* includes layers of quartzofeldspathic leucosome (lc). Sheets of massive *white granitic orthogneiss* (w) occur as layers of irregular thickness and as large boudins. Person (arrowed) for scale; the width of the view is about 20m.

c) Mafic orthogneiss, with a well developed S_1 foliation and nebulitic patches of orthopyroxene + plagioclase leucosome that transgress the foliation. *Proclamation gneiss*. Width of view is 30cm.

d) *Metapelitic gneiss*, with a strong $S_{2/1}$ gneissosity that has been crenulated into F_3 folds. Width of view is 1m.

e) *Felsic paragneiss*, with a strong $S_{2/1}$ gneissosity defined by alternating layers of biotite + garnet-rich gneiss and quartzofeldspathic leucosome (lc). Width of view is 1.1m.

f) Nebulitic leucosome (lc) nucleating around porphyroblasts of garnet. *Felsic paragneiss*. Width of view is 25cm.



laminations 5-15mm wide interspersed through felsic and metapelitic gneiss layers, which are continuous over 0.3-20m. The boundaries of most leucosome layers are irregular and some transgress the S_1 gneissosity in adjacent gneiss at low angles. Many of the leucosome layers are broken in boudins that are wrapped around by an S_2 foliation in the adjacent gneiss.

Layers of mafic orthogneiss are present in the *proclamation gneiss* group, immediately to the south of Proclamation Point (Plate 2.1a, Fig. 2.1). They are mostly composed of orthopyroxene, clinopyroxene and plagioclase, with lesser amounts of K-feldspar, quartz and hornblende, and minor amounts of biotite, ilmenite, rutile and apatite. Mafic orthogneiss has a granoblastic S_1 gneissosity, transgressed by irregular patches of coarser grained plagioclase and orthopyroxene (Plate 2.1c). It occurs as layers up to 5m thick and 10 to 150m in length. The layers are broken in many places into podiform or angular boudins 5 to 10m across. Where layers of mafic orthogneiss lie adjacent to metapelitic gneiss or leucosome, they are separated from the latter by 1-10cm thick selvages rich in biotite and garnet. Garnet porphyroblasts also occur in mafic orthogneiss near contacts with metapelitic gneiss. Samples of garnet-bearing mafic orthogneiss are described in Chapter 3 (Metamorphism).

Layers of felsic orthogneiss make up approximately 2% of the total outcrop of *proclamation gneiss*. They contain plagioclase, orthopyroxene, quartz and biotite, with or without clinopyroxene, K-feldspar and ilmenite. The layers are fine-grained and lie parallel to the dominant S_1 gneissosity.

Outcrops of calc-silicate gneiss occur around and 800m to the south-east of Proclamation Point. They occur as pods 0.1-2m across that are distributed along strike. Calc-silicate gneiss is fine-grained (0.2-1mm), granoblastic, and pale pink or green in colour. It is composed mostly of plagioclase, quartz, diopside and sphene, with lesser amounts of calcite and scapolite. Grossular is present in pink-coloured samples, varying in modal proportion from 0 to 15%.

The pre-tectonic relationship of the units in *proclamation gneiss* is obscured by the intensity of S_1 gneissosity. Inferences on the nature of protoliths can be made from the composition of the gneiss units. The presence of metapelitic and calc-silicate gneiss is consistent with sedimentary protoliths for most of the *proclamation gneiss*. The mafic and felsic orthogneisses may represent recrystallized dykes or sills, or may have been extrusive rocks.

Metapelitic gneiss

Approximately 25% of the outcrop at Cape Bruce is made up of *metapelitic gneiss* (Fig. 2.1), interlayered on a metre to tens of metres scale with garnet-bearing *felsic paragneiss*. *Metapelitic gneiss* is mostly composed of K-feldspar, quartz, biotite, garnet and sillimanite, with lesser amounts of plagioclase, ilmenite, magnetite, monazite, apatite and zircon. Cordierite and spinel form up to 20 modal % of *metapelitic gneiss* in a few localities. A strong S_1 gneissosity is defined by laminations 3 to 50mm wide, that are alternatively rich in biotite, garnet, sillimanite or feldspar (Plate 2.1d). Biotite, sillimanite and some garnet grains define an S_2 foliation that mostly lies parallel to the gneissosity. Garnet occurs as equant or oblate grains 5-10mm across in sillimanite-rich layers, or as irregular poikiloblasts up to 40mm across in quartz-bearing layers.

Stromatic and nebulitic leucosome may form up to 20% of the volume of *metapelitic gneiss*. Leucosome is massive and contains porphyroblasts of K-feldspar in a matrix of K-feldspar, quartz and plagioclase, with or without cordierite. Garnet and biotite are common on the contacts of leucosome with metapelitic layers.

The petrography of samples of *metapelitic gneiss* is described in detail in Chapter 3 (Metamorphism).

Felsic paragneiss

This lithology is distinguished from other felsic units on the basis of being garnet-bearing and having gradational boundaries with the *metapelitic gneiss*. *Felsic paragneiss* consists of quartz, K-feldspar, garnet, biotite, and lesser amounts of plagioclase,

ilmenite and magnetite. Layers 1-10cm thick are distinguished by differing modal proportions of biotite and garnet (Plate 2.1e). Small amounts of *metapelitic gneiss* rich in sillimanite, biotite and garnet occur as discontinuous laminations 3-15mm wide. The gneiss has a granoblastic fabric with a strongly to poorly defined S_1 gneissosity, and an S_2 foliation defined by biotite grains. Garnet occurs as poikiloblasts 10 to 40mm across, that are commonly flattened parallel to S_1 and S_2 .

Leucosome is abundant in *felsic paragneiss*, occurring as laminations parallel to the S_1 gneissosity, or as nebulitic patches and veinlets. It is composed of megacrystic K-feldspar grains set in a matrix of K-feldspar, quartz and plagioclase. Nebulitic leucosome is commonly found enclosing garnet porphyroblasts (Fig. 2.1f). Large layers of leucosome are often difficult to distinguish from the host felsic gneiss where the two are of a similar composition.

The aluminous composition of *felsic paragneiss*, and the intimate relationship it shows with *metapelitic gneiss*, are consistent with a sedimentary protolith.

Charnockitic orthogneiss

This lithology outcrops as a single monotonous unit in the north-eastern corner of Cape Bruce, to the north and south of Proclamation Point (Plate 2.1a, Fig. 2.1). It is an orthopyroxene-bearing felsic gneiss of charnockitic to enderbitic composition. Quartz, plagioclase, orthopyroxene and K-feldspar are present, with lesser amounts of clinopyroxene, biotite, and ilmenite. *Charnockitic orthogneiss* contains a strong S_1 gneissosity and a granoblastic fabric (1-3mm grain size). Layers 0.1 to 5m thick are distinguished by differing modal proportions of orthopyroxene, clinopyroxene and biotite. Layers containing distinct clusters of orthopyroxene up to 10mm in diameter are interleaved with fine-grained *charnockitic orthogneiss* (Plate 2.2a).

On the large island north of Proclamation Point, *charnockitic orthogneiss* contains 0.3 to 5m wide pods of ultramafic orthogneiss (Plate 2.2b). These pods occur in trains along the strike of S_1 gneissosity in the host gneiss. The pods consist of granoblastic

Plate 2.2

a) *Charnockitic orthogneiss*. Layers with a strong $S_{2/1}$ gneissosity alternate with layers containing clusters, 0.5 to 1cm in diameter, composed of fine-grained orthopyroxene. Width of view is 35cm.

b) Boudins of dark green ultramafic orthogneiss in *charnockitic orthogneiss*. Width of view in foreground is about 5m.

c) A dyke of *biotite-pyroxene orthogneiss* that intruded into *felsic paragneiss*. The edges of the dyke are rich in biotite. The dyke contains a weak S_2 foliation and has been deformed into an F_4 fold. Mitten for scale; width of view is about 90cm.

d) *Biotite-pyroxene orthogneiss*, containing an S_2 foliation defined by thin sheets of leucosome. The S_2 foliation is deformed into F_3 folds, with sheets of leucosome lying parallel to the axial planes of F_3 folds. Notebook for scale (arrowed); width of view is 1.5m.



grains of magnesian orthopyroxene, chromian clinopyroxene, and lesser amounts of plagioclase and phlogopite. Most of the pods are mantled by 5-10mm thick rims of orthopyroxene.

Charnockitic to enderbitic layers occur within *proclamation gneiss* immediately south of Proclamation Point (Plate 2.1a). These are probably *charnockitic orthogneiss* that have been tectonically interleaved with *proclamation gneiss*. Patches of *felsic paragneiss* up to 3m across are found in *charnockitic orthogneiss*. Parts of the host charnockite adjacent to these patches contain garnet and biotite along with orthopyroxene. The petrography of garnet-bearing *charnockitic orthogneiss* is discussed in Chapter 3 (Metamorphism).

Pre-tectonic textures within *charnockitic orthogneiss* and contacts of the unit with other lithologies have been obliterated by the pervasive S₁ gneissosity. The pyroxene-rich composition of *charnockitic orthogneiss*, along with the presence of ultramafic pods and the monotonous nature of the unit, is consistent with an igneous protolith for the unit. Patches of felsic paragneiss in *charnockitic orthogneiss* may have been entrained when the charnockite intruded into adjacent sedimentary or metasedimentary lithologies.

Biotite-pyroxene orthogneiss

The S₁ gneissosity in *metapelitic* and *felsic paragneisses* is truncated by biotite and pyroxene-bearing dykes (Plate 2.2c). These dykes are particularly common near the western edge of the Cape Bruce area, where they lie near a large body of dark brown gneiss of similar composition. The dykes and gneiss body are grouped together in this study as *biotite-pyroxene orthogneiss* (Fig. 2.1). The main body of the orthogneiss contains a strong S₂ gneissosity (Plate 2.2d), and has a dark colour imparted by the high proportions of biotite (7-20 modal %) and ilmenite (2-5 modal %) in the rock. The rest of the gneiss is composed of 1-2mm wide grains of plagioclase and K-feldspar in a matrix of finer-grained (0.3-0.6mm) quartz, with minor amounts of

monazite (up to 1 modal %), apatite and zircon. Orthopyroxene is present in small amounts (less than 2 modal %), as grains 0.5-1mm across.

Dykes of *biotite-pyroxene orthogneiss* are typically 0.5-1m across and up to 200m in length. Many dykes have been dismembered into trains of fragments 0.5 to 10m long. They are lighter coloured and finer grained (0.3-1mm) than the large body, but have a similar mineralogy, containing biotite, orthopyroxene, monazite, ilmenite and apatite. The dykes have a pervasive anastomosing S_2 foliation defined by biotite, enveloping augens of quartz and feldspar. Where the dykes are hosted by *metapelitic* or *felsic paragneiss*, they have selvages of biotite and garnet-rich gneiss (Plate 2.2c).

Leucosome composed of K-feldspar, plagioclase and quartz, is common in *biotite-pyroxene orthogneiss*. In the main body of the orthogneiss, laminations of leucosome are found parallel to the S_2 foliation and in the axial planes of F_3 folds (Plate 2.2d). Laminations parallel to S_2 are found in the dykes.

White granitic orthogneiss

White granitic orthogneiss intrudes *charnockitic orthogneiss* and *proclamation gneiss* in the north-eastern quarter of Cape Bruce (Fig. 2.1). The largest outcropping body truncates the S_1 gneissosity in *charnockitic orthogneiss* (Plate 2.3a). The granite is leucocratic and massive, with an inequigranular fabric and a weak S_2 foliation defined by flattened quartz and feldspar grains. K-feldspar (antiperthitic orthoclase) grains 1-5mm across are set in a matrix of quartz and plagioclase grains less than 0.5mm across. Garnet occurs in small amounts (less than 3 modal %) as round inclusion-free grains 1-5mm across. Biotite is sometimes present in minor amounts (less than 1 modal %). Opaque minerals (magnetite, ilmenite) are only present in accessory amounts, along with monazite and zircon.

White granitic dykes occur in *charnockitic orthogneiss* (Plate 2.3b) and in *proclamation gneiss* around Proclamation Point (Plate 2.1b). The dykes truncate the S_1 gneissosity in the host gneisses at a low angle, and are attenuated parallel to the S_2 foliation in the

Plate 2.3

- a) *White granitic orthogneiss* that has been intruded into *charnockitic orthogneiss*, transgressing the S_1 gneissosity in the latter. Mitten for scale; width of view in the foreground is about 35cm.
- b) View of the north face of the island 400m north of Proclamation Point (see Fig 2.1). Sheets of *white granitic orthogneiss* have been intruded into dark grey *charnockitic orthogneiss* and are transposed into an S_2 foliation. Person on bike (arrowed) for scale; width of view in the foreground is about 40m.
- c) *Proclamation gneiss* possessing a strong S_1 gneissosity has been intruded by *red granitic orthogneiss* (rgo). Although *red granitic orthogneiss* truncates the S_1 gneissosity, a weak S_2 foliation in *red granitic orthogneiss* lies sub-parallel to the S_1 gneissosity. Width of view is about 1m.
- d) An irregular network of felsic pegmatite that has been intruded into *red granitic orthogneiss*. Camera case for scale; width of view in the foreground is about 1.7m.
- e) A dyke of felsic orthogneiss truncates $S_{2/1}$ gneissosity in *felsic paragneiss*. Compass for scale; width of view is about 1m.
- f) Pre- D_4 felsic pegmatite that has been intruded into *metapelitic gneiss*, truncating the $S_{2/1}$ gneissosity in the latter. Width of view is 80cm.



gneisses. The dykes are up to 3m thick and extend for tens to hundreds of metres. Dykes hosted by *proclamation gneiss* are commonly dismembered into chains of boudins (Plate 2.1b). The dykes only occur within 1km of the main body of *white granitic orthogneiss*, have the same mineralogy and appearance as the latter.

Red granitic orthogneiss

This lithology is distinguished from others on the basis of a distinctive red colour when weathered and a massive coarse-grained fabric. It dominates exposures in south-east quarter of the Cape Bruce area (Fig. 2.1). The orthogneiss is composed mostly (50-60 modal %) of grains of antiperthitic K-feldspar 2-15mm across, set in a matrix of quartz, plagioclase, and biotite. Garnet is uncommonly present as irregular porphyroblasts 1-10mm across. Monazite, ilmenite, magnetite and zircon are present in minor amounts. *Red granitic orthogneiss* may be distinguished from *white granitic orthogneiss* on the basis of a higher mode of K-feldspar and biotite and a lesser amount of quartz.

Red granitic orthogneiss truncates S_1 gneissosity in a small outcrop of felsic *proclamation gneiss* that lies 1km south-south-east of Proclamation Point (Plate 2.3c). An S_2 foliation is present in the *red granitic orthogneiss*, defined by biotite grains, flattened quartz-rich segregations and mafic schlieren. Isolated remnants of biotite-rich *felsic paragneiss* and biotite-orthopyroxene-bearing dykes are present enclosed within *red granitic orthogneiss*.

Bodies of *biotite-pyroxene*, *red granitic* and *white granitic orthogneisses* truncate S_1 gneissosity in *proclamation gneiss*, *felsic paragneiss*, *metapelitic paragneiss* and *charnockitic orthogneiss*. The composition and transgressive contacts of the orthogneisses is consistent with an intrusive origin. None of these three orthogneisses, however, have mutual contacts, and the relative timing of their emplacement is ambiguous.

Felsic dykes and pegmatites

Two suites of felsic dykes transgress S_1 and S_2 foliations in all of the lithologies described above. A suite of pale pink felsic dykes occur in the middle of the Cape Bruce area (Fig. 2.1). Individual dykes are continuous over distances of 5 to 800m, and have widths between 0.3m and 2m. In places, the dykes develop web-like offshoots of felsic pegmatite (Plate 2.3d). Near the western edge of Cape Bruce, a large mass of pink felsic orthogneiss approximately 50m across intrudes *biotite-pyroxene orthogneiss* and *felsic paragneiss* (Fig. 2.1). Dykes of a similar composition intrude *red granitic orthogneiss* and are mostly oriented sub-parallel to the S_2 foliation in the latter. The felsic dykes are massive and composed of antiperthitic K-feldspar, quartz and plagioclase, with minor amounts of biotite, garnet, magnetite and zircon. Unaligned porphyroblasts of K-feldspar are common. Biotite grains define a weak S_3 foliation. Unlike the leucosome present in most of the earlier lithologies, contacts between the felsic dykes and the host rocks tend to be sharp and do not show evidence of migmatization or rehydration (Plate 2.3e).

A separate generation of white felsic dykes and pegmatites transgress all previously described gneisses and dykes (Plate 2.3f). They are typically 0.3-1m wide and extend for distances of 10-100m. The dykes are for the most part unfolded, sub-vertical and east-trending. They contain antiperthitic K-feldspar grains 2-30mm across, in a matrix of quartz, plagioclase, and minor amounts of biotite, ilmenite and magnetite. Garnet occurs in minor amounts (less than 1 modal %), as porphyroblasts up to 20mm across. Where these felsic dykes have intruded metapelitic and biotite-rich gneisses, the host rock near the contact commonly contains leucosome. The dykes have similar compositions and structural relationships to pegmatite dykes in the Colbeck Archipelago, which were interpreted by White & Clarke (1993) as offshoots from the Mawson Charnockite intrusion at Chapman Ridge.

2.2 Structural relationships

Tectonic processes that produced the high-grade gneisses of Cape Bruce can be divided into a series of events, distinguished by structural styles and orientations. Structural relationships are shown in the map in Fig. 2.2; structural data associated with each event is shown in Fig. 2.3.

D₁ event

The earliest tectonic event recognised at Cape Bruce produced a penetrative gneissosity in *proclamation gneiss*, *metapelitic gneiss*, *felsic paragneiss* and *charnockitic orthogneiss* (Plates 2.1a & b). S_1 is a gneissosity defined by layers of differing compositions that are repeated across strike. Quartz, feldspar and pyroxene grains are flattened into an S_1 foliation. The majority of leucosome found in the metasedimentary gneisses has also been flattened parallel to S_1 . F_1 folds are rare, occurring as rootless or intrafolial isoclinal folds that deform veinlets of leucosome (Plate 2.4a).

The original orientation of structures produced by D_1 cannot be determined, due to the pervasive effects of later events. In most places the S_1 gneissosity lies parallel to an S_2 gneissosity and foliation, and both S_1 and S_2 have been reoriented by folding during D_3 and D_4 (Fig. 2.3a). No lineations were found that could be unambiguously identified as L_1 . Kinematic indicators were also absent.

D₂ event

The D_2 event produced an intense gneissosity and mineral foliation in all lithologies, except for post- D_2 felsic dykes and pegmatites. S_2 is the earliest recognised foliation in *biotite-pyroxene*, *white granitic*, and *red granitic orthogneisses*. The foliation is defined by flattened and aligned mineral grains, and is particularly well developed in biotite and sillimanite-rich gneisses (Plate 2.4b). Leucosome in *biotite-pyroxene orthogneiss* and *red granitic orthogneiss* is deformed into S_2 -parallel laminations (Plate 2.2d) and F_2 folds (Plate 2.4c). In metasedimentary gneisses and *charnockitic*

Fig. 2.2. Structural map of the Cape Bruce locality. The markers N-S and W-E indicate the locations of the vertical cross sections in Fig. 2.4.

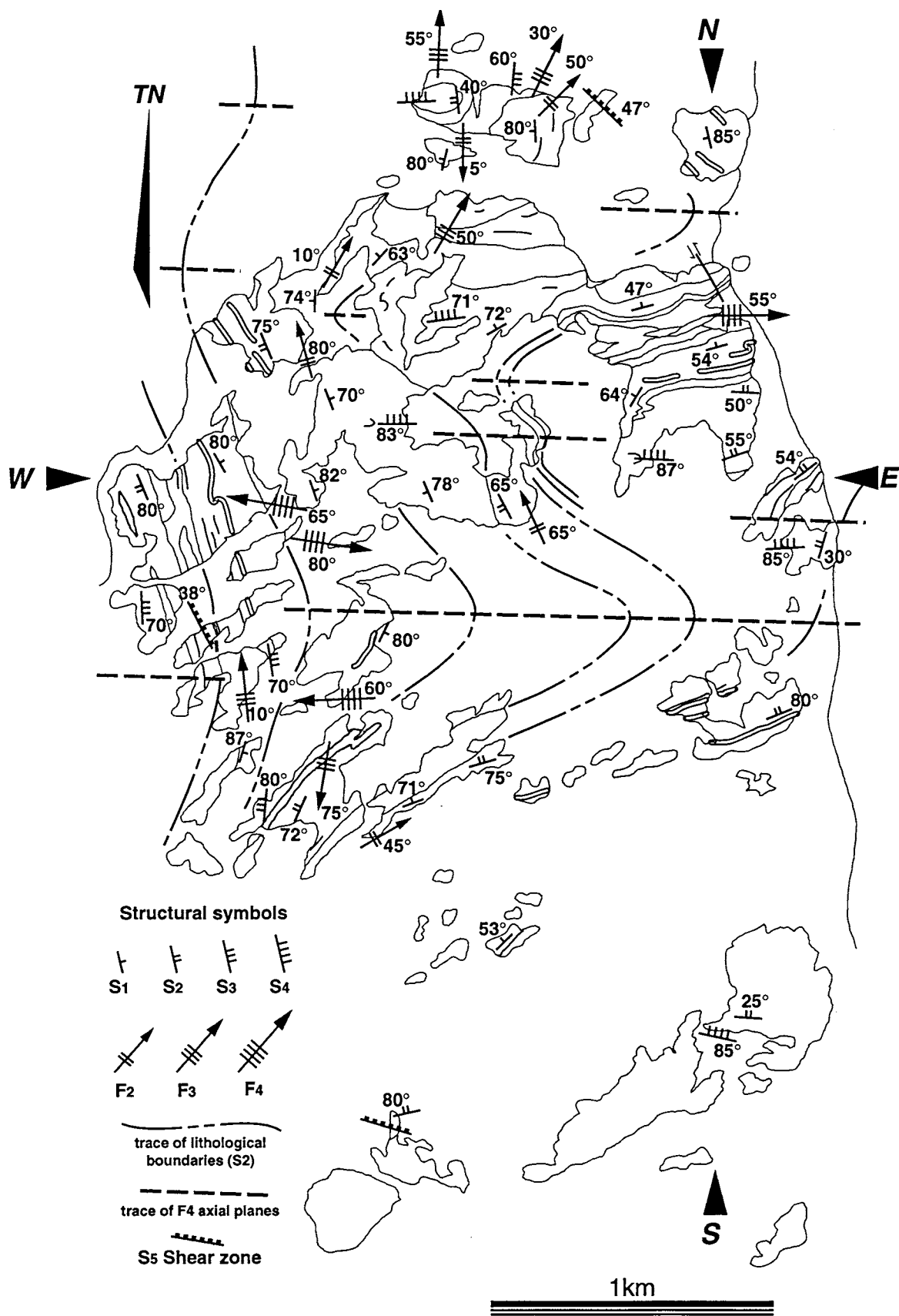
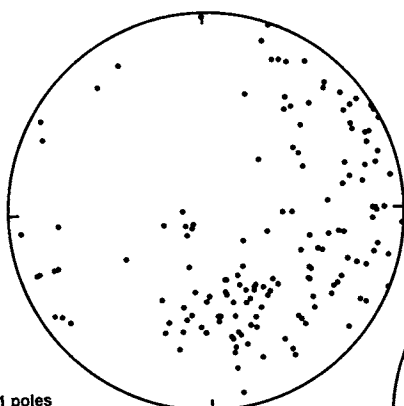


Fig. 2.3. Structural data for the Cape Bruce locality, plotted on Wulff nets. North is towards the top of the page.

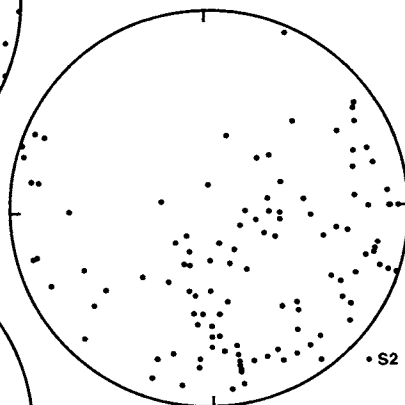
- a) 161 poles to S_1 planes.
- b) 109 poles to S_2 planes.
- c) 14 L_2 stretching and mineral lineations and 22 fold axes to F_2 folds.
- d) 22 poles to S_3 planes and 44 fold axes to F_3 folds.
- e) 23 poles to S_4 planes, 3 L_4 mineral lineations and 8 fold axes to F_4 folds.
- f) 11 poles to S_5 planes and 5 L_5 stretching lineations.

a



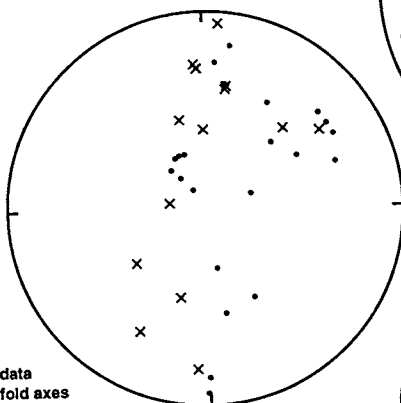
• S1 161 poles

b



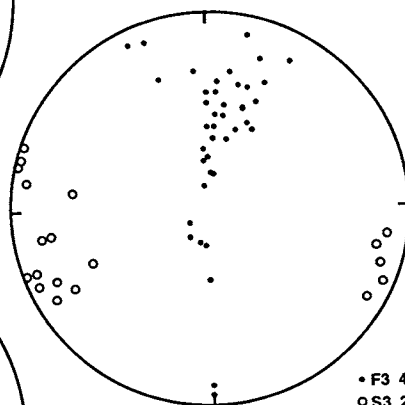
• S2 109 poles

c



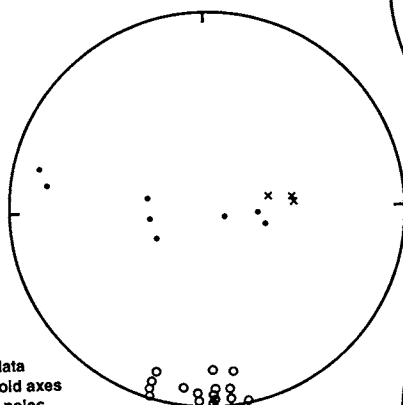
x L2 14 data
• F2 22 fold axes

d



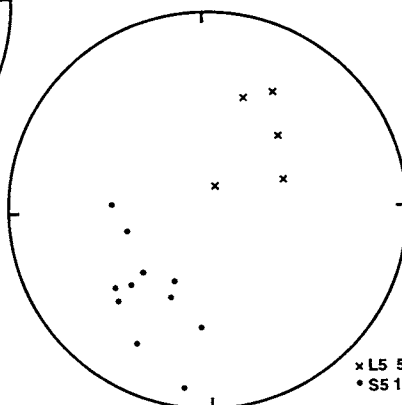
• F3 44 fold axes
o S3 22 poles

e



x L4 3 data
• F4 8 fold axes
o S4 23 poles

f



x L5 5 data
• S5 11 poles

Plate 2.4

a) Intrafolial F_1 fold, defined by leucosome (lc) in *proclamation gneiss*. Lens cap for scale; width of view is 40cm.

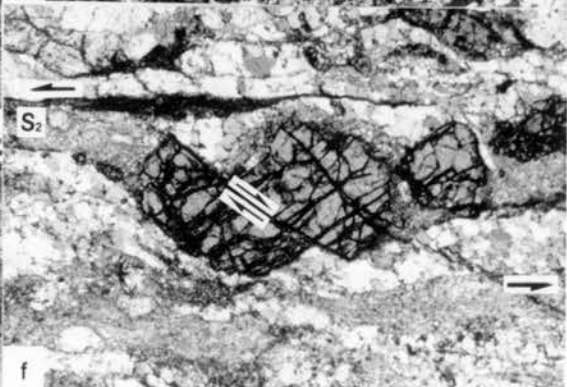
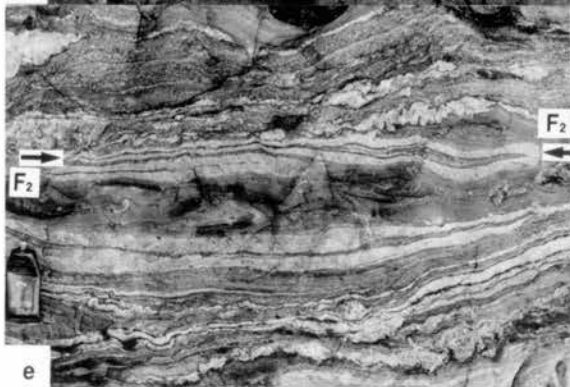
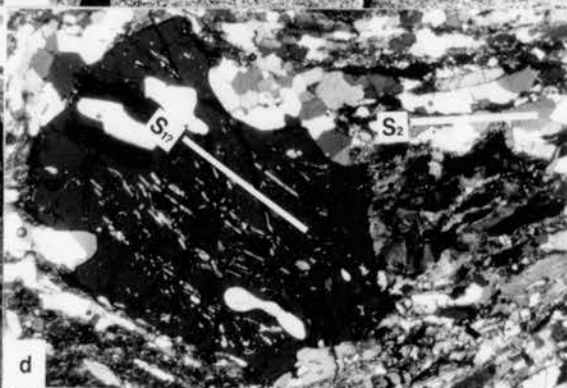
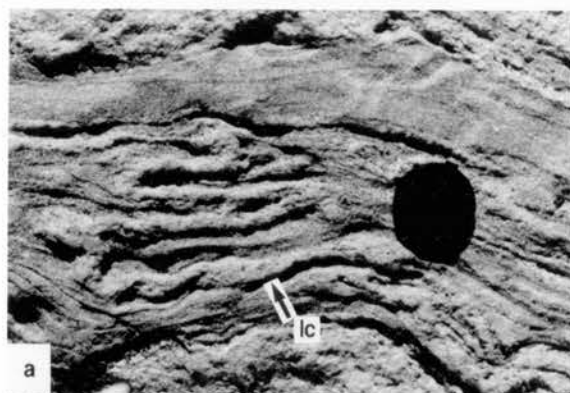
b) Strong S_2 foliation in *biotite-pyroxene orthogneiss*. The S_2 foliation wraps around K-feldspar porphyroblasts and lenses of less deformed orthogneiss. Chisel for scale; width of view is 35cm.

c) Recumbent isoclinal F_2 folds in *red granitic orthogneiss*, defined by deformed layers of leucosome. Width of view is 4m.

d) Sillimanite inclusions in a garnet grain, preserving an S_1 or early S_2 foliation that is oblique to the dominant S_2 foliation in the rock matrix. Thin section of sample P9, *felsic paragneiss*. Width of view is 4mm.

e) Attenuated F_2 sheath fold in *metapelitic gneiss*. Dual fold closures are indicated by arrows. Width of view is 70cm.

f) S_2 mylonitic shear planes, with a sinistral sense of shear indicated by normal 'faulting' (*sic* Simpson, 1986) of a garnet grain. Thin section of sample 921311b, *white granitic orthogneiss*. Width of view is 4mm.



orthogneiss, some leucosome transgresses S_1 and is deformed into S_2 laminations 5–50mm thick. S_2 is also defined by compositional layering on a millimetre to centimetre scale. S_2 is most clearly distinguished from S_1 gneissosity in the noses of isoclinal F_2 folds, that have an axial planar S_2 foliation defined by aligned biotite grains. Garnet porphyroblasts in *metapelitic gneiss* may also contain inclusions that define a foliation which is oblique to the S_2 foliation in the matrix (Plate 2.4d).

In almost all places S_1 has been completely transposed into the S_2 gneissosity, so that the two planar fabrics have identical ranges of orientation (Figs 2.3a & b). At locations where an orthogneiss intrudes across an S_1 gneissosity in an older unit, the S_2 gneissosity in the orthogneiss lies parallel to the truncated S_1 fabric (Plate 2.3c).

The D_2 event produced mesoscopic to macroscopic isoclinal F_2 folds, with axes that plunge shallowly to steeply north or south (Fig. 2.3c) and axial planes that are inclined to varying degrees towards the north-west and west (Fig. 2.3b). Sheath-type and doubly-plunging F_2 folds are common (Plate 2.4e), and many F_2 folds possess a strong asymmetry with both sinistral and dextral senses of shear. On sub-horizontal surfaces, the traces of most F_2 folds have a dextral asymmetry.

A well developed L_2 mineral and stretching lineation is present within S_2 in some lithologies. In felsic gneisses, L_2 is defined by elongate aggregates of quartz and feldspar. This forms the dominant fabric in some exposures of *white granitic orthogneiss*. In *metapelitic gneiss*, L_2 is defined by aligned grains of sillimanite and biotite. Measured L_2 lineations plunge to varying degrees towards the north or south (Fig. 2.3c). Where L_2 is observed near F_2 folds, it is usually parallel to F_2 fold axes. Boudinage structures occur in S_2 -parallel felsic layers and mafic dykes that are hosted by less competent gneisses (Plate 2.1b). The boudins are typically oriented with their longest axis approximately parallel to L_2 .

Mylonitic planes less than 0.5mm wide occur parallel to S_2 in *white granitic orthogneiss* and *biotite-pyroxene orthogneiss* (Plate 2.4f). These mylonites occur as diffuse shear

bands rather than as narrow zones of high strain, consistent with the highly ductile style of deformation seen in other D₂ fabrics. The presence of S₂ mylonites, F₂ sheath folds, and L₂ mineral lineations that lie parallel to F₂ fold axes, all indicate that D₂ strain contained a strong shear component. A consistent sense of shear, however, could not be inferred from D₂ structures. The distribution of co-linear F₂ fold axes and L₂ lineations (Fig. 2.3c) is consistent with a north or south-plunging direction of transport for the shear component of D₂ strain.

D₃ event

F₃ folding is recognised in all gneissic lithologies. Folding is upright, open to tight and chevronic, with wavelengths of centimetres to metres (Plate 2.1d). Dykes of felsic orthogneiss that truncate S₂/S₁ gneissosity were also folded by D₃ (Plate 2.5a). Post-D₃ felsic pegmatites truncate F₃ structures.

A weak S₃ foliation is best developed near the axial planes of F₃ folds. S₃ is defined by aligned biotite and flattened quartz and feldspar grains. Sheets of leucosome are also found parallel to the axial planes of F₃ folds (Plate 2.2d). S₃ foliations, and the axial planes of F₃ folds, are sub-vertical and trend northwards (Fig. 2.3d). Most F₃ fold axes plunge at moderate angles towards the north, but some plunge shallowly or steeply south (Fig. 2.3d). F₃ folding produced a east-west redistribution of S₂ and S₁ foliations (Figs 2.3a & b). The intensity of F₃ folding increases towards the south-western corner of Cape Bruce, where the S₂/S₁ gneissosity has been rotated into a steeply dipping orientation (Fig. 2.4a).

Progressive deformation during the D₃ event can be seen in the co-axial refolding of F₃ chevronic folds (Plate 2.5b). Where mesoscopic F₂ folds are refolded by D₃, the F₂ and F₃ fold axes are sub-parallel, producing type-III interference patterns (Plate 2.5c; Ramsey, 1967).

Plate 2.5

- a) A dyke of felsic orthogneiss (fog) truncates S_2 gneissosity in *metapelitic gneiss*. The dyke and the S_2 gneissosity are deformed into an upright F_3 fold. Vertical view towards the north-west. Sledge hammer head for scale; width of view is 50cm.
- b) Progressive refolding of F_3 folds in migmatized *felsic paragneiss*. The fold marked F_{3a} is refolded by fold F_{3b} ; the folds have parallel fold axes. Vertical view towards the north; width of view is 5m.
- c) Type-III (Ramsey, 1967) interference pattern between isoclinal F_2 and chevronic F_3 folds. F_2 and F_3 have parallel fold axes. *Metapelitic gneiss*. Width of view is about 1m.
- d) Upright F_4 fold in *felsic paragneiss*. A thin felsic pegmatite (peg) has been intruded parallel to the axial plane of the F_4 fold. Height of view is about 60cm.

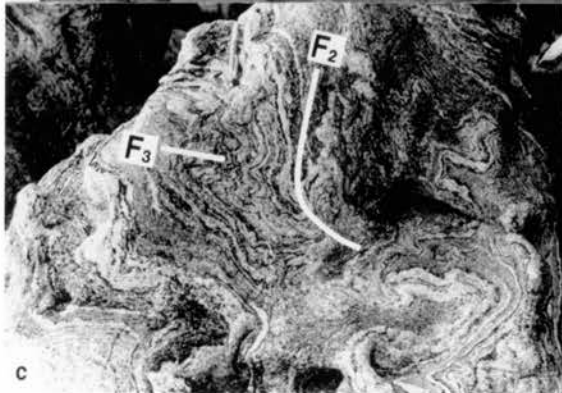
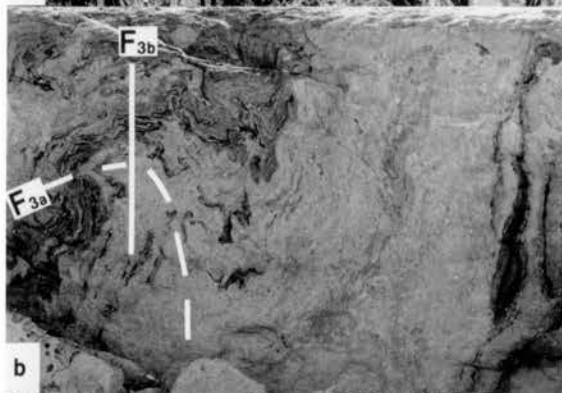
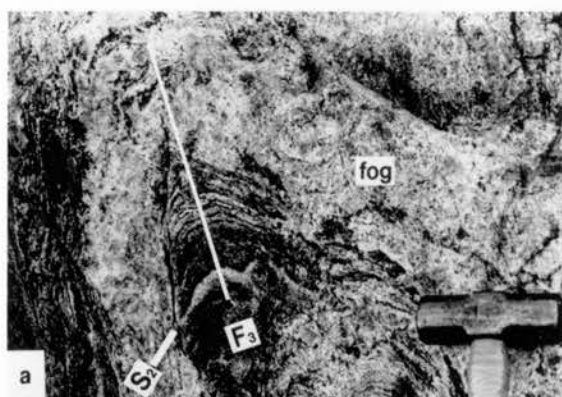
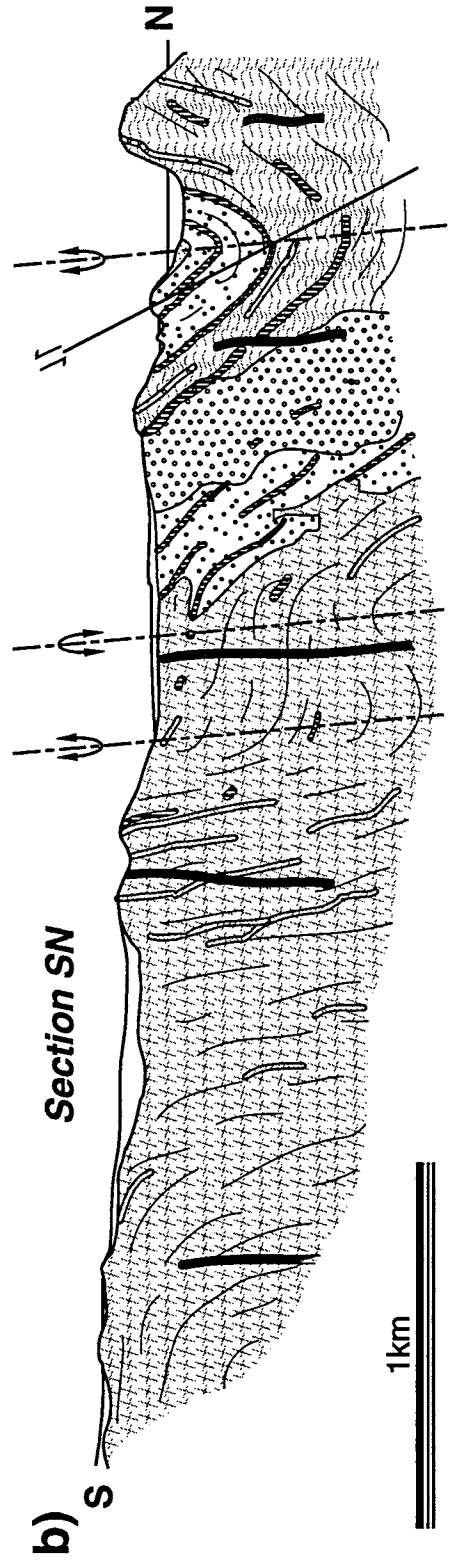
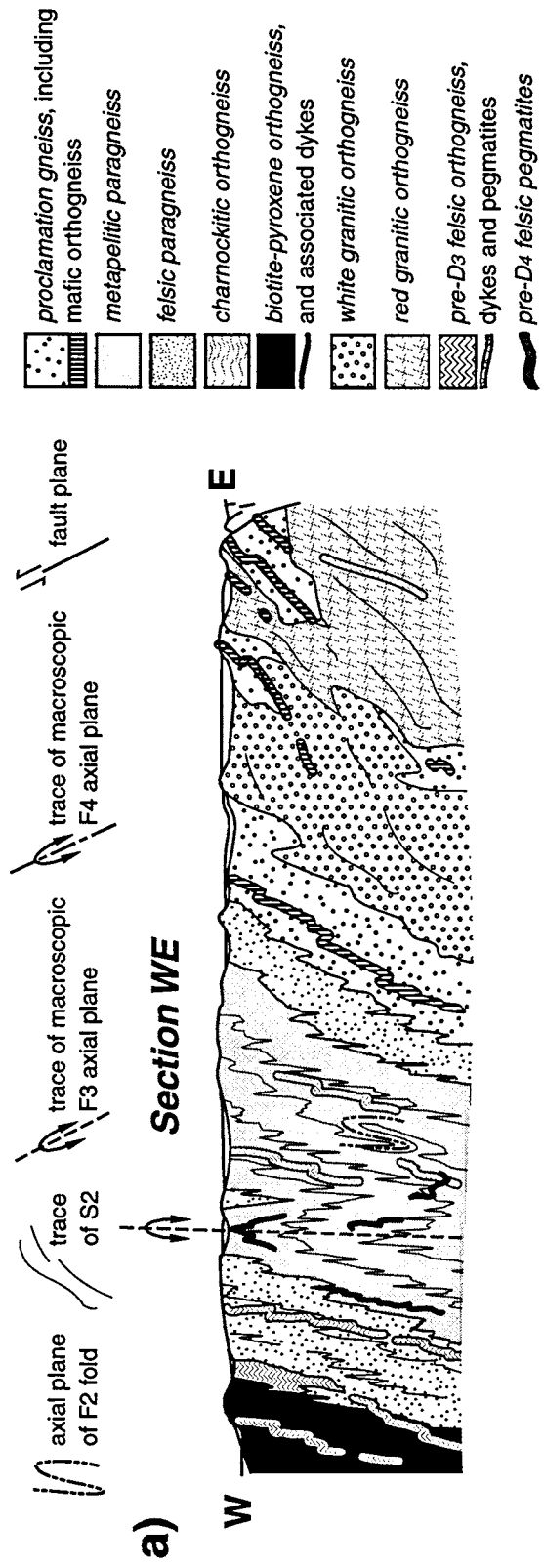


Fig. 2.4. Geological cross sections of the Cape Bruce locality. Positions of the cross sections are shown on the maps in Figs 2.1 and 2.2. There is no vertical exaggeration.

a) W-E geological section, looking north. The intensity of D_3 strain increases towards the western side of the area.

b) S-N geological section, looking west.



D₄ event

The distribution in the Cape Bruce area of gneisses layered by D₁ and D₂ is controlled by an open syncline-anticline pair, that has a wavelength of 2km and an east-trending axial trace (Figs 2.2, 2.4b). This macroscopic structure is correlated with mesoscopic open F₄ folds, that have upright east-trending axial planes (Fig. 2.3e). F₄ folds occur sparsely, and are commonly associated with felsic pegmatites that lie parallel to the axial planes of F₄ folds (Plate 2.5d). F₄ fold axes vary in plunge from east to westward (Fig. 2.3e). This distribution is due to the undulose S₂/S₁ form surface consequent to D₃. The limb trace of the macroscopic F₄ becomes more open towards the western edge of Cape Bruce (Fig. 2.2). This is due to the rotation of S₂/S₁ gneissosity into a zone of higher D₃ strain (Fig. 2.3b), so that the macroscopic F₄ fold axis is nearly vertical in the west.

An S₄ mineral foliation is commonly found in all lithologies. It is particularly well developed in post-D₃ pegmatites and in the hinges of F₄ folds. The foliation is defined by aligned biotite grains. Small amounts of leucosome lie along the axial planes of several F₄ folds. The S₄ foliation is sub-vertical and east-west trending (Fig. 2.3e).

D₅ event

A number of narrow, discontinuous ultramylonite-pseudotachylite zones are distributed across the Cape Bruce area (Fig. 2.2). These zones transect all earlier fabrics. Networks of D₅ mylonite are common, with a width of 0.5-50cm, and can be traced for distances of tens to hundreds of metres (Plate 2.6a). Mylonitic veinlets contain finely recrystallized grains of host rock minerals, as well as newly grown biotite and garnet (White & Clarke, 1994). Wider shear zones up to 3 metres across have planar boundaries, that bind an anastomosing network of shear planes, brecciated rock and pseudotachylite (Plate 2.6b). The dip direction of the shear zones varies from northward to eastward (Fig. 2.3f). Exposed mylonite planes contain a L₅ stretching lineation that plunges to the north-east (Fig. 2.3f). Offsets along the shear

Plate 2.6

a) Horizontal displacement of approximately 30cm on an S₅ mylonite. The absolute direction of movement is unknown. *Felsic paragneiss*. Width of view in foreground is about 50cm.

b) A D₅ ultramylonite-pseudotachylite zone transects *metapelitic gneiss*. The sense of movement is indeterminate. Notebook for scale; width of view is 1.1m.



a



b

zones are mostly indeterminate, but where seen with the stretching lineation indicate a north-east-downward normal movement of centimetres to tens of metres. Reverse movement is seen in a few shear zones.

A few macroscopic faults are found in the vicinity of Proclamation Point (Fig. 2.1, Plate 2.1a). These faults are north-trending and dip steeply to the east, and have a east-upward sense of movement. Leucosome is found in the fault plane where the faults transect felsic lithologies (Plate 2.1a). The faults cannot be related to any other deformation events.

2.3 The sequence of geological events at Cape Bruce

The relative timing of structural and intrusive events in the history of the Cape Bruce area is summarised in Table 2.1. The timing of events has been refined by dating work that is discussed in Chapter 4. Periods of metamorphism, and their relationship to the history of the area, are discussed in Chapter 3.

The oldest rocks recognised in the Cape Bruce area are the sedimentary and igneous lithologies that were the precursors of the *proclamation gneiss*, *metapelitic gneiss*, *felsic paragneiss* and the *charnockitic orthogneiss* (Table 2.1). The relative timing of these units cannot be determined solely from field relationships, as all lithological contacts have been transposed into an intense S_1 gneissosity. The crystalline nature of the gneisses, and the mineral assemblages they contain, are consistent with D_1 having occurred under granulite facies conditions (M_1). Large amounts of leucosome were generated by extensive partial melting, immediately before and/or during the D_1 event.

The original orientation of S_1 cannot be determined, due to the transposition of S_1 into an intense S_2 gneissosity (Table 2.1). The D_1 and D_2 events are separated in time by the intrusion of igneous precursors to the *biotite-pyroxene*, *white granitic* and *red*

Table 2.1. The sequence of geological events in the Cape Bruce area. Comparisons with the geological history of the Colbeck Archipelago (White & Clarke (1993) are included. All dates quoted are derived from new data that are discussed in Chapter 4 (Geochronology). References: *White & Clarke, 1993; °Metamorphic events defined in White (1993) and White & Clarke (1994).

Age	Structural	Metamorphic	Intrusive	Characteristics
?				<i>Metapelitic gneiss, felsic paragneiss, proclination gneiss, calcareous & cherty metasediments</i>
≥ 995 Ma			<i>Charnockitic orthogneiss</i>	Charnockite-enderbite with ultramafic xenoliths
? c. 995 Ma		M1 (M1 ^o)		Pre to syn-D1 granulite-facies metamorphism and extensive partial melting
	D1 (D1 [*])			Production of dominant S1 gneissosity in paragneisses and charnockite
992±10 Ma			<i>Biotite-pyroxene orthogneiss and related dykes</i>	Stock and tonalitic to mafic dykes. Similar composition to the <i>biotite-pyroxene gneiss</i> in the Colbeck Archipelago*
			<i>White granitic orthogneiss</i>	Leucocratic Pl-Ksp-Grt granitoid stock and related felsic dykes. Contains remnants of post-D1 mafic dykes
? 995-989 Ma			<i>Red granitic orthogneiss</i>	Massive Ksp-rich granitoid stock. Contains remnants of mafic dykes.
? 994-945 Ma		M2 (M2 ^o)		Pre to syn-D2 transitional granulite facies rehydration and retrogression, possibly during cooling after M1. Syn-D2 partial melting in felsic orthogneisses and paragneisses
	D2 (D2 [*])			Deformation of S1 into inclined N-plunging isoclinal F2, development of intense S2/L2 and boudinage of dykes and leucocratic layers. Possibly a continuation of D1.
944±10 Ma			<i>Felsic orthogneiss dykes and pegmatites</i>	Ksp -rich Bt-Grt-bearing granitoid dykes, pegmatites and web-like injected networks. Most dykes are sub-parallel to S2. Late-post-D2 intrusion.
	D3			Upright chevronic folding of S2/S1, with N-plunging F3 axes and weak N-trending S3. Higher strain deformation in the SW parts of the Cape Bruce outcrop. Syn-D3 partial melting in felsic rocks. Possibly the waning stage of a continuous D1 to D3 event.
c. 963 Ma			Felsic pegmatites	Ksp-rich Grt-bearing felsic pegmatites. Pre-syn-D4 intrusion. Similar composition and timing to pegmatites in the Colbeck Archipelago*, that are related to the intrusion of <i>Mawson Charnockite</i> .
924-889 Ma		M3 (M3 ^o)		Upper amphibolite to transitional granulite -facies metamorphism, with partial melting in felsic rocks. Possibly syn-D4.
	D4 (D3 [*])			Macroscopic to mesoscopic open to tight E-trending F4. Vertical S3 axial-planar mineral foliation and E to W-plunging F4 axes.
?		M4 (M5 ^o)		Garnet growth in D5 shear zones at granulite facies conditions
	D5 (D5 [*])			Ultramylonite/pseudotachylite shear zones, variably E-NE trending.

granitic orthogneisses. It is possible that these orthogneisses were intruded during a continuous D₁-D₂ tectonic episode. This is supported by the observation of sheets of *white granitic orthogneiss* which were emplaced sub-parallel to the S₁ gneissosity, and by the parallel orientations of S₁ and S₂ gneissosities near pre-D₂ intrusive contacts that have not been transposed into S₂.

Orthogneisses that intruded after D₁ contain leucosome that is deformed by D₂. This is consistent with partial melting having continued after D₁ and before or during D₂. S₂ is defined by the alignment of granulite facies minerals, which included abundant biotite grains. The mineral assemblages provide evidence that D₂ occurred under granulite facies conditions (M₂), with a high water activity than that present during D₁. These conditions may have resulted from the addition of hydrous fluids, that were released by felsic melts that solidified during D₂/M₂.

Due to re-orientation of S₂ by the D₃ and D₄ events, the original orientation of D₂ structures cannot be inferred. The D₃ event is temporally separated from D₂ by the intrusion of garnet-bearing granitic dykes and pegmatites. Most of these dykes were intruded sub-parallel to the S₂ gneissosity, and may have been emplaced during a late stage of the D₂ event. The intensity of deformation during D₃ is much lower than during D₂ or D₁, and increases towards the west of Cape Bruce. S₃ foliations and F₃ axial planes are sub-vertical and north-trending, consistent with a sub-horizontal east-west direction of maximum compression. There is no evidence for a simple shear component to D₃ strain. Since measured F₃ axes are sub-parallel to F₂ axes and L₂ stretching lineations, D₃ may represent a lower strain stage of continuous D₁-D₂ deformation. If this is the case, the transition from inclined S₂ fabrics to a sub-vertical S₃ involved a rotation of the direction of maximum shortening.

D₄ tectonism resulted in the gentle macroscopic warping of all previous fabrics. This event was accompanied by the production of minor amounts of leucosome, which developed along the axial planes of mesoscopic F₄ folds. The presence of garnet in the

leucosomes, along with evidence of partial melting, indicates that granulite or amphibolite facies conditions prevailed during D₄ (M₃, Table 2.1). The orientation of S₃ mineral foliations and F₃ axial planes is consistent with a sub-horizontal north-south direction of maximum shortening during D₄. This marks a significant transition in the tectonic development of the area.

The D₅ event involved semi-brittle deformation as opposed to the ductile deformation styles of the previous tectonic events. Most of the D₅ shear zones have a south-west-upward sense of movement and a north-east-plunging direction of maximum extension. Displacements along the shear zones have had little apparent effect on the local distribution of lithologies. Garnet growth in D₅ mylonites from Cape Bruce indicate that granulite facies conditions (M₄) were present during the D₅ event (White & Clarke, 1994).

2.4 Correlations with the Colbeck Archipelago

The geology of Cape Bruce is very similar to that of the Colbeck Archipelago and Chapman Ridge, which lie approximately 10km to the east of Cape Bruce. Correlations can be made between the localities, based on similarities in the composition of the lithologies and on the similar sequence of geological events. The structural and metamorphic events that were inferred by White & Clarke (1993) for the Colbeck Archipelago, are shown in Table 2.1, where they correspond to events in the history of the Cape Bruce gneisses.

Proclamation gneiss was recognised by White & Clarke (1993) at both Cape Bruce and the Colbeck Archipelago. The orthopyroxene±clinopyroxene gneiss of White & Clarke (1993) has a similar petrography to the charnockitic orthogneiss of Cape Bruce. In both localities, the aforementioned lithologies were metamorphosed into granulitic gneisses by a peak metamorphic event (D₁-M₁; Table 2.1). Orthogneisses

were intruded after the formation of the S₁ gneissosity (Tables 1.1 & 2.1). The biotite±pyroxene gneiss in the Colbeck Archipelago (White & Clarke, 1993) has a very similar composition to that of the biotite-pyroxene orthogneiss at Cape Bruce. The granitic gneiss in the Colbeck Archipelago, along with the red and white granitic orthogneisses at Cape Bruce, were also emplaced after D₁. These intrusions were deformed by D₂, which in both localities was associated with a strong S₂ foliation and isoclinal F₂ folds. In metapelitic gneisses from both areas, S₂ is defined by the alignment of biotite and sillimanite grains. After D₂, felsic dykes and pegmatites intruded gneisses at Cape Bruce and the Colbeck Archipelago.

Later structural events are also comparable between the two localities. At Cape Bruce, upright F₄ folds with east-trending vertical axial planes have a similar orientation and morphology to the F₃ folds of White & Clarke (1993), that have east-plunging axes and steeply south-dipping axial planes. In both localities, D₅ shear zones containing ultramylonite and pseudotachylite transect all earlier fabrics. These shear zones share a north-east to north-west dip direction and a dominant south over north sense of movement.

2.5 Comparisons with other parts of the Mawson Coast

Geological correlations of Cape Bruce and the Colbeck Archipelago with other localities on the Mawson Coast need to be re-assessed, in light of the new data. The separation of gneisses at Cape Bruce into the Colbeck Gneiss and the Stillwell Gneiss by Trail (1970) is not considered valid by the author, since it groups together lithologies of diverse composition and history. It is possible that paragneisses at Cape Bruce and the Colbeck Archipelago correlate with the Colbeck Gneiss in the Stillwell Hills (Trail, 1970); however, insufficient work has been done on the metasediments of the Stillwell Hills to verify this possibility.

Structural correlations by White & Clarke (1993) and Clarke (1988) between different localities on the Mawson Coast were made on a tentative basis, due to the lack of geochronological constraints on the formation of lithologies and structural fabrics. White & Clarke (1993) suggested, only as a possibility, that their D₁ event occurred during the Archaean, on the basis of the *c.* 2700 Ma age derived from pre- to syn-tectonic charnockite from the Stillwell Hills (Clarke, 1988). The correlation of 1190±200 Ma mafic dykes on the Mawson Coast with the Amundsen dykes in Napier Land is also suspect, due to the lack of geochemical evidence for the correlation.

The author suggests an alternative interpretation for the timing of events at Cape Bruce and the Colbeck Archipelago. The lack of differentiation between D₁ and D₂ fabrics in the area is consistent with the two events forming a continuous deformation episode. This is supported by new geochronological work, described in Chapter 4, which implies that the D₁ and D₂ events cannot be temporally resolved. All of the deformation events observed in the Cape Bruce area, with the exception of the D₅ shear zones, can be assigned with confidence to the *c.* 1000 Ma Rayner Structural Episode. In view of the new geochronological data, new structural correlations between localities on the Mawson Coast are made in Chapter 6 (Synthesis).

Chapter 3

Metamorphism

3. Metamorphism

Introduction

The gneisses at Cape Bruce contain an array of textures and mineral relationships that provide evidence for metamorphism under granulite facies conditions. In this chapter, an investigation of the development of mineral assemblages reveals complexity in the metamorphic history of the Neoproterozoic Rayner Structural Episode. Samples containing several stages of mineral growth are described. The results of chemical analysis of stable mineral assemblages are described to provide a quantitative basis for constraints on the conditions of metamorphism. Finally, the paragenesis of mineral assemblages are explained by changes in ambient conditions during metamorphism.

3.1 Previous Work

Gneisses at Cape Bruce and other localities along the Mawson Coast contain textures and mineral assemblages representative of lithologies metamorphosed under transitional (or hornblende) granulite to granulite facies conditions, as defined by Turner & Verhoogen (1960). White (1993) used samples from the Colbeck Archipelago (Fig. 1.2) to infer a metamorphic history for that area that involved three distinct events (M_{1-3}). Granulite facies assemblages (M_1) had no hydrous minerals. 'Peak' assemblages were inferred to have formed before or during the early stages of the D_1 event, with subsequent cooling having controlled the growth of anhydrous reactants (M_1) during D_1 (White, 1993). Hydrous minerals, such as biotite and hornblende, were interpreted to have grown as a result of lower temperature metamorphism and the influx of water during a D_2 - M_2 event. Further biotite growth

and recrystallization of gneiss (M₃, White & Clarke 1994) occurred during the D₃ event.

Geothermobarometry on M₁ assemblages in samples from the Colbeck Archipelago yielded conditions of $P=5.1\pm0.8$ kbar at $T\geq 750^\circ\text{C}$ (White & Clarke, 1993). The significance of this result was considered to be ambiguous by White (1993), as assemblages containing products of M₂ metamorphism produce similar estimates using the same methods ($P\approx 5\text{--}6$ kbar at $T\approx 700^\circ\text{C}$; White, 1993).

Comparable results have been derived from Painted Gneiss in the Framnes Mountains (Fig. 1.2). In samples from the Casey Range, Clarke & Norman (1993) obtained an estimate of M₂ conditions of $P=5.6\pm0.4$ kbar at $T=700^\circ\text{C}$. These samples contain reaction textures that record near-isobaric cooling. At Mt. Henderson, the migmatization of paragneiss in contact with Mawson Charnockite provided a useful estimate of conditions at the time of intrusion. Clarke & Norman (1993) derived an estimate for pre-D₃ (M₃) metamorphism of $P=3.9\pm1.0$ kbar at $T=700^\circ\text{C}$ for the assemblage in the migmatized Painted Gneiss.

Metamorphic conditions within mylonite zones (M₅ of White & Clarke, 1994) in the Colbeck Archipelago and at Cape Bruce (Fig. 1.2) were estimated by White & Clarke (1994) at $P\approx 6.5$ kbar and $T\approx 550\text{--}800^\circ\text{C}$. This estimate applies to recrystallized assemblages within the mylonites and does not necessarily reflect the ambient conditions that accompanied the development of the shear zones (Clarke & Norman, 1993).

3.2 Petrography

A limited number of rock types from the Cape Bruce locality were used to constrain the conditions and history of metamorphism during the Rayner Structural Episode. The most useful rock types for this purpose were garnet-pyroxene-bearing orthogneiss

and garnet-sillimanite-bearing paragneiss. Samples of these rock types contain textures that can be attributed to sequential stages of mineral growth during metamorphism. Sample used for this purpose are divided into the following groups: 1) garnet-bearing *charnockitic orthogneiss*; 2) garnet-bearing mafic orthogneiss, included in the *proclamation gneiss* group; 3) Post-D₁ orthogneisses; 4) cordierite-free, garnet-sillimanite-bearing paragneiss; 5) cordierite-bearing paragneiss; 6) spinel-quartz-bearing paragneiss; 7) spinel-corundum-bearing paragneiss.

3.2.1 Orthogneiss

Garnet-bearing charnockitic orthogneiss

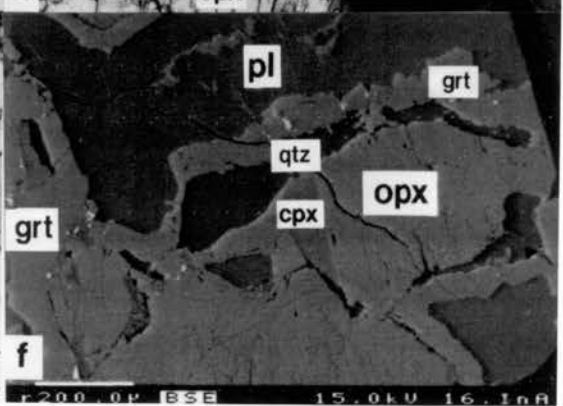
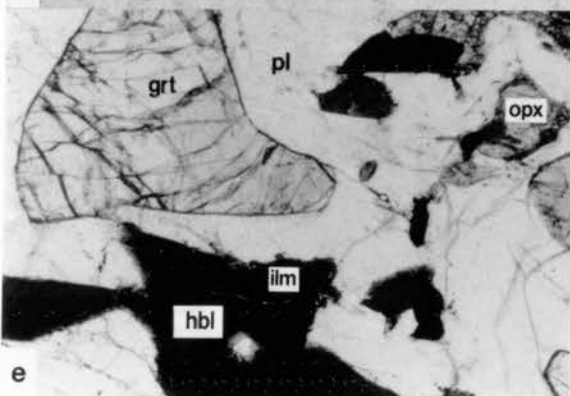
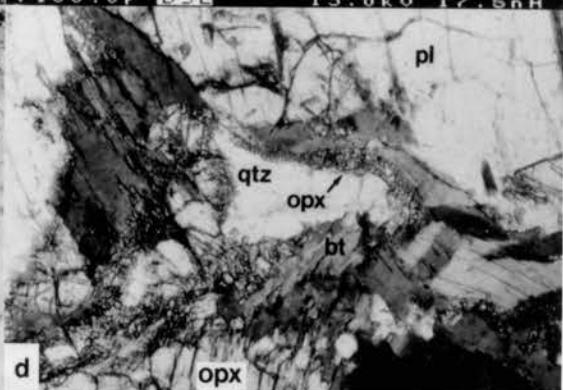
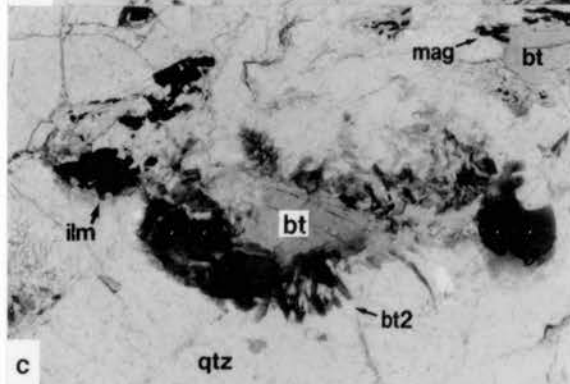
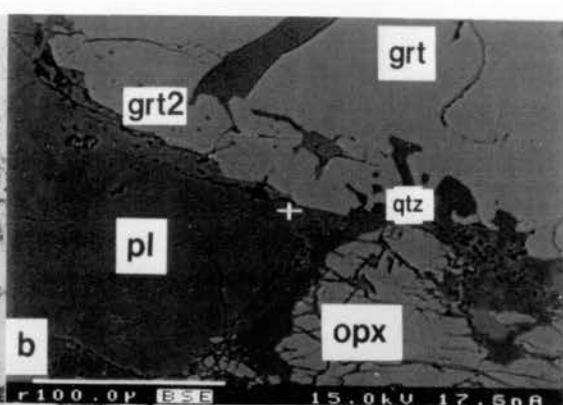
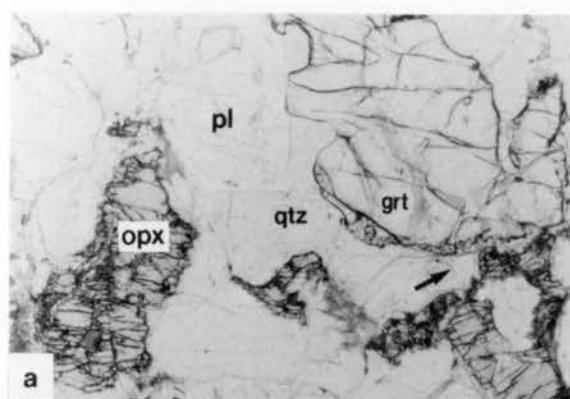
Garnet is found in only a few exposures of *charnockitic orthogneiss*, where it occurs adjacent to patches of *felsic paragneiss* that have been partially incorporated into the orthogneiss. Samples O2, O3 and O5 contain garnet porphyroblasts up to 1cm across, that appear to be in textural equilibrium with a granoblastic matrix of orthopyroxene, plagioclase, K-feldspar, quartz and ilmenite (Plate 3.1a). Layers of granoblastic rock and flattened biotite, quartz and feldspar define S₁. In some samples, biotite grains define a weak S₂ foliation at a small angle to S₁.

The rims of garnet porphyroblasts are commonly overgrown by subhedral crystals of garnet less than 0.2 mm across. Some overgrowths comprise garnet intergrown with ilmenite or quartz. Where garnet overgrowths lie adjacent to orthopyroxene, thin rims of quartz separate the two minerals (Plate 3.1b).

Biotite, garnet and orthopyroxene grains are overgrown by fine-grained intergrowths of biotite, ilmenite and/or quartz (Plate 3.1c). In sample O3, some biotite grains are rimmed by very fine-grained orthopyroxene (Plate 3.1d), or by symplectites of orthopyroxene, K-feldspar and ilmenite.

Plate 3.1.

- a) Coarse-grained garnet, orthopyroxene, plagioclase and quartz. The arrow indicates the site of Plate 3.1b. Sample O5, *charnockitic orthogneiss*. Plane-polarised transmitted light, width of photo = 2.3mm.
- b) Fine-grained garnet (grt2) overgrowing coarse-grained garnet, and quartz replacing orthopyroxene and plagioclase. Sample O5, *charnockitic orthogneiss*. Backscattered secondary electron image, width of photo = 0.3mm.
- c) Biotite grains partially recrystallised to fine-grained biotite (bt2) and magnetite. Sample O2, *charnockitic orthogneiss*. Plane-polarised transmitted light, width of photo = 1.0mm.
- d) Coarse-grained orthopyroxene partially replaced by biotite, which is in turn overgrown by fine-grained orthopyroxene. Sample O3, *charnockitic orthogneiss*. Plane-polarised transmitted light, width of photo = 0.9mm.
- e) Co-existing garnet, plagioclase, orthopyroxene, ilmenite and hornblende. Sample O6, mafic orthogneiss, *proclamation gneiss*. Plane-polarised transmitted light, width of photo = 2.2mm.
- f) Orthopyroxene and clinopyroxene separated from plagioclase by a dual corona of quartz and garnet. Sample O7, mafic orthogneiss, *proclamation gneiss*. Backscattered secondary electron image, width of photo = 1.5mm.



Garnet-bearing mafic orthogneiss

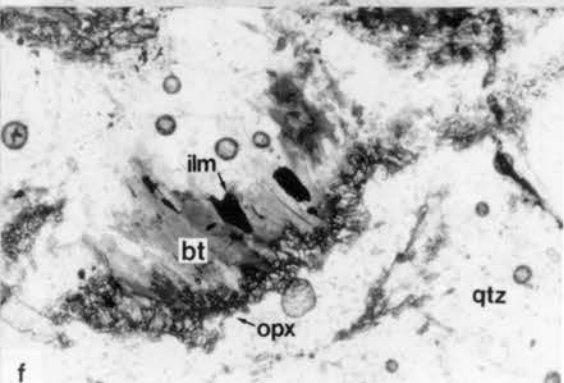
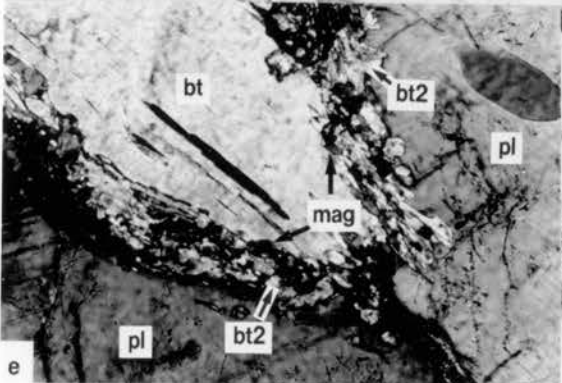
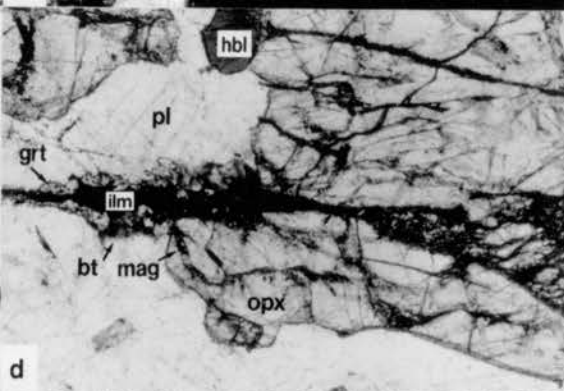
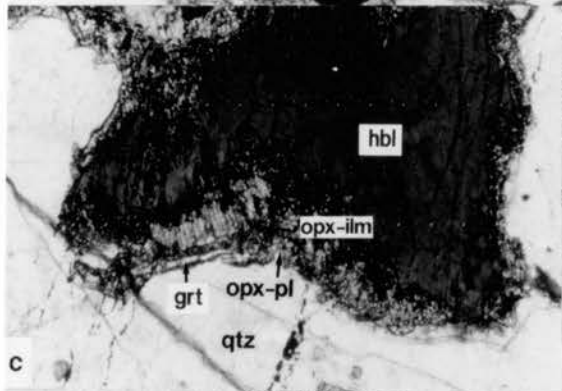
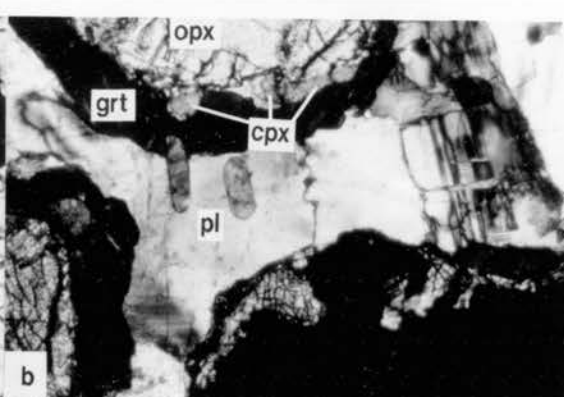
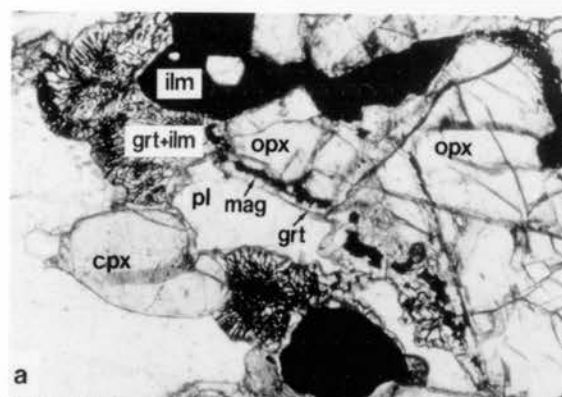
Two-pyroxene-bearing orthogneiss occurs in *proclamation gneiss* as layers parallel to S_1 (Fig. 2.1). These layers are commonly separated from adjacent layers of felsic and metapelitic gneiss by 1-50cm thick selvages containing biotite, quartz, garnet, ilmenite and feldspar. Garnet occurs in the mafic orthogneiss adjacent to these selvages, as porphyroblasts up to 1cm across in textural equilibrium with a granoblastic (0.5-1mm) matrix. The matrix contains abundant orthopyroxene, clinopyroxene and plagioclase, with variable amounts of brown-green hornblende and lesser quartz, ilmenite and biotite. In sample O6, hornblende is a major component, and the thin section contains a patch of very coarse-grained (5-15mm across) hornblende, garnet, plagioclase and orthopyroxene (Plate 3.1e). S_1 is defined by compositional layering and by the alignment of flattened pyroxene and ilmenite grains.

Coronas of garnet, that overgrow grains of garnet, hornblende, ilmenite, orthopyroxene and clinopyroxene are present in many samples. Pyroxene grains are commonly separated from plagioclase by coronas that are up to 0.3 mm thick and composed of garnet and magnetite, with or without a rim of quartz on the pyroxene contact (Plates 3.1f, 3.2a). Ilmenite grains are commonly overgrown by splay-shaped symplectites of garnet and ilmenite, with or without minor amounts of magnetite (Plate 3.2a). In sample O7, coronas of garnet, quartz, magnetite and clinopyroxene separate orthopyroxene and plagioclase (Plate 3.2b). Clinopyroxene occurs in the coronas as discrete grains up to 0.3mm across.

In samples O7 and O8, green-brown hornblende has overgrown ilmenite, orthopyroxene and/or clinopyroxene. In sample O6, ilmenite is commonly overgrown by small grains of biotite, and some garnet porphyroblasts contain cracks filled with finely intergrown biotite and magnetite. In all samples, some hornblende grains have rims less than 0.1mm thick of symplectic orthopyroxene, ilmenite and plagioclase (Plate 3.2c).

Plate 3.2

- a) Fe³⁺-rich ilmenite separated from plagioclase by symplectites of garnet and Fe³⁺-poor ilmenite. Orthopyroxene is also separated from plagioclase by coronas of garnet and magnetite. Sample O10, mafic orthogneiss, *proclamation gneiss*. Plane-polarised transmitted light, width of photo = 0.9mm.
- b) Orthopyroxene separated from plagioclase by a dual corona of clinopyroxene and garnet. Sample O7, mafic orthogneiss, *proclamation gneiss*. Crossed nichols, width of photo = 0.6mm.
- c) Coarse-grained hornblende replaced on it's rims by very fine-grained symplectites of orthopyroxene, ilmenite, quartz and plagioclase. Sample O6, mafic orthogneiss, *proclamation gneiss*. Plane-polarised transmitted light, width of photo = 0.8mm.
- d) Coarse-grained orthopyroxene and plagioclase recrystallized by a D₅ mylonite, that also contains fine-grained garnet, biotite and magnetite. Sample O6, mafic orthogneiss, *proclamation gneiss*. Plane-polarised transmitted light, width of photo = 2.1mm.
- e) Coarse-grained biotite with a rim composed of fine-grained biotite (bt2) and magnetite. Sample O11, pre-D₄ felsic pegmatite. Plane-polarised transmitted light, width of photo = 1.5mm.
- f) Biotite grains partially replaced by fine-grained orthopyroxene. Sample O1, *biotite-pyroxene orthogneiss* dyke. Plane-polarised transmitted light, width of photo = 0.8mm.



Most samples contain anastomosing sheets of recrystallized grains less than 0.1mm wide. In the thin section of Sample O6, a mylonite contains recrystallized quartz, pyroxene, ilmenite and garnet (Plate 3.2d).

Post-D₁ Orthogneisses

Reaction textures and limited recrystallization of granoblastic textures are also observed in various post-D₁ orthogneisses. *Red granitic orthogneiss* and *biotite-pyroxene orthogneiss* contain biotite grains that are mostly aligned with S₂. Biotite and adjacent grains are partially recrystallized into fine-grained (less than 0.2 mm) intergrowths of biotite, quartz, magnetite and/or ilmenite. In samples of pre-D₃ and pre-D₄ felsic orthogneiss, fine-grained biotite and magnetite occur as overgrowths on larger biotite grains (Plate 3.2e).

Samples of pre-D₃ and pre-D₄ felsic dykes contain sheets, less than 0.5mm wide, of finely recrystallized biotite, quartz, feldspar, magnetite and/or ilmenite. These sheets form anastomosing networks of variable orientation that transect all other mineral fabrics.

In a thin section from sample O1 of *biotite-pyroxene orthogneiss*, biotite grains aligned with S₂ are rimmed by fine-grained intergrowths of orthopyroxene, quartz and ilmenite (Plate 3.2f). These textures lie near a 0.1 mm wide D₅ mylonite that traverses the thin section. A sample from a similar dyke contains rare brown-green hornblende that is partially replaced by symplectites of orthopyroxene, ilmenite and plagioclase.

3.2.2 Paragneiss

Garnet and sillimanite-bearing metapelites record several stages of metamorphic evolution. Almost all of the samples described below contain leucosome with textural relationships that are consistent with partial melting before and during D₁ and D₂

(see Chapter 2). For the sake of clarity, paragneiss assemblages are categorised by the presence of diagnostic phases (cordierite, spinel and corundum). In a few samples, more than one of these assemblages are present in the same thin section. Where this is the case, the assemblages are present in discrete domains within the thin section.

Cordierite-free assemblage

These samples contain abundant quartz, K-feldspar, plagioclase, sillimanite, garnet, and biotite, with lesser amounts of magnetite, ilmenite and monazite. S_1 and S_2 are defined by layers of alternating compositions. Leucosome occurs abundantly as layers and patches that transgress S_1 . It contains K-feldspar, quartz and plagioclase. K-feldspar occurs as porphyroblasts up to 3cm across, that contain plagioclase exsolution lamellae. Myrmekitic intergrowths of quartz and feldspar are common (Plate 3.3a). Most of the layers are composed of inequigranular grains of garnet, K-feldspar, plagioclase, magnetite, ilmenite, sillimanite and biotite. Layers up to 0.7mm wide that are rich in sillimanite and biotite define an intense S_2/S_1 foliation. In sample P10, garnet grains up to 1mm across possess inclusion-free cores, along with rims that contain inclusions of fibrolitic sillimanite and biotite. Inclusions in the garnet rims are oriented sub-parallel to S_2 , defined in the surrounding matrix by sillimanite (Plate 3.3b).

Garnet grains contain cracks that are oriented at a large angle to the dominant S_2/S_1 foliation. In these cracks, garnet is commonly replaced by very fine-grained biotite and magnetite, with or without sillimanite. Most samples contain planes parallel to S_2 that contain finely recrystallized biotite, magnetite, feldspar, quartz and sillimanite.

Cordierite-bearing assemblage

This assemblage comprises garnet, K-feldspar, quartz, cordierite, sillimanite, magnetite and biotite, with or without small amounts of plagioclase and ilmenite. The samples are granoblastic with an intense S_2/S_1 foliation defined by laminations rich in sillimanite, biotite and magnetite grains up to 2mm long. Most garnet grains are

Plate 3.3

a) Quartz-feldspar myrmekitic intergrowths developed in leucosome, adjacent to a grain of ilmenite. Porphyroblastic K-feldspar grains in the leucosome contain plagioclase exsolution lamellae. Sample P5, cordierite-free paragneiss. Crossed Nichols, width of photo = 2.1mm.

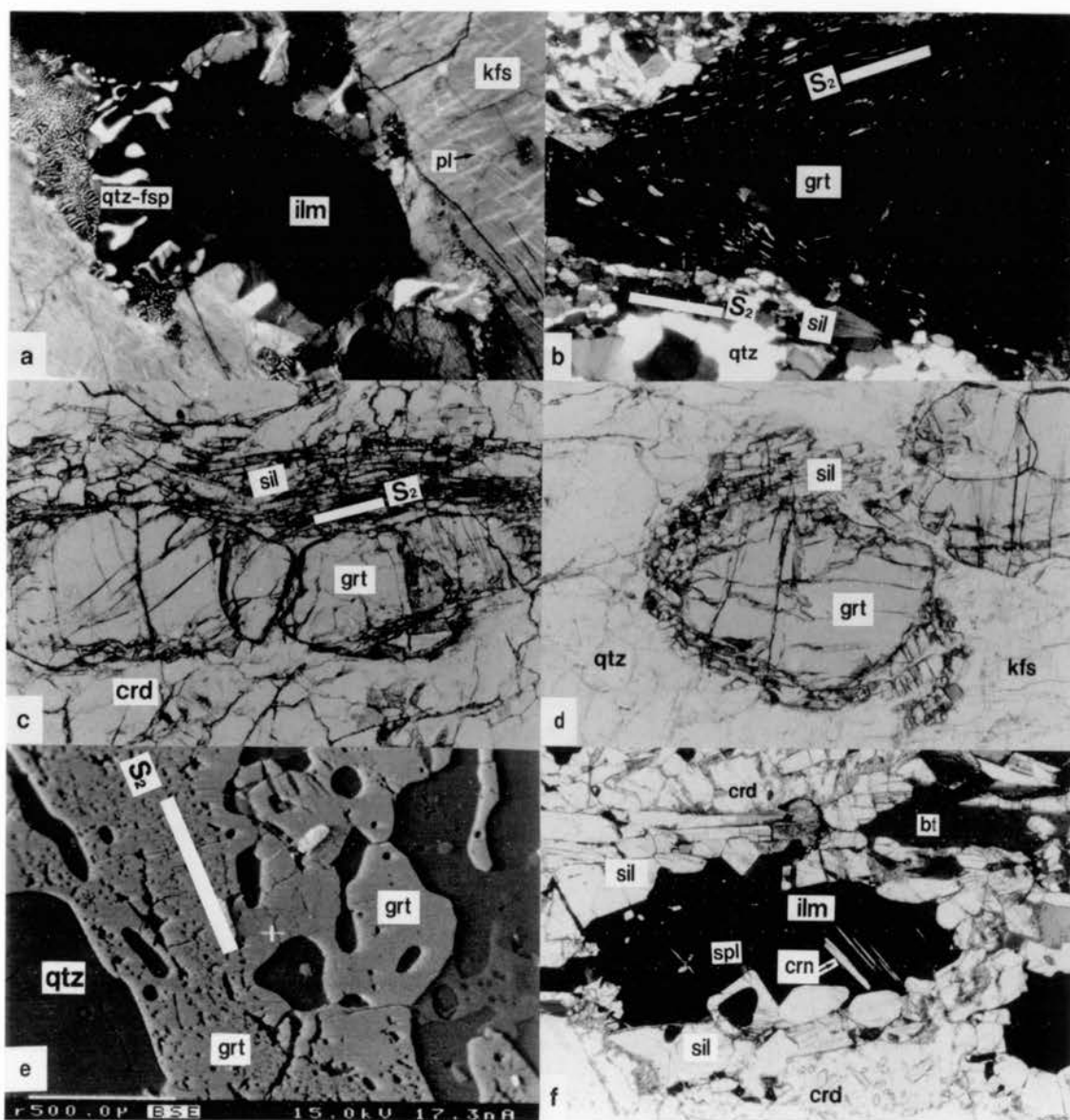
b) Garnet porphyroblast that is elongate parallel to $S_{2/1}$, with rims that contain S_2 -parallel inclusions of fibrolitic sillimanite that wrap around the core of the porphyroblast. Sample P10, cordierite-free paragneiss. Crossed Nichols, width of photo = 2.5mm.

c) Garnet grains that are elongate parallel to $S_{2/1}$ and that contain a few S_2 -parallel sillimanite inclusions in their rims. Sample P4, cordierite-bearing paragneiss. Plane-polarised transmitted light, width of photo = 2.3mm.

d) Sillimanite grains clustered around a garnet grain, enclosed in quartzofeldspathic leucosome. Sample P13, spinel-bearing *metapelitic gneiss*. Plane-polarised transmitted light, width of photo = 2.3mm.

e) Co-genetic inclusion-rich garnet and inclusion-poor garnet, elongated parallel to S_2 . The inclusions are S_2 -aligned sillimanite and quartz. Sample P14, spinel-bearing *metapelitic gneiss*. Backscattered secondary electron image, width of photo = 1.7mm.

f) Spinel intergrown with ilmenite grains that contain thin sheets of corundum. Coarse crystals of sillimanite separate the spinel + ilmenite intergrowth from cordierite. Sample P13, spinel-bearing *metapelitic gneiss*. Plane-polarised transmitted light, width of photo = 2.3mm.



equant, 0.5-2mm in diameter, and contain few inclusions. Several garnet grains have small cores that contain round inclusions of quartz and feldspar. Other garnet grains are flattened parallel to S_2 (Plate 3.3c), and may contain numerous inclusions of sillimanite aligned with S_2 .

The edges of some biotite grains have been recrystallized into very fine-grained intergrowths of biotite, quartz and magnetite. Cordierite is commonly replaced by pale yellow pinites.

Spinel-bearing assemblage

Spinel-bearing paragneiss differs from the above assemblages in that it contains higher modal proportions of sillimanite, biotite and cordierite. A typical sample (*e.g.* P12) contains two types of layers that define S_1 . The first is composed of granoblastic quartz, K-feldspar, and garnet, with minor amounts of sillimanite, plagioclase, biotite, cordierite, magnetite, ilmenite and spinel. The second is inequigranular, and contains abundant cordierite, garnet and sillimanite, with much less quartz and K-feldspar than the former type. Leucosome is also common, occurring as layers parallel to S_1 and as nebulitic patches which truncate S_1 (Plate 3.3d).

Sillimanite-rich layers have an intense S_2/S_1 foliation, defined by aligned sillimanite and biotite grains up to 2mm long. Grains of cordierite and garnet and clots of spinel + magnetite + ilmenite are flattened parallel to S_1 . Garnet occurs as sub-equant or flattened grains. Sub-equant garnet grains are up to 10 mm across and have inclusion-free cores, along with occasional thin rims that contain fibrolitic sillimanite and biotite inclusions aligned with S_2 . Flattened porphyroblasts of garnet, up to 5 cm long, lie parallel to S_2 and are rich in inclusions of fibrolitic sillimanite, biotite and quartz (Plate 3.3e).

Spinel occurs as irregular grains 0.1-3 mm across, that form clots of grains along with magnetite and ilmenite (Plate 3.3f). These clots are flattened into S_2 . In samples P12, P13 and P15, ilmenite grains up to 0.5 mm across contain very thin slivers of

corundum, especially in quartz-poor areas of the thin sections (Plate 3.3f). All spinel is dark green and contains varying amounts of magnetite exsolution. Where a spinel grain is found adjacent to a large magnetite grain, the number and size of magnetite blebs exsolved in the spinel decreases towards the adjacent magnetite grain (Plate 3.4a). Many spinel grains are partially to completely replaced by rims and patches of magnetite (Plate 3.4b).

Evidence for the co-existence of spinel and quartz under granulite-grade conditions is found in many samples. Tiny grains of spinel in contact with quartz are found in several samples (Plate 3.4c). Most spinel is separated from quartz by one or more other minerals. Spinel grains near quartz are surrounded by single crystals of sillimanite up to 1 mm long (Plate 3.4b), or, less commonly, coronas of cordierite (Plate 3.4d). Where spinel is found near garnet porphyroblasts, it is commonly rimmed by garnet coronas 0.1 to 0.5 mm thick, with or without symplectic magnetite (Plate 3.4e). Clots of spinel, magnetite and ilmenite are separated from cordierite by clusters of sillimanite crystals, that are 0.3-1 mm long and aligned with S_2 (Plate 3.4f).

Near garnet porphyroblasts, coronas of garnet with or without symplectic magnetite surround grains of magnetite, spinel, ilmenite, and sillimanite (Plate 3.5a). Irregular garnet-magnetite symplectites also enclose the edges of garnet grains, replacing adjacent quartz or cordierite (Plate 3.5b).

In leucosome from sample P13, garnet is partially separated from adjacent plagioclase by very fine-grained coronas of symplectic garnet and quartz, along with clusters of fibrolitic sillimanite (Plate 3.5c). Randomly orientated fibrolitic sillimanite is seen in all samples, which has replaced cordierite along grain boundaries and has overgrown coarser grained sillimanite, ilmenite and garnet (Plate 3.5d). In sample P8, patches of S_2 fibrolitic sillimanite are crenulated by F_4 (Plate 3.5e); this may be a 'ghost' effect due to the growth of fibrolitic sillimanite crystals along planes within deformed

Plate 3.4

- a) Spinel and magnetite surrounded by sillimanite. Note that the spinel grains become lighter towards adjacent magnetite grains, as the amount of exsolved magnetite in the spinel decreases. Sample P13, spinel-bearing *metapelitic gneiss*. Plane-polarised transmitted light, width of photo = 1.0mm.
- b) A grain of spinel partially replaced by magnetite, and enclosed in a single crystal of sillimanite. Sample P7, spinel-bearing *metapelitic gneiss*. Plane-polarised transmitted light, width of photo = 1.0mm.
- c) Adjacent grains of spinel and quartz. Sample P9, spinel-bearing *metapelitic gneiss*. Backscattered secondary electron image, width of photo = 0.3mm.
- d) Spinel separated from quartz by thin coronas of cordierite. Sample P7, spinel-bearing *metapelitic gneiss*. Crossed Nichols, width of photo = 1.3mm.
- e) Coarse-grained garnet adjacent to spinel and ilmenite grains, which are separated from quartz by a thin coronas of garnet. Parts of the coronas also contain magnetite. Sample P13, spinel-bearing *metapelitic gneiss*. Plane-polarised transmitted light, width of photo = 2.3mm.
- f) Spinel and ilmenite grains, flattened parallel to S_1 , are separated from cordierite by medium-sized grains of sillimanite, in a quartz-poor area of the thin section. Sample P2, spinel-bearing *metapelitic gneiss*. Crossed Nichols, width of photo = 2.3mm.

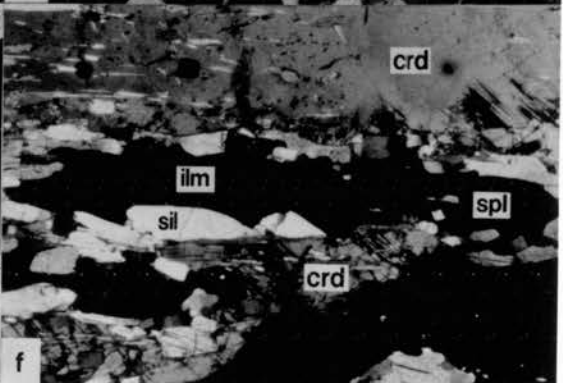
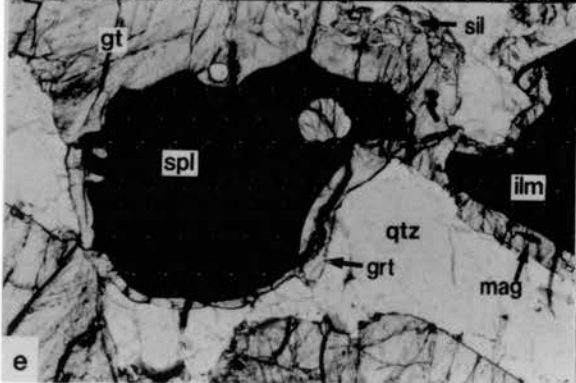
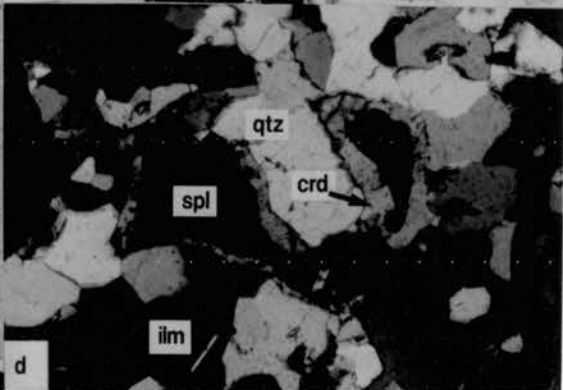
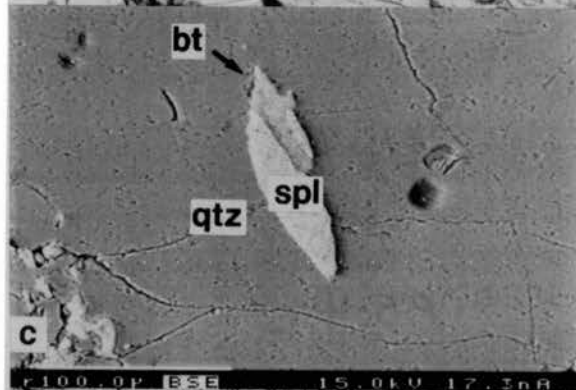
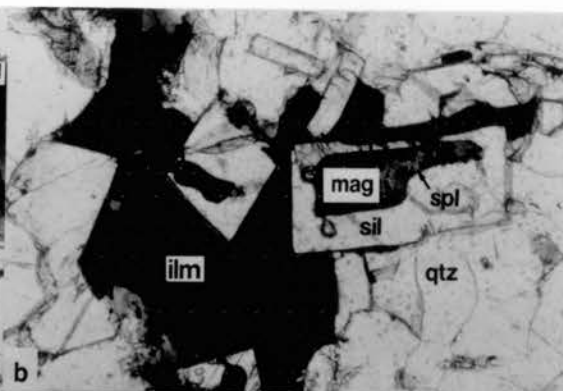
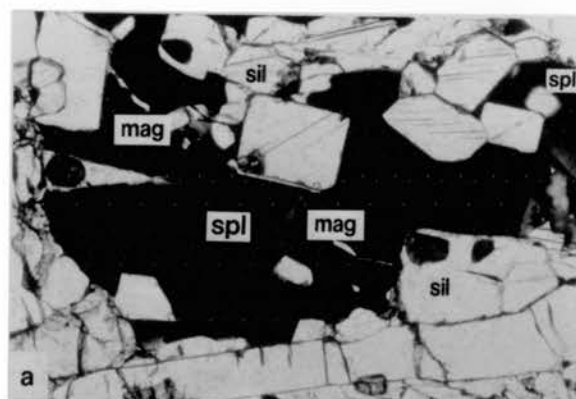
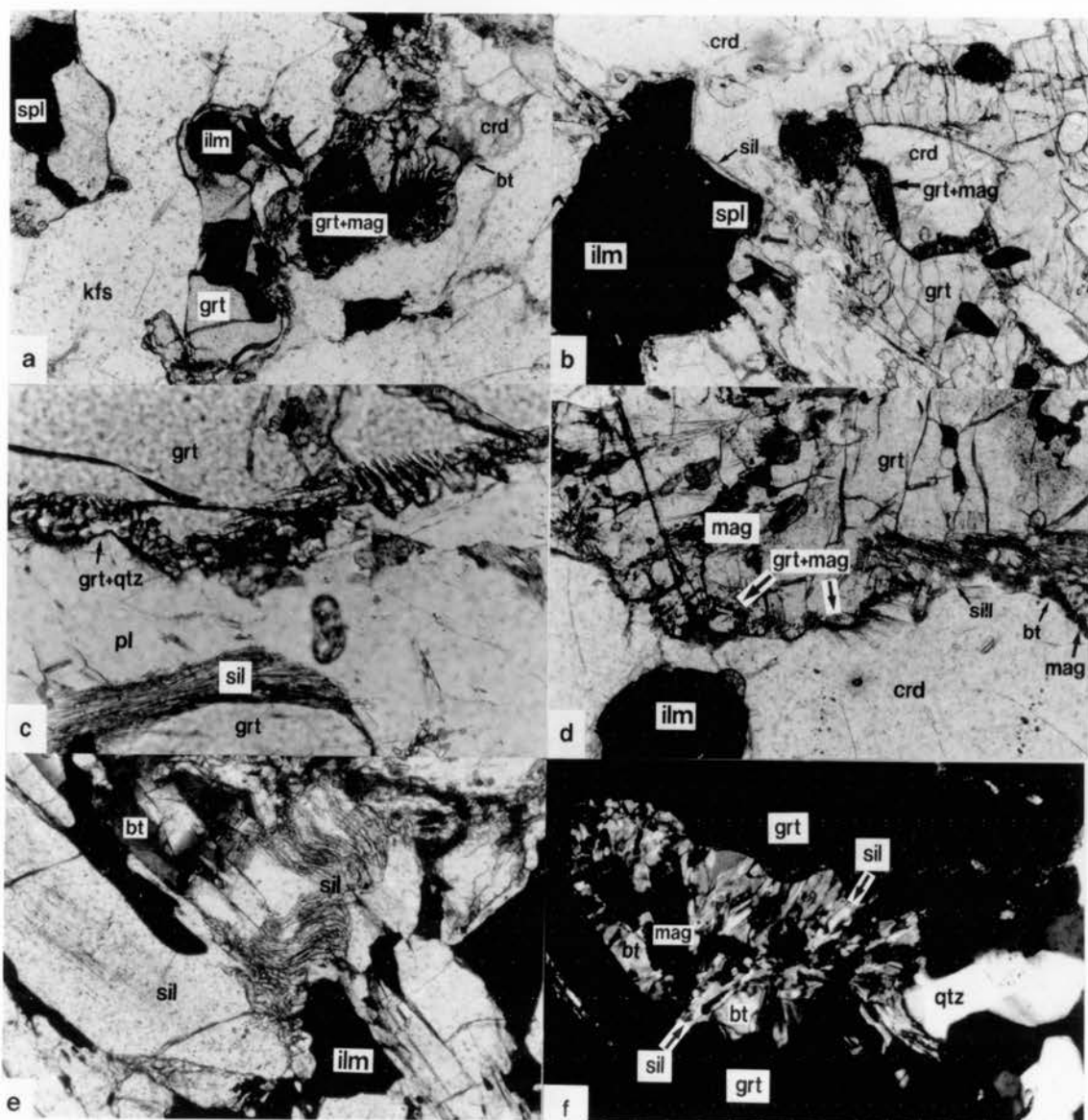


Plate 3.5

- a) Ilmenite and spinel grains partially replaced by symplectic intergrowths of garnet and magnetite. Sample P12, spinel-bearing *metapelitic gneiss*. Plane-polarised transmitted light, width of photo = 0.9mm.
- b) Cordierite adjacent to a garnet porphyroblast is partially replaced by symplectic garnet and magnetite. Thin rims of sillimanite separate cordierite from spinel and ilmenite. Sample P12, spinel-bearing *metapelitic gneiss*. Plane-polarised transmitted light, width of photo = 2.4mm.
- c) Plagioclase replaced by fibrolitic sillimanite and garnet-quartz symplectites, adjacent to garnet grains in quartzofeldspathic leucosome. Sample P13, spinel-bearing *metapelitic gneiss*. Plane-polarised transmitted light, width of photo = 0.5mm.
- d) Coarse-grained garnet overgrown by fine-grained intergrowths of garnet and magnetite. In places, fibrolitic sillimanite, biotite and magnetite partially replace the garnet-magnetite intergrowths where they lie adjacent to cordierite. Sample P12, spinel-bearing *metapelitic gneiss*. Plane-polarised transmitted light, width of photo = 0.9mm.
- e) Coarse-grained sillimanite is overgrown by fibrolitic sillimanite, that is crenulated into an F₄ fold. Fine-grained biotite encloses ilmenite grains. Sample P8, spinel-bearing *metapelitic gneiss*. Plane-polarised transmitted light, width of photo = 0.7mm.
- f) An embayment in a garnet porphyroblast, in which the garnet is replaced by a fine-grained intergrowth of magnetite, sillimanite and green Ti-poor biotite. Sample P13, spinel-bearing *metapelitic gneiss*. Plane-polarised transmitted light, width of photo = 0.5mm.



cordierite or sillimanite. Finely intergrown masses of biotite, magnetite and/or fibrolitic sillimanite enclose ilmenite and garnet grains, and replace larger biotite grains. Cracks in garnet porphyroblasts contain very fine-grained magnetite, green titanium-free biotite, and fibrolitic sillimanite (Plate 3.5f).

Biotite grains up to 0.5mm long define a weak S_3 or S_4 foliation in several samples. Mylonites (S_5) and fine-grained sheets are present in most of the thin sections examined. The latter are typically less than 0.2mm wide, and have offshoots along pre-existing grain boundaries. In sample P15, very fine-grained quartz, K-feldspar, sillimanite, biotite and magnetite occur within and adjacent to an S_5 mylonite. Fibrolitic sillimanite is common near the mylonite, replacing sillimanite, ilmenite, spinel, and K-feldspar along grain boundaries (Plate 3.6a).

Corundum-bearing assemblage

In some samples of spinel-bearing paragneiss, corundum is present in minor amounts as slivers enclosed in ilmenite (Plate 3.3f). Only one sample examined (P11) contains abundant corundum, as blocky crystals up to 1mm across, along with garnet, cordierite, spinel, ilmenite, magnetite, sillimanite, and small amounts of biotite and monazite. Quartz and K-feldspar in the thin section of sample P11 are restricted to a patch of leucosome, which also contains cordierite and small amounts of sillimanite and biotite. Corundum crystals occur in clumps with ilmenite and/or magnetite, that are surrounded by sillimanite crystals up to 0.5mm long (Plate 3.6b). Clots of spinel, magnetite and ilmenite grains up to 2mm across tend to cluster around the edges of garnet porphyroblasts. Garnet grains are equant, up to 1cm across, and contain randomly orientated inclusions of cordierite, biotite and fibrolitic sillimanite.

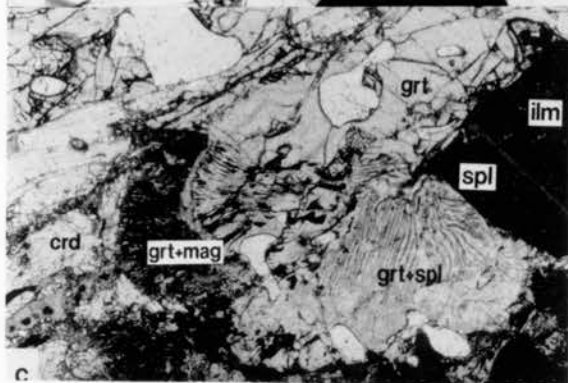
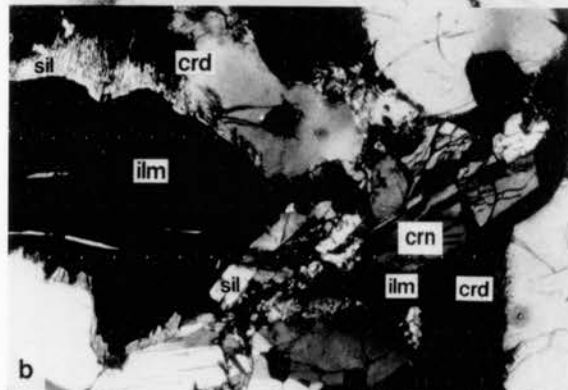
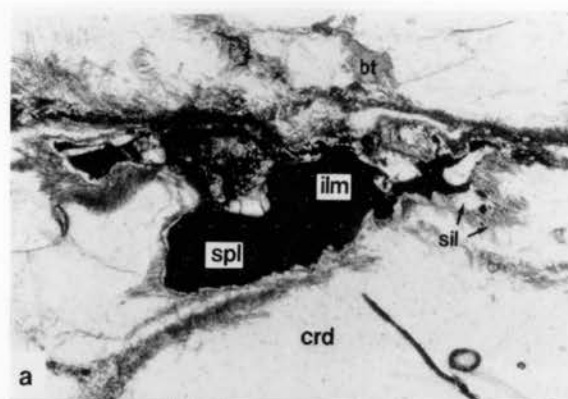
Garnet porphyroblasts adjacent to cordierite are surrounded by coronas up to 2mm thick, composed of symplectic garnet, green spinel and magnetite (Plate 3.6c). Fibrolitic sillimanite replaces cordierite along grain boundaries, and overgrows larger sillimanite, spinel, ilmenite and magnetite grains (Plate 3.6b). Very fine-grained

Plate 3.6

a) Spinel and ilmenite grains enclosed by fibrolitic sillimanite and biotite, that also formed along D₅ mylonitic planes that recrystallise quartz, feldspar and cordierite. Sample P15, spinel-bearing *metapelitic gneiss*. Plane-polarised transmitted light, width of photo = 1.1mm.

b) Coarse-grained corundum, ilmenite and cordierite. Corundum crystals are separated from cordierite by overgrowths of ilmenite. Large ilmenite grains are separated from cordierite by coarse-grained and fibrolitic overgrowths of sillimanite. Sample P11, quartz-free *metapelitic gneiss*. Crossed Nichols, width of photo = 2.5mm.

c) Porphyroblastic garnet adjacent to coarse-grained spinel, ilmenite and cordierite, that is overgrown by splayed symplectites of garnet, magnetite and spinel. Sample P11, quartz-free *metapelitic gneiss*. Plane-polarised transmitted light, width of photo = 1.1mm.



planes less than 0.2mm wide traverse the thin section, and contain cordierite, fibrolitic sillimanite, ilmenite, magnetite and biotite.

3.3 Sequence of Metamorphism

There is sufficient textural and mineralogical evidence in the samples described to define a series of stages of mineral growth. Each stage may represent a unique metamorphic event, or may have been a part of a single continuous event. Constraints on the nature and timing of these stages are considered below, and are summarised, along with the associated mineral assemblages, in Table 3.1.

M₁ involved granoblastic recrystallization in all pre-D₁ lithologies, producing most of the coarse-grained mineral textures in the gneisses. Garnet porphyroblasts contain randomly orientated inclusions in their cores, and inclusions parallel to S₂ in their rims. These are consistent with the garnet grains having grown from before D₁ to after D₂. In metapelitic gneiss, coarse-grained sillimanite and biotite aligned with S₂ are in textural equilibrium with other coarse-grained minerals, and therefore probably grew during the same stage of metamorphism. In spinel-bearing paragneiss, the separation of coarse-grained spinel and quartz by reaction coronas indicate that they were stable during M₁ metamorphism. Coarse-grained biotite in metapelite and in garnet-bearing *charnockitic orthogneiss* is inferred to have grown before and/or during D₂, and is also considered to be part of the M₁ metamorphic assemblage.

M₂ mineral growth is defined by the appearance of coarse-grained reaction textures that partially replace M₁ minerals. In metapelitic gneiss, this stage is represented by the growth of sillimanite, magnetite, garnet, cordierite, and coarse-grained biotite on spinel or ilmenite grains, along with the formation of garnet-magnetite symplectites on M₁ garnet grains. The M₂ event is inferred to have involved the exsolution and replacement of spinel by magnetite. In garnet-bearing orthogneiss, M₂ metamorphism

Table 3.1. Summary of the timing, textures and mineral assemblages associated with the four stages of mineral growth.

<i>Stage</i>	<i>M₁</i>	<i>M₂</i>	<i>M₃</i>	<i>M₄</i>
Timing	Pre-D ₁ to pre-D ₂	Syn-D ₂ to post-D ₂	Late-D ₂ / post-D ₃	Syn-D ₅
Textural Characteristics	Granoblastic crystallisation	Coarse-grained reaction textures	Fine-grained reaction textures	Very fine recrystallisation along and adjacent to mylonites
Assemblages (minerals grown during each stage)				
Metapelite	<i>Quartz-bearing:</i> Qtz -kfs -grt -sil -mag -ilm -bt ±pl ±crd ±spl <i>Quartz-free:</i> Crn -grt -crd -spl -mag -ilm -sil	<i>Quartz-bearing:</i> Sil -mag -grt -bt ±crd <i>Quartz-free:</i> Spl -mag -grt -sil	<i>Both types:</i> Sil -mag -bt ±qtz	<i>Both types:</i> Qtz -kfs -sil -mag -ilm -bt ±pl ±crd
Garnet-pyroxene-bearing orthogneiss	<i>Felsic:</i> Qtz -pl -opx -grt -kfs -ilm ±mag ±bt <i>Mafic:</i> Pl -opx -cpx -grt -qtz -ilm ±hbl ±kfs ±mag	<i>Felsic:</i> Grt -mag -qtz -bt ±ilm <i>Mafic:</i> Grt -mag -qtz -cpx -ilm ±hbl	<i>Both types:</i> Mag -bt	<i>Felsic:</i> Qtz -pl -opx -mag -bt -ilm <i>Mafic:</i> Pl -opx -cpx -qtz -ilm -mag -bt
Post-D₃ felsic dykes			Bt-mag	Qtz -kfs -grt -mag -ilm -bt

is represented by the growth of garnet, with or without quartz, magnetite, ilmenite and clinopyroxene, on pre-existing garnet, pyroxene and ilmenite, and also by the formation of biotite rims on ilmenite and garnet.

Sillimanite and biotite grains that are inferred to have grown during M_2 are aligned parallel to S_2 . M_2 mineral growth occurred before or during D_2 . Inclusion trails in the rims of M_1 garnet porphyroblasts are also aligned with S_2 . In this case, M_2 mineral growth cannot be clearly separated from late-stage M_1 mineral growth. The simplest explanation for these textures invokes M_2 mineral growth as a later stage of M_1 metamorphism.

M_3 mineral growth is characterised by fine-grained reaction textures that partially replace coarser grained minerals. They are distinguished from M_2 reaction textures by their finer grain size (less than 0.3mm), and by the replacement of M_2 minerals. In metapelitic gneiss, fine-grained biotite, magnetite and fibrolitic sillimanite have overgrown larger grains. In the garnet-bearing orthogneisses, fine-grained biotite and magnetite replace garnet, ilmenite and larger biotite grains. Similar textures are seen in post- D_1 orthogneisses, such as the *biotite-pyroxene orthogneiss*, and in post- D_2 and post- D_3 felsic dykes. Hence the metamorphism that produced M_3 mineral growth may have occurred later than D_3 . Alternatively, the reaction textures in pre- D_2 and post- D_2 gneisses may be unrelated, and the formation of M_3 minerals may represent a later stage of M_2 mineral growth.

M_4 textures comprise fine-grained minerals that occur in very narrow planes, which traverse most of the thin sections examined. These planes are accompanied by the recrystallization of host minerals along pre-existing grain boundaries. The planes are also associated with clots of fine-grained biotite, magnetite and fibrolitic sillimanite, which partially replace pre-existing ilmenite, garnet and cordierite. Similar recrystallized planes are also present in post- D_2 and post- D_3 felsic dykes. M_4 mineral

growth is probably related to the formation of D₅ mylonite and pseudotachylite, which occur throughout the Cape Bruce area (see Chapter 2).

Other reaction textures that do not clearly relate to any of the stages described, may have occurred at any time after M₁ mineral growth. The replacement of hornblende and biotite grains by fine-grained intergrowths of orthopyroxene, plagioclase, ilmenite and quartz occurs in pre-D₁ mafic orthogneiss and pre-D₂ *biotite-pyroxene orthogneiss*. In one sample, the proximity of these textures to a M₄ (or D₆) mylonite is consistent with co-development, however, the evidence is ambiguous as the two textures are not always found together.

3.4 Mineral Chemistry

Mineral compositions were determined using the Cameca Camebax SX-50 electron microprobe at the University of New South Wales, with an operating voltage of 15kV and 20nA. Fe³⁺ contents in garnet, spinel, ilmenite, magnetite, cordierite and pyroxene were estimated using charge balance calculations for ideal stoichiometry. Since the Fe³⁺ content of most minerals is small, $X_{Fe} = \text{total Fe} / (\text{total Fe} + \text{Mg})$.

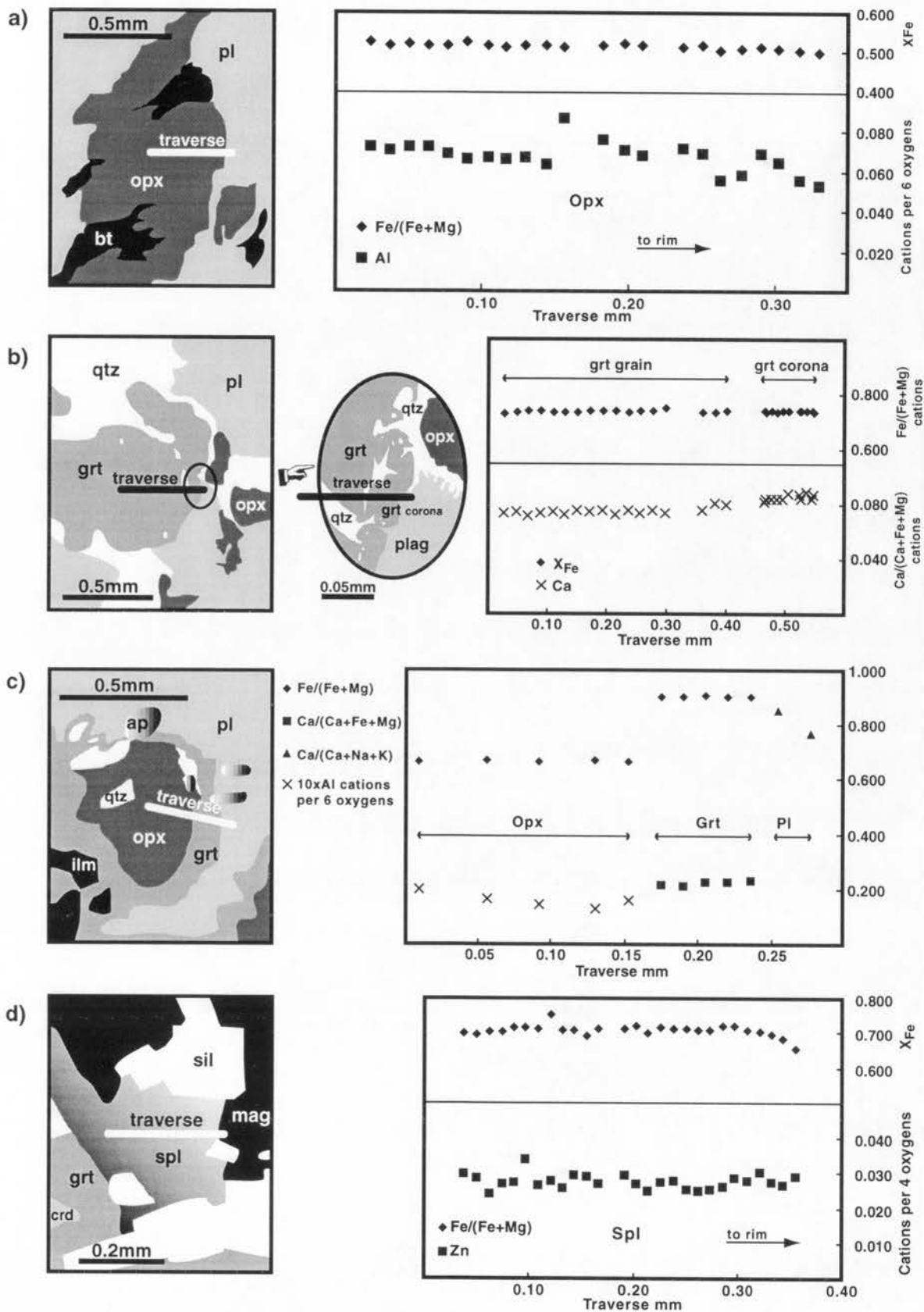
The ranges in composition of almost all of the minerals described below reflect differences in bulk rock composition between the samples analysed, since this variation is larger than the variation within samples and individual grains.

The **orthopyroxene** in the orthogneisses is hypersthene, with a higher X_{Fe} in mafic orthogneiss (0.57-0.71) than in *charnockitic orthogneiss* (0.45-0.53). In *charnockitic orthogneiss*, small amounts of Al are substituted for Si (Al = 0.05-0.13 cations per 6 oxygens). Mn in orthopyroxene is less than 0.02 cations per 6 oxygens. Compositional zoning occurs rarely; for example, a 0.4mm wide grain from sample O4 decreases in X_{Fe} from core (0.53) to rim (0.50) (Fig. 3.1a).

Fig. 3.1. Compositional profiles of mineral grains. $X_{Fe} = \text{Total Fe}/(\text{Fe} + \text{Mg})$.

- a) Core to rim traverse across an orthopyroxene grain in sample O4, garnet-bearing *charnockitic orthogneiss*.
- b) Core to rim traverse across an M_1 garnet grain with an M_2 garnet corona in sample O5, *charnockitic orthogneiss*. The same garnet is shown in Plate 3.1a, and the oval close-up corresponds to the area shown in Plate 3.2b.
- c) Traverse across an M_2 garnet corona between orthopyroxene and plagioclase. Sample O7, garnet-bearing mafic orthogneiss, *proclamation gneiss*.
- d) Traverse across a spinel grain containing magnetite exsolution textures, adjacent to a large magnetite grain. These grains are shown in Plate 3.4a. Sample P13, spinel-bearing *metapelitic gneiss*.

Fig. 3.1



Clinopyroxene in the orthogneisses is augitic, with an $X_{Fe} = 0.45-0.59$. Al is a minor component (0.03-0.07 cations per 6 oxygens). Clinopyroxene grains do not exhibit any compositional zoning.

Amphiboles in mafic orthogneisses are hornblende-rich, with a smaller pargasitic component. X_{Fe} ranges from 0.60 to 0.76. K and Na contents are similar (K = 0.39-0.52 cations per 23 oxygens, and Na = 0.38-0.41 cations per 23 oxygens). Ti content ranges from 0.27 to 0.35 cations per 23 oxygens. There is a small amount of F present in the hydroxyl ion site (F = 0.19 anions per 23 oxygens). Hornblende grown as coronas on pyroxene has a similar composition to coarse-grained hornblende.

Garnet in all samples is almandine-rich. Mafic orthogneiss has a higher X_{Fe} (0.84-0.91) than *charnockitic orthogneiss* (0.74-0.82) and the paragneisses (0.68-0.76). Grossular is a minor component, with the highest Ca contents occurring in mafic orthogneiss (0.56-0.69 cations per 12 oxygens), and lower amounts in *charnockitic orthogneiss* (0.21-0.36 cations per 12 oxygens). Ca in garnet from the paragneisses occurs in very small amounts (0.06-0.09 cations per 12 oxygens). Spessartine is a minor component in all garnet (Mn = 0.03-0.15 cations per 12 oxygens). Ferric iron, calculated from stoichiometry, is less than 0.13 cations per 8 cations.

Garnet porphyroblasts up to 1cm across were analysed, and all showed almost no compositional zoning. A few analysed grains showed a slight increase in Ca content toward their rims (Fig. 3.1b). In *charnockitic orthogneiss*, garnet in coronas on garnet porphyroblasts and other grains is slightly yet consistently higher in Ca content (0.66-0.78 cations per 12 oxygens) and X_{Fe} (0.90-0.91) than garnet porphyroblasts. In mafic orthogneiss and paragneiss, garnet coronas have the same composition as porphyroblastic garnet. An exception is found in paragneiss sample P11, where a garnet corona on a garnet porphyroblast contains much less Ca (less than 0.01 cations per 12 oxygens) than the host garnet (Ca = 0.08-0.09 cations per 12 oxygens).

In *charnockitic orthogneiss*, garnet coronas that separate orthopyroxene and plagioclase may be zoned, with Ca content increasing towards the contact with plagioclase (Fig. 3.1c). In *paragneiss*, garnet coronas that separate spinel or ilmenite from quartz show a slight decrease in X_{Fe} towards the quartz contact.

Plagioclase compositions differ greatly between rock types. *Charnockitic orthogneiss* contains labradorite, in the range An_{45-69} (where An = total Ca/(total Ca + Na)). Plagioclase in mafic orthogneiss has a compositional range of An_{59-88} , but is mostly bytownitic ($An_{0.71-0.80}$). The plagioclase in paragneisses is an oligo-andesine, with a compositional range of An_{18-34} .

K-feldspar is present in the gneisses and in leucosome as microperthitic orthoclase, of the composition Or_{70-84} (where Or = total K/(total K + Na + Ca)).

Biotite is phlogopitic in all samples. The range of values for X_{Fe} is similar in the orthogneisses (0.36-0.45) and the paragneisses (0.26-0.48). Ti in the orthogneisses occurs in similar amounts (0.40-0.65 cations per 22 oxygens) to Ti in the paragneisses (0.37-0.61 cations per 22 oxygens). All biotite contains a significant amount of F in the hydroxyl ion site (F = 0.45-1.26 anions per 22 oxygens). The composition of biotite grains is independent of grain size. Some samples of paragneiss contain blue-green biotite, which is found replacing garnet. This biotite has X_{Fe} = 0.22-0.26 and a Ti content of less than 0.03 cations per 22 oxygens.

Ilmenite contains a small but significant haematite component. The amount of Fe^{3+} present was calculated for perfect stoichiometry. In the orthogneisses, Fe^{3+} contents varied from 0 to 0.14 cations per 2 cations. In sample O10, an ilmenite grain was observed with an overgrowth of symplectic garnet and ilmenite (Plate 3.2a). In this case, the coarse-grained ilmenite contained 0.08 cations of Fe^{3+} per 2 cations, whereas the ilmenite in the symplectite had negligible Fe^{3+} . In the paragneisses, all ilmenite contains a significant amount of Fe^{3+} (0.16-0.21 cations per 2 cations).

Cordierite in the paragneisses has an $X_{\text{Fe}} = 0.20\text{-}0.25$. Calculated Fe^{3+} ranges from 0.06 to 0.09 cations per 11 cations.

Magnetite contains less calculated Fe^{3+} (1.90-1.97 cations per 3 cations) than twice the amount of Fe^{2+} (0.99-1.03 cations per 3 cations), and minor amounts of Cr (0.02-0.04 cations per 4 oxygens).

Spinel is hercynitic ($X_{\text{Fe}} = 0.70\text{-}0.81$), with a significant amount of gahnitic substitution ($\text{Zn} = 0.03\text{-}0.41$ cations per 4 oxygens) that varies greatly between samples and between spinel grains within samples. Chromium content is small ($\text{Cr} = 0.03\text{-}0.10$ cations per 4 oxygens). Calculated Fe^{3+} content varies from 0.03 to 0.09 cations per 3 cations. Iron contents were measured for spinel grains that contained exsolved magnetite. Analyses derived by scanning the electron beam over small and large areas were compared and found to have negligible differences in Fe content. Spinel compositions before exsolution may have contained more Fe^{2+} and Fe^{3+} than measured. In spinel grains adjacent to magnetite, the total Fe content is lowest near the contact with the magnetite (Fig. 3.1d). No zoning was detected in spinel grains with garnet or sillimanite coronas. In sample P11, symplectic spinel in coronas on garnet porphyroblasts has a slightly lower X_{Fe} (0.68-0.71) than coarse-grained spinel ($X_{\text{Fe}} = 0.73\text{-}0.75$).

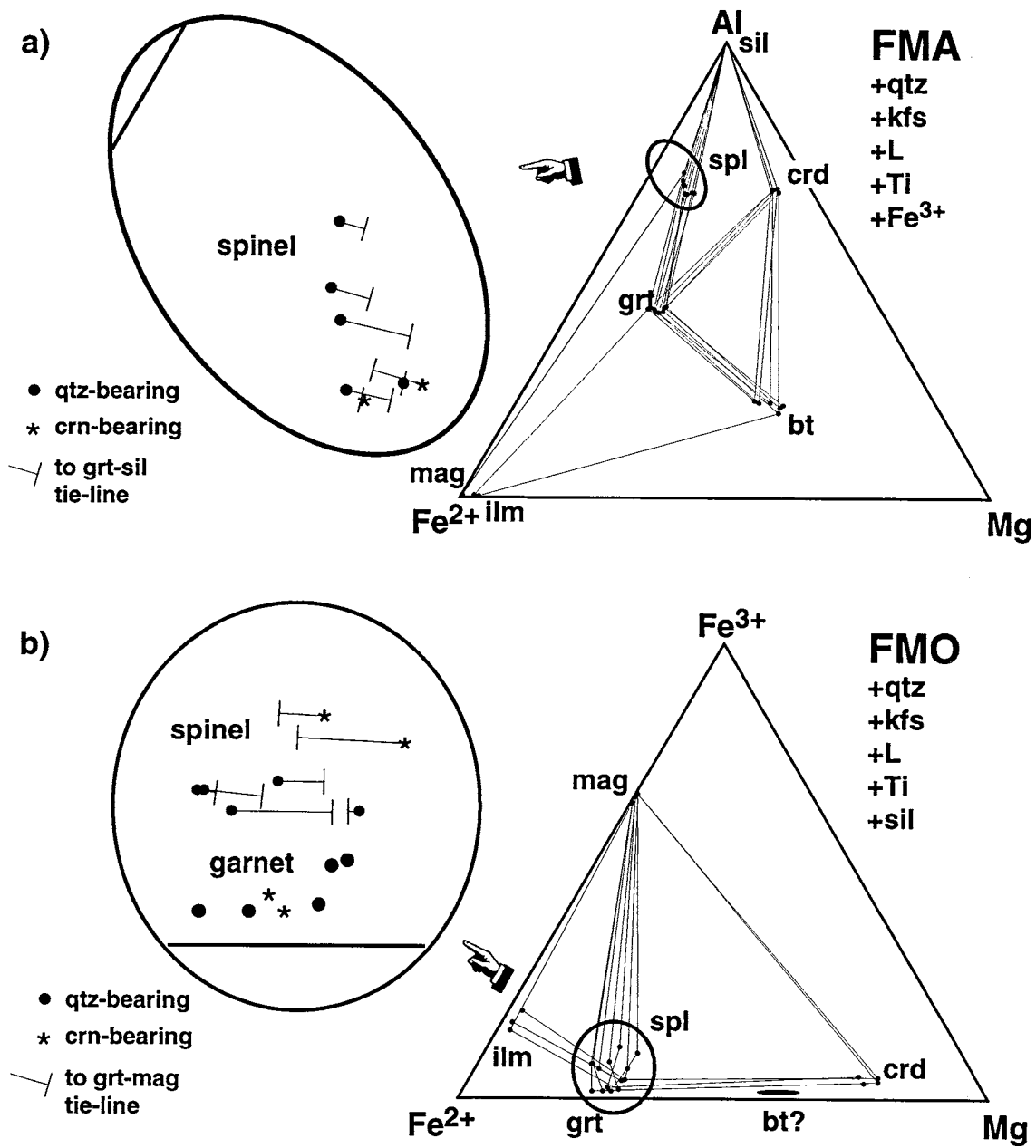
For the purpose of determining the stability of metapelitic mineral assemblages in compositional space, it is important to know whether spinel or garnet is richer in Fe^{2+} relative to Fe^{3+} and Mg contents (see Paragenesis). This is difficult, as the ranges of X_{Fe} values of garnet and spinel overlap. For averaged compositions of garnet and spinel from each sample, Fe^{3+} was calculated using charge balancing for perfect stoichiometry. On an AFM diagram, most spinel compositions lie on the high- Fe^{2+} side of the garnet-sillimanite tie-line (Fig. 3.2a). On an FMO diagram, most spinel lies on the high- Fe^{2+} of the garnet-magnetite tie-line (Fig. 3.2b). In both diagrams,

Fig. 3.2. Compositional diagrams for analysed spinel-bearing *metapelitic gneiss*, showing tie-lines between constituent phases. FeO and FeO_{3/2} were estimated from total analysed Fe by calculating for ideal stoichiometry.

a) FeO-MgO-AlO_{3/2} compositional diagram. Crossing tie-lines between phases indicate that the compositional differences between phase pairs cannot be simply accounted for by shifts in bulk rock composition. The oval close-up shows that, for quartz-bearing *metapelitic gneiss*, spinel compositions have a slightly higher X_{Fe} than the corresponding garnet-sillimanite tie-lines, and slightly less X_{Fe} than the tie-lines for corundum-bearing samples.

b) FeO-MgO-FeO_{3/2} compositional diagram. In this case, for quartz-bearing samples, spinel compositions have a higher X_{Fe} than the corresponding garnet-magnetite tie-lines. For corundum-bearing samples, spinels have a lower X_{Fe} than the corresponding garnet-magnetite tie-lines.

Fig. 3.2



coarse-grained and symplectic spinel from sample P11 (corundum-bearing assemblage) lie on the low-Fe²⁺ side of the appropriate tie-line.

3.5 Geothermobarometry

From the information provided by mineral textures and compositions in the samples analysed, geothermobarometric calculations were used to constrain the pressures and temperatures involved in the metamorphic history. Only M₁ assemblages contained coexisting phases upon which geobarometric calculations could be completed. Geothermometry calculations were completed for M₁ and M₂ assemblages. A variety of geothermobarometric and phase activity models were used; the relative applicability of different calculations to the samples will be discussed below.

3.5.1 Spinel - quartz stability

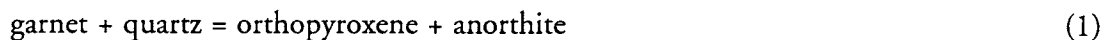
Boundary constraints on metamorphic conditions can be established from a consideration of the phases observed to be in textural equilibrium. In M₁ assemblages in metapelitic gneiss, hercynitic spinel and quartz are inferred to have coexisted in the presence of garnet, sillimanite and/or cordierite. The stability of spinel with quartz has been used as an indicator of high-temperature, low-pressure metamorphic conditions (Vielzeuf, 1983; Waters, 1991). In the compositional system FeO-Al₂O₃-SiO₂ (FAS), the minimum pressures and temperatures at which hercynite and quartz coexist has been experimentally determined at approximately 770°C and 3 kbar (Bohlen *et al.*, 1986). Adding MgO and H₂O to this system contracts the stability range of spinel with quartz towards higher temperatures and pressures (Waters, 1991). Inversely, adding Zn, Cr and Fe³⁺ to the composition of spinel shifts the spinel-quartz stability field to higher pressure and lower temperature conditions. For a zinc content in spinel involving $X_{Zn} = 0.4$, where $X_{Zn} = Zn/(Zn + Fe + Mg +$

Cr), the stability of spinel + quartz extends to below 650°C at low pressures (3-5kbar, Nichols *et al.*, 1992).

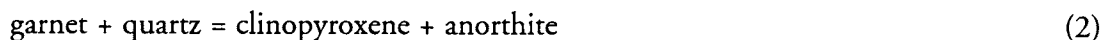
Spinel compositions are highly variable in the samples analysed, with X_{Zn} values up to 0.40 and a significant degree of (later exsolved) magnetite substitution. However, some samples contain spinel grains that are hercynitic and contain as little as 0.03 cations of Zn + Cr + Fe³⁺ per 3 cations, with only a small amount of magnetite exsolution. These hercynitic grains occur in contact with quartz, or are only separated from quartz by thin coronas. The co-existence of hercynite and quartz along with garnet and cordierite in M₁ assemblages provides a minimum temperature constraint on 'peak' metamorphic conditions of about 800°C. This temperature was used in geobarometry calculations for M₁ assemblages.

3.5.2 Orthogneiss assemblages

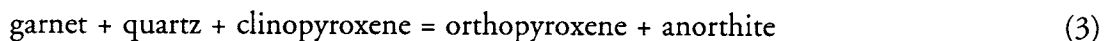
Results of geothermobarometry on mineral assemblages from garnet-bearing orthogneiss are summarised in Tables 3.2 and 3.3. Geobarometric calculations were completed assuming a temperature of 800°C, and are based on calibrated equilibria involving minerals present in the M₁ mineral assemblage. These equilibria, and their associated calibrations, are:



(Newton & Perkins, 1982; Bohlen *et al.*, 1983; Perkins & Chipera, 1985; Moecher *et al.*, 1988; Bhattacharya *et al.*, 1991; Eckert *et al.*, 1991)



(Newton & Perkins, 1982; Moecher *et al.*, 1988; Eckert *et al.*, 1991)



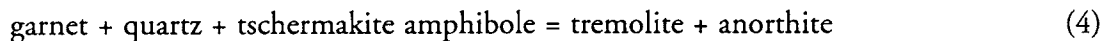
(Paria *et al.*, 1988)

Table 3.2. Pressure and temperature estimates for M₁ equilibrium assemblages in garnet-bearing *charnockitic orthogneiss* (samples O2, O3, O5 and O4). A temperature estimate for a dyke of *biotite-pyroxene orthogneiss* (sample O1) is included for comparison. Def = calculations done using the default garnet mixing model used by the author of each published calibration; AE = using the garnet mixing model of Anovitz & Essene (1987), which is modified from Ganguly & Saxena (1984). 'Average P' and 'Average T' calculations for Powell & Holland (1988) are completed for a water activity of $a_{\text{H}_2\text{O}}=0.5$ and without water and biotite (anhyd). Estimates from the four samples of *charnockitic orthogneiss* are averaged for each calibration. Errors are quoted at two standard deviations (2σ).

Table 3.2. P-T estimates, felsic orthogneiss

Sample Comment Garnet Model	O2		O3		O5		O4		O1		AVERAGES		
	Def	AE	Def	AE	Def	AE	Def	AE	Def	AE	Av	2sd Def	Av AE
Barometry @ 800°C (kb)													
Equilibria phases													
Author													
Grt-Opx	Harley 1984b	Fe-Mg	6.3		2.6		3.3		1.2		3.3	4.3	
Grt-Opx-Pl-Qtz (1)	Bhattacharya et al 1991	Fe	5.8	5.9	6.1	6.2	6.1	6.2	6.3	6.4	6.0	0.4	6.2 0.4
	Bhattacharya et al 1991	Mg	6.7	6.8	6.2	6.4	6.0	6.2	5.6	6.4	6.1	0.8	6.5 0.6
	Bohlen et al 1983b	Fe	5.8	5.6	6.2	6.0	6.2	6.0	5.5	5.9	5.9	0.7	5.9 0.4
	Eckert et al 1991	Mg	5.6	7.5	5.0	7.0	4.7	6.7	3.8	7.1	4.8	1.5	7.1 0.7
	Moescher et al 1988	Fe	7.1	7.1	7.3	7.3	7.4	7.4	7.0	7.0	7.2	0.4	7.2 0.4
	Newton & Perkins 1982	Mg	6.0	7.9	5.4	7.5	5.2	7.1	4.3	7.6	5.2	1.5	7.5 0.7
	Perkins & Chipera1985	Fe	6.9	7.6	6.9	7.9	7.1	7.9	6.8	7.9	6.9	0.2	7.8 0.3
	Perkins & Chipera1985	Mg	6.6	6.9	6.1	6.5	5.7	6.1	5.9	6.6	6.0	0.7	6.5 0.7
	Powell & Holland 1988	H ₂ O	5.9		6.1		6.0		6.3		6.1	1.1	
Average P		2σ			2.3		2.2		2.1				
		anhyd	6.0		6.2		6.2		6.3		6.2	1.3	
		2σ	2.2		2.4		3.0		2.7				
Thermometry (Fe-Mg exchange) @7kb (°C)													
Equilibria phases													
Author													
Grt-Opx	Harley 1984a		700		641		659		578		667	60	
Grt-Opx	Perchuk 1983		502		473		493		512		489	30	
Grt-Bt	Perchuk & Lavrenteva 1983		654		660		635		557		650	26	
Grt-Bt	Ferry & Spear 1978		753		763		710		560		742	56	
Opx-Bt	Sengupta et al. 1990		667		739		681		618	682	696	76	
	Powell & Holland 1988	H ₂ O	810		865		820		785		832	44	
	Average T		2σ		80		104		80				
		anhyd	808		775		733		764		772	110	
		2σ	182		210		248		234				

Table 3.3. Pressure and temperature estimates for M₁ equilibrium assemblages in garnet-bearing mafic orthogneisses from the *proclamation gneiss* (samples O6, O8, O10, O9 and O7). Also included are temperature estimates for an M₂ corona in sample O7 that contains garnet and clinopyroxene. Def = calculations done using the default garnet mixing model used by the author of each published calibration; AE = using the garnet mixing model of Anovitz & Essene (1987), as modified from Ganguly & Saxena (1984). 'Average P' and 'Average T' calculations for Powell & Holland (1988) are done for a water activity of $a_{\text{H}_2\text{O}}=0.5$ and without water, biotite and hornblende (anhyd). Estimates from the five M₁ assemblages are averaged for each calibration. Errors are quoted at two standard deviations (2 σ).



(Kohn & Spear, 1990)

The aluminium solubility garnet-orthopyroxene geobarometer of Harley (1984b) was also applied, along with the THERMOCALC 'Average P' and 'Average T' methods of Powell & Holland (1988), which uses the internally consistent thermodynamic dataset of Holland & Powell (1990). 'Average P' and 'Average T' calculations were undertaken with and without the presence of H₂O and H₂O-bearing minerals in the assemblage. A water activity ($a_{\text{H}_2\text{O}}$) of 0.5 was assumed for assemblages containing H₂O-bearing minerals.

There are significant differences in the pressure estimates obtained. Fe end-member calibrations tended to give higher pressure estimates (around 5-9kbar) than Mg end-member calibrations (around 3-7kbar). The greatest divergence between Fe and Mg results occurred with the Paria *et al.* (1988), Eckert *et al.* (1991) and Newton & Perkins (1982) geobarometers. Harley (1984b) gave consistently low estimates (1.2-6.3kbar). Powell & Holland (1988), calculated with or without water vapour, gave a narrow range of pressure estimates comparable to those obtained for the Fe end-member calibrations (5.6-7.0kbar). The Kohn & Spear (1990) calibration, involving hornblende, gave similar results for mafic orthogneiss assemblages to the calibrations involving pyroxenes (4.7-6.5kbar).

The variance in the results can be partially attributed to the use of different garnet-mixing models in different calibrations. In order to assess the effect of different garnet mixing models on the results, calculations were made for all calibrations using the mixing model of Anovitz & Essene (1987), as modified from Ganguly & Saxena (1984). For Fe end-member calibrations, the results obtained using Anovitz & Essene (1987) mixing did not differ significantly from those using default mixing models. For Mg end-member calibrations, most of the results obtained with Anovitz & Essene

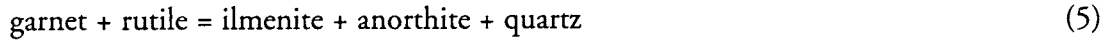
(1987) mixing were higher than those obtained using default mixing models (about 5-9kbar).

Based on the pressures estimates obtained above, geothermometry calculations were completed assuming a pressure of 7kbar. All of the calibrations used, excepting the 'Average T' technique of Powell & Holland (1988), are based on cationic Fe-Mg exchange between garnet, orthopyroxene, clinopyroxene amphibole and/or biotite. Temperature estimates obtained for M₁ assemblages vary between 473 and 840°C. For orthopyroxene-clinopyroxene exchange equilibria, the Wells (1977) calibration produced consistently higher estimates (814-840°C) than the Fonarev & Graphchikov (1983) calibration (678-718°C). All other estimates were less than the 800°C limit assumed for the stability of spinel with quartz. Estimates obtained from Powell & Holland (1988) are notably higher, with 'Average T' results for 7kbar ranging from 764 to 889°C, with errors in the order of $\pm 150^{\circ}\text{C}$ (2σ). 'Average T' estimates for *charnockitic orthogneiss* were slightly lower in anhydrous assemblages than in H₂O-bearing assemblages, and had much larger errors (about $\pm 200^{\circ}\text{C}$, as opposed to $\pm 90^{\circ}\text{C}$ for H₂O-bearing assemblages).

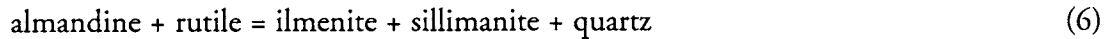
Thermometry was also undertaken for a mineral assemblage containing garnet and clinopyroxene, from an M₂ corona in mafic orthogneiss sample O7. The temperature estimates were only about 40°C lower than those for the M₁ assemblage from the same sample. In sample O1 of a *biotite-pyroxene orthogneiss* dyke (Table 3.2), coexisting orthopyroxene and biotite yielded a temperature estimate of 682°C (Sengupta *et al.*, 1990).

3.5.3 Paragneiss assemblages

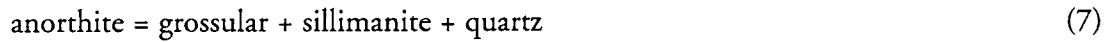
Geobarometry for the spinel-free (Table 3.4) and spinel-bearing (Table 3.5) assemblages was based on a number of equilibria, as listed with their attendant calibrations:



(Bohlen & Liotta, 1986)



(Bohlen *et al.*, 1983a)



(Newton & Haselton, 1981; Koziol & Newton 1988)



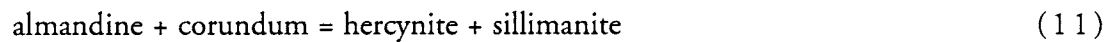
(Nichols *et al.*, 1992)



(Bohlen *et al.*, 1986; Nichols *et al.*, 1992)



(Nichols *et al.*, 1992)



(Shulters & Bohlen, 1989)

The 'Average P' technique of Powell & Holland (1988) was also used, with $a_{\text{H}_2\text{O}}=0.5$ and either quartz or corundum present in the equilibrium assemblage.

The results obtained were more varied than those obtained for orthogneiss assemblages. Pressure estimates obtained from equilibrium (7) were highly divergent, and could not be obtained at all for many plagioclase-bearing assemblages. Pressures

Table 3.4. Pressure and temperature estimates for M_1 equilibrium assemblages in spinel-free paragneiss. Samples P10 and P5 are M_1 assemblages from cordierite-free paragneiss. Samples P2, P3, P6 and P4 are M_1 assemblages from cordierite-bearing paragneiss, with or without plagioclase. For sample P4, P-T estimates were made using the rim and core compositions of a garnet porphyroblast. Def = calculations done using the default garnet mixing model used by the author of each published calibration. 'Average P' and 'Average T' calculations for Powell & Holland (1988) were completed for a water activity of $a_{H_2O}=0.5$. Estimates from the M_1 assemblages are averaged for each calibration. Errors are quoted at two standard deviations (2σ).

Table 3.4. P-T estimates, spinel-free paragneiss. (Def = default model for published barometer; x = no result)

Slide Comment Garnet Model		P10 Crd-free Def	P5 Crd-free Def	P2 Crd-bearing Def	P3 Crd-bearing Def	P6 Crd-bearing Def	P4 Crd-Grt core Def	AVERAGES	
								Av	2sd Def
Barometry @ 800°C (kb)									
Equilibria phases									
Grt-Pl-Sil-Qtz (7)		Newton & Haselton 1981	Ca	5.4		4.5	5.3	4.6	5.1 1.0
		Koziol & Newton 1988	Ca	5.7		4.8	5.6	4.9	5.4 1.0
Grt-Crd-Sil-Qtz (8)		Nichols et al 1992	Fe		6.9	6.8	6.8	6.8	6.8 0.2
		Nichols et al 1992	Mg		4.4	5.0	4.6	4.6	4.8 0.6
Powell & Holland 1988		H ₂ O	5.9	7.4	5.7	6.4	6.8	6.3	6.4 1.2
Average P		2σ	4.0	3.5	1.0	0.8	0.9	0.8	
Thermometry (Fe-Mg exchange) @7kb (°C)									
Equilibria phases									
Grt-Crd		Nichols et al 1992		555	618	607	574	575	589 58
Grt-Crd		Perchuk & Lavrenteva 1983		527	577	572	543	558	555 48
Grt-Bt		Perchuk & Lavrenteva 1983	632	712	641	633	591	628	632 92
Grt-Bt		Ferry & Spear 1978	705	882	718	705	615	679	704 202
Powell & Holland 1988		H ₂ O	729	736	754	760	737	755	745 25
Average T		2σ	78	118	58	42	38	42	

Table 3.5. Pressure and temperature estimates for M_1 equilibrium assemblages in spinel-bearing paragneisses. Assemblages from samples P16 and P13 contain cordierite, plagioclase and traces of rutile. Samples P8, P12 and P1 contain cordierite without plagioclase, and samples P7 and P9 contain plagioclase without cordierite. The M_1 assemblage from sample P11 contains corundum instead of quartz. For comparison, P-T estimates for sample P9 were done using the rim and core compositions of a garnet porphyroblast. Def = calculations done using the default garnet mixing model used by the author of each published calibration. 'Average P' and 'Average T' calculations for Powell & Holland (1988) were completed for a water activity of $a_{H_2O}=0.5$. Results are shown for spinel-bearing (+Spl), spinel-free (-Spl), and corundum-bearing/quartz-free (+Crn) assemblages. Estimates from the M_1 assemblages are averaged for each calibration. Errors are quoted at two standard deviations (2σ).

estimated from equilibria (9) and (10) were also varied, especially for assemblages containing Zn-rich spinel (P16 and P7).

Most of the estimates fall within the range 4-7kbar. The Nichols *et al.* (1992) calibration of equilibrium (8) yielded consistent results for mineral compositions from spinel-free and spinel-bearing assemblages. Estimates for Mg end-member activities (4.4-5.1kbar) were lower than estimates for Fe end-member activities (6.6-7.2kbar). This was also the case for the Nichols *et al.* (1992) calibration of equilibrium (10).

'Average P' calculations (Powell & Holland, 1988) produced results with a small range of mean values (5.6-7.4kbar), although estimates for samples P16 and P8 could be obtained only when calculated without spinel. The range of estimates for spinel-free assemblages was similar to that for spinel-bearing assemblages.

For the sake of comparison, pressure estimates for sample P9 were calculated for the subtly different core and rim compositions of a garnet porphyroblast. The estimates do not differ significantly from each other. Estimates for sample P11, which contains corundum as an M_1 phase instead of quartz, do not differ greatly from other estimates.

Temperature calculations for the paragneiss assemblages were completed assuming $P=7$ kbar. Estimates vary within the range 514-882°C, with the Ferry & Spear (1978) garnet-biotite geothermometer yielding consistently higher results than the other calibrations. Almost all of the estimates are much lower than the 800°C limit inferred from the stability of spinel with quartz.

The 'Average T' method of Powell & Holland (1988) yielded higher estimates than the other methods, with mean values ranging from 729 to 809°C. These results are similar to those obtained using orthogneiss assemblages. The errors on 'Average T' estimates are smaller for the paragneisses assemblages (± 40 -120°C, 2σ) than for the orthogneiss assemblages.

3.5.4 Powell & Holland (1988) 'Average P-T' estimates

The THERMOCALC approach of Powell & Holland (1988) includes an 'Average P-T' technique which, unlike the calibrations above, does not involve an assumed pressure or temperature. However, the presence or absence of water must be assumed. The low abundance of hydrous minerals (biotite, hornblende) in M_1 assemblages implies that these rocks are water-undersaturated. 'Average P-T' estimates were therefore derived for a variety of water activities in the range $0.9 > a_{H_2O} > 0.1$. Water-free calculations were also undertaken for garnet-bearing orthogneisses, as biotite and/or hornblende are not always present in the M_1 assemblages of these samples. The results quoted in Table 3.6 are given for $a_{H_2O} = 0.5$, as well as for water activities that produced the best statistical fit in the estimates (see Appendix 3.4).

The estimates obtained are strongly dependant on the water activity chosen. In all samples, decreasing the water activity tends to lower both pressure and temperature estimates. Estimates from orthogneiss samples for anhydrous M_1 assemblages are typically 1-2kbar and 150-250°C lower than those for $a_{H_2O} = 0.5$. For orthogneiss samples, results with the best fit were obtained from water-free assemblages; however, the errors on these results are often larger than for water-bearing estimates, and there are fewer constraining reactions (see Appendix 3.4). Best fits for water-bearing assemblages were mostly obtained at $a_{H_2O} = 0.5$ for *charnockitic orthogneiss* and at $a_{H_2O} = 0.1$ for mafic orthogneiss. This is consistent with the abundant presence of biotite in the former assemblages and the absence of biotite in the latter. Best statistical fits for paragneiss assemblages were obtained for a wide variety of water activities ($a_{H_2O} = 0.1$ -0.9).

For $a_{H_2O} = 0.5$, M_1 assemblages from orthogneiss and paragneiss samples yielded similar P-T estimates. Estimates for orthogneiss assemblages range from $P=5.2$ to 6.1kbar and from $T=764$ to 819°C, with large 95% confidence errors ($P \pm 2.9$ -5.0kbar and $T \pm 113$ -384°C). Pressure estimates from paragneiss assemblages ranged from

Table 3.6. Pressure and temperature estimates using the ‘Average P-T’ approach of Powell & Holland (1988). For assemblages containing hydrous minerals (biotite, hornblende), results are shown for $a_{\text{H}_2\text{O}}=0.5$, $a_{\text{H}_2\text{O}}=0$, and for the value of $a_{\text{H}_2\text{O}}$ that produced the result with the best statistical fit (bf). ‘Average P-T’ errors are quoted with 95% confidence. Estimates for each calibration are averaged separately for samples of garnet-bearing *charnockitic orthogneiss*, garnet-bearing mafic orthogneiss, spinel-free paragneiss and spinel-bearing paragneiss. Errors on averages are quoted at two standard deviations (2σ).

a) Estimates for M_1 equilibrium assemblages in garnet-bearing orthogneisses. Assemblages are the same as those used in Tables 3.2 & 3.3.

b) Estimates for M_1 equilibrium assemblages in garnet-bearing paragneisses. Assemblages are the same as those used Tables 3.4 & 3.5. Results for spinel-bearing samples are shown for assemblages with (+spl) and without (-spl) spinel.

Table 3.6a. Average P-T estimates, garnet-bearing orthogneiss.

(bf = best fit for $1 > a_{H_2O} > 0$)

Sample Comment	O2		O3		O5		O4		AVERAGES		
	Grt-charnockitic og	Grt-charnockitic og	Grt-charnockitic og	Grt-charnockitic og	Grt-charnockitic og	Grt-charnockitic og	Grt-charnockitic og	Grt-charnockitic og	Av	2sd	Av 2sd Best Fit
a_{H_2O}	0.5	0	0.5	0	0.5	0.2	0	0.5	0.1	0	0
T (°C)	788	794	819	668	806	700	562	765	639	553	795
95%con.	113	298	151	338	142	113	254	108	245	283	644
P (kbar)	5.7	5.8	6.1	5.0	5.9	4.9	4.1	5.9	4.6	4.0	5.9
95%con.	2.9	4.1	3.8	4.1	3.8	3.4	2.9	2.6	3.6	3.1	0.3
fit	0.81	1.00	1.25	1.19	1.23	1.19	0.63	0.99	1.32	0.96	5.7
	bf		bf		bf		bf		+kfs; bf -kfs; bf -bt		

Sample Comment	O6		O8		O10		O9		AVERAGES		
	Grt-hbl-bt mafic og	Grt-hbl mafic og	Grt-hbl mafic og	Grt-hbl mafic og	Grt-hbl mafic og	Grt-hbl mafic og	Grt-mafic	Grt-mafic	Av	2sd	Av 2sd Best Fit
a_{H_2O}	0.5	0.1	0	0.5	0.1	0	0	0	0.5	0	0
T (°C)	735	579	506	800	676	493	764	671	766	65	642
95%con.	134	70	149	384	202	163	283	170	541	170	109
P (kbar)	5.9	4.3	3.6	6.1	4.9	3.2	5.2	4.5	5.7	0.9	4.6
95%con.	2.9	1.7	1.9	5.0	2.6	1.9	3.8	2.4	1.9	1.9	0.6
fit	1.46	0.90	0.83	1.51	0.88	0.45	1.39	0.86	0.49	0.47	3.6
	bf		bf		bf		bf		bf		

Table 3.6b. Average P-T estimates, garnet-bearing paragneiss.

(bf = best fit for $1 > a_{H_2O} > 0$)

Sample Comment	P10		P5		P2		P3		P6		P4		AVERAGES		
	Crd-free	Crd-free	Crd-free	Crd-free	Crd-free	Crd-free	Crd-free	Crd-free	Crd-free	Crd-free	Crd-free	Crd-free	Av	2sd	Av 2sd Best Fit
a_{H_2O}	0.5	0.7	0.5	0.9	0.5	0.3	0.5	0.3	0.5	0.1	0.5	0.3	0.5	0.5	0
T (°C)	718	752	734	799	707	657	746	694	728	591	724	673	726	27	694
95%con.	103	113	190	158	142	122	118	101	103	77	122	103	5.6	1.0	1.8
P (kbar)	5.0	5.4	6.2	7.0	5.2	4.6	5.9	5.3	6.0	4.5	5.5	4.9	5.3	1.0	5.3
95%con.	4.1	4.3	7.2	5.8	1.7	1.4	1.4	1.2	1.4	1.0	1.4	1.2	1.1	1.0	1.2
fit	0.64	0.63	1.81	1.34	1.34	1.27	1.06	0.98	0.85	0.23	1.11	1.03	1.11	0.99	bf
	bf		bf		bf		-spl		-spl; bf		bf		bf		

Sample Comment	P16		P13		P8		P12		P1		P7		P9		P11		AVERAGES		
	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Spl-Pl-Crd-Rt	Av	2sd	Av 2sd Best Fit
a_{H_2O}	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0
T (°C)	773	755	730	734	769	704	720	709	742	734	672	709	734	672	709	734	728	61	737
95%con.	132	118	118	130	137	96	98	101	106	149	242	240	149	242	240	240	5.3	1.6	5.5
P (kbar)	6.1	6.3	5.8	5.2	5.6	5.5	3.8	5.5	5.9	4.9	5.0	5.9	4.9	5.0	5.9	5.9	5.3	1.6	1.6
95%con.	1.7	1.4	1.4	1.7	1.7	1.2	2.4	4.3	4.6	4.1	1.9	1.9	4.1	1.9	1.9	1.9	1.1	0.99	1.1
fit	1.21	1.19	1.17	1.26	1.22	0.89	0.68	0.16	0.09	1.34	1.10	0.99	1.34	1.10	0.99	1.10	1.1	0.99	1.1
	-spl; bf		bf		bf		+spl; bf		-spl		-spl; bf		+spl; bf		+cm		+cm; bf		

$P=4.9\text{--}6.3\text{ kbar}$, and temperatures from $T=672\text{--}773^\circ\text{C}$, most with smaller 95% confidence errors than for the orthogneiss assemblages ($P\pm 1.2\text{--}7.2\text{ kbar}$, $T\pm 96\text{--}242^\circ\text{C}$). The most precise results were obtained using M_1 assemblages from cordierite-bearing and spinel-bearing paragneisses (approx. $P=5.5\pm 1.5\text{ kbar}$ and $T=730\pm 120^\circ\text{C}$).

For M_1 assemblages from all sample types, pressures obtained using the 'Average P-T' method have slightly lower mean values than those obtained using the 'Average P' method. For each sample, the 'Average P-T' result that were obtained assuming $a_{H_2O} = 0.5$ lies within error of the 'Average P' estimate. Temperatures estimated by the 'Average P-T' method assuming $a_{H_2O} = 0.5$ are similar to those obtained by the 'Average T' method.

3.5.5 Discussion of geothermobarometric results

There is significant variation between the P-T estimates described above. This variation can be attributed a number of factors, including the composition of minerals in the assemblages, and the disturbance of equilibrium assemblages by later processes. The reliability of estimates can be assessed by a consideration of these factors.

One potential source of error in the calibrations is the use of end-member components with low concentrations. In the orthogneiss samples, inferred M_1 assemblages include garnet that has a composition of approximately $\text{Alm}_{73}\text{Pyp}_{17}\text{Gro}_7\text{Sps}_3$ for *charnockitic orthogneiss* and $\text{Alm}_{68}\text{Pyp}_9\text{Gro}_{20}\text{Sps}_3$ for mafic orthogneiss (Alm = almandine, Pyp = pyrope, Gro = grossular, Sps = spessartine). Due to low pyrope content of garnet, especially in that from mafic orthogneisses, Mg-end-member estimates are more sensitive to the method used to calculate activities than Fe-end-member estimates. This sensitivity is indicated by the differences in pressure estimates obtained using the garnet mixing model of Anovitz & Essene (1987), as opposed to those used in the references for each calibration. In Fe-rich samples, such as the orthogneisses used in

this study, Fe end-member calibrations are likely to be more reliable than their Mg counterparts (Perkins & Chipera, 1985).

In pressure estimates for paragneiss samples that contain garnet with a composition of approximately $\text{Alm}_{66}\text{Pyp}_{30}\text{Gro}_2\text{Sps}_2$, Mg-end-member estimates also diverge significantly from their Fe counterparts. Due to the Fe-rich compositions of garnet and spinel ($X_{\text{Fe}} = 0.70\text{--}0.81$), Fe end-member calibrations are likely to be more reliable than their Mg counterparts. Garnet has a very low grossular content, and this is a possible cause of the scattered results obtained from calibrations of equilibrium (7).

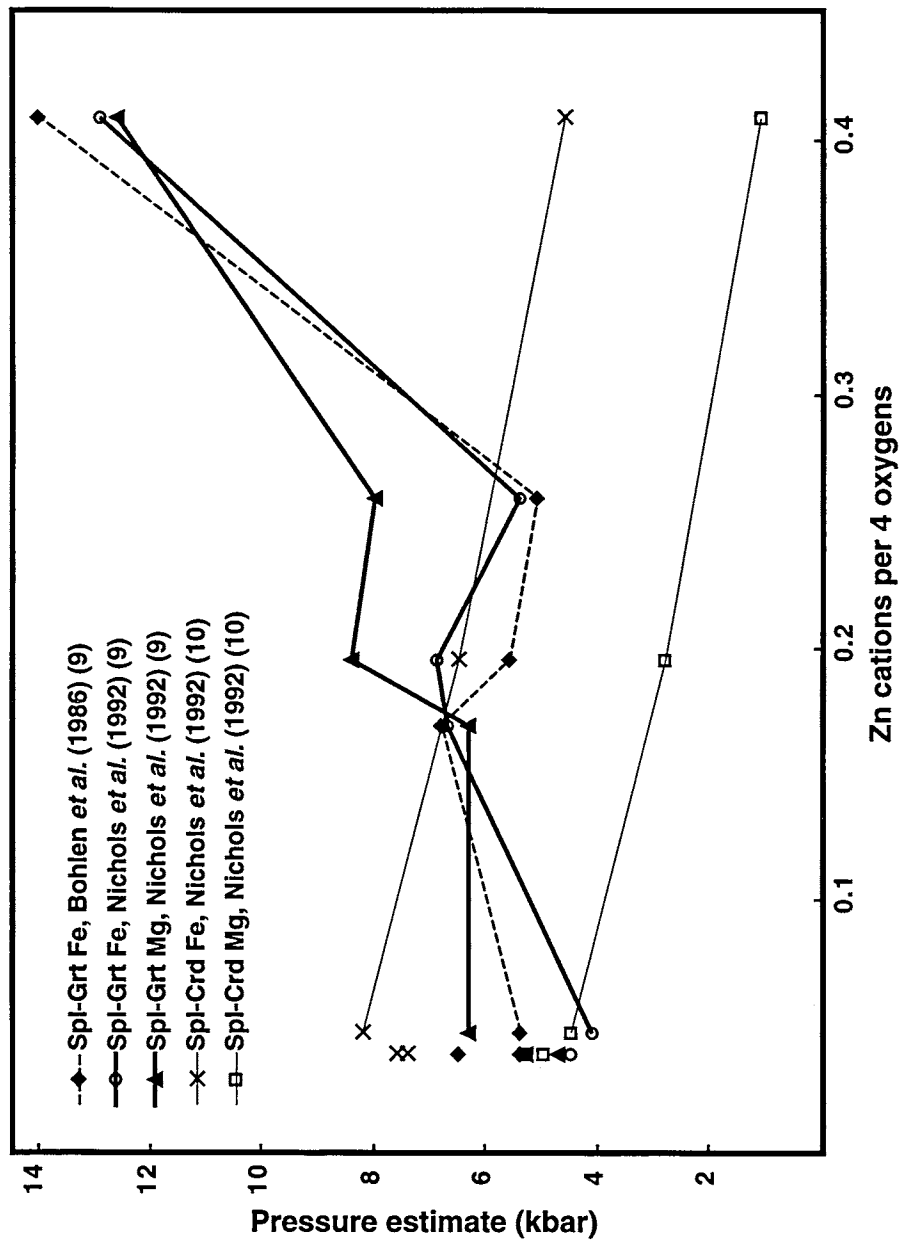
The presence of Zn has an effect on the pressure estimates obtained from calibrations involving spinel. Higher Zn contents in spinel tend to produce lower estimates for calibrations involving equilibrium (10), and higher estimates for calibrations involving equilibrium (9) (Fig. 3.3). This divergence reflects the difficulty of determining the activities of components in zincian spinel.

There is an abundance of reaction textures in the samples used for geothermobarometry. These textures show that the M_1 'peak' assemblage was disturbed by changes in ambient conditions, especially towards lower temperatures than those present during M_1 (see section 3.6, Paragenesis).

Reaction textures preserved in the samples indicate that post- M_1 changes have taken place, particularly at lower temperatures than those present during M_1 mineral growth. The compositions of minerals inferred to belong to M_1 assemblages are likely to have been disturbed by post- M_1 changes. Fe and Mg are very susceptible to re-equilibration at lower temperatures, either during cooling from peak temperatures or by reheating during later metamorphic events. In the samples analysed, minerals such as garnet and orthopyroxene show little or no Fe-Mg zoning. The lack of zoning may be the result of mineral growth during M_1 under constant P-T conditions. Alternatively, Fe-Mg exchange may have occurred after mineral growth, under

Fig. 3.3. Pressure estimates derived from calibrations of spinel-involving equilibria (9) and (10), versus the zinc content of the spinel used in the calculations. There is a loose correlation between zinc content and divergence between estimates using equilibrium (9) and equilibrium (10).

Fig. 3.3



different conditions to those present during M_1 . The lack of Fe or Mg zoning in garnet porphyroblasts that have weak Ca zoning (Fig. 3.1a) can be attributed to the post- M_1 re-equilibration of Fe and Mg.

Temperatures of 600-700°C given by geothermometry calculations probably represent closure temperatures of Fe-Mg exchange between minerals, rather than M_1 conditions. Similar temperature estimates were derived from M_2 garnet-clinopyroxene thermometry and from the *biotite-orthopyroxene orthogneiss* dyke, which is consistent with the possibility that Fe-Mg exchange continued into the D_2 event.

Fe-Mg re-equilibration is also a likely source of error in some pressure estimates. Calibrations of equilibria (8), (9) and (10) rely entirely on the activity of Fe and Mg end-members in garnet, cordierite and/or spinel. This makes the pressures estimates from these equilibria susceptible to Fe-Mg cationic exchange that may have occurred after M_1 mineral growth. The Harley (1984b) aluminium solubility geobarometer produced scattered results which vary with the Al content of orthopyroxene analyses. This barometer is very susceptible to Fe-Mg cationic re-equilibration, as described by Fitzsimons & Harley (1994) and Bégin & Patterson (1994). Techniques described by these authors to 'recover' peak P-T estimates from the results of the Harley (1984a) thermometer and the Harley (1984b) barometer were attempted with several of the samples analysed, but the results failed to account for the variability and low values of these estimates.

The THERMOCALC method of Powell & Holland (1988) has been touted as a more reliable approach to P-T estimation than directly calibrated equilibria, since it make use of a number of equilibria to obtain an averaged result with errors. 'Average P' estimates, calculated with and without water, compare favourably with those obtained by most other methods. The 'Average T' method gave higher temperature estimates than most of the other calibrations. This may be attributed to the use in the

'Average T' method of equilibria that do not involve Fe-Mg exchange (see Appendix 3.4). Consequently, the 'Average T' results are the most reliable estimates of temperatures during M_1 .

'Average P-T' estimates are for the most part lower in pressure and temperature than 'Average T' estimates (at an assumed $P=7\text{kbar}$) and 'Average P' estimates (at an assumed $T=800^\circ\text{C}$). This is particularly the case for best fit results with low water activities ($a_{\text{H}_2\text{O}} < 0.5$). The strong dependence of the estimates on assumed water activity is due to the dominance of water-bearing equilibria in the reaction bundles (see Appendix 3.4). Large errors involved with these calculations are caused by equilibria that involve end-members with small activities, such as grossular, Ca-tschermakitic pyroxene, and pargasite. These calculations, particularly the 'Average T' and 'Average P-T' methods, are also susceptible to the effects of cationic exchange, since Fe and Mg end-member activities are used in the equilibria.

Metamorphic conditions during M_1 can now be inferred. Pressure estimates derived from the calibrations of Fe-end-member equilibria and from the 'Average P' method of Powell & Holland (1988) mostly fall within a range of 6-7kbar. The 'Average T' method provides a useful lower estimate of M_1 temperatures at about 750°C . These P-T estimates are consistent with the stability of spinel with quartz in M_1 assemblages. Water activity during M_1 was dependant on the rock type undergoing metamorphism, but was in most situations greater than zero.

The conditions of M_2 and M_3 mineral growth are not constrained by geobarometry. Temperatures estimated from Fe-Mg exchange for M_2 assemblages (approximately 650°C) were comparable to those for M_1 assemblages. These temperatures are probably that of the closure of Fe-Mg exchange in the minerals analysed. Biotite, orthopyroxene and garnet were also produced during M_3 , and had similar compositions to minerals present in M_2 and M_1 assemblages in the same sample. This

provides evidence that temperatures in excess of 650°C were present during the M₁, M₂ and M₃ events.

3.6 Paragenesis

3.6.1 M₁ assemblages

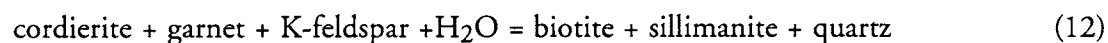
The production of M₁ assemblages involved deformation, melting and the recrystallization of pre-D₁ fabrics, to such an extent that any evidence for earlier events was obliterated. As all textures produced after M₁ metamorphism modified earlier fabrics to a relatively minor degree, M₁ conditions represent the 'peak' of metamorphism. Geothermobarometry provided a range of P=6-7kbar and T≥750°C for the conditions at which M₁ assemblages grew. These estimates are slightly higher than those obtained from similar 'peak' assemblages from the Colbeck Archipelago (P=5-6kbar at T≥750°C, White & Clarke 1993).

The role of water in the M₁ event is demonstrated by the production of leucosome in most pre-D₁ lithologies. Large amounts of leucosome are present in paragneiss, where it is both deformed into and transgresses across S₁ and S₂ foliations. The leucosome provides evidence for a large degree of partial melting during M₁ and M₂. Partial melting would have been responsible for the dehydration of pre-M₁ mineral assemblages. In similar lithologies described at the Colbeck Archipelago, White & Clarke (1993) distinguished 'peak' metamorphic assemblages (called M₁ by White, 1993) by the lack of hydrous minerals, and suggested that most of the biotite in metapelitic gneiss was formed by the hydrous retrogression of M₁ assemblages during D₂ (M₂; White, 1993). In paragneiss samples from Cape Bruce, coarse-grained biotite is present and appears to be in textural equilibrium with other minerals in M₁ assemblages. Despite the large degree of partial melting in the paragneisses, biotite may have persisted at 'peak' temperatures of over 750°C. The stability of biotite at

such temperatures is consistent with the high titanium and fluorine contents of analysed grains (Patino Douce, 1993; Le Breton & Thompson, 1988).

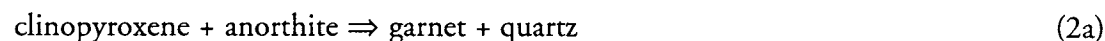
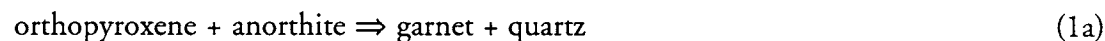
3.6.2 M₂ mineral assemblages

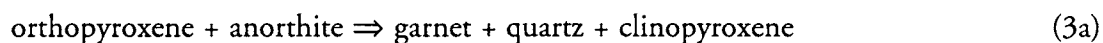
M₂ mineral growth involved a change in metamorphic conditions from those that accompanied M₁ growth. M₂ occurred either between the D₁ and D₂ events or during the D₂ event. In the Colbeck Archipelago, White (1993) attributed the breakdown of spinel with quartz to post-M₁ cooling (M_{1b}), and the growth of garnet, sillimanite and biotite to a separate episode of hydrous retrogression during D₂ (M₂; White, 1993). White (1993) attributes the appearance of biotite to the hydration reaction:



In this study, the growth of garnet in reaction textures is grouped together with coronas on spinel as products of a single (M₂) event. This is done for three reasons: 1) garnet grown as a reaction mineral is more abundant and coarser-grained than reaction textures involving the growth of biotite; 2) garnet coronas do not contain intergrown biotite, but are occasionally overgrown by fine-grained biotite; and 3) the biotite-forming reaction (12) suggested by White (1993) consumes garnet, rather than producing it.

The development of M₂ assemblages can be attributed to reactions caused by the crossing of some of the equilibria already described (see Geothermobarometry). Reactions causing mineral growth in garnet-bearing orthogneisses during M₂ include:

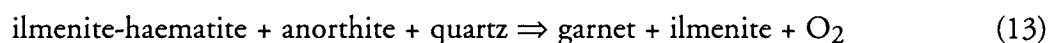




Reactions during M_2 in quartz-bearing metapelitic gneisses include:



All of these reactions, with the possible exception of (10a), cross equilibria that have positive gradients in P-T space, in an up-pressure/down-temperature direction. The overgrowth of garnet-ilmenite symplectites on haematite-containing ilmenite grains can be accounted for by the reaction:



In this reaction, the haematitic component of M_1 ilmenite is being consumed and excess ilmenite is recrystallized in the symplectite. Ellis & Green (1985) observed similar reaction textures involving ilmenite in basic gneiss from Enderby Land, and suggested that reaction (13) occurred in response to cooling.

The limited development of M_2 textures suggests that either the kinematics of mineral growth were slower than those during M_1 growth, and/or that less time was available for M_2 growth. Both possibilities are consistent with lower temperatures having prevailed during M_2 . The similar grossular contents in M_2 and M_1 garnets is consistent with growth having occurred under comparable pressures.

The alignment of some M_2 minerals by S_2 constrains this metamorphic event to having occurred before the end of the D_2 deformation. As some M_1 assemblages grew after D_1 , it is likely that M_2 mineral growth represents retrograde re-equilibration directly after the peak of metamorphism. This corresponds to the M_{1b} stage of metamorphism defined by White (1993).

Abundant amounts of biotite were also produced during M_2 , indicating an increase in water activity in the gneisses after M_1 . The most likely source of water for this

increase is the crystallization of leucosome. Leucosome in *metapelitic gneiss* is both deformed into and transgresses across S_1 and S_2 foliations, indicating that it crystallized before D_2 and after D_2 . Biotite-rich gneiss with an S_2 foliation is commonly observed adjacent to leucosome.

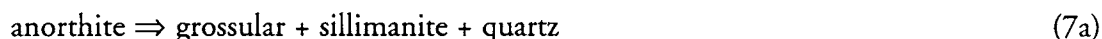
Hydrous retrogression of gneisses in the Colbeck Archipelago was considered by White (1993) to have involved water derived from *biotite $\pm$ pyroxene gneiss*, an 'auto-retrogressed charnockitic gneiss' (*sic*) that was intruded between D_1 and D_2 . This unit has been correlated with the *biotite-pyroxene orthogneiss* of this study (Chapter 2). At Cape Bruce, dykes of *biotite-pyroxene orthogneiss* are enriched in biotite adjacent to migmatized paragneiss. Hydration of the orthogneisses occurred before or during D_2 , and was probably caused by the release of water from solidifying partial melts in the paragneiss. There is little evidence of a causal link between biotite growth in *metapelitic gneiss* and the emplacement after D_1 of water-rich orthogneisses, such as the *biotite-pyroxene orthogneiss*.

3.6.3 M_3 assemblages

The third stage of mineral growth is defined by fine-grained reaction textures and recrystallization of previously grown grains. The minor degree of recrystallization does not significantly change the mode of minerals present; so that an increase in water activity need not be invoked for this event. It is not clear if these textures represent a single well-defined episode of re-equilibration, or whether recrystallization occurred in different rocks at different times during the Rayner Structural Episode. In *metapelitic gneisses* from the Colbeck Archipelago, the formation of fine-grained intergrowths of sillimanite, biotite and quartz on earlier sillimanite coronas were attributed by White (1993), to M_2 metamorphism during the D_2 event (*e.g.* Fig. 3.1b in White, 1993). White also related minor recrystallization and retrogression in gneisses to the intrusion of the Mawson Charnockite during the D_4 event. In the

gneisses of Cape Bruce, similar syn-D₄ recrystallization and growth of biotite is seen, although there is no clear causal connection between this and M₃ mineral growth.

M₃ reaction textures include the breakdown of garnet to biotite and magnetite and of cordierite to fibrolitic sillimanite. These reactions can be accounted for by a change in metamorphic conditions in a multi-component chemical system (see section 3.6.5). In the leucosome of sample P13 (Plate 3.5c), the formation of garnet can be attributed to the reaction:



This reaction texture is attributed to M₃ metamorphism, as it is very fine-grained, and was grown in solid leucosome that intersects the S₂ foliation in the sample. The reaction involves crossing equilibrium (7) down-temperature and/or up-pressure. The very fine-grained nature and limited extent of M₃ reaction textures is consistent with formation under conditions with slower growth kinematics (*i.e.* lower temperatures), and/or briefer periods, than those that governed M₁ and M₂ mineral growth.

The presence of M₃ recrystallization in samples of all pre-D₄ lithologies can be explained as one of two events: 1) a pulse of recrystallization after D₃ and before or during D₄; or 2) limited recrystallization (at less than M₂ temperatures and at constantly elevated pressures) from the D₂ event to the D₄ event. The possibility that elevated metamorphic conditions were maintained though the period of time between the D₂ and D₄ events is supported by the growth of garnet in post-D₂ and post-D₃ felsic dykes and pegmatites.

The compositions of M₃ biotite grains are essentially the same as those in M₁ and M₂ assemblages. As M₃ mineral growth was generated by reactions in earlier assemblages, the similarity in biotite compositions cannot be attributed to identical conditions of growth. An alternative explanation is that Fe and Mg have re-equilibrated during or after the M₃ event. Other minerals, such as garnet, pyroxene and cordierite, have

similar compositions in M₁ and M₂ assemblages. It is possible that Fe and Mg were also re-equilibrated in these minerals during M₃.

3.6.4 M₄ assemblages

The growth of M₄ minerals is restricted to recrystallization in and adjacent to narrow ultramylonite planes that traverse all lithologies. The ultramylonites have been grouped together as a single D₅ event, and the recrystallization corresponds to the M₅ mylonite-ultramylonite event described by White (1993). Estimates of metamorphic conditions during M₅ (P≈6.5kbar at T=550-800°C) were calculated by White & Clarke (1994) for samples of mafic *proclamation gneiss* and *metapelitic gneiss* from Cape Bruce (samples 91071, 92151 and 92156 in White & Clarke, 1994). M₄ produced similar mineral assemblages to M₃ mineral growth, so that both were probably generated at lower temperatures than the M₁ and M₂ assemblages.

3.6.5 Metapelite phase compatibility in KFMASHTO

The reaction textures in *metapelitic gneiss* can be constrained in P-T space with a consideration of phase stability in a multi-component chemical system. Qualitative determinations in the system K₂O-FeO-MgO-Al₂O₃-SiO₂-H₂O-TiO₂-O₂ (KFMASHTO) by Clarke *et al.* (1989) and Clarke & Powell (1991) have demonstrated the usefulness of this technique in deriving mineral parageneses for ilmenite, magnetite and rutile-bearing assemblages. The presence of Ti and Fe³⁺ in minerals can have a critical effect on the stability of assemblages in P-T space.

Phase compatibilities for a representative spinel and quartz-bearing metapelite (sample P13) were used to reconstruct the orthopyroxene + rutile-absent invariant point (herein abbreviated as the [opx, rt] invariant), originally constructed by Clarke *et al.*

(1989). This reconstruction was undertaken since the Cape Bruce *metapelitic gneiss* has a composition that is inappropriate to the earlier grid. Firstly, all assemblages derived from the *metapelitic gneiss* are free of both orthopyroxene and rutile. Secondly, most spinel is slightly more Fe-rich than co-existing garnet. Both the Clarke *et al.* (1989) and the Clarke & Powell (1991) constructions of the [opx] invariant in KFMASHTO are constructed for the case where spinel is less Fe-rich than co-existing garnet. These grids also ignore the presence of a haematitic component to ilmenite, which can have a significant effect on the topology of the compatibility diagrams. In addition, garnet is consumed by M_3 reaction phases; if this is to be accounted for, the system cannot be projected from excess garnet, as is done in Clarke *et al.* (1989).

Before phase compatibility can be assessed, the amount of Fe^{3+} in minerals must be estimated. Excepting biotite, Fe^{3+} contents were estimated by recalculating analyses to ideal stoichiometry. This technique has limitations, especially in the case of spinel, due to the presence of exsolved magnetite. Despite this, each type of mineral contained an amount of Fe^{3+} that was usually consistent between several analyses. Fe^{3+} contents in garnet and cordierite are low and variable, but in almost all cases cordierite contains more Fe^{3+} than garnet. Ilmenite consistently contained more Fe^{3+} than spinel. Due to the volatile-bearing structure of biotite, Fe^{3+} content cannot be calculated simply from stoichiometry. Mössbauer spectral analysis of biotite crystals in metapelitic gneisses by Dyer *et al.* (1992) suggests that not more than 13% of the total Fe content of a biotite will be ferric. For the sake of constructing compatibility diagrams, biotite was assumed to have 13% Fe^{3+} . It was found that the uncertainty in the Fe^{3+} content of biotite, when properly projected from excess phases, did not effect the topology of the compatibility diagrams.

For the sake of clarity, the [opx, rt] invariant was constructed using phase compatibility diagrams with three, four and five components, sequentially. Complexity in KFMASHTO was reduced by projecting from phases in excess. Minerals that are present in all of the metamorphic assemblages (M_1 - M_4), and that

can therefore be considered as excess phases, include quartz, K-feldspar, sillimanite and magnetite. H_2O is also present in biotite and cordierite in all assemblages, and is projected as 'L' in the compatibility diagrams, which represents a liquid containing H_2O . Since there is evidence that partial melting occurred during the M_1 , M_2 and M_3 events, liquid 'L' corresponds to the melt present during metamorphism.

Ternary compatibility diagrams and their corresponding [opx] invariants are shown in Figs 3.4a-c. Univariant equilibria emanating from the [opx] invariant are denoted according to the phase absent in the equilibria (*e.g.* [spl] is the spinel-free univariant). The [spl], [bt], [sil] and [grt] univariant equilibria are positioned similarly to those constructed by Clarke *et al.* (Fig. 6b in Clarke *et al.*, 1989). The [crd] univariant has a positive gradient in P-T space that is less than that of the [bt] univariant (Fig. 3.4a). Since only biotite and ilmenite contain appreciable Ti, the [ilm] invariant equilibrium is the same as the [bt] univariant, and is therefore degenerate (Fig. 3.4b). This is not the case in the construction of Clarke *et al.* (1989), because the latter involves rutile as a non-excess phase. The [mag] univariant in Fig. 3.4c differs from the construction of Clarke *et al.* (1989), in that it has a steeper positive gradient than both the [grt] and [spl] univariants. This reflects the presence of haematite in solid solution with ilmenite, from which the FMO diagram is projected. In this projection, biotite shifts to the low- Fe^{3+} side of the garnet-cordierite join, regardless of the Fe^{3+} content of biotite.

Four-phase compatibility diagrams (Figs 3.5a & 3.6a) were constructed using isometric projections of the quadrivariant tetrahedron. Although the exact positions of the phases are difficult to visualise, phase relations are analogous to those in ternary compatibility diagrams. The positions of univariant equilibria emanating from the [opx] invariant (Figs 3.5b & 3.6b) are also analogous to those constructed from ternary diagrams. Because X_{Fe} in spinel and garnet are similar, the [opx] invariant was also constructed for the case where the X_{Fe} of spinel is less than that of garnet

Fig. 3.4. Three-component compatibility diagrams for the system $\text{K}_2\text{O}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{H}_2\text{O}-\text{TiO}_2-\text{O}_2$ (KFMASHTO). Complexity in the system is reduced by projecting from five excess phases. Each compatibility diagram is used to construct reactions emanating from the [opx, rt] invariant point in P-T space. The invariant points are based on those done for similar systems by Clarke *et al.* (1989), Clarke & Powell (1991), and Carson *et al.* (1996).

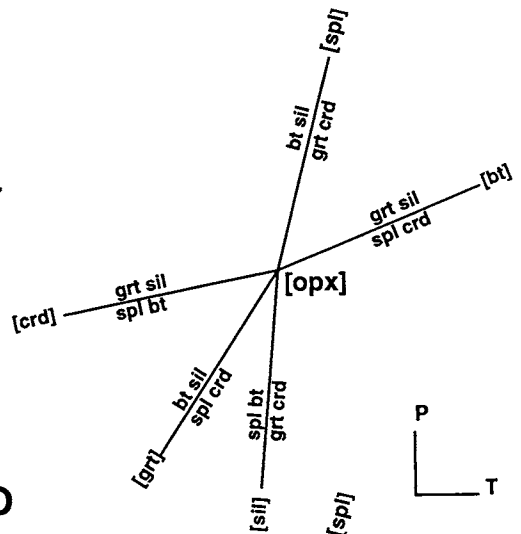
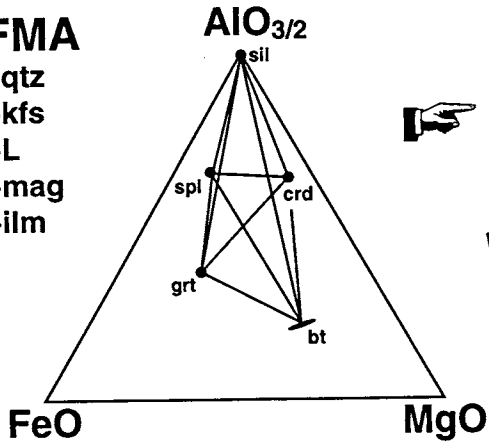
a) Compatibility diagram with apices $\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3$, projected from quartz, K-feldspar, magnetite, ilmenite and a hypothetical liquid (L) composed mostly of H_2O . Note that spinel has higher X_{Fe} than the garnet-sillimanite tie-line (see Fig. 5.2a).

b) Compatibility diagram with apices $\text{FeO}-\text{MgO}-\text{TiO}_2$, projected from quartz, K-feldspar, magnetite, sillimanite and hydrous liquid (L). Since only biotite and ilmenite contain appreciable Ti, the [bt] and [ilm] univariant equilibria combine as a degenerate [ilm, bt] univariant.

c) Compatibility diagram with apices $\text{FeO}-\text{MgO}-\text{Fe}_2\text{O}_3$, projected from quartz, K-feldspar, sillimanite, ilmenite and a hydrous liquid (L). Spinel has a higher X_{Fe} than the garnet-magnetite tie-line (see Fig. 3.2b). Due to the presence of haematite in solid-solution with ilmenite, projection from ilmenite displaces the position of biotite onto the low- Fe_2O_3 side of the garnet-cordierite tie-line.

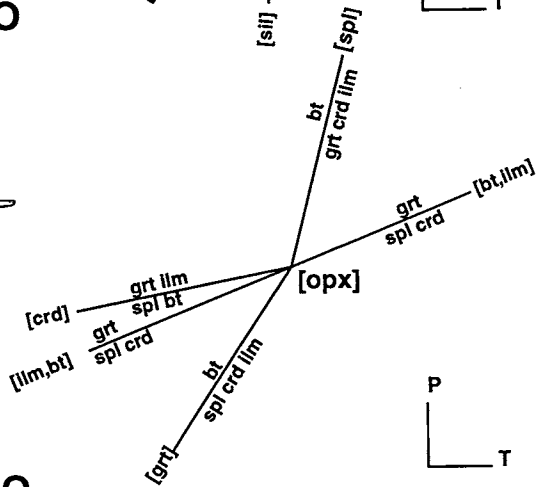
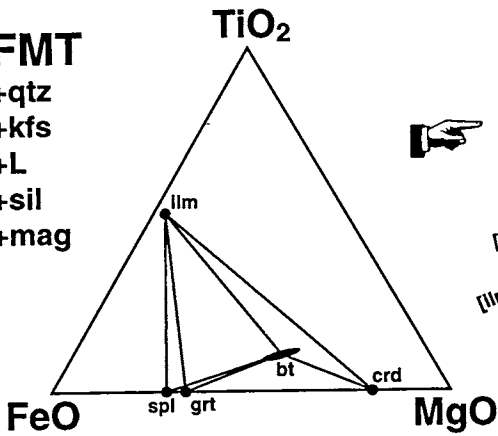
a) FMA

+qtz
+kfs
+L
+mag
+ilm



b) FMT

+qtz
+kfs
+L
+sil
+mag



c) FMO

+qtz
+kfs
+L
+sil
+ilm

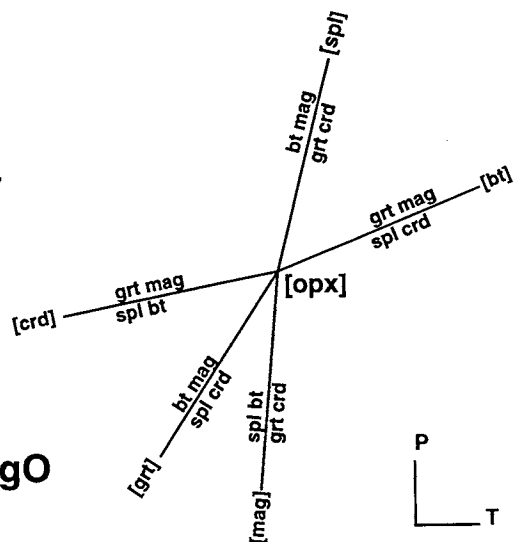
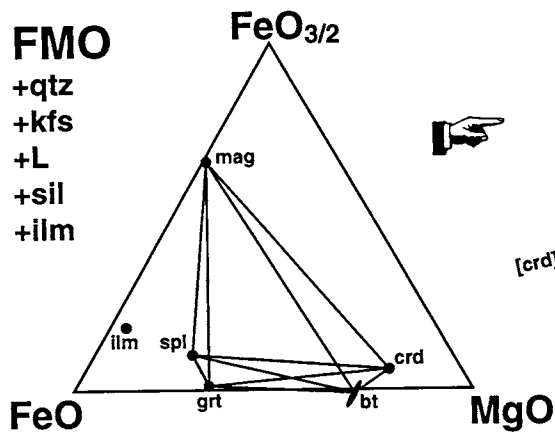


Fig. 3.5. Four-component compatibility diagram for the system K_2O - FeO - MgO - Al_2O_3 - SiO_2 - H_2O - TiO_2 - O_2 (KFMASHTO). Complexity in the system is reduced by projecting from four excess phases.

a) Obliquely oriented isometric projection of a compatibility tetrahedron with the apices $FeO_{3/2}$ - FeO - MgO - TiO_2 , projected from the excess phases quartz, K-feldspar, sillimanite and a hypothetical liquid (L) composed mostly of H_2O . The position of spinel is equivalent to that in Fig. 3.4. Dotted lines below each of the phases are used to show their relative Fe_2O_3 contents. Biotite is shown for a range of Fe_2O_3 contents (Fe_2O_3 = 0 to 13% of total FeO content).

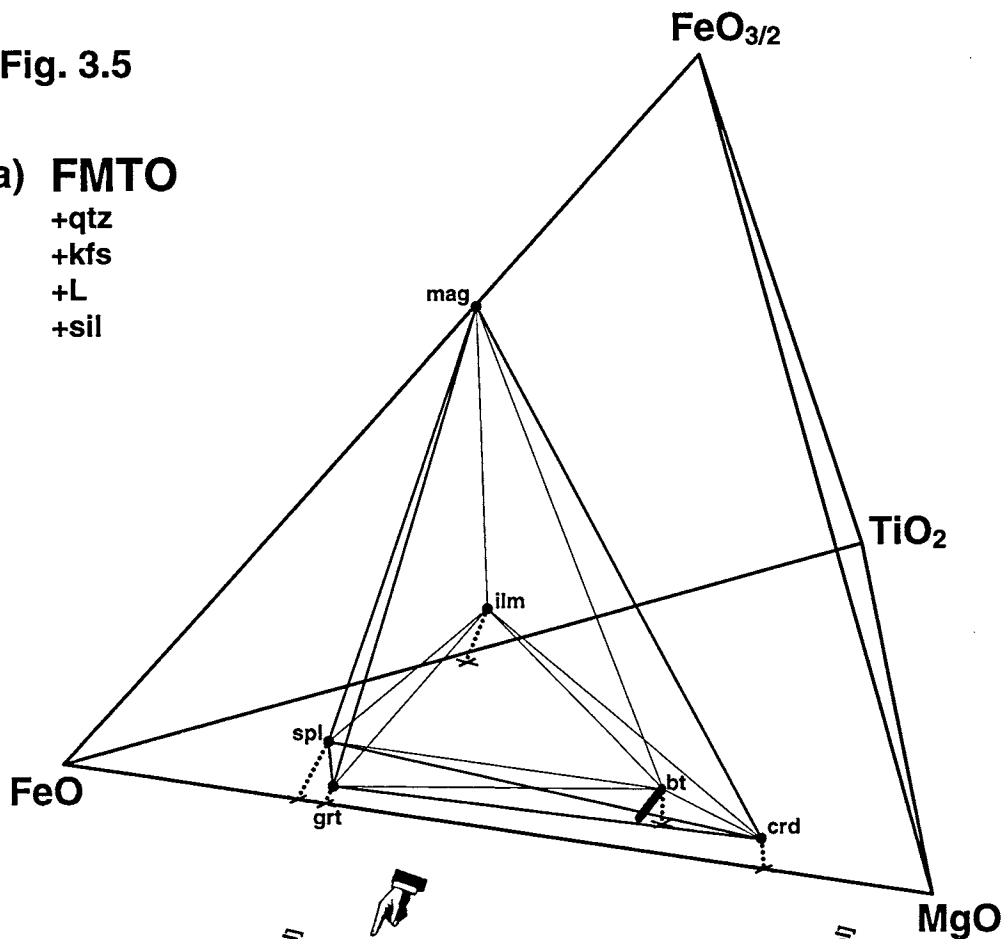
b) Reactions emanating from the [opx, rt] invariant, constructed from Fig. 3.5a. Since only biotite and ilmenite contain appreciable Ti, the [bt] and [ilm] univariant equilibria collapse into a degenerate [ilm, bt] univariant.

c) An alternative construction of the [opx, rt] invariant, with spinel containing less Fe^{2+} than garnet (*cf.* Fig. 3.2b). The positions of the [crd] and [bt] univariant equilibria have been swapped.

Fig. 3.5

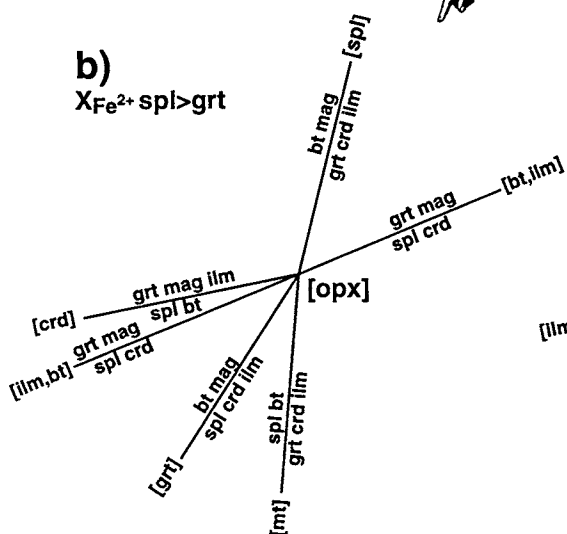
a) FMTO

+qtz
+kfs
+L
+sil



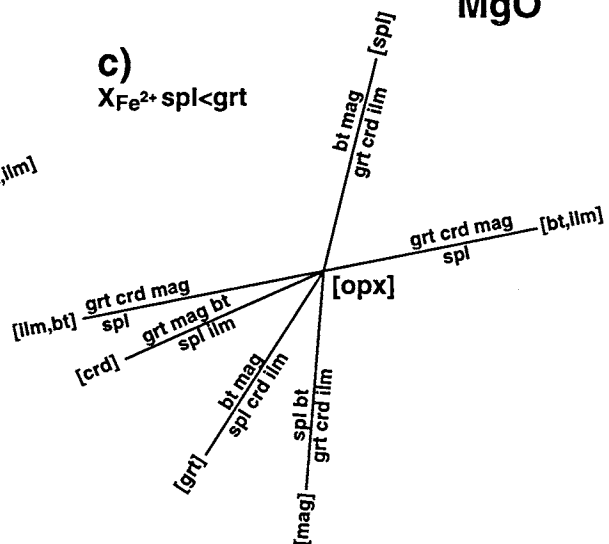
b)

$X_{\text{Fe}^{2+}} \text{ spl} > \text{grt}$



c)

$X_{\text{Fe}^{2+}} \text{ spl} < \text{grt}$



(Figs 3.5c & 3.6c). In this case, the [crd] invariant has a steeper positive gradient in P-T space than the [bt] invariant.

The [opx] invariant projected from three excess phases was then constructed by analogy to the quadrivariant projections (Fig. 3.7). As rutile is absent in the assemblage used to construct the diagram, the [opx] invariant is equivalent to the [opx, rt] invariant constructed by Clarke *et al.*, 1989. This construction was then related to the stages of metamorphic growth observed in *metapelitic gneiss*. M₁ assemblages lie up-temperature and probably up-pressure of the [opx, rt] invariant, consistent with the observed stability of spinel with quartz, ilmenite and cordierite (Fig. 3.7b). The breakdown during M₂ of spinel with quartz into sillimanite, garnet and magnetite can be attributed to the down-temperature (and possibly up-pressure) crossing of the [bt] univariant (Fig. 3.7b). The consumption of cordierite by this reaction is consistent with the absence of cordierite in M₂ reaction textures.

M₃ assemblages (and, to a lesser degree, M₄) can be explained if mineral growth occurred down-temperature (and/or up-pressure) of the [spl] equilibrium line. This would cause biotite, sillimanite and magnetite to grow at the expense of garnet and cordierite, as observed in the samples. The down-temperature progression from M₁ to M₃ metamorphism is supported by the decrease in the degree of mineral development along this series (due to decreasing growth rates), and by other down-temperature reactions identified above. Ambiguities still remain as to whether this progression represents a single history of near-isobaric cooling, or a series of decreasingly intense periods of heating, under nearly constant pressures.

3.7 Comparisons with previous work on the Mawson Coast

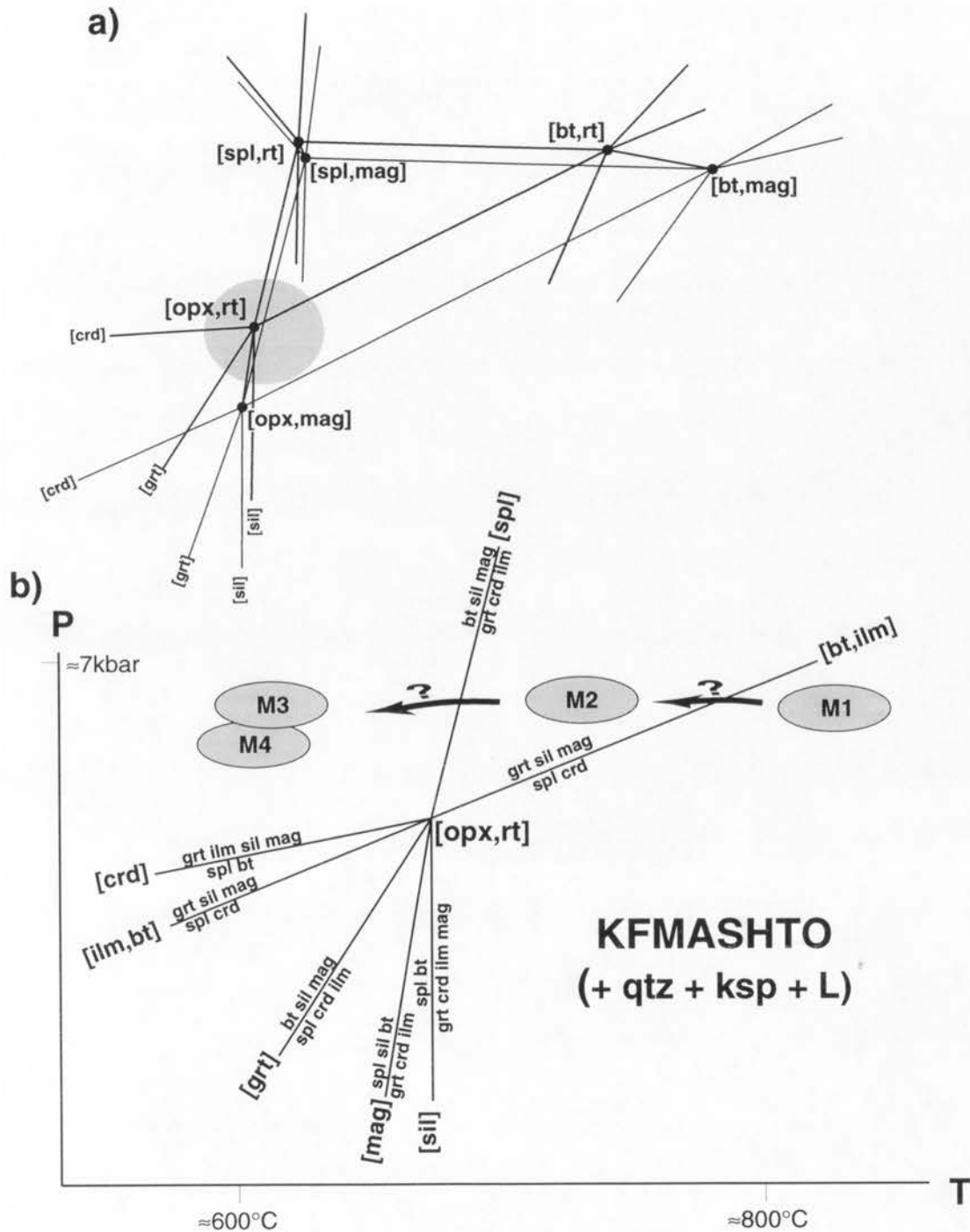
The metamorphic textures preserved at Cape Bruce and at other localities on the Mawson Coast preserve strong evidence for a history of real or apparent isobaric

Fig. 3.7. Construction of the [opx, rt] invariant point in P-T space, for the system $\text{K}_2\text{O}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{H}_2\text{O}-\text{TiO}_2-\text{O}_2$ (KFMASHTO).

a) Relationships between the [rt] and [mag] equivalents of the [opx], [spl] and [bt] invariant points in P-T space. This diagram is purely qualitative, and the positions of the [spl] and [bt] invariant points are conjectural.

b) Reactions emanating from the [opx, rt] invariant, constructed from Figs. 3.5 and 3.6. Transitions between the three stages of mineral growth described in the text (M_1 - M_3) are characterised by the down-temperature crossing of the [opx, rt]-[bt, rt] and [opx, rt]-[spl, rt] univariant equilibria. Whether these transitions occurred as a series of discrete heating events or as a single episode of near isobaric cooling is ambiguous.

a)



cooling after the peak of metamorphism. Although the cooling textures observed may be the product of a series of transient metamorphic events, these events occurred at similar pressures, and it is likely that each event ended with isobaric cooling back to the average continental geotherm. There is no evidence for decompression having occurred after the peak of metamorphism, which is observed in other parts of the Rayner Complex, including Edward VIII Gulf (Fig. 1.2, Sheraton *et al.*, 1987), Enderby Land (Sheraton *et al.*, 1987), Jetty Peninsula (Hand *et al.*, 1994a) and Prydz Bay (*e.g.* Fitzsimons & Harley, 1992).

White & Clarke (1993) suggested that metamorphism in the Colbeck Archipelago occurred in discrete stages, during the Archaean (M_1), the Mesoproterozoic (M_2) and the early Neoproterozoic (M_3). M_2 and M_3 were assigned to the Rayner Structural Episode, and involved the rehydration of Archaean granulites with an unknown source of fluids. However, if the D_1 - M_1 event is also considered as part of the Rayner Structural Episode, no exotic source of fluids need be inferred, as rehydration would have occurred as a result of the crystallisation of felsic melts generated by migmatization during M_1 . This possibility is supported by the new geochronological data presented in Chapter 4, that indicate that M_1 to M_3 occurred during the period *c.* 1000-900 Ma.

3.8 Conclusion

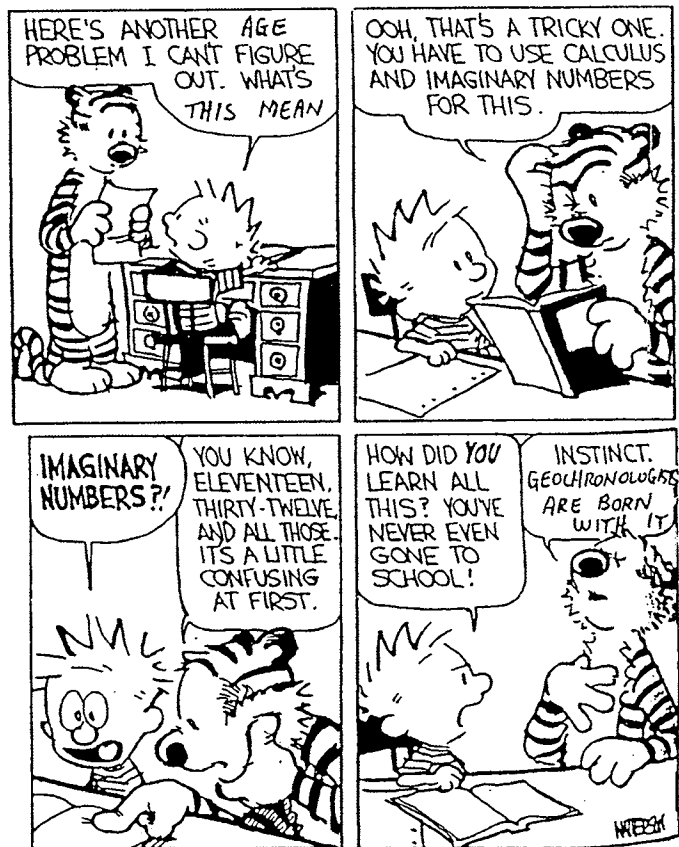
Granulite facies metamorphism accompanied multiple deformation and intrusive events that define the Rayner Structural Episode at Cape Bruce. The rocks record a history of mineral growth under elevated conditions that can be broken down into a series of texturally defined stages. Lithological precursors of *proclamation gneiss*, *felsic paragneiss* and *metapelitic gneiss* were pervasively recrystallized into a granoblastic fabric during the D_1 and D_2 events. Partial melting of felsic orthogneiss and all paragneiss units occurred during the same period, but did not totally dehydrate the

terrain. The equilibration of minerals in a granoblastic fabric (M_1 mineral growth) occurred under water-undersaturated conditions involving $P=6-7\text{ kbar}$ and $T\geq 750^\circ\text{C}$. Cooling under near-isobaric conditions caused the crystallization of most of the partial melt and some rehydration of the terrain, along with the incomplete breakdown and re-equilibration of granoblastic assemblages (M_2 mineral growth). Limited amounts of partial melting and mineral growth occurred after the D_2 event and during the D_3 and D_4 events. Limited recrystallization and re-equilibration of earlier mineral assemblages occurred before and/or during D_4 (M_3 mineral growth). This may have been the result of isobaric cooling from M_1/M_2 metamorphic conditions, or alternatively was due to reheating of the terrain after post- D_2 cooling, to temperatures above 550°C at continuously elevated pressures. All lithologies were partially recrystallized within D_5 ultramylonite - pseudotachylite shear zones. Recrystallization and mineral growth during D_5 (M_4) occurred at similar pressures ($P=6-7\text{ kbar}$) to M_1 and M_2 mineral growth, but at lower temperatures ($T\geq 550^\circ\text{C}$).

The timing of metamorphism during the Rayner Structural Episode was further constrained by the dating of stages of zircon growth (Chapter 4). The causes of elevated metamorphic conditions, including sources of heating and tectonic setting, will be considered in the Synthesis (Chapter 6).

Chapter 4

Geochronology



-courtesy of Dr. Chris Carson

4. Geochronology

Introduction

Since the Rayner Complex was defined by Kamenev (1970) in Enderby Land, knowledge of its age, extent and construction has been refined by numerous studies in geochronology (*e.g.* Sheraton & Black, 1981; Black *et al.*, 1987; Clarke, 1987; Grew *et al.*, 1988; Young & Black, 1991; Zhao *et al.*, 1992; Kinny *et al.*, 1993; Hensen & Zhou, 1995; Thost, 1996; Kinny *et al.*, 1997). Evidence that the Rayner Complex represents a Proterozoic metamorphic belt, involving a *c.* 1000 Ma tectonic episode (the *Rayner Structural Episode* of Sandiford & Wilson, 1984), is provided by numerous age estimates from magmatic and metamorphic lithologies (Black *et al.*, 1987; Grew *et al.*, 1988; Young & Black, 1991; Kinny *et al.*, 1993; Kinny *et al.*, 1997).

Complexity within the *c.* 1000 Ma event has only been addressed recently. Studies of high-grade gneisses that were produced during the Rayner Structural Episode have provided evidence for a period of tectonic and metamorphic activity that can be broken down into a number of discrete events (Fitzsimons & Thost, 1992; White & Clarke, 1993; Hand *et al.*, 1994b). The temporal separation of these events was previously uncertain, due to a lack of structurally constrained geochronology.

Recent geochronological work in several localities has provided important constraints on the timing of tectonic and metamorphic events. SHRIMP dating of syn-tectonic intrusions in the northern Prince Charles Mountains and on the Mawson Coast has supported the interpretation that the Rayner Structural Episode was a polyphase event (Young & Black, 1991; Kinny *et al.*, 1997). Studies in other regions, especially in the Prydz Bay area, have re-interpreted events previously assigned to the Rayner Structural Episode, as belonging to a widespread *c.* 500 Ma event (*e.g.* Zhao *et al.*, 1992; Hensen & Zhou, 1995). However, it is difficult to correlate these results to

other parts of the Rayner Complex, due to the lack of geochronological data. In particular, the timing of the later stages of geological activity in localities such as the Jetty Peninsula (Hand *et al.*, 1994a) have been inferred from structural relationships only.

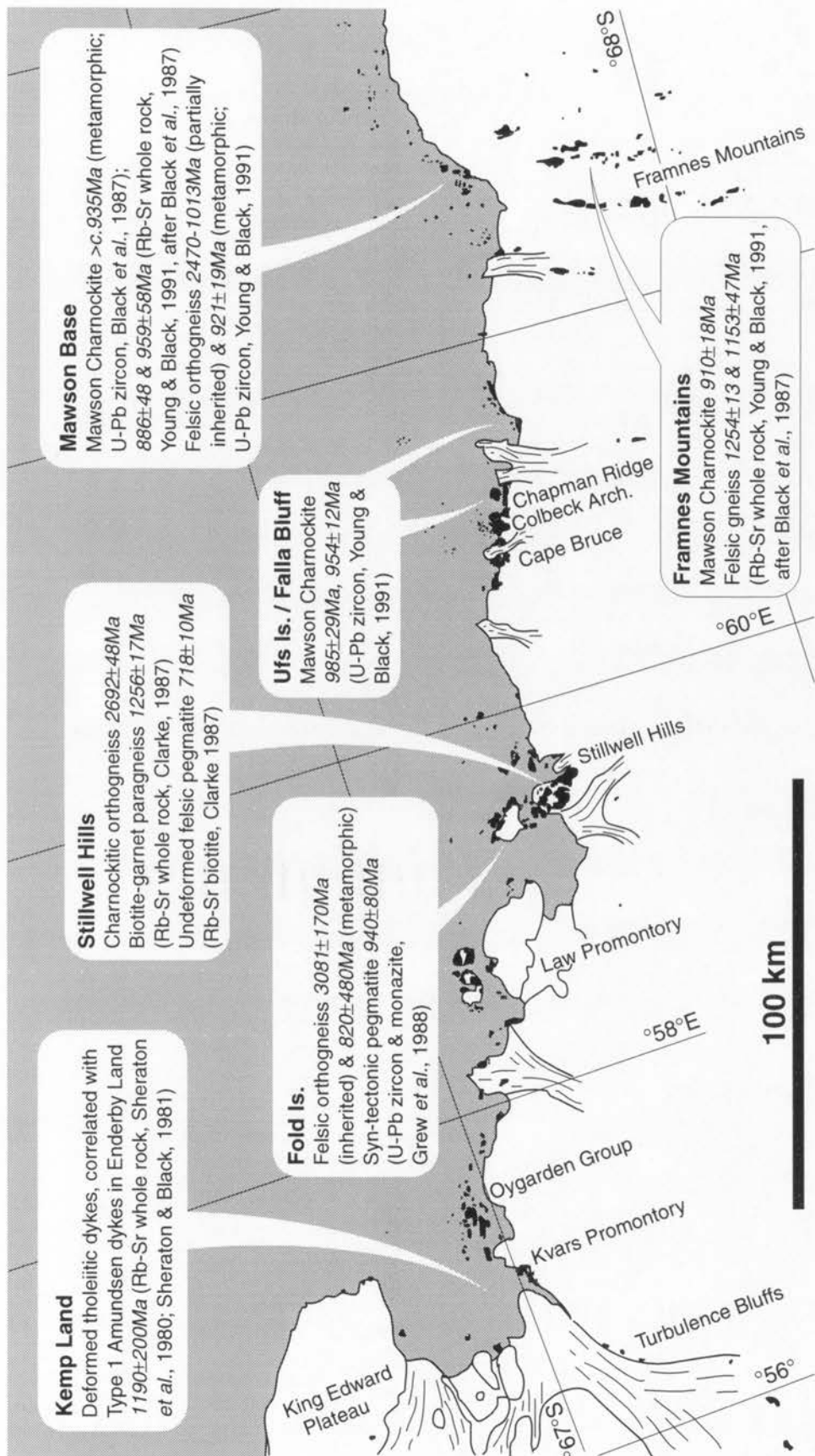
In this chapter, new geochronological data for the Cape Bruce locality are presented. Age estimates from structurally constrained samples demonstrate that the complex history of deformation, metamorphism and magmatism described in Chapters 2 and 3 can be related to a series of discrete events, that occurred within the period 1000-900 Ma. Age data also provide evidence for Archaean and Paleoproterozoic lithologies within the lower crust, which provided a partial source for granitic bodies that intruded supercrustal lithologies at Cape Bruce.

4.1 Previous geochronology for the Mawson Coast

Geochronological studies of the Rayner Complex have already been summarised in Chapter 1. Published age data that are relevant to this chapter are discussed below.

The only lithology from the Mawson Coast with an Archaean age is a charnockitic orthogneiss from the Stillwell Hills (Fig. 4.1), which has a Rb-Sr whole rock age of 2692 ± 48 Ma (Clarke, 1987). This age was interpreted by Clarke (1987) as that of granulite facies metamorphism and deformation that affected the charnockite. This event was tentatively correlated by White & Clarke (1993) with D₁ deformation and granulite facies metamorphism, which affected gneiss at Chapman Ridge and the Colbeck Archipelago (Fig. 4.1). However, the involvement of pre-Mesoproterozoic lithologies in the formation of the Mawson Coast is demonstrated by a number of inherited age data. Zircon from felsic orthogneiss in the Stillwell Hills yielded discordant U-Pb data with an upper intercept age of 3081 ± 170 Ma (Grew *et al.*, 1988). The intercept age was interpreted by Grew *et al.* (1988) as that of xenocrystic

Fig. 4.1. Map of the Mawson Coast in Kemp Land and MacRobertson Land, showing the locations of dated lithologies. Age data are discussed in the text.



zircon in the orthogneiss. At Mawson Base, cores in zircon grains from a felsic orthogneiss yielded concordant U-Pb ages between 2460 and 1700 Ma (Young & Black, 1991). A Sm-Nd T_{CHUR} model age of 2020 Ma was obtained from Mawson Charnockite at Mawson Base (Black *et al.*, 1987).

Metasedimentary gneiss from the Stillwell Hills yielded a Rb-Sr whole rock ages of 1256 ± 17 Ma (Clarke, 1987). This age was interpreted by Clarke (1987) as the product of partial resetting of the Rb-Sr system by metamorphism during the Rayner Structural Episode. In the Framnes Mountains, Rb-Sr whole rock ages of 1153 ± 47 and 1254 ± 13 Ma were derived from Painted Gneiss (Young & Black, 1991, recalculated from unpublished data by Paul Arriens). These *c.* 1200 Ma ages provided a loose maximum constraint for the timing of the Rayner Structural Episode.

In Kemp Land (Fig. 4.1), tholeiitic dykes that were deformed during the Rayner Structural Episode were correlated by Sheraton *et al.* (1980) with undeformed dykes that cut Archaean gneisses in Enderby Land. The latter group of dykes were dated using Rb-Sr whole rock isochrons at 1190 ± 200 Ma (Sheraton & Black 1981). These dykes were tentatively correlated with bands of mafic gneiss in the Stillwell Hills (Clarke, 1988) and deformed dolerite dykes in the Colbeck Archipelago (White & Clarke, 1993). The latter two lithologies were inferred by White & Clarke (1993) to have been deformed by the same event (D_2), which was considered to have occurred at peak granulite conditions (M_2) during the Rayner Structural Episode.

Age estimates have been obtained for orthogneisses that have been interpreted to have intruded during the Rayner Structural Episode. Syn-tectonic felsic pegmatites from the Stillwell Hills were dated by the U-Pb zircon method at 940 ± 80 Ma (Grew *et al.*, 1988). The Mawson Charnockite, which constitutes most of the outcrop in the Framnes Mountains and along the Mawson Coast between Mawson Base and Chapman Ridge, is deformed by events that have been interpreted to represent the later stages of the Rayner Structural Episode (D_3 , Clarke, 1988; D_{3-4} , White &

Clarke, 1993). SHRIMP dating of U-Pb in zircon yielded ages of 985 ± 29 Ma and 954 ± 12 Ma for two samples of Mawson Charnockite from the Mawson Coast (Young & Black, 1991). U-Pb zircon dating of Mawson Charnockite from the Framnes Mountains yielded a minimum age of emplacement of 935 Ma (Black *et al.*, 1987). Rb-Sr whole rock dating of the charnockite gave slightly younger ages of 959 ± 58 , 910 ± 18 and 886 ± 48 Ma (Black *et al.*, 1987; Young & Black, 1991, recalculated from unpublished data by Paul Arriens).

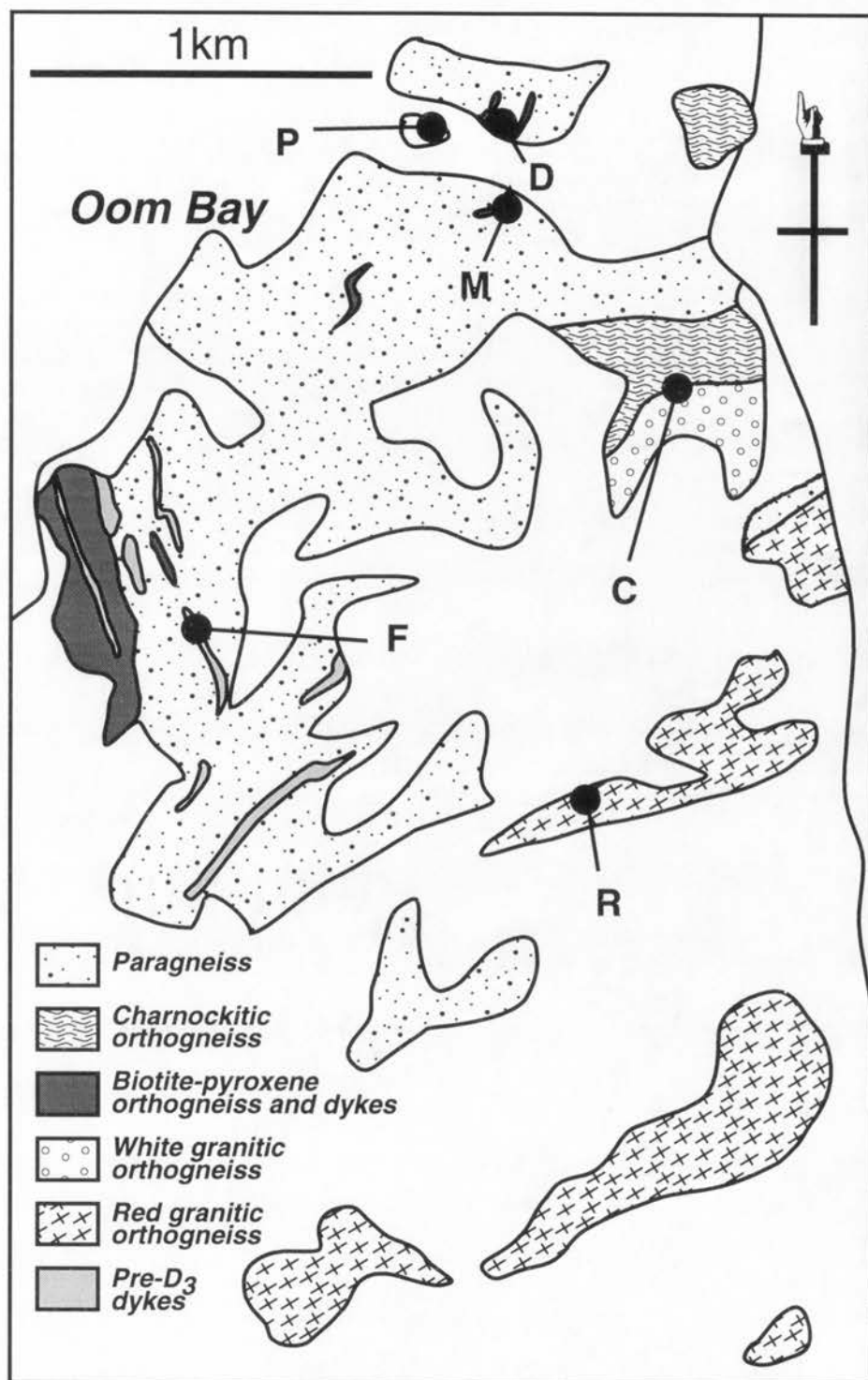
Ages have also been obtained from zircon that was inferred to have a metamorphic origin. Metamorphic zircon in an enclave of layered orthogneiss in the Mawson Charnockite at Mawson station was dated at 921 ± 19 Ma (Young & Black, 1991). This age is comparable to the Rb-Sr whole rock ages obtained from the Mawson Charnockite. Young & Black (1991) interpret the zircon and Rb-Sr ages as representing metamorphism during the D₃ event of Clarke (1988).

A Rb-Sr biotite age of 718 ± 10 Ma was obtained by Clarke (1987) from an undeformed pegmatite from the Stillwell Hills. This date provides a lower constraint on the age of tectonism and metamorphism in this area. However, no dates have been obtained from the Mawson Coast or the Framnes Mountains that unequivocally represent geological events that occurred after *c.* 900 Ma.

4.2 Dated samples from Cape Bruce

Six fresh cleaned samples at least 1000cm³ in size were taken from various localities at Cape Bruce (Fig. 4.2). The samples were selected as being representative of lithologies that have clear relationships to the deformation events inferred in the history of Cape Bruce geology (Chapter 2). The samples were trimmed to remove weathered faces and cracks, and were then crushed by hydraulic press. Fragments were briefly ground in a tungsten carbide ring-mill, and the powder was sieved to separate fractions 300-

Fig. 4.2. Simplified geological map of Cape Bruce. The distribution of the major lithologies, and the locations of samples used for geochronology are shown.



150 μm and 150-60 μm in size. Heavy mineral fractions were separated by settling in methylene iodide and magnetic fractions were removed. Zircon grains were hand-picked from these separates. Approximately 100-150 grains were picked from each sample, except for sample P, from which around 60 grains were picked. The grains were mounted in epoxy disks with fragments of the CZ3 standard zircon, and were then machine polished.

Paragneiss sample P

The sample is of *metapelitic gneiss*, with a strong S_1/S_2 foliation defined by gneissic layering and aligned grains of sillimanite and biotite. Garnet is present as porphyroblasts, which commonly contain inclusions aligned with S_1/S_2 . The gneissosity was gently crenulated by F_3 folds, with a weak axial planar S_3 defined by biotite grains. About 5-10% of the sample is composed of leucosome, that contains K-feldspar, quartz and plagioclase. The leucosome occurs as 3-10 mm thick layers parallel to S_1/S_2 , and as nebulitic patches that cut S_1/S_2 . Zircon grains occur in minor amounts in both the leucosome and the pelitic layers.

Charnockitic orthogneiss sample C

The sample was taken from an outcrop of *charnockitic orthogneiss* that was a few metres from the contact with *white granitic orthogneiss*, which intruded after D_1 (Fig. 4.2). It is a massive, even-grained grey rock of enderbitic composition. A weak S_1 gneissosity is defined by slightly flattened and aligned grains of orthopyroxene, plagioclase, quartz and biotite. Zircon grains are distributed evenly through a thin section of the sample.

Biotite-pyroxene orthogneiss sample B

The sample was taken from a dyke that cuts S_1 in *felsic paragneiss* (Fig. 4.2). The dyke is approximately 1m thick and was broken into fragments a few metres long by the D_2 event. The dyke has biotite and garnet-rich borders, which were avoided by taking the sample from the centre of the dyke. The sample is a biotite-rich monzodiorite,

with a strong S_2 foliation defined by aligned biotite and orthopyroxene grains. Abundant monazite and fewer zircon grains are distributed throughout the sample.

Red granitic orthogneiss sample R

The sample was taken from the main stock of *red granitic orthogneiss*, that dominates outcrops in the south-east of the Cape Bruce area (Fig. 4.2). The sample is massive and sub-equigranular, with sub-equal amounts of orange K-feldspar and quartz, and smaller amounts of biotite, garnet and plagioclase. A strong S_2 foliation is defined by flattened quartz and biotite grains.

Pre-D₃ felsic orthogneiss sample F

The sample was taken from a granitic dyke which cuts S_2/S_1 in *felsic paragneiss* (Fig. 4.2). The dyke is part of a suite that was openly folded by the D_3 event. The sample is fine-grained and pinkish-grey, and contains quartz, K-feldspar, plagioclase and small amounts of biotite and garnet. A well developed S_3 foliation is defined by aligned biotite grains, and by sheets less than 0.5 mm thick composed of very fine grains.

Pre-D₄ felsic pegmatite sample D

The sample is from an pegmatite that cuts F_3 crenulations in *metapelitic gneiss* (Fig. 4.2). The sample is granitic and coarse-grained, with K-feldspar phenocrysts up to 2cm across in a matrix of K-feldspar, quartz and plagioclase. Lesser amounts of biotite and garnet are also present. Biotite grains define a weak S_4 . Zircon occurs in very small amounts.

4.3 Complexity in U-Pb isotopic data

Due to the high closure temperatures of U and Pb diffusion in zircon, grains can preserve evidence of the age of their crystallization, along with evidence for subsequent high-temperature metamorphic or magmatic events. Such information

may be recorded in the multi-stage growth of a zircon grain, in zircon that has been recrystallized, or in zircon that has suffered Pb loss or gain via diffusion. How well a 'closed' isotopic system in zircon is protected against the effects of Pb diffusion and recrystallization is dependant on several factors. These factors may include the extent of damage to the crystal lattice caused by radioactive decay, and the exposure of the zircon to weathering or penetrative fluids at low temperatures (Black 1987; Mezger & Krogstad, 1997). The combined effects of such processes can introduce great complexity into the interpretation of isotopic data, even within a single zircon grain.

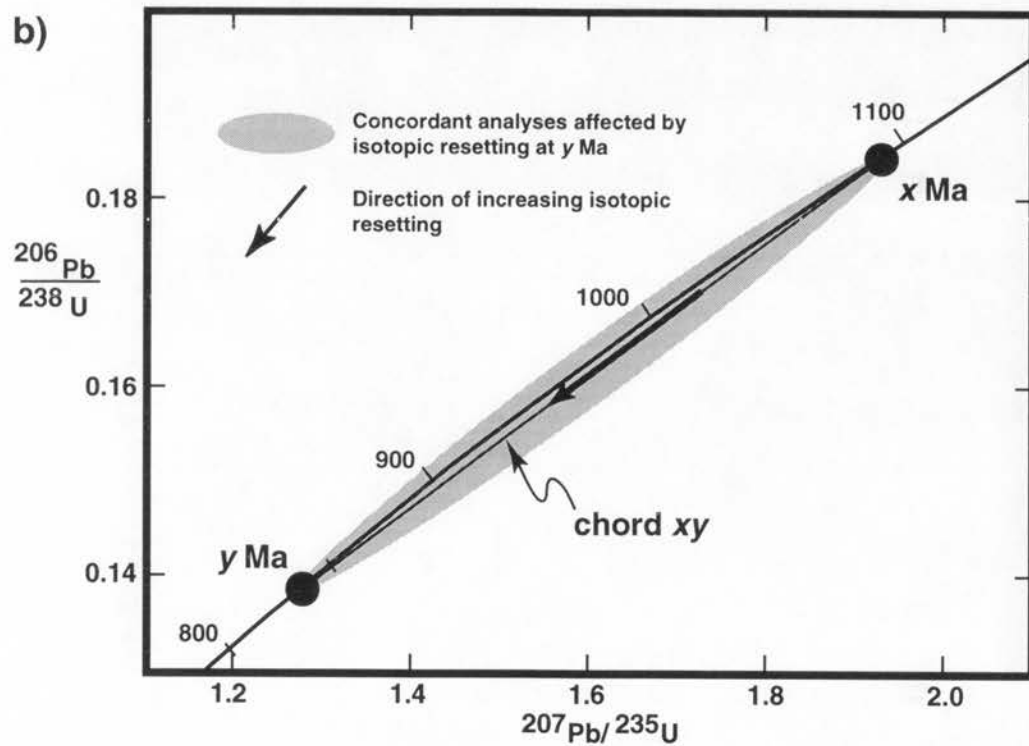
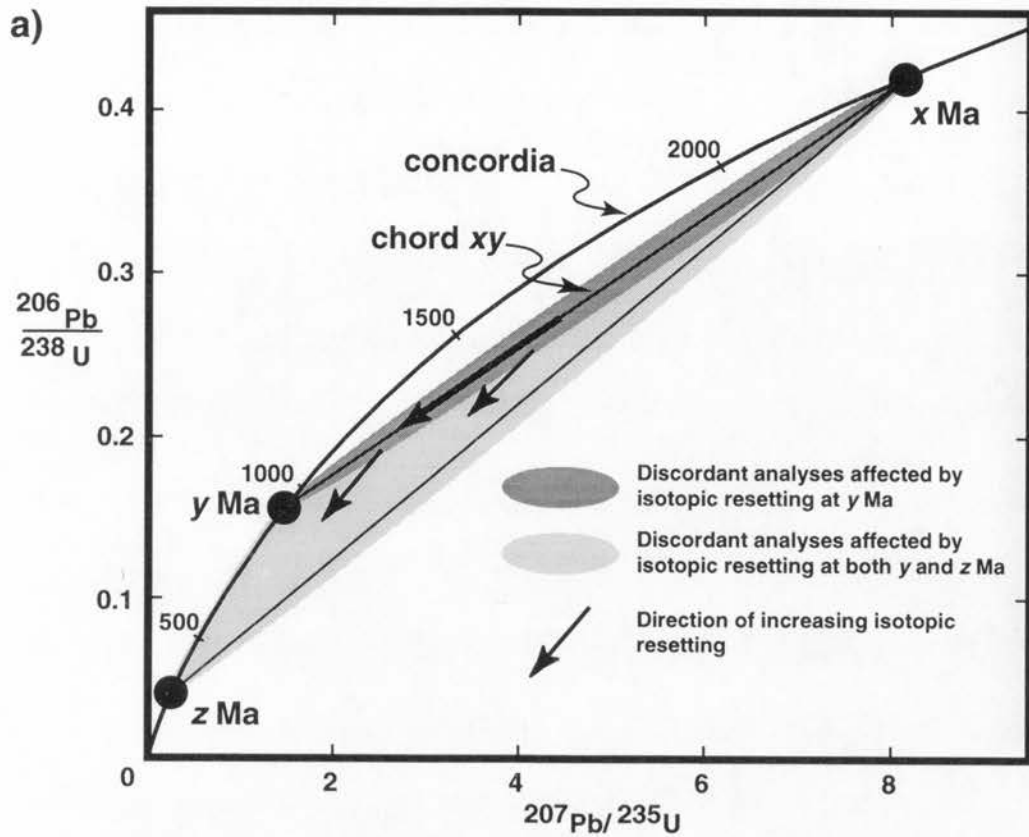
In this study, U-Pb isotope ratios in zircon grains were interpreted with the aid of the *concordia diagram* of Wetherill (1956), which compares the ratios of three parent-daughter pairs, $^{206}\text{Pb}/^{238}\text{U}$, $^{207}\text{Pb}/^{235}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ (Fig. 4.3a). In the ideal case of zircon that has not been isotopically disturbed since crystallization, the age estimates for these ratios will be *concordant*, i.e. within error of each other. Such ages lie along a curved line on the Wetherill diagram, known as the *concordia* (Fig. 4.3a). *Discordance* between age estimates for each ratio may occur as a result of Pb loss, or the partial recrystallization of a zircon grain. Mixed analyses of two or more generations of zircon will also produce discordant age estimates, which is particularly a problem in U-Pb dating by whole zircon digestion. *Reverse discordance*, where the $^{207}\text{Pb}/^{206}\text{Pb}$ age is significantly younger than the $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$ ages, is a special case. Reverse discordant data, which lie above the concordia, have been explained as a result of Pb gain in small localities in a zircon grain (Williams *et al.*, 1984), or as an analytical problem during the SHRIMP analysis of zircon that is structurally damaged (Black *et al.*, 1991; McLaren *et al.*, 1994).

Discordance in analysed zircon may be the result of effects caused by one or more geological events. For example, consider a population of zircons that were crystallized at x Ma and were subsequently affected by a geological event at y Ma (Fig. 4.3a). Analyses of zircons that were isotopically disturbed by the event at y Ma will lie along a chord xy that intersects the concordia at x and y Ma. The simplest way of

Fig. 4.3. Hypothetical distribution of U-Pb age data on the concordia diagram of Wetherill (1956). *Concordant* age data, such as that derived from a single event of zircon crystallisation, lie along the concordia; *discordant* data lie away from the concordia.

a) Age data from zircon that was crystallised at x Ma and isotopically disturbed by an event at y Ma, will lie along the chord xy , at a point that depends on the degree of isotopic resetting. If zircon containing isotopes with discordant data is subsequently disturbed by an event at z Ma, then the discordant data will lie within the field shown. A chord drawn through data in this field will intercept the concordia at points that do not represent any event.

b) If the y Ma event occurred less than about 500 My after the crystallisation of zircon at x Ma, then partially reset isotopic age data will lie along a chord xy that is close to the concordia. Age data along this chord will remain concordant, even though each age does not represent a single real event.



interpreting discordant data is to fit a chord through the analyses. However, the significance of the intercepts of the chord with the concordia may be ambiguous. If another event at z Ma also affected the zircon analyses, then the intercepts of any chord fitted to the discordant analyses may be meaningless (Mezger & Krogstad, 1997). In such a case, concordant analyses at x , y and z Ma are needed to reinforce any interpretation of the discordant data (Fig. 4.3a).

Concordant analyses do not always represent the age of a single episode of zircon crystallization or isotopic resetting. If the difference between the ages of two events is less than about 500 Ma, analyses which lie along a chord between the two events may still be concordant (Fig. 4.3b). Thus, an array of analyses that are spread over a small section of the concordia needs to be interpreted carefully. In the example in Fig. 4.3b, zircons crystallised at x Ma have isotopic systems that have been partially to completely reset by a metamorphic event at y Ma. Analyses that lie along the chord xy are all concordant. Such a distribution lacks clear intercepts with the concordia, and thus the ages of the causal events will be difficult to determine precisely.

A few simple rules can be applied to ambiguous discordant and concordant age data, so that the age of events that produced the distribution can be defined with confidence. An examination of the internal structure of the zircon grains is needed to determine if multiple generations of zircon growth are present. This can be compared to the distribution of measured age estimates to determine their significance. For each event that has affected the zircon, whether by recrystallization, total Pb loss or fresh growth, a distinct population of concordant age data is required to define that event with any accuracy. This is particularly the case where the population data of two events lie close together in age and may overlap within error. If a geological event has affected zircon in samples from lithologies of differing ages, then age populations which have been attributed to this event should correspond across samples.

4.4 Zircon morphology

Cathodoluminescence (CL) imaging has recently been recognised as a powerful tool for revealing multi-stage growth and recrystallization in zircon grains (Vavra, 1990; Hanchar & Miller, 1993; Koschek, 1993). The mechanisms for cathodoluminescent emission are complex. Zircon itself luminesces under electron bombardment with a signature emission spectrum (Koschek, 1993). The intensity of the emission is affected by a number of factors, including the preservation of zircon structure after radioactive decay (Koschek, 1993; McLaren *et al.*, 1994) and the concentration of trace elements that luminesce, especially Dy, Hf, Y and U (Hanchar & Miller, 1993; Koschek, 1993).

Structural features in zircon crystals are often visible in CL images. Variations in emission intensity can indicate differing compositions, growth habits, and degrees of radiation damage in the stages of growth of a zircon grain. CL imaging can therefore be used as a non-destructive way of selecting areas on polished zircon crystals for ion microprobe analysis.

Prior to SHRIMP analysis, mounted zircons were examined and photographed using a Siemens 505 Scanning Electron Microscope, with an operating voltage of 20 kV, which was equipped with polychromatic CL detectors. The facility is located in the Electron Microscopy Unit at the University of Sydney.

Paragneiss (P)

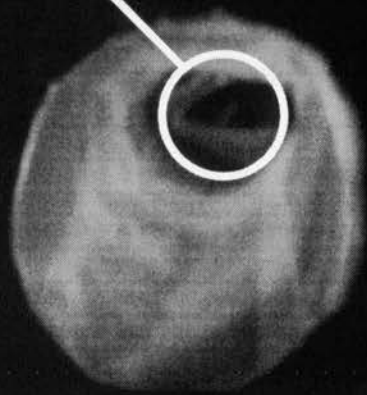
Zircon grains extracted from this sample are small (40-70 μm across), equant and subhedral to euhedral. Less than half of the grains contain round cores of zircon with growth zoning defined by fine bands of varied CL response. These growth zones are truncated by the round edges of the core (Plate 4.1a). The rest of the zircon is strongly luminescent, with little or no visible zonation (Plate 4.1b). Several of the

Plate 4.1. Cathode luminescence (CL) images of analysed zircon grains. White circles indicate the position and size of analysed spots. $^{206}\text{Pb}/^{238}\text{U}$ ages are shown, except for discordant ages (disc.), which are quoted for $^{207}\text{Pb}/^{206}\text{Pb}$ ages. All ages are quoted with 1σ errors.

- a) Grain P8 from sample P, *paragneiss*.
- b) Grain P21 from sample P, *paragneiss*.
- c) An unanalysed grain from sample P, *paragneiss*.
- d) Grain C1 from sample C, *charnockitic orthogneiss*.

Plate 4.1

a)



P8.1
967±50Ma
(disc.)

50μ

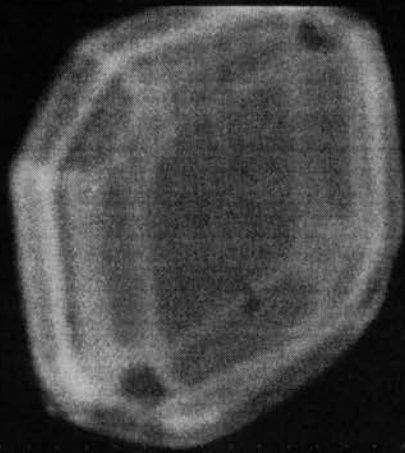
b)



P21.1 1340±37Ma
(disc.)

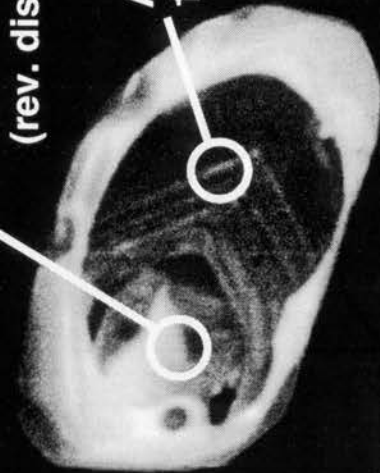
50μ

c)



50μ

d)



C1.1 763±105Ma
(rev. disc.)

A1.2
1129±18Ma

50μ

grains have distinctive facets, which are visible in the CL image due to the reflection of light off the crystal faces (Plate 4.1c).

Charnockitic orthogneiss (sample C)

Most zircon crystals from this sample are elongate, 120-250 μm long and are subhedral to anhedral. Squat and equant anhedral grains 60-150 μm make up approximately 20% of mounted zircons. Most grains contain anhedral round or irregular cores, that for the most part have a low CL response. Where growth zones are visible in these cores, they are truncated at the edges of the core by rims with different CL response (Plate 4.1d). Other cores show no internal structure (Plate 4.2a). Most elongate zircons have a high CL response and strong growth zoning (Plate 4.2b). Several elongate grains have narrow rims of unzoned weakly luminescent zircon (Plate 4.2b). Equant zircon grains are typically anhedral, and have a uniformly low CL response (Plate 4.2c).

Biotite-pyroxene orthogneiss dyke (sample B)

Zircons from this sample are elongate (150-600 μm long), anhedral and irregular. Many of the grains are fractured. The grains have low CL responses, with irregular patches of moderately luminescent zircon (Plate 4.2d). No regular growth structures are visible.

Red granitic orthogneiss (sample R)

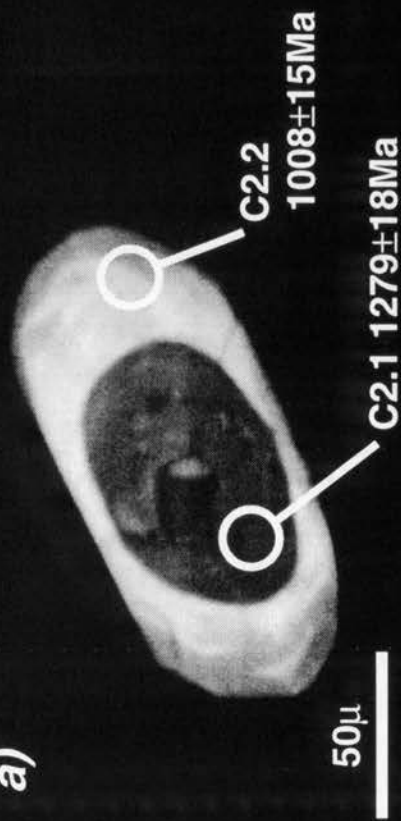
Zircon from this sample occurs as crystals 200-400 μm long, together with a few smaller anhedral grains with squat or round forms. Most elongate grains contain round or irregular cores, with truncated growth zones (Plate 4.3a) or no CL emission (Plate 4.3b). The cores are surrounded by strongly zoned zircon, which typically grades from strongly luminescent to weakly luminescent from core to rim (Plate 4.3c). Almost all zircons are anhedral, and several grains have overgrowths of weakly luminescent zircon with no visible zoning (Plate 4.3b). The equant zircons consist of unzoned weakly luminescent zircon, that is similar to the overgrowths.

Plate 4.2. Cathode luminescence (CL) images of analysed zircon grains. White circles indicate the position and size of analysed spots. $^{206}\text{Pb}/^{238}\text{U}$ ages are quoted with 1σ errors.

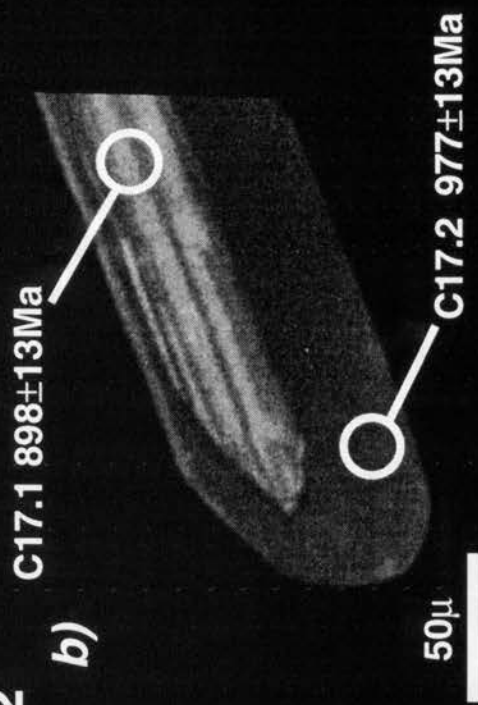
- a) Grain C2 from sample C, *charnockitic orthogneiss*.
- b) Grain C17 from sample C, *charnockitic orthogneiss*.
- c) Grain C3 from sample C, *charnockitic orthogneiss*.
- d) Grain B1 from sample B, *biotite-pyroxene orthogneiss*.

Plate 4.2

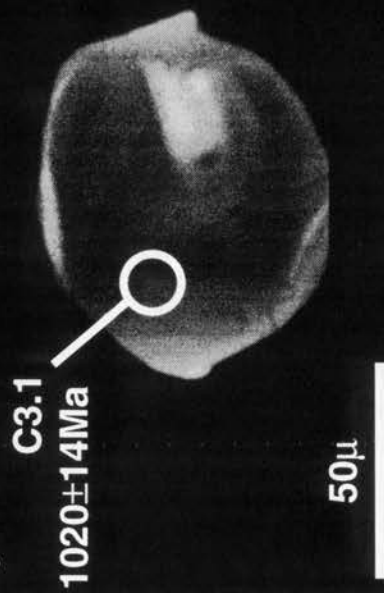
a)



b)



c)



d)

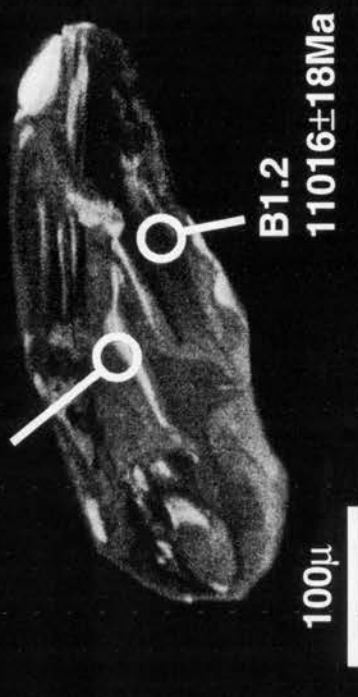
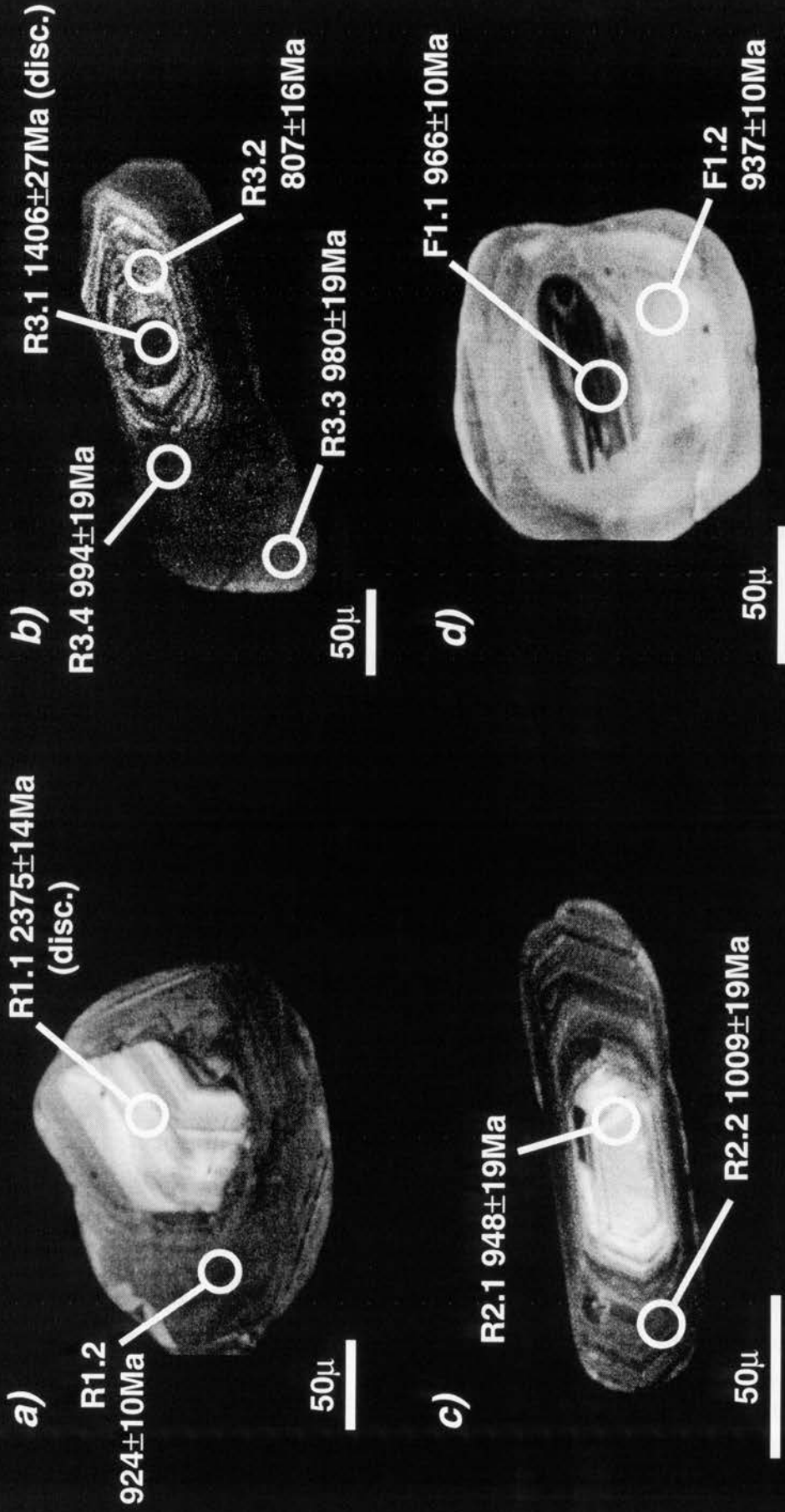


Plate 4.3. Cathode luminescence (CL) images of analysed zircon grains. White circles indicate the position and size of analysed spots. $^{206}\text{Pb}/^{238}\text{U}$ ages are shown, except for discordant ages (disc.), which are quoted for $^{207}\text{Pb}/^{206}\text{Pb}$ ages. All ages are quoted with 1σ errors.

- a) Grain R1 from sample R, *red granitic orthogneiss*.
- b) Grain R3 from sample R, *red granitic orthogneiss*.
- c) Grain R2 from sample R, *red granitic orthogneiss*.
- d) Grain F1 from sample F, a dyke of *pre- D_3 felsic orthogneiss*.

Plate 4.3



Pre-D₃ felsic orthogneiss dyke (sample F)

Zircon grains are equant and 100-300 μm across. They are subhedral to anhedral, with round and irregular forms. Many grains contain anhedral cores with truncated growth zoning (Plate 4.3d). The rest of the zircon is strongly luminescent, and has little or no visible zoning (Plates 4.3d & 4.4a).

Pre-D₄ felsic pegmatite (sample D)

This sample contains equal amounts of equant zircon (Plate 4.4b) and elongate zircon (Plate 4.4c). Grains are 70-300 μm long and have anhedral to subhedral forms. Almost all of the grains show strong oscillatory zoning (Plates 4.4b & 4.4c). These growth zones are commonly truncated at the ends of grains (Plate 4.4b). A few grains have anhedral cores with truncated growth zones (Plates 4.4b & 4.4d). Overgrowths of strongly luminescent zircon are present on a few grains (Plate 4.4d). These overgrowths are narrow and have no growth zoning.

4.5 Isotopic techniques and data reduction

Isotopic analysis of the zircon grains was conducted with Dr. David Nelson using the Sensitive High Resolution Ion Microprobe (SHRIMP II) facility at Curtin University, Perth, Western Australia. The procedures used for isotopic analysis and data reduction are based on those outlined in Compston *et al.* (1984) and Williams *et al.* (1984), and are described in greater detail in Nelson (1997). The ion microprobe was operated with a 30 μm diameter primary beam of O_2^- ions at 10 keV. Pb-U ratios for analysed zircon were determined relative to a standard Sri Lankan zircon, CZ3, with an assigned $^{206}\text{Pb}/^{238}\text{U}$ value of 0.0914, corresponding to an age of 564 Ma. The relationship between Pb^+/U^+ and UO^+/U^+ , and between $\text{Zr}_2\text{O}^+/\text{U}^+$ and UO^+/U^+ , was assumed to have followed the power law described by Claoue Long *et al.* (1995), with an exponent of 2. The relationship between Th^+/U^+ and UO^+/U^+ was assumed to have followed the empirical relationship described in Williams *et al.* (1996). The

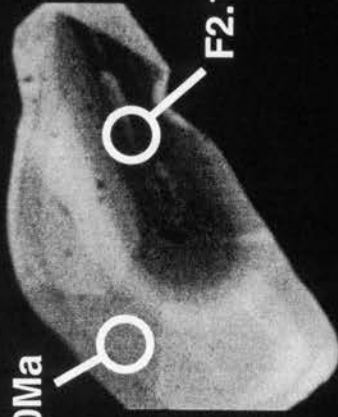
Plate 4.4. Cathode luminescence (CL) images of analysed zircon grains. White circles indicate the position and size of analysed spots. $^{206}\text{Pb}/^{238}\text{U}$ ages are shown, except for discordant ages (disc.), which are quoted for $^{207}\text{Pb}/^{206}\text{Pb}$ ages. All ages are quoted with 1σ errors.

- a) Grain F2 from sample F, a dyke of *pre-D₃ felsic orthogneiss*.
- b) Grain D3 from sample D, a *pre-D₄ felsic pegmatite*.
- c) Grain D6 from sample D, a *pre-D₄ felsic pegmatite*.
- d) Grain D8 from sample D, a *pre-D₄ felsic pegmatite*.

Plate 4.4

a)

F2.2 $932 \pm 10 \text{ Ma}$



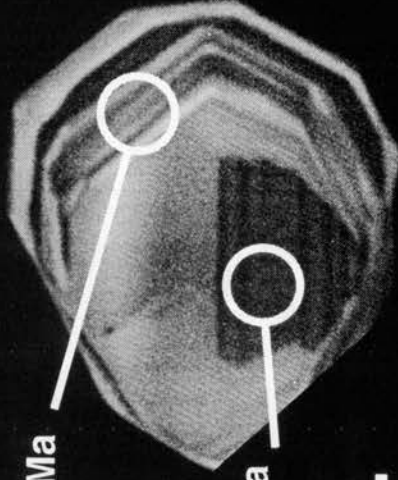
50 μ

F2.1 $944 \pm 9 \text{ Ma}$



b)

D3.1 $937 \pm 9 \text{ Ma}$



D3.2 $1151 \pm 31 \text{ Ma}$

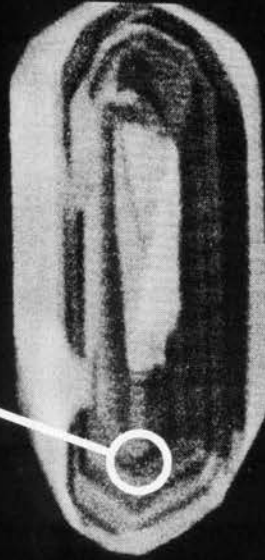
(disc.)



50 μ

c)

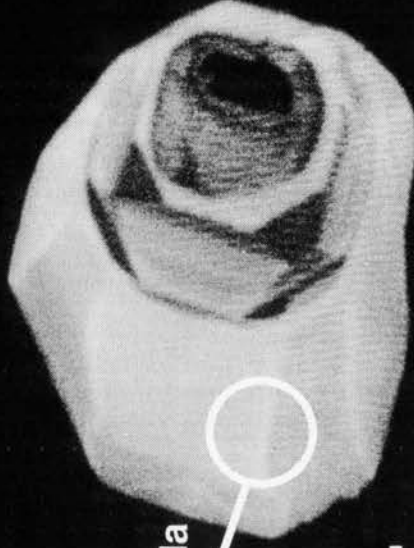
D6.1 $975 \pm 10 \text{ Ma}$



50 μ

d)

D8.1 $946 \pm 10 \text{ Ma}$



50 μ

measured concentrations of ^{204}Pb were comparably low between analyses of the CZ3 standard and of the samples. The common Pb correction of Compston *et al.* (1984) was applied, using the isotopic composition of Broken Hill common Pb ($^{204}\text{Pb}/^{206}\text{Pb} = 0.0625$, $^{207}\text{Pb}/^{206}\text{Pb} = 0.9618$, $^{208}\text{Pb}/^{206}\text{Pb} = 2.2285$). Common ^{206}Pb was accounted for by using a ^{207}Pb correction in all cases. The isotopic decay constants used are those recommended by Steiger & Jager (1978).

Concordance is defined as the overlap of $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ ages for each analysis, within 2σ error. The age estimates for the samples were obtained by pooling analyses to produce a weighted mean ratio, where individual analyses are weighted according to the inverse square of their analytical error. Data with the greatest deviance from the weighted mean ratio were separated from the population until the remaining data lay within 2σ of the weighted mean. This method was applied to $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ ratios independently. A chi-square test was also applied to each dataset, to distinguish geologically significant variance from analytical error. The chi-square parameter is given by:

$$\chi^2 = \frac{1}{(n-1)} \times \sum_{i=1}^n \frac{(x_i - w_n)^2}{(\sigma_n)^2 + (\sigma_i)^2}$$

where x_i is the value of analysis i , n is the number of pooled analyses, w_n is the weighted mean of all pooled x values, σ_n is the error in w_n calculated by addition of all weighted individual errors in quadrature, and σ_i is the error in x_i . Any outlying data were removed from the population until the chi-square parameter was reduced to a value as near as possible to, but not below, unity. A chi-square of unity indicates that the population has a normal distribution, with errors that can be attributed to analytical error.

The resolution of multiple age components from a complex isotopic dataset is a simple process if the data are distributed in discrete populations, between which there is little or no overlap, within error. If this is not the case, it must be determined if the distribution represents a set of discrete ages with overlapping errors, or if the data are

spread indeterminately, with each age analysis being controlled by more than one process. The latter would be the case if, for example, the zircon has been partially recrystallized, or has lost or gained Pb without re-equilibrating.

A conventional approach to this problem is to place data into separate groups, according to zircon morphology and/or composition, and attempt to obtain statistically valid distributions for each group using the ' 2σ ' and 'chi-square' methods discussed above. However, the criteria for defining groups are often ambiguous, especially if the zircon has been affected by Pb diffusion or recrystallization. Zircon that has been isotopically disturbed may yield age data that are not related to the time of zircon growth. Such data is typically discarded with this method of age estimation.

An alternative method for resolving statistically valid populations in a dataset is the mixture modelling approach by Sambridge & Compston (1994). This study applied the technique to four samples, utilising the MIX program for Macintosh (Sambridge & Gallagher, 1995). A given number of model age components, n_c , with symmetrical error distributions, are fitted to the dataset. Best fit to the dataset for n_c components is determined by maximising a likelihood function, L , which is derived from the probability density of each model component at each age datum included in the set. The maximum likelihood function is obtained by solving for L through a series of reductive iterations, starting with n_c random component ages and their relative proportions and shifting them until they converge on a solution with the maximum L .

The maximum likelihood function is expressed in this study as the function $Misfit = -\ln(L)$. The number of components that can be resolved by this technique is indicated by two factors:

- 1) The model age components do not overlap each other within the 95% confidence limit.

2) The *Misfit* for n_c components is significantly smaller than that for n_c-1 components, but is not significantly larger than the *Misfit* for n_c+1 components.

In this study, the addition of another model component is not considered significant if it reduces the *Misfit* by less than 20%. MIX was applied to four samples, with model ages determined for 50 iterations from 20 random starting solutions in the range 800 to 1100 Ma. As the errors on individual isotope analyses are analytical, Gaussian error distributions were assumed for the model age components.

Datasets from the four samples were used that contained $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ ages with 1σ errors. All four datasets contained a spread of overlapping age data with mean values in the range 850-1050 Ma, along with data with widely varied isotopic values. To reduce the analyses to a dataset to which the mixture modelling technique could be applied, the following selection criteria were applied:

1) Analyses that were derived from cores with anhedral forms and/or truncated growth zones were excluded. This reduced the likelihood of including analyses of xenocrystic zircon in the dataset.

2) To reduce the effect of discordance caused by recent Pb loss or gain, data were excluded that had $^{206}\text{Pb}/^{238}\text{U}$ ages which differed by more than 2σ from corresponding $^{207}\text{Pb}/^{206}\text{Pb}$ ages.

3) To eliminate data that were widely divergent from the main group, the few $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ ages with mean values outside the range 850-1050 Ma were excluded. Most of these outliers represent zircon inherited as xenocrystal cores.

The geological significance of the model ages obtained by the mixture modelling technique is determined from zircon morphologies and compositions, and from structural and stratigraphic relationships of the analysed samples in the Cape Bruce locality.

4.6 Analytical results

Results are summarised in Tables 4.1 to 4.6 and Figs 4.4 to 4.9. All errors on individual analyses are quoted at 1σ . Analyses are labelled by sample letter, grain number and number of analysis within the grain (*e.g.* analysis C3.2 is the second analysis of grain 3 from sample C).

Paragneiss (sample P)

Twenty nine zircons from the paragneiss were analysed (Table 4.1, Fig. 4.4). The majority of analyses lie within 2σ of the section of the concordia between 850 and 1050 Ma. Of the six analyses of zircon cores with truncated growth zones, three are discordant and older than the other analyses, the oldest (P21.1) with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1340 ± 74 Ma. Among the analyses of strongly luminescent zircon, four are normally discordant, with $^{207}\text{Pb}/^{206}\text{Pb}$ ages slightly older than most of the data.

Charnockitic orthogneiss (sample C)

Fifty six analyses are spread over a wide range of ages, with most of the data being concordant between 850 and 1100 Ma (Table 4.2, Figs 4.5a & 4.5b). Zircon cores with anhedral forms and/or truncated zoning were analysed, yielding fourteen concordant $^{206}\text{Pb}/^{238}\text{U}$ ages between 850 and 1600 Ma, and three discordant ages, with analysis C19.1 having a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2102 ± 16 Ma. Nineteen rims and grains of weakly luminescent zircon without growth zones were analysed, yielding concordant analyses with ages around 980 Ma. Three of these analyses are reverse discordant, with analysis C24.2 having a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1192 ± 12 Ma. Twenty three strongly zoned rims and grains were analysed, and have concordant ages between 850 and 1100 Ma.

Table 4.1. Ion-microprobe analyses of zircon from sample P, *paragneiss*. All errors are quoted at 1σ .

Common $^{206}\text{Pb}(\%)$ = percentage of total ^{206}Pb that is common ^{206}Pb .

* Corrected for common Pb using ^{204}Pb and the Pb composition of Broken Hill Feldspar (surface contaminant).

Conc.(%) = % concordance = $100 \times (^{206}\text{Pb}/^{238}\text{U}) / (^{207}\text{Pb}/^{206}\text{Pb})$.

Table 4.1. Zircon Analyses from Sample P *Paragneiss*

Grain.spot	U (ppm)	Th (ppm)	Pb (ppm)	Common								Conc* (%)	Apparent Ages (Ma)			
				206Pb (%)	207Pb* 206Pb	±1σ	206Pb* 238Pb	±1σ	207Pb* 235Pb	±1σ	206Pb* 238Pb		±1σ	207Pb* 206Pb	±1σ	
Unzoned and zoned , strongly luminescent grains																
P1.1	252	92	37	0.880	0.07206	0.00313	0.1386	0.0042	1.377	0.077	85	837	24	988	89	
P2.1	173	87	29	1.419	0.06902	0.00345	0.1524	0.0046	1.451	0.089	102	915	26	899	103	
P3.1	188	90	30	0.263	0.07372	0.00485	0.1490	0.0045	1.515	0.115	87	896	26	1034	133	
P4.1	313	47	51	0.935	0.07097	0.00248	0.1621	0.0049	1.586	0.077	101	969	27	957	72	
P5.1	337	78	41	0.056	0.06956	0.00179	0.1217	0.0036	1.167	0.049	81	741	21	915	53	
P6.1	309	82	48	0.705	0.07021	0.00184	0.1544	0.0046	1.495	0.063	99	926	26	935	54	
P10.1	204	96	32	1.529	0.07097	0.00330	0.1438	0.0043	1.407	0.082	91	866	25	957	95	
P12.1	196	91	33	1.277	0.07248	0.00266	0.1536	0.0016	1.535	0.060	92	922	9	1000	75	
P13.1	299	69	49	1.033	0.06802	0.00229	0.1597	0.0017	1.498	0.054	110	955	10	869	70	
P14.1	184	88	31	1.520	0.06815	0.00393	0.1542	0.0017	1.449	0.087	106	925	10	873	120	
P15.1	644	25	103	0.428	0.07391	0.00128	0.1674	0.0017	1.706	0.036	96	998	10	1039	35	
P16.1	308	61	51	3.135	0.06691	0.00330	0.1538	0.0016	1.419	0.073	110	922	9	835	103	
P17.1	346	44	57	1.747	0.06890	0.00227	0.1631	0.0017	1.550	0.055	109	974	10	896	68	
P18.1	283	63	44	1.178	0.06831	0.00215	0.1541	0.0016	1.452	0.050	105	924	9	878	65	
P19.1	183	85	30	0.754	0.07637	0.00314	0.1514	0.0016	1.594	0.070	82	909	9	1105	82	
P20.1	283	82	35	0.157	0.07614	0.00220	0.1237	0.0012	1.299	0.041	68	752	7	1099	58	
P23.1	152	90	25	1.191	0.06706	0.00261	0.1506	0.0017	1.393	0.058	108	905	10	840	81	
P24.1	248	78	37	0.713	0.06764	0.00161	0.1478	0.0016	1.379	0.037	104	889	9	858	50	
P25.1	174	95	29	0.965	0.06803	0.00235	0.1521	0.0017	1.427	0.053	105	913	10	870	72	
P26.1	155	145	28	0.570	0.07100	0.00208	0.1546	0.0017	1.513	0.049	97	927	10	958	60	
P27.1	313	58	48	0.638	0.07062	0.00155	0.1542	0.0016	1.501	0.038	98	925	9	947	45	
P28.1	91	135	19	1.997	0.07192	0.00517	0.1497	0.0019	1.484	0.110	91	900	11	984	147	
P29.1	222	72	36	0.366	0.07337	0.00152	0.1575	0.0017	1.594	0.039	92	943	10	1024	42	
Anhedral cores, with or without truncated growth zones																
P7.1	304	8	40	0.044	0.07554	0.00103	0.1402	0.0042	1.460	0.050	78	846	24	1083	28	
P8.1	452	53	58	0.052	0.07130	0.00175	0.1337	0.0040	1.315	0.054	84	809	23	967	50	
P9.1	380	28	56	0.393	0.07143	0.00222	0.1529	0.0046	1.506	0.069	95	917	26	970	64	
P11.1	517	76	80	0.449	0.07229	0.00173	0.1579	0.0016	1.573	0.042	95	945	9	994	49	
P21.1	251	87	54	0.640	0.08607	0.00163	0.1956	0.0020	2.321	0.053	86	1152	11	1340	37	
P22.1	361	20	53	0.915	0.07655	0.00180	0.1487	0.0015	1.570	0.042	81	894	9	1110	47	

Fig. 4.4. Concordia plot of zircon analyses from sample P, *paragneiss*. Error hexagons are 1σ .

Sample P *Paragneiss*

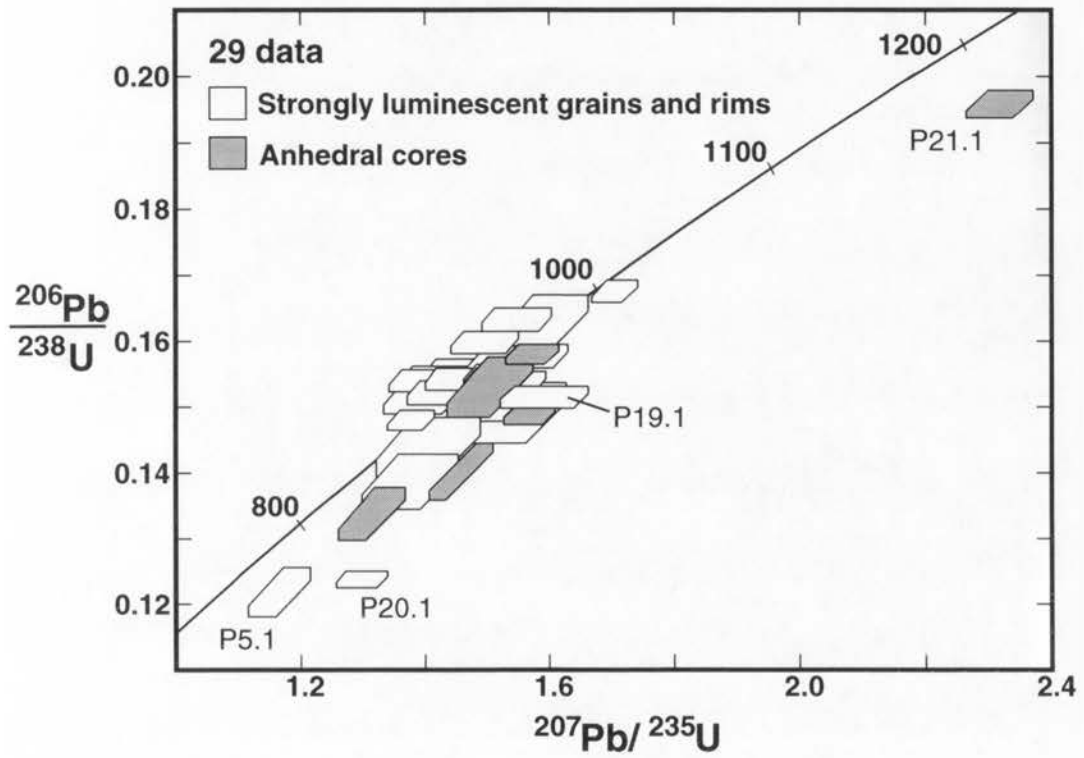


Table 4.2. Ion-microprobe analyses of zircon from sample C, *charnockitic orthogneiss*. All errors are quoted at 1σ .

Common $^{206}\text{Pb}(\%)$ = percentage of total ^{206}Pb that is common ^{206}Pb .

* Corrected for common Pb using ^{204}Pb and the Pb composition of Broken Hill Feldspar (surface contaminant).

Conc.(%) = % concordance = $100 \times (^{206}\text{Pb}/^{238}\text{U}) / (^{207}\text{Pb}/^{206}\text{Pb})$.

Table 4.2. Zircon Analyses from Sample C *Charnockitic othogneiss*

Grain.spot	Common										Conc* (%)	Apparent Ages (Ma)			
	U (ppm)	Th (ppm)	Pb (ppm)	²⁰⁶ Pb (%)	$\frac{^{207}\text{Pb}^*}{^{206}\text{Pb}}$	$\pm 1\sigma$	$\frac{^{206}\text{Pb}^*}{^{238}\text{Pb}}$	$\pm 1\sigma$	$\frac{^{207}\text{Pb}^*}{^{235}\text{Pb}}$	$\pm 1\sigma$		$\frac{^{206}\text{Pb}^*}{^{238}\text{Pb}}$	$\pm 1\sigma$	$\frac{^{207}\text{Pb}^*}{^{206}\text{Pb}}$	$\pm 1\sigma$
Zoned , weakly luminescent cores and rims															
C9.1	799	224	134	0.230	0.07093	0.00059	0.1682	0.0025	1.645	0.029	105	1002	14	955	17
C15.1	674	226	112	0.718	0.07116	0.00098	0.1606	0.0024	1.576	0.034	100	960	13	962	28
C20.3	693	190	99	0.005	0.07093	0.00034	0.1442	0.0022	1.410	0.024	91	868	13	955	10
C28.1	290	242	59	2.392	0.06959	0.00239	0.1653	0.0018	1.586	0.059	108	986	10	916	71
Zoned and unzoned, strongly luminescent cores and rims															
C2.2	177	108	32	0.000	0.07329	0.00107	0.1693	0.0027	1.711	0.039	99	1008	15	1022	29
C4.1	141	95	25	1.511	0.07112	0.00315	0.1562	0.0024	1.532	0.075	97	936	14	961	90
C5.2	208	98	34	1.025	0.07031	0.00221	0.1506	0.0023	1.460	0.053	96	904	13	938	65
C7.2	256	109	41	1.082	0.06880	0.00203	0.1490	0.0022	1.413	0.049	100	895	13	893	61
C8.2	222	138	40	1.202	0.07156	0.00217	0.1618	0.0025	1.596	0.057	99	967	14	973	62
C10.1	565	180	94	0.554	0.07238	0.00104	0.1616	0.0024	1.613	0.035	97	966	13	997	29
C13.1	362	199	60	0.661	0.06825	0.00134	0.1538	0.0023	1.447	0.038	105	922	13	876	41
C14.2	202	124	41	0.062	0.07327	0.00139	0.1873	0.0030	1.892	0.050	108	1107	16	1021	38
C17.1	240	160	41	1.045	0.07201	0.00205	0.1495	0.0023	1.484	0.050	91	898	13	986	58
C17.3	179	142	32	1.760	0.06766	0.00314	0.1485	0.0023	1.385	0.070	104	892	13	858	96
C18.2	232	134	43	1.894	0.07006	0.00250	0.1636	0.0025	1.581	0.064	105	977	14	930	73
C20.2	174	123	31	0.034	0.07310	0.00138	0.1578	0.0025	1.590	0.041	93	944	14	1017	38
C21.2	151	90	28	1.316	0.07305	0.00256	0.1666	0.0026	1.678	0.067	98	993	14	1015	71
C23.1	186	150	36	1.840	0.07156	0.00268	0.1608	0.0018	1.587	0.064	99	961	10	973	76
C24.1	172	152	33	1.475	0.06787	0.00290	0.1589	0.0025	1.487	0.070	110	951	14	865	89
C26.1	157	123	31	1.306	0.07382	0.00253	0.1696	0.0026	1.727	0.068	97	1010	14	1037	69
C27.1	129	97	26	1.634	0.07443	0.00361	0.1697	0.0027	1.742	0.092	96	1011	15	1053	98
C30.2	161	114	32	2.706	0.07141	0.00356	0.1659	0.0019	1.634	0.086	102	990	11	969	102
C31.1	169	105	32	0.000	0.07063	0.00077	0.1733	0.0028	1.688	0.034	109	1030	15	947	22
Unzoned , weakly luminescent rims and grains															
C3.1	4390	54	698	0.089	0.07229	0.00019	0.1714	0.0025	1.708	0.026	103	1020	14	994	5
C4.2	1057	171	164	0.328	0.07001	0.00057	0.1601	0.0024	1.545	0.027	103	957	13	929	17
C6.1	3571	101	557	0.064	0.07136	0.00020	0.1676	0.0025	1.649	0.025	103	999	14	968	6
C9.2	1089	145	174	0.278	0.07086	0.00052	0.1664	0.0025	1.626	0.028	104	992	14	953	15
C11.2	2827	53	431	0.091	0.07175	0.00026	0.1640	0.0024	1.622	0.025	100	979	13	979	7
C13.2	2484	29	367	0.225	0.07187	0.00032	0.1585	0.0023	1.570	0.025	97	948	13	982	9
C15.2	1793	167	282	0.145	0.07163	0.00034	0.1658	0.0024	1.637	0.026	101	989	13	975	10
C17.2	2205	397	351	0.125	0.07190	0.00030	0.1637	0.0024	1.623	0.026	99	977	13	983	8
C22.2	2488	32	427	0.001	0.07236	0.00020	0.1856	0.0029	1.852	0.030	110	1098	16	996	6
C23.2	1061	163	176	0.339	0.07228	0.00054	0.1698	0.0017	1.692	0.023	102	1011	9	994	15
C24.2	2459	25	757	0.181	0.07978	0.00023	0.3283	0.0048	3.611	0.055	154	1830	23	1192	6
C25.1	4264	60	669	0.068	0.07187	0.00018	0.1692	0.0025	1.677	0.025	103	1008	14	982	5
C26.2	2266	31	343	0.113	0.07005	0.00028	0.1633	0.0024	1.578	0.025	105	975	13	930	8
C27.2	4031	53	667	0.091	0.07189	0.00022	0.1782	0.0026	1.767	0.027	108	1057	14	983	6
C29.1	2700	41	413	0.142	0.07119	0.00027	0.1644	0.0017	1.613	0.018	102	981	9	963	8
C30.1	1099	22	164	0.411	0.07043	0.00057	0.1590	0.0016	1.544	0.021	101	951	9	941	17
Anhedral and Irregular cores, with or without truncated growth zones															
C1.2	511	256	97	0.249	0.07731	0.00070	0.1787	0.0026	1.905	0.035	94	1060	14	1129	18
C2.1	1151	504	264	0.008	0.08331	0.00027	0.2195	0.0034	2.522	0.041	100	1279	18	1277	6
C5.1	497	284	91	0.383	0.07454	0.00090	0.1693	0.0025	1.740	0.035	95	1008	14	1056	24
C8.1	235	122	43	1.244	0.07214	0.00211	0.1667	0.0025	1.658	0.057	100	994	14	990	60
C16.1	1087	307	196	0.325	0.07539	0.00053	0.1792	0.0026	1.862	0.032	98	1062	14	1079	14
C18.1	496	155	98	0.409	0.07969	0.00081	0.1936	0.0029	2.127	0.040	96	1141	16	1189	20
C21.1	484	160	101	0.350	0.08114	0.00076	0.2032	0.0030	2.273	0.042	97	1192	16	1225	19
C1.1	189	112	31	2.185	0.06466	0.00322	0.1426	0.0016	1.271	0.066	113	859	9	763	105
C7.1	418	227	75	0.969	0.07047	0.00133	0.1649	0.0025	1.602	0.041	104	984	14	942	39
C11.1	1020	149	168	0.326	0.07300	0.00059	0.1702	0.0025	1.713	0.030	100	1013	14	1014	16
C12.1	495	164	77	0.532	0.07043	0.00099	0.1513	0.0022	1.469	0.032	97	908	13	941	29
C14.1	636	384	191	0.042	0.09342	0.00043	0.2724	0.0043	3.508	0.059	104	1553	22	1496	9
C19.1	519	48	185	0.236	0.13032	0.00056	0.3546	0.0053	6.372	0.102	93	1956	25	2102	8
C19.2	287	111	52	0.111	0.07821	0.00084	0.1754	0.0028	1.891	0.038	90	1042	15	1152	21
C20.1	1243	449	210	0.102	0.07486	0.00032	0.1651	0.0026	1.704	0.028	93	985	14	1065	9
C22.1	875	184	129	0.504	0.06804	0.00077	0.1492	0.0015	1.399	0.023	103	896	9	870	23
C31.2	1305	169	217	0.042	0.07298	0.00034	0.1739	0.0027	1.750	0.029	102	1033	15	1013	9

Fig. 4.5. Concordia plots of zircon analyses from sample C, *charnockitic orthogneiss*. Error hexagons are 1σ .

a) All age data.

b) Age data in the range 800-1300Ma.

Sample C *Charnockitic orthogneiss*

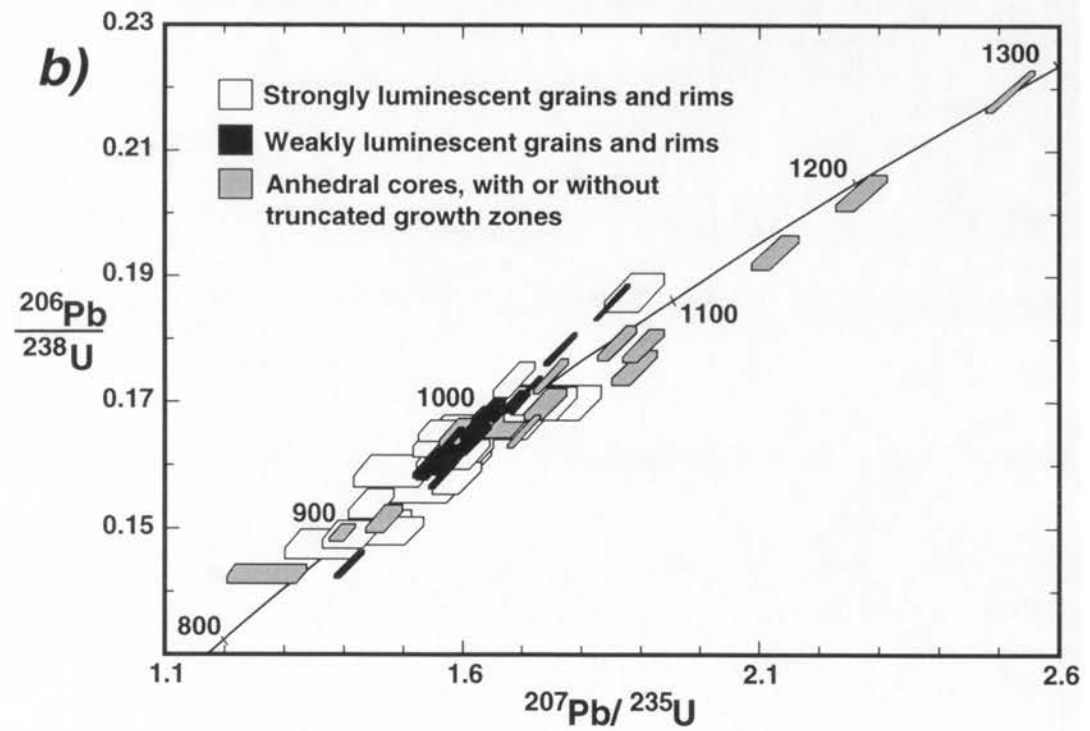
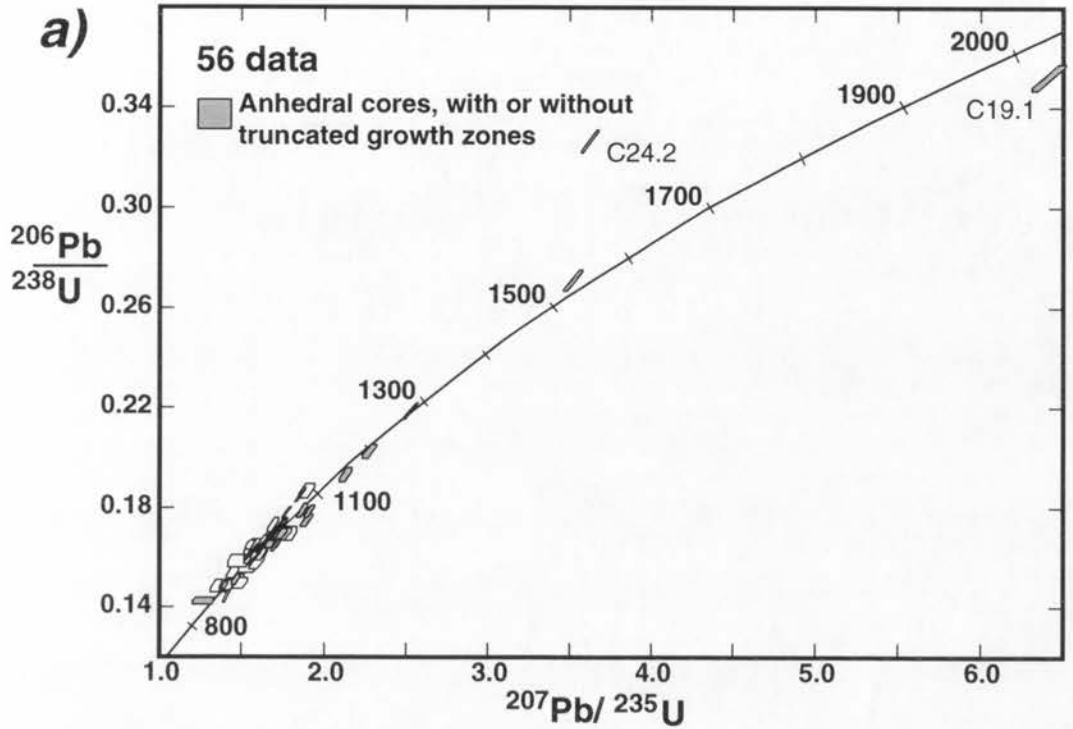


Table 4.3. Ion-microprobe analyses of zircon from sample B, *biotite-pyroxene orthogneiss*. All errors are quoted at 1σ .

Common $^{206}\text{Pb}(\%)$ = percentage of total ^{206}Pb that is common ^{206}Pb .

* Corrected for common Pb using ^{204}Pb and the Pb composition of Broken Hill Feldspar (surface contaminant).

Conc.(%) = % concordance = $100 \times (^{206}\text{Pb}/^{238}\text{U}) / (^{207}\text{Pb}/^{206}\text{Pb})$.

Table 4.3. Zircon Analyses from Sample B *Biotite-pyroxene orthogneiss dyke*

Grain.spot	U (ppm)	Th (ppm)	Pb (ppm)	Common							Conc* (%)	Apparent Ages (Ma)			
				²⁰⁶ Pb (%)	²⁰⁷ Pb* / ²⁰⁶ Pb	± 1σ	²⁰⁶ Pb* / ²³⁸ Pb	± 1σ	²⁰⁷ Pb* / ²³⁵ Pb	± 1σ		²⁰⁶ Pb* / ²³⁸ Pb	± 1σ	²⁰⁷ Pb* / ²⁰⁶ Pb	± 1σ
B1.1	526	26	80	0.157	0.07358	0.00103	0.1602	0.0030	1.626	0.040	93	958	17	1030	28
B1.2	535	1413	150	0.074	0.07633	0.00260	0.1707	0.0033	1.797	0.074	92	1016	18	1104	68
B2.1	953	561	179	0.221	0.07254	0.00060	0.1702	0.0032	1.702	0.037	101	1013	18	1001	17
B3.1	434	28	72	3.158	0.05776	0.00286	0.1631	0.0031	1.299	0.072	187	974	17	521	109
B3.2	526	32	81	0.786	0.07394	0.00149	0.1591	0.0030	1.622	0.048	92	952	17	1040	41
B4.1	698	1310	160	0.336	0.07259	0.00109	0.1662	0.0031	1.664	0.042	99	991	17	1002	30
B5.1	314	342	62	0.003	0.07816	0.00069	0.1664	0.0032	1.793	0.039	86	992	18	1151	17
B6.1	942	63	152	0.454	0.07142	0.00073	0.1687	0.0032	1.661	0.037	104	1005	18	970	21
B7.1	667	18	107	0.653	0.07077	0.00095	0.1686	0.0032	1.646	0.040	106	1005	18	951	27
B8.1	3307	1382	593	0.146	0.07160	0.00027	0.1736	0.0033	1.714	0.034	106	1032	18	974	8
B9.1	475	26	77	1.043	0.07186	0.00156	0.1674	0.0032	1.658	0.051	102	998	18	982	44
B10.1	2805	354	461	0.056	0.07222	0.00029	0.1707	0.0032	1.700	0.034	102	1016	18	992	8
B11.1	2964	8272	852	0.150	0.07195	0.00027	0.1741	0.0033	1.727	0.034	105	1035	18	985	8
B11.2	1870	2544	418	0.053	0.07209	0.00052	0.1706	0.0032	1.696	0.036	103	1016	18	989	15
B12.1	2624	10640	883	0.171	0.07299	0.00038	0.1693	0.0032	1.704	0.034	99	1008	18	1014	10
B13.1	562	107	99	2.318	0.06939	0.00204	0.1663	0.0032	1.591	0.059	109	992	18	910	61

Fig. 4.6. Concordia plot of zircon analyses from sample B, *biotite-pyroxene orthogneiss*. Error hexagons are 1σ .

Sample B *Biotite-pyroxene orthogneiss*

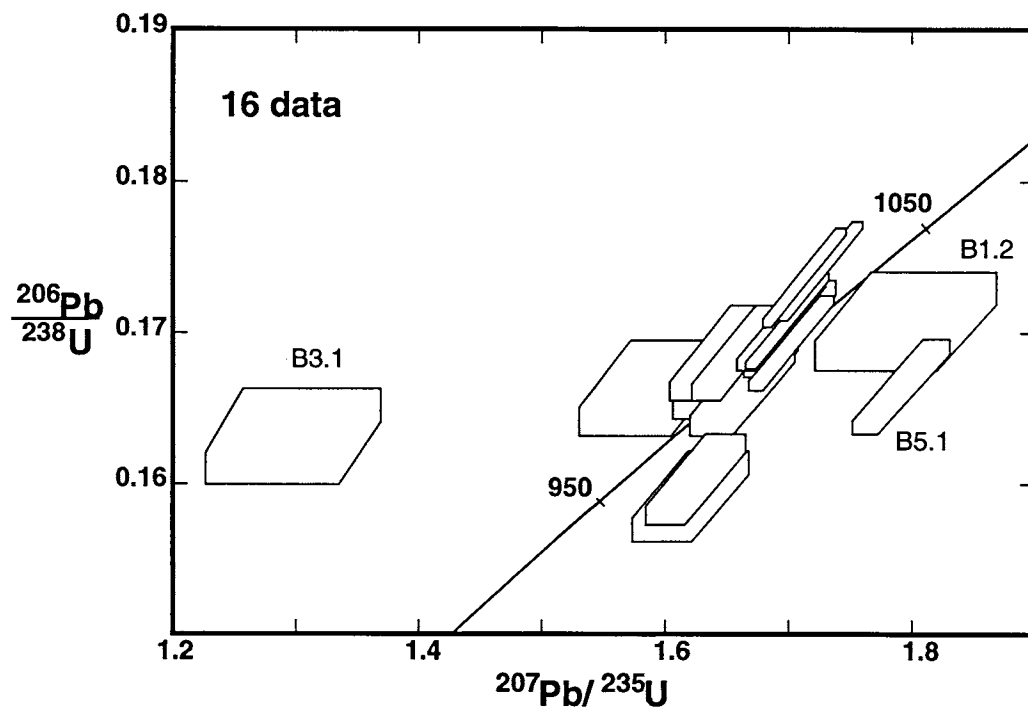


Table 4.4. Ion-microprobe analyses of zircon from sample R, *red granitic orthogneiss*. All errors are quoted at 1σ .

Common $^{206}\text{Pb}(\%)$ = percentage of total ^{206}Pb that is common ^{206}Pb .

* Corrected for common Pb using ^{204}Pb and the Pb composition of Broken Hill Feldspar (surface contaminant).

Conc.(%) = % concordance = $100 \times (^{206}\text{Pb}/^{238}\text{U}) / (^{207}\text{Pb}/^{206}\text{Pb})$.

Table 4.4. Zircon Analyses from Sample R *Red granitic orthogneiss*

Grain.spot	Common											Apparent Ages (Ma)			
	U (ppm)	Th (ppm)	Pb (ppm)	206Pb (%)	207Pb* 206Pb	± 1σ	206Pb* 238Pb	± 1σ	207Pb* 235Pb	± 1σ	Conc* (%)	206Pb*	± 1σ	207Pb*	± 1σ
												238Pb	206Pb	235Pb	206Pb
Zoned , weakly luminescent cores and rims															
R1.2	636	185	98	0.093	0.07071	0.00065	0.1541	0.0018	1.503	0.024	97	924	10	949	19
R2.2	694	420	128	0.301	0.07150	0.00069	0.1695	0.0035	1.671	0.040	104	1009	19	972	20
R9.2	3686	335	627	0.081	0.07252	0.00023	0.1796	0.0018	1.796	0.020	106	1065	10	1000	6
R10.3	1570	425	264	0.183	0.07168	0.00036	0.1682	0.0017	1.662	0.020	103	1002	9	977	10
R12.1	1193	1754	258	0.157	0.07224	0.00041	0.1648	0.0017	1.642	0.020	99	984	9	993	12
R13.1	456	913	113	0.530	0.07164	0.00098	0.1716	0.0018	1.695	0.031	105	1021	10	976	28
R14.2	1759	274	292	0.114	0.07141	0.00029	0.1727	0.0017	1.700	0.019	106	1027	10	969	8
R15.1	1293	363	214	0.001	0.07204	0.00029	0.1666	0.0019	1.655	0.021	101	993	11	987	8
R19.2	1243	229	191	0.120	0.07008	0.00042	0.1580	0.0018	1.527	0.021	102	946	10	931	12
R21.1	730	248	116	0.175	0.07053	0.00065	0.1577	0.0018	1.534	0.024	100	944	10	944	19
R21.3	1225	286	200	0.129	0.07198	0.00040	0.1653	0.0019	1.641	0.022	100	986	11	985	11
R22.1	578	189	95	0.175	0.07199	0.00067	0.1632	0.0019	1.620	0.026	99	975	11	986	19
R25.2	911	142	132	0.127	0.06900	0.00050	0.1511	0.0017	1.438	0.021	101	907	10	899	15
R26.2	2021	459	333	0.049	0.07158	0.00025	0.1678	0.0019	1.656	0.021	103	1000	11	974	7
Zoned and unzoned , strongly luminescent cores and rims															
R2.1	165	148	32	1.925	0.06648	0.00317	0.1584	0.0034	1.452	0.079	115	948	19	822	100
R3.2	355	219	53	0.847	0.06705	0.00175	0.1333	0.0028	1.233	0.044	96	807	16	839	54
R4.1	403	334	77	0.386	0.07127	0.00099	0.1658	0.0017	1.629	0.030	102	989	9	965	28
R5.2	266	155	48	1.266	0.06852	0.00207	0.1629	0.0018	1.539	0.051	110	973	10	884	62
R6.1	166	106	31	1.582	0.06700	0.00309	0.1629	0.0019	1.505	0.074	116	973	11	838	96
R8.1	206	243	45	1.360	0.06813	0.00237	0.1710	0.0019	1.606	0.061	117	1018	11	873	72
R9.1	284	311	57	1.221	0.06752	0.00200	0.1629	0.0018	1.516	0.050	114	973	10	854	62
R10.2	153	128	30	1.395	0.07429	0.00278	0.1649	0.0019	1.689	0.068	94	984	10	1049	75
R18.1	164	132	31	1.589	0.06364	0.00270	0.1637	0.0020	1.437	0.066	134	978	11	730	90
R20.1	334	283	54	0.429	0.06497	0.00135	0.1418	0.0017	1.270	0.032	111	855	9	773	44
R22.2	259	170	48	0.706	0.07163	0.00165	0.1684	0.0020	1.663	0.045	103	1003	11	975	47
R23.1	96	99	19	0.985	0.06862	0.00361	0.1647	0.0021	1.558	0.087	111	983	12	887	109
R24.1	279	248	50	0.320	0.06842	0.00116	0.1548	0.0018	1.460	0.032	105	928	10	881	35
Unzoned , weakly luminescent rims and grains															
R3.3	1175	243	189	0.217	0.07025	0.00050	0.1642	0.0034	1.590	0.036	105	980	19	936	15
R3.4	1240	270	203	0.151	0.07151	0.00044	0.1666	0.0034	1.643	0.036	102	994	19	972	12
R4.2	486	54	79	0.412	0.07270	0.00096	0.1694	0.0017	1.698	0.030	100	1009	10	1006	27
R7.2	1135	242	181	0.277	0.06960	0.00058	0.1621	0.0017	1.555	0.022	106	968	9	917	17
R11.2	1631	308	267	0.205	0.07190	0.00035	0.1675	0.0017	1.660	0.020	102	998	9	983	10
R17.2	1132	185	180	0.126	0.07163	0.00041	0.1645	0.0019	1.625	0.022	101	982	10	975	12
R21.2	2348	701	407	0.062	0.07183	0.00025	0.1736	0.0020	1.719	0.021	105	1032	11	981	7
R24.2	1352	287	217	0.086	0.07126	0.00038	0.1644	0.0019	1.615	0.021	102	981	10	965	11
Anhedral and irregular cores, with or without truncated growth zones															
R5.1	74	35	26	1.576	0.15085	0.00341	0.3038	0.0042	6.319	0.176	73	1710	21	2356	39
R7.1	1106	365	216	0.318	0.07677	0.00058	0.1894	0.0019	2.004	0.027	100	1118	11	1115	15
R10.1	471	155	107	0.473	0.11141	0.00097	0.2148	0.0023	3.300	0.048	69	1254	12	1823	16
R25.1	513	172	101	0.312	0.09523	0.00083	0.1896	0.0022	2.490	0.038	73	1119	12	1533	16
R1.1	167	79	76	0.317	0.15257	0.00122	0.4107	0.0052	8.639	0.137	93	2218	24	2375	14
R3.1	419	180	69	0.463	0.08908	0.00124	0.1553	0.0032	1.908	0.050	66	931	18	1406	27
R11.1	164	120	51	1.331	0.13742	0.00210	0.2576	0.0029	4.880	0.098	67	1477	15	2195	27
R14.1	4760	1056	853	0.042	0.07181	0.00015	0.1831	0.0018	1.813	0.019	111	1084	10	980	4
R16.1	1751	114	245	0.653	0.06867	0.00058	0.1456	0.0017	1.378	0.021	99	876	9	889	17
R17.1	1294	162	191	0.103	0.06998	0.00036	0.1544	0.0018	1.490	0.020	100	926	10	928	11
R19.1	1177	59	321	0.057	0.10741	0.00032	0.2825	0.0033	4.184	0.051	91	1604	16	1756	5
R26.1	1827	77	272	0.081	0.06975	0.00033	0.1598	0.0018	1.537	0.020	104	956	10	921	10

Fig. 4.7. Concordia plots of zircon analyses from sample R, *red granitic orthogneiss*. Error hexagons are 1σ .

a) All age data.

b) Age data in the range 800-1200Ma.

Sample R Red granitic orthogneiss

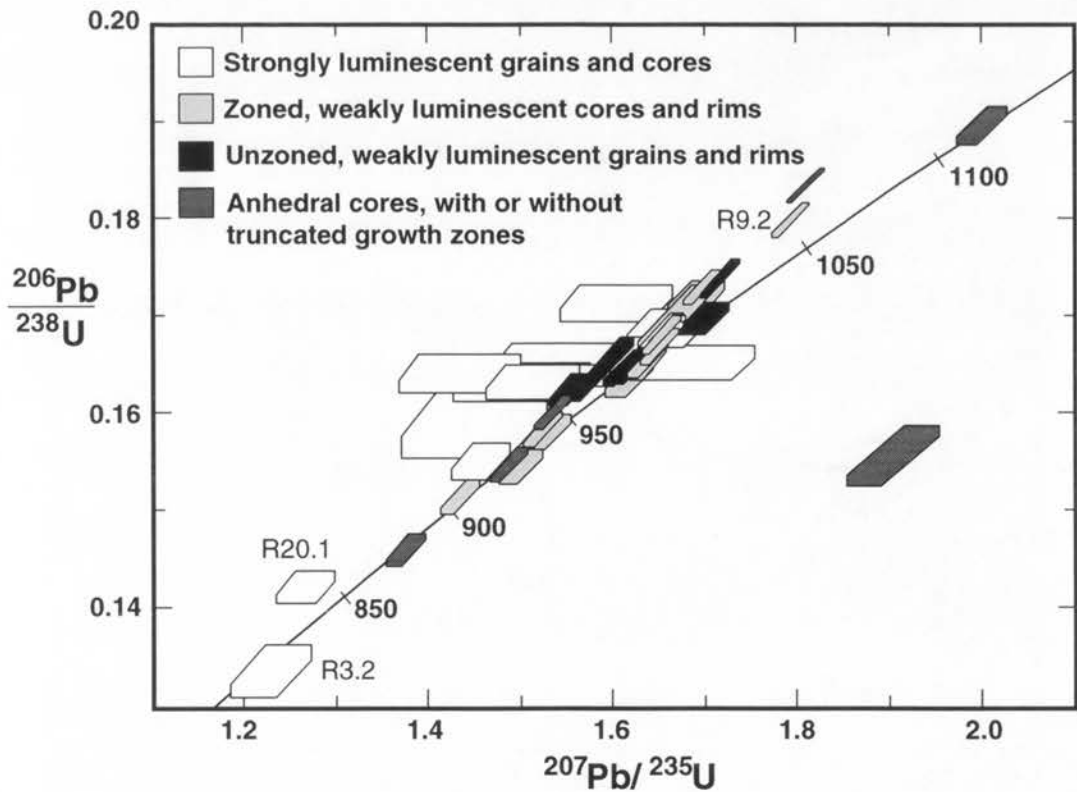
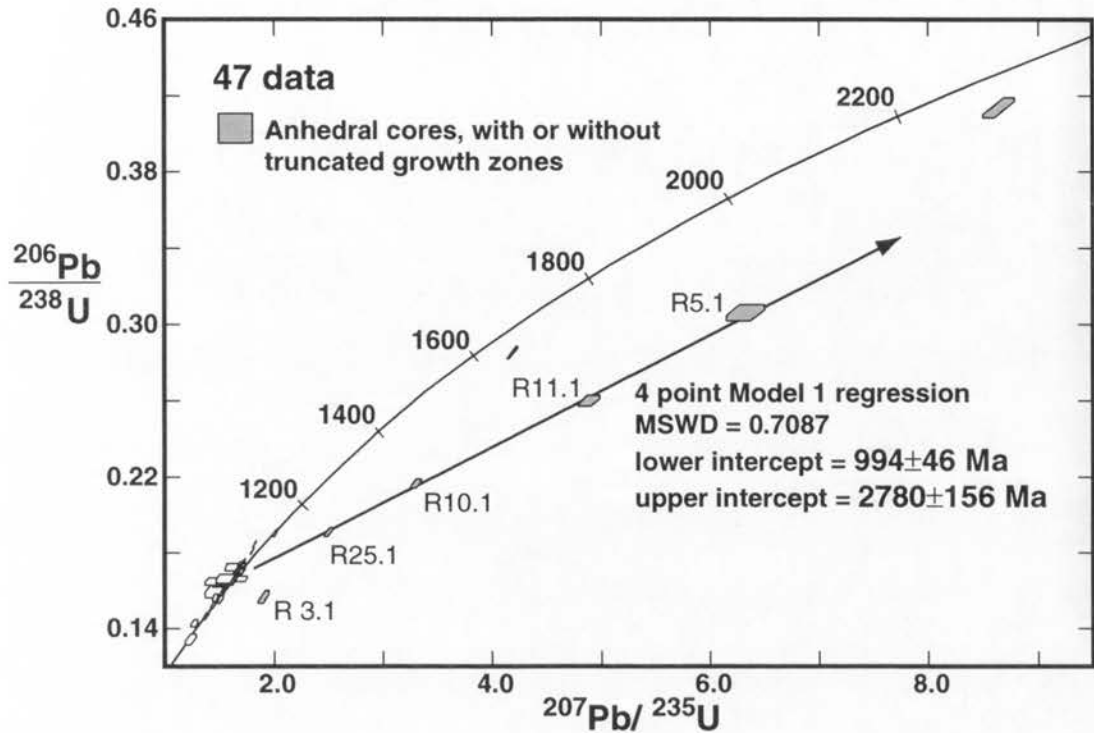


Table 4.5. Ion-microprobe analyses of zircon from sample F, a dyke of *pre-D₃ felsic orthogneiss*. All errors are quoted at 1σ .

Common $^{206}\text{Pb}(\%)$ = percentage of total ^{206}Pb that is common ^{206}Pb .

* Corrected for common Pb using ^{204}Pb and the Pb composition of Broken Hill Feldspar (surface contaminant).

Conc.(%) = % concordance = $100 \times (^{206}\text{Pb}/^{238}\text{U}) / (^{207}\text{Pb}/^{206}\text{Pb})$.

Table 4.5. Zircon Analyses from Sample F *Pre-D₃ felsic orthogneiss dyke*

Grain.spot	Common										Conc* (%)	Apparent Ages (Ma)			
	U (ppm)	Th (ppm)	Pb (ppm)	206Pb (%)	207Pb*	+/- 1σ	206Pb*	+/- 1σ	207Pb*	+/- 1σ		206Pb*	+/- 1σ	207Pb*	+/- 1σ
					206Pb		238Pb		235Pb			238Pb		206Pb	
Unzoned and zoned , strongly luminescent grains and rims															
F1.2	120	40	21	3.548	0.06479	0.00475	0.1564	0.0019	1.397	0.106	122	937	10	768	155
F2.2	139	49	24	0.060	0.06755	0.00412	0.1556	0.0018	1.449	0.092	109	932	10	855	127
F3.1	120	40	20	2.871	0.06623	0.00396	0.1529	0.0018	1.396	0.087	113	917	10	814	125
F4.1	155	58	27	2.356	0.07135	0.00229	0.1619	0.0018	1.592	0.056	100	967	10	967	66
F4.2	136	54	23	0.015	0.06540	0.00389	0.1566	0.0018	1.412	0.087	119	938	10	787	125
F4.4	109	35	17	0.096	0.06915	0.00132	0.1591	0.0029	1.517	0.042	105	952	16	903	39
F5.2	131	44	22	5.829	0.07137	0.00153	0.1590	0.0029	1.565	0.047	98	951	16	968	44
F7.2	128	42	21	5.227	0.06977	0.00142	0.1594	0.0029	1.533	0.044	103	953	16	922	42
F9.1	403	55	63	0.099	0.06847	0.00127	0.1592	0.0017	1.503	0.034	108	952	9	883	38
F9.2	135	118	26	3.520	0.06946	0.00340	0.1588	0.0018	1.521	0.078	104	950	10	913	101
F9.3	132	44	23	2.584	0.07108	0.00400	0.1574	0.0019	1.542	0.091	98	942	10	960	115
F10.1	111	43	18	2.649	0.07123	0.00105	0.1566	0.0028	1.538	0.038	97	938	16	964	30
F11.1	108	44	20	0.077	0.07482	0.00406	0.1601	0.0019	1.651	0.094	90	957	11	1064	109
F12.1	132	46	21	0.522	0.07155	0.00084	0.1573	0.0029	1.552	0.035	97	942	16	973	24
F12.2	121	39	19	0.184	0.07040	0.00119	0.1546	0.0028	1.501	0.040	99	927	16	940	35
F14.1	173	67	29	0.780	0.06831	0.00370	0.1540	0.0017	1.451	0.082	105	924	10	878	112
F14.2	128	52	25	2.127	0.06413	0.00719	0.1564	0.0020	1.383	0.158	126	937	11	746	239
F15.1	124	46	23	2.602	0.06494	0.00575	0.1562	0.0019	1.399	0.127	121	936	11	772	187
Anhedral and irregular cores, with or without truncated growth zones															
F1.1	176	100	32	1.770	0.06609	0.00273	0.1616	0.0018	1.473	0.065	119	966	10	809	86
F2.1	230	197	44	2.661	0.06997	0.00203	0.1576	0.0017	1.521	0.049	102	944	9	927	59
F4.3	214	120	36	1.438	0.06995	0.00085	0.1569	0.0028	1.513	0.035	101	939	16	927	25
F5.1	200	142	56	1.216	0.06384	0.00757	0.1562	0.0021	1.375	0.166	127	936	11	736	253
F6.1	195	71	32	0.018	0.07092	0.00116	0.1593	0.0029	1.558	0.040	100	953	16	955	33
F7.1	175	64	29	16.047	0.07090	0.00094	0.1598	0.0029	1.562	0.037	100	955	16	955	27
F8.1	233	357	58	0.512	0.07177	0.00388	0.1592	0.0018	1.575	0.089	97	952	10	980	110
F13.1	177	120	34	5.883	0.06934	0.00312	0.1586	0.0018	1.516	0.072	104	949	10	909	93

Table 4.6. Ion-microprobe analyses of zircon from sample D, a *pre-D₄ felsic pegmatite*. All errors are quoted at 1 σ .

Common $^{206}\text{Pb}(\%)$ = percentage of total ^{206}Pb that is common ^{206}Pb .

* Corrected for common Pb using ^{204}Pb and the Pb composition of Broken Hill Feldspar (surface contaminant).

Conc.(%) = % concordance = $100 \times (^{206}\text{Pb}/^{238}\text{U}) / (^{207}\text{Pb}/^{206}\text{Pb})$.

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Table 4.6. Zircon Analyses from Sample D *Pre-D4 felsic pegmatite*

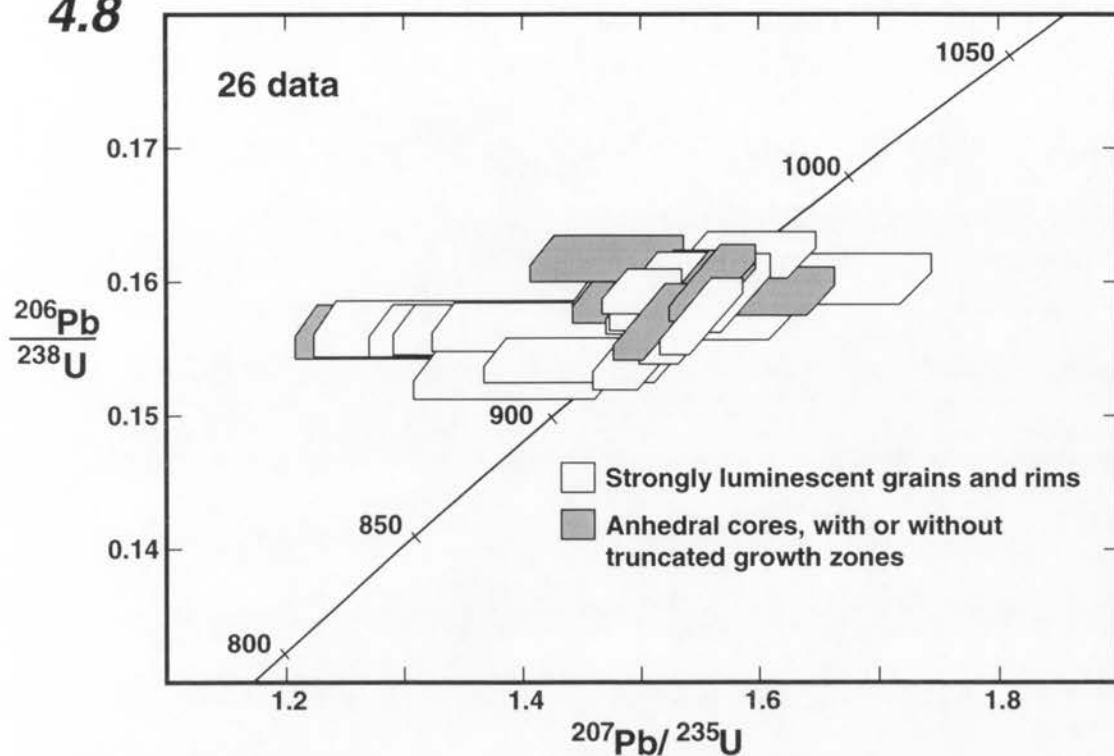
Grain.spot	Common											Apparent Ages (Ma)			
	U (ppm)	Th (ppm)	Pb (ppm)	²⁰⁶ Pb (%)	²⁰⁷ Pb* ²⁰⁶ Pb	+/- 1σ	²⁰⁶ Pb* ²³⁸ Pb	+/- 1σ	²⁰⁷ Pb* ²³⁵ Pb	+/- 1σ	Conc* (%)	²⁰⁶ Pb* ²³⁸ Pb	+/- 1σ	²⁰⁷ Pb* ²⁰⁶ Pb	+/- 1σ
Zoned and unzoned , weakly luminescent cores and grains															
D4.1	306	247	56	0.276	0.07206	0.00122	0.1608	0.0017	1.598	0.034	97	961	9	988	35
D5.1	342	240	63	0.526	0.07061	0.00132	0.1643	0.0017	1.600	0.036	104	981	10	946	38
D11.1	317	227	55	0.147	0.07435	0.00140	0.1562	0.0016	1.602	0.036	89	936	9	1051	38
D15.1	573	413	93	0.085	0.07253	0.00134	0.1464	0.0015	1.464	0.033	88	881	8	1001	38
D16.1	413	364	75	0.231	0.07176	0.00091	0.1576	0.0017	1.559	0.027	96	943	9	979	26
D18.1	632	231	104	0.117	0.07209	0.00100	0.1615	0.0017	1.605	0.029	98	965	9	988	28
D19.1	478	257	84	0.066	0.07346	0.00078	0.1653	0.0017	1.674	0.026	96	986	10	1027	21
D21.1	305	253	53	0.315	0.07521	0.00237	0.1512	0.0016	1.568	0.054	85	908	9	1074	63
D24.1	237	156	43	1.732	0.06946	0.00243	0.1599	0.0018	1.531	0.058	105	956	10	912	72
D25.1	313	251	59	0.749	0.07114	0.00154	0.1648	0.0018	1.616	0.041	102	983	10	961	44
D28.1	556	311	98	0.521	0.07203	0.00111	0.1623	0.0047	1.612	0.056	98	969	26	987	31
Zoned and unzoned, strongly luminescent rims and grains															
D1.1	143	77	26	1.768	0.07358	0.00300	0.1599	0.0018	1.622	0.071	93	956	10	1030	83
D2.1	342	232	62	0.660	0.07090	0.00147	0.1609	0.0017	1.573	0.038	101	962	9	955	42
D6.1	301	183	53	0.578	0.06753	0.00216	0.1633	0.0017	1.520	0.053	114	975	10	854	67
D7.1	236	152	42	1.191	0.06859	0.00208	0.1579	0.0017	1.494	0.050	107	945	10	886	63
D8.1	157	144	29	0.874	0.06704	0.00381	0.1580	0.0018	1.461	0.087	113	946	10	839	119
D9.1	221	114	35	0.904	0.07099	0.00308	0.1475	0.0016	1.443	0.066	93	887	9	957	89
D10.1	215	110	35	0.089	0.07193	0.00123	0.1524	0.0016	1.512	0.032	93	915	9	984	35
D12.1	255	116	40	0.594	0.06863	0.00236	0.1497	0.0016	1.416	0.053	101	899	9	888	71
D13.1	220	328	46	0.668	0.06989	0.00195	0.1600	0.0017	1.542	0.048	103	957	10	925	57
D15.2	237	153	39	0.526	0.06964	0.00220	0.1494	0.0016	1.435	0.050	98	898	9	918	65
D17.1	218	64	34	0.322	0.07346	0.00193	0.1539	0.0017	1.559	0.046	90	923	9	1027	53
D20.1	178	94	30	0.770	0.07061	0.00308	0.1550	0.0017	1.509	0.070	98	929	10	946	89
D22.1	175	109	30	0.796	0.07068	0.00293	0.1547	0.0017	1.507	0.067	98	927	9	948	85
D23.1	230	158	41	1.167	0.06230	0.00271	0.1623	0.0017	1.394	0.064	142	970	10	684	93
D26.1	172	100	28	0.705	0.06977	0.00240	0.1494	0.0016	1.437	0.054	97	897	9	922	71
D1.2	150	83	26	1.816	0.07204	0.00378	0.1518	0.0045	1.508	0.095	92	911	25	987	107
D30.2	219	75	36	0.668	0.07580	0.00194	0.1558	0.0045	1.628	0.067	86	933	25	1090	51
Anhedral and irregular cores, with or without truncated growth zones															
D3.1	272	542	61	0.426	0.07432	0.00218	0.1565	0.0017	1.604	0.052	89	937	9	1050	59
D14.1	271	182	40	0.000	0.07707	0.00980	0.1350	0.0019	1.435	0.186	73	816	11	1123	256
D27.1	155	91	26	1.628	0.07314	0.00360	0.1482	0.0044	1.494	0.090	88	891	24	1018	100
D28.2	169	124	29	0.056	0.07563	0.00265	0.1516	0.0044	1.581	0.076	84	910	25	1085	70
D29.1	439	195	72	0.205	0.07715	0.00120	0.1545	0.0045	1.643	0.057	82	926	25	1125	31
D29.2	129	66	24	3.965	0.07269	0.00492	0.1515	0.0045	1.519	0.117	90	909	25	1005	138
D30.1	446	517	85	0.801	0.07052	0.00140	0.1531	0.0044	1.489	0.055	97	918	25	943	41
D31.1	221	398	51	0.834	0.07592	0.00243	0.1606	0.0047	1.681	0.077	88	960	26	1093	64
D32.1	223	388	49	0.670	0.07583	0.00233	0.1585	0.0046	1.658	0.075	87	949	26	1091	62
D3.2	299	606	56	0.120	0.07815	0.00245	0.1302	0.0038	1.403	0.063	69	789	22	1151	62

Fig. 4.8. Concordia plot of zircon analyses from sample F, a dyke of *pre-D₃* felsic orthogneiss . Error hexagons are 1σ .

Fig. 4.9. Concordia plot of zircon analyses from sample D, a *pre-D₄* felsic pegmatite. Error hexagons are 1σ .

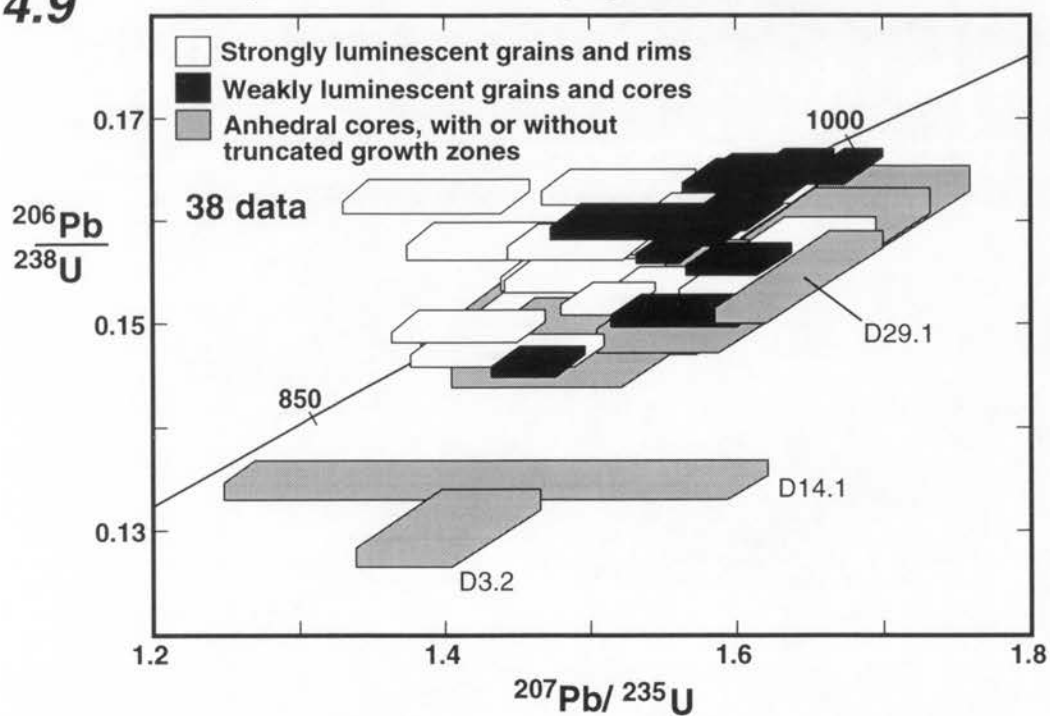
Sample F *Pre-D₃ felsic orthogneiss dyke*

4.8



Sample D *Pre-D₄ felsic pegmatite*

4.9



Biotite-pyroxene orthogneiss (sample B)

Sixteen analyses were obtained for this sample. These analyses cluster on the concordia at around 1000 Ma (Table 4.3, Fig. 4.6). Two of the analyses are reverse discordant, and one analysis, B5.1, is normally discordant, with a significantly older $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1151 ± 34 Ma.

Red granitic orthogneiss (sample R)

Most of the forty seven analyses from this sample (Table 4.4, Figs 4.7a & 4.7b) lie on the concordia between 850 and 1020 Ma. Twelve analyses from truncated cores include data that plot on the concordia, along with seven discordant analyses with $^{207}\text{Pb}/^{206}\text{Pb}$ ages between 1400 and 2400 Ma. Data for eight unzoned, weakly luminescent rims and grains mostly lie on the concordia at around 980 Ma. Thirteen analyses of strongly luminescent rims and grains are mostly concordant, with means slightly lower in age but within 2σ of most analyses of weakly luminescent zircon. Two strongly luminescent analyses, R3.2 and R20.1, are significantly younger. Fourteen analyses of zoned, weakly luminescent grains and cores gave ages that spread between 850 and 1020 Ma. One of these analyses, R9.2, has a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1000 ± 12 Ma and is reverse discordant.

Pre-D₃ felsic orthogneiss dyke (sample F)

All twenty six analyses from this sample plot on the concordia at around 950 Ma (Table 4.5, Fig. 4.8). There is no observable difference in distribution between the eighteen analyses of strongly luminescent grains and rims, and the eight analyses of cores with truncated growth zones.

Pre-D₄ felsic pegmatite (sample D)

Most of the thirty eight analyses for this sample lie on the concordia between 850 and 1050 Ma (Table 4.6, Fig. 4.9). Ten analyses of truncated cores include two discordant analyses, with $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 1125 ± 62 Ma (D29.1) and 1151 ± 124 Ma (D3.2). Eleven analyses of zoned, weakly luminescent cores and grains mostly plot on the

concordia at around 970 Ma. Strongly luminescent rims and grains lie scattered along the concordia between 850 and 1050 Ma.

4.7 Interpretation

4.7.1 General interpretations

Paragneiss (sample P)

Zircon grains in this sample are all predominantly composed of strongly luminescent zircon. Most of the analysed grains contain 100-400 ppm U and 50-100 ppm Th (Fig. 4.10). Many grains have equant faceted habits, which would not be preserved if the grains were of detrital origin. As the paragneiss preserves both metamorphic and migmatitic textures, the zircon grains are interpreted to have crystallized from felsic melt and/or hydrothermal fluids present during metamorphism. Similar zircons, with high cathodoluminescent response and low Th and U contents, were observed by Williams *et al.* (1997) in migmatized patches of granulite facies paragneiss from the Reynolds Range, Australia. Cores in zircons from the paragneiss, with truncated zoning and/or anhedral forms, are interpreted as detrital zircon.

Most of the six detrital cores analysed have similar $^{206}\text{Pb}/^{238}\text{U}$ ages to those of the metamorphic zircon. This is attributed to recrystallization and/or Pb loss in the detrital cores during metamorphism. One detrital core, P21.1, yielded a discordant $^{207}\text{Pb}/^{206}\text{Pb}$ age of *c.* 1340 Ma. As this age was probably disturbed by metamorphism, the analysis provides only tentative evidence that deposition of the paragneiss occurred before the end of the Mesoproterozoic.

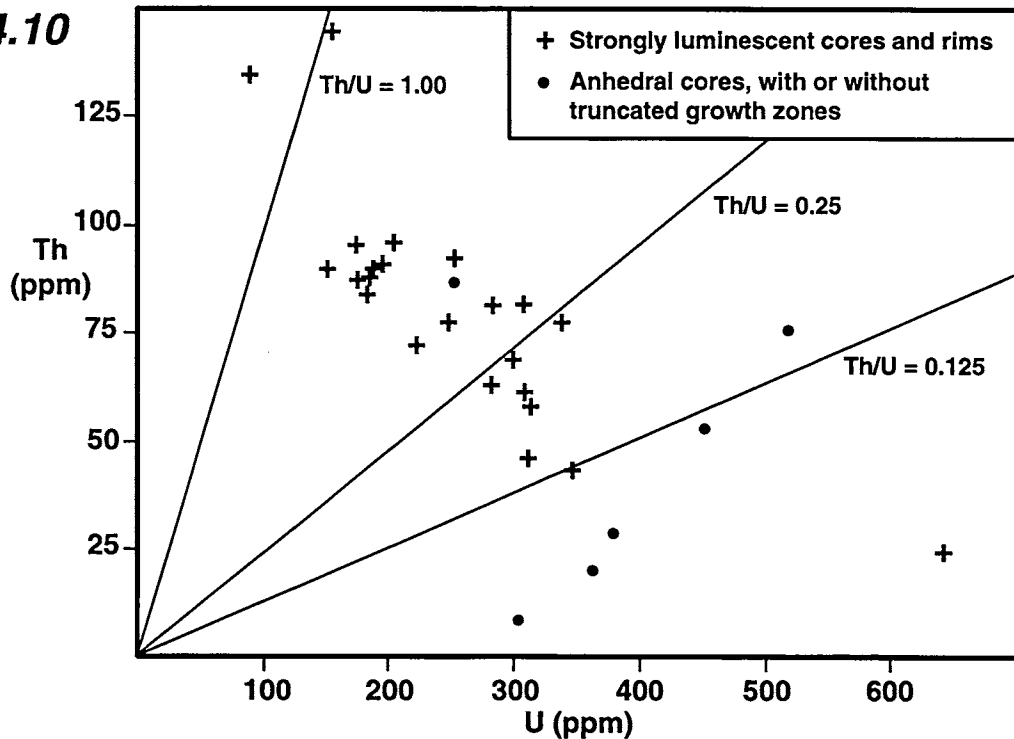
Discordant ages for analyses P5.1, P19.1 and P20.1 are attributed to the later loss of radiogenic Pb from the analysis sites. The remaining twenty of the analyses form a continuous distribution with a variance greater than that which can be attributed to analytical error. The modelling of this distribution is described in section 4.7.2.

Fig. 4.10. Th-U contents of analysed zircon from sample P, *paragneiss*.

Fig. 4.11. Th-U contents of analysed zircon from sample C, *charnockitic orthogneiss*.

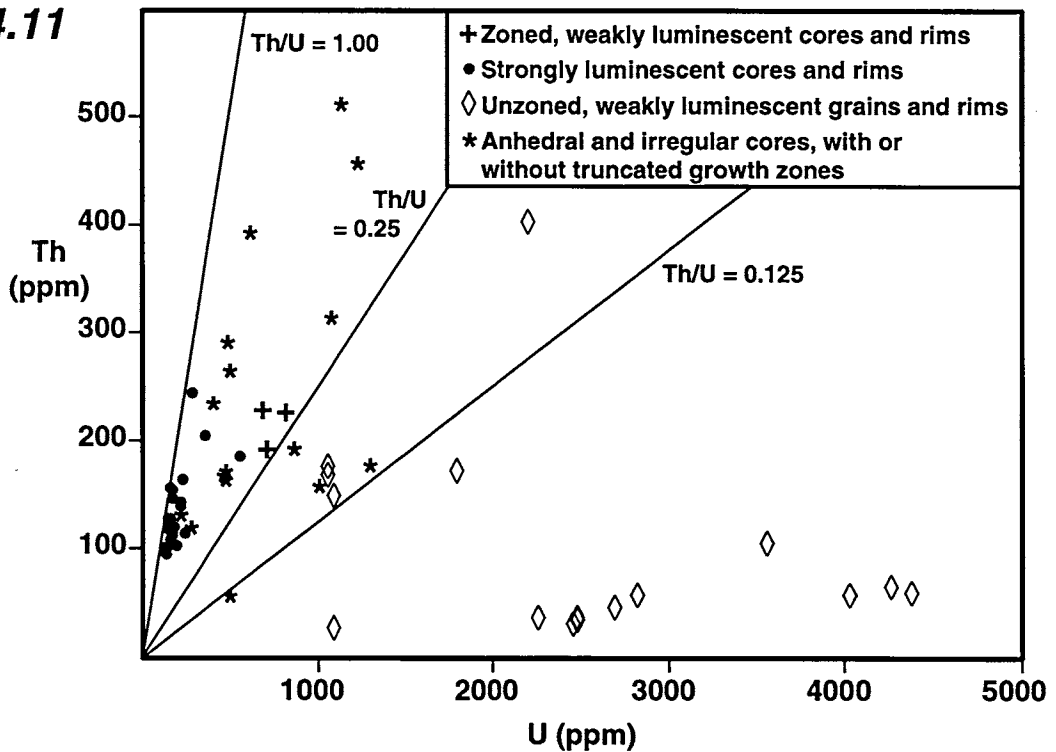
Sample P *Paragneiss*

4.10



Sample C *Charnockitic orthogneiss*

4.11



Charnockitic orthogneiss (sample C)

Zircon cores with anhedral forms and/or truncated zoning provide analyses that vary widely in age and U and Th content (Fig. 4.11). These cores are interpreted as being xenocrysts inherited from crust that was incorporated into a magma during the formation or emplacement of the protolith. The anhedral forms are interpreted as the product of partial resorption of the inherited grains within the magma, or the result of abrasion during detrital transport. The oldest inherited cores (*c.* 2100 Ma) provide evidence for the reworking of Paleoproterozoic and/or Archaean crust in the region.

The strongly luminescent zircons, whether zoned or unzoned, consistently contain approximately 200 ppm U and 110 ppm Th, and may therefore have shared a common origin. Zoned, weakly luminescent zircons contain 300-800 ppm U and 180-250 ppm Th. The elongate forms and oscillatory zoning in these grains are consistent with the zircon having grown from a magma.

Unzoned, weakly luminescent zircon rims and grains have U contents up to 1000 ppm and Th contents lower than 100 ppm. Low Th and high U contents are characteristic of metamorphic zircon (Williams & Claesson, 1987). The occurrence of unzoned, weakly luminescent zircon as rims and equant grains supports the interpretation that this stage of zircon growth occurred during metamorphism (Bibikova & Williams, 1990).

Populations in the age data cannot be resolved by placing analyses into groups according to CL response. The spread in age data can be attributed in part to the redistribution of Pb in the zircon by diffusion or recrystallization. Additionally, zircon growth may have occurred at more than one time. As the majority of the data lie on the concordia, the partial resetting of isotopic systems in the zircon probably occurred within a few hundred million years of the growth of disturbed zircon grains (*cf.* Fig. 4.3b). Mixture modelling was applied to this data to establish whether distinct stages of zircon resetting and growth could be resolved (section 4.7.2).

Biotite-pyroxene orthogneiss dyke (sample B)

The elongate shape and simple structure of the grains are consistent with zircon growth having occurred during the crystallization of the dyke. Discordant $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ ages in a few analyses are attributed to radiogenic Pb loss from the analysis sites. The concordant analyses, excepting analysis B1.2, define a single population with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 992 ± 10 Ma (2σ error, $\chi^2=1.21$), and a $^{206}\text{Pb}/^{238}\text{U}$ age of 998 ± 20 Ma ($\chi^2=1.76$). The former estimate is used as the crystallization age of the dyke in the following discussion.

Red granitic orthogneiss (sample R)

Zircon cores with anhedral forms and/or truncated zoning have a wide range of age estimates and U-Th contents (Fig. 4.12). This is consistent with a xenocrystic origin for the cores. Anhedral cores that have little or no CL response contain more than 1000 ppm U. Some of these cores have younger $^{207}\text{Pb}/^{206}\text{Pb}$ ages than the zircon that encloses them (*e.g.* analyses R17.1 *vs* R17.2). It is likely that these cores have been damaged by radioactive decay (McLaren *et al.*, 1994), and that the ages they give are the product of recrystallization which occurred after the growth of the zircon rims.

Analyses R5.1, R10.1, R25.1 and R11.1 are interpreted as age estimates of xenocrystic cores. A Model 1 regression was fitted to these analyses which intercepts the concordia at 2780 ± 156 Ma and 994 ± 46 Ma (2σ , Fig. 4.7b). Assuming that these cores crystallized at the same time, the upper intercept is interpreted as representing the crystallization age of the zircon, and the lower intercept as the age of an event which has partially reset the ages of the cores. The xenocrystic cores were probably derived from crustal material that was incorporated into the igneous protolith.

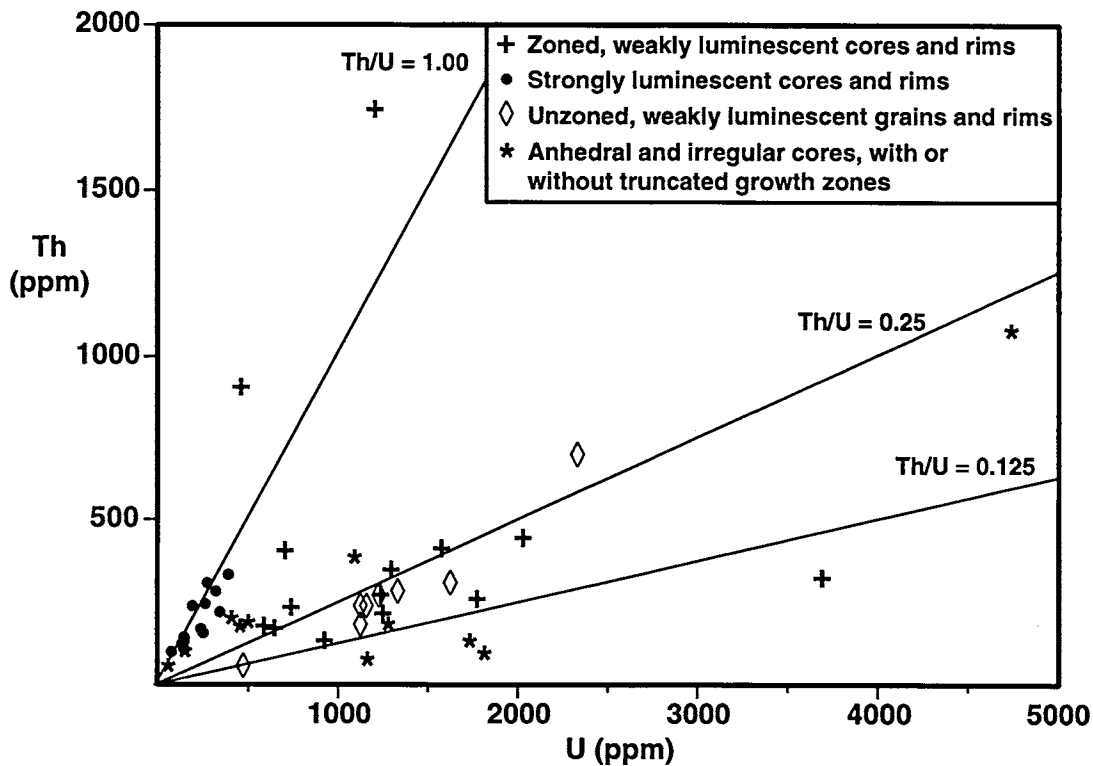
Most of the analysed zircon occurs as zoned, elongate grains. Strongly luminescent zircon contains 100-400 ppm U and Th (Fig. 4.12). Weakly luminescent zircons with growth zoning have U contents from 500-2000 ppm. The elongate crystal forms,

Fig. 4.12. Th-U contents of analysed zircon from sample R, *red granitic orthogneiss*.

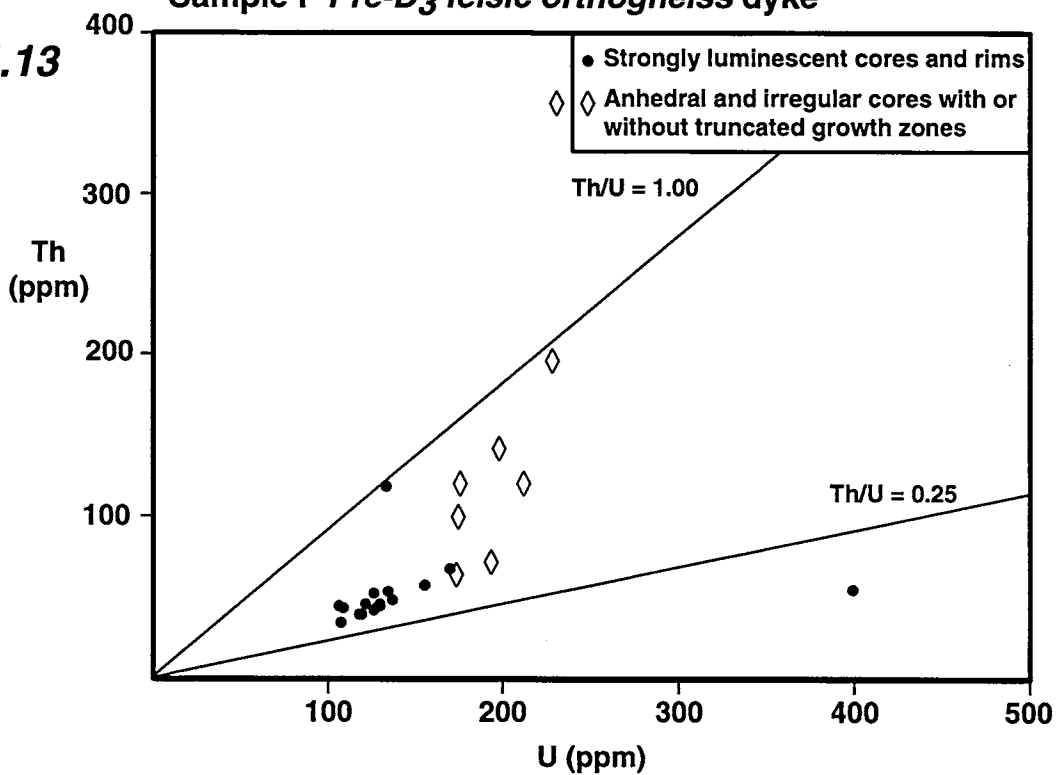
Fig. 4.13. Th-U contents of analysed zircon from sample F, a dyke of *pre-D₃ felsic orthogneiss*.

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4.12 Sample R *Red granitic orthogneiss*



4.13 Sample F *Pre-D₃ felsic orthogneiss dyke*



strong oscillatory zoning and U-Th contents are consistent with the crystallization of these zircons from a magma.

Unzoned, weakly luminescent zircon, which occurs as rims and equant grains, contains more than 1000 ppm U and 200-300 ppm Th. The anhedral form of these rims and grains, and their high U content, are consistent with growth during metamorphism (Williams & Claesson, 1987).

Distinct populations could not be resolved for analyses that were separated into groups according to morphology and CL response. Modelling of the $^{206}\text{Pb}/^{238}\text{U}$ age variation was applied to these analyses, and is discussed in section 4.7.2.

Pre-D₃ felsic orthogneiss dyke (sample F)

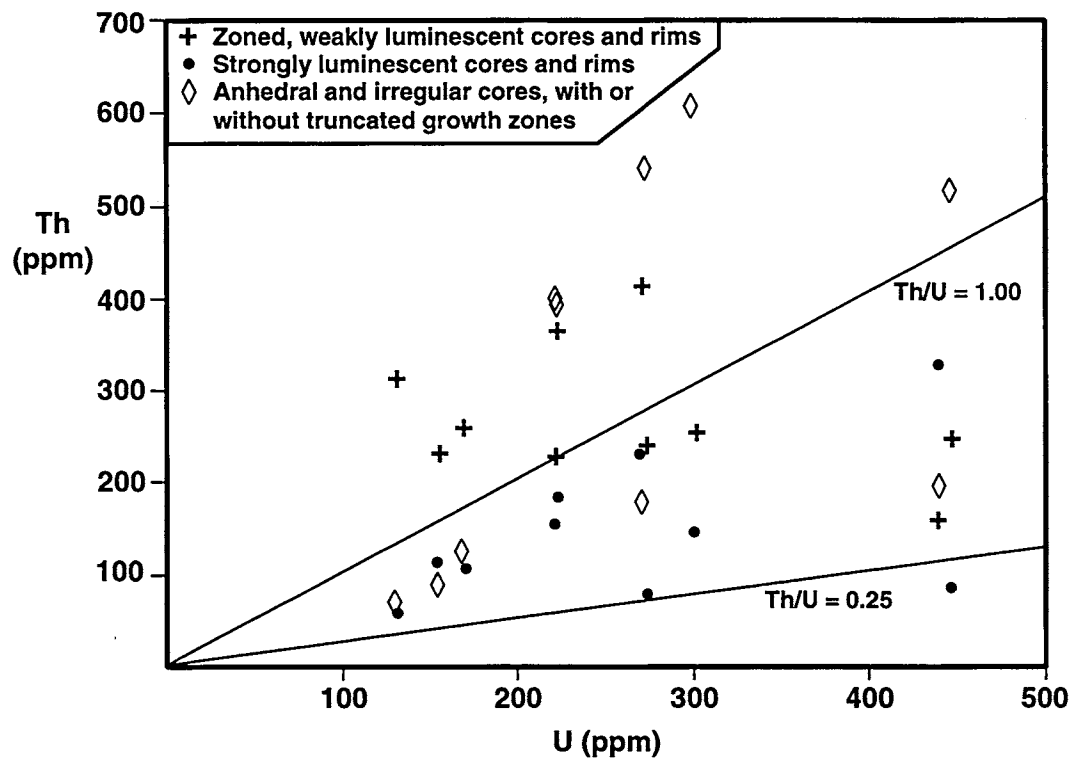
Analyses of zircon cores with truncated growth zoning have variable U and Th contents (Fig. 4.13). These cores are interpreted as being xenocrystal zircons of varied origin. Most of the strongly luminescent zircon analyses have a U content of approximately 130 ppm, and a Th content of approximately 50 ppm. The low Th content is consistent with the strongly luminescent zircon having formed during the crystallization of the dyke. All age data from the dyke fall within a single population with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 937 ± 19 Ma (2σ , $\chi^2=0.72$), and a $^{206}\text{Pb}/^{238}\text{U}$ age of 944 ± 10 Ma ($\chi^2=1.24$). The latter is the most precise estimate, and is taken as the age of crystallization for the dyke. This population includes age estimates from xenocrystal cores, that are interpreted as having been isotopically reset at the same time as the crystallization of the dyke.

Pre-D₄ felsic pegmatite (sample D)

Zircon cores with truncated growth zoning are interpreted as having a xenocrystic origin. Weakly luminescent zircon contains slightly more Th (150-400 ppm, Fig. 4.14) than most analyses of strongly luminescent zircon (50-300 ppm). All analyses contain 100-500 ppm U. The continuity of growth zoning between weakly and strongly luminescent zircon is consistent with both types of zircon having grown in a

Fig. 4.14. Th-U contents of analysed zircon from sample D, a *pre-D₄ felsic pegmatite*.

4.14 Sample D *Pre-D₄* felsic pegmatite



single stage. Elongate crystal habits and oscillatory zoning are commonly used as evidence of an igneous origin for zircon grains. The anhedral form of most grains is attributed to zircon dissolution and recrystallization at some stage after the crystallization of the dyke.

Ages obtained from high and weakly luminescent zircon were not resolvable into statistical populations, and modelling of the $^{206}\text{Pb}/^{238}\text{U}$ age variation was applied.

4.7.2 Mixture modelling

SHRIMP analytical data from four samples were investigated using the mixing program of Sambridge and Compston (1994). All four datasets have a spread of $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ age estimates, in the range 850-1050 Ma. Mixture modelling was applied to determine if the spread distribution of ages could be attributed to the overlap of two or more statistical populations. To minimise complexities arising from other sources of variation, the following selection criteria were applied to the analyses:

- 1) All analyses of cores with a probable xenocrystic origin were excluded.
- 2) Analyses with $^{207}\text{Pb}/^{206}\text{Pb}$ or $^{206}\text{Pb}/^{238}\text{U}$ ages outside the range 850-1050 Ma were excluded.
- 3) To minimise the effect of radiogenic Pb diffusion on the data, $^{206}\text{Pb}/^{238}\text{U}$ ages that differed by more than 2σ from the corresponding $^{207}\text{Pb}/^{206}\text{Pb}$ ages were excluded.

The Sambridge and Compston (1994) program produces model distributions with errors quoted at 2σ . These errors differ from those derived by the technique of Nelson (1997), in that the calibration error determined from duplicate analyses of the

zircon standard is included in the individual errors for each analysis, but has not been added again in quadrature to the errors on the weighted means.

Paragneiss (sample P)

Twenty concordant analyses were modelled with respect to $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ age estimates (Table 4.7). Due to the large errors in the $^{207}\text{Pb}/^{206}\text{Pb}$ age estimates, only one population of 940 ± 31 Ma (2σ error) could be clearly resolved, whereas two distinct populations were resolved in the $^{206}\text{Pb}/^{238}\text{U}$ age data (914 ± 6 and 975 ± 12 Ma; Fig. 4.15). The *c.* 914 Ma age comprises approximately 80% of the data, and represents a significant episode of zircon growth. The *c.* 975 Ma population is poorly constrained, but corresponds to ages modelled for data from the *charnockitic orthogneiss* and the *red granitic orthogneiss* (see below).

Charnockitic orthogneiss (sample C)

Thirty seven $^{207}\text{Pb}/^{206}\text{Pb}$ age data were modelled to infer a two-component distribution of 941 ± 16 and 982 ± 5 Ma (Table 4.8, Fig. 4.16a). Thirty two concordant $^{206}\text{Pb}/^{238}\text{U}$ age data were resolved into three components with ages 902 ± 13 , 957 ± 12 and 995 ± 7 Ma (Table 4.8, Fig. 4.17a). In both cases, the oldest component has the greatest proportion.

To assess the significance of these results, the model populations were compared to the distribution of analyses classified according to their CL response (Figs 4.16b & 4.17b). In both the $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ age data, strongly and weakly luminescent zircon with growth zoning is present in all of the model populations. If this zoned zircon is igneous in origin, then the spread in the $^{206}\text{Pb}/^{238}\text{U}$ age data indicates that Pb diffusion and/or recrystallization has occurred at *c.* 902 Ma, and possibly also at *c.* 957 Ma. Weakly luminescent, unzoned zircon, interpreted as having a metamorphic origin, has age estimates in the *c.* 957 Ma and *c.* 995 Ma model populations. The dominant population at *c.* 995 Ma probably represents a metamorphic event that occurred during or after the crystallization of the igneous

Tables 4.7-4.10. Mixture modelling experiments, using the methods of Sambridge & Compston (1994) and the MIX program of Sambridge & Gallagher (1995). All analyses used for modelling had mean ages in the range 850-1050 Ma, and had $^{206}\text{Pb}/^{238}\text{U}$ ages within 2σ error of their corresponding $^{207}\text{Pb}/^{206}\text{Pb}$ ages. MIX was run for both the $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ age datasets. Best fit results are listed in bold (*cf.* Figs 4.15-4.20), and were chosen using the criteria given in the text.

Isotopes: 6/38 = $^{206}\text{Pb}/^{238}\text{U}$ age data; 7/6 = $^{207}\text{Pb}/^{206}\text{Pb}$ age data.

Run = number of components modelled in the MIX run.

Prop = proportion of each age component.

Misfit = $-\ln(L)$, where L = the Likelihood function of Sambridge & Compston (1994); see text for explanation.

95%c = 95% confidence test: *pass* = each modelled age component lies beyond the 95% confidence limit of the other components. *fail* = one or more modelled age components are not resolvable at 95% confidence limits.

Table 4.7. MIX modelling experiments for sample P, *paragneiss*.

Table 4.8. MIX modelling experiments for sample C, *charnockitic orthogneiss*.

Table 4.9. MIX modelling experiments for sample R, *red granitic orthogneiss*.

Table 4.10. MIX modelling experiments for sample D, a *pre-D₄ felsic pegmatite*.

Table 4.7. Paragneiss , sample P

Isotopes	Data	Run	Component	Age	1 σ	Prop	1 σ	Misfit	95% c
6/38	20	1	1	926.6	2.7	1.00	0.00	67.93	pass
		2	1	975.4	5.8	0.20		29.99	pass
	3	2	2	913.9	2.9	0.81	0.21		
		3	1	975.7	6.1	0.19		26.24	fail
7/6	20	2	2	889.2	10.2	0.22	0.15		
		3	3	920.1	4.2	0.59	0.21		
	4	1	1	996.0	11.6	0.06		24.43	fail
		2	2	964.2	7.5	0.13	0.09		
7/6	20	3	3	889.0	10.2	0.21	0.15		
		4	4	919.9	4.0	0.59	0.21		
	3	1	1	939.6	15.3	1.00	0.00	9.38	pass
		2	1	1024.4	36.7	0.11		7.57	fail
7/6	30	2	2	914.5	18.0	0.89	0.22		
		3	1	914.5	94.9	0.07		7.57	fail
	3	2	2	1024.4	40.6	0.11	0.13		
		3	3	914.5	19.9	0.82	0.24		

Table 4.9. Red granitic orthogneiss , sample R

Isotopes	Data	Run	Component	Age	1 σ	Prop	1 σ	Misfit	95% c
6/38	30	1	1	972.1	2.0	1.00	0.00	189.61	pass
		2	1	915.4	4.7	0.21		71.98	pass
	3	2	2	989.4	2.3	0.79	0.16		
		3	1	855.0	9.0	0.03		46.57	pass
7/6	30	2	2	930.5	4.4	0.19	0.08		
		3	3	989.6	2.2	0.78	0.16		
	4	1	1	855.0	9.0	0.03		41.66	fail
		2	2	930.3	4.5	0.18	0.08		
7/6	30	3	3	1005.9	5.9	0.25	0.14		
		4	4	981.7	3.9	0.53	0.17		
	3	1	1	974.1	2.3	1.00	0.00	53.04	pass
		2	1	923.7	7.1	0.28		30.70	pass
7/6	30	2	2	981.1	2.5	0.72	0.16		
		3	1	997.8	6.3	0.07		27.98	fail
	3	2	2	923.1	7.4	0.27	0.11		
		3	3	976.8	2.9	0.66	0.17		

Table 4.8. Charnokitic orthogneiss , sample C

Isotopes	Data	Run	Component	Age	1 σ	Prop	1 σ	Misfit	95% c
6/38	32	1	1	969.7	2.3	1.00	0.00	122.04	pass
		2	1	911.6	7.8	0.22		53.87	pass
	3	2	2	984.4	3.0	0.78	0.16		
		3	1	902.2	6.4	0.16		42.38	pass
7/6	37	2	2	957.2	6.1	0.31	0.12		
		3	3	995.2	3.7	0.53	0.14		
	4	1	1	902.1	6.5	0.16		41.59	fail
		2	2	1005.2	8.0	0.24	0.18		
7/6	37	3	3	955.6	6.3	0.28	0.12		
		4	4	985.8	7.9	0.32	0.19		
	4	1	1	977.0	1.9	1.00	0.00	51.96	pass
		2	1	941.4	8.1	0.27		33.22	pass
7/6	4	2	2	982.2	2.3	0.73	0.16		
		3	1	931.6	7.9	0.14		28.92	fail
	4	2	2	964.8	5.2	0.33	0.15		
		3	3	986.2	2.5	0.53	0.17		
7/6	4	1	1	931.6	7.9	0.14		28.48	fail
		2	2	992.6	5.4	0.18	0.16		
	4	3	3	964.4	5.5	0.30	0.17		
		4	4	981.3	4.7	0.39	0.21		

Table 4.10. Pre-D4 felsic pegmatite , sample D

Isotopes	Data	Run	Component	Age	1 σ	Prop	1 σ	Misfit	95% c
6/38	23	1	1	939.8	2.1	1.00	0.00	105.44	pass
		2	1	908.9	3.3	0.40		38.39	pass
	3	2	2	962.7	2.8	0.60	0.17		
		3	1	897.6	5.3	0.22		32.71	pass
7/6	23	2	2	933.0	5.6	0.32	0.14		
		3	3	968.0	4.0	0.46	0.16		
	4	1	1	896.4	5.0	0.21		31.14	fail
		2	2	978.4	7.6	0.21	0.13		
7/6	23	3	3	925.0	6.4	0.21	0.11		
		4	4	955.1	5.0	0.37	0.16		
	3	1	1	979.0	9.2	1.00	0.00	8.98	pass
		2	1	1011.5	24.2	0.13		8.68	fail
7/6	23	2	2	969.1	12.9	0.87	0.22		
		3	3						
	3	1	1						
		2	2						

Fig. 4.15. Histogram of the mean values of $^{206}\text{Pb}/^{238}\text{U}$ age data for sample P, *paragneiss*. The curve represents the distribution of best-fit model age components obtained using the MIX program of Sambridge & Gallagher (1995). Model ages are quoted with 2σ errors. The inset compares the ages of the model components. A 95% confidence ellipse is plotted for the model ages, to show that the two ages lie well outside error of each other.

Sample P Paragneiss

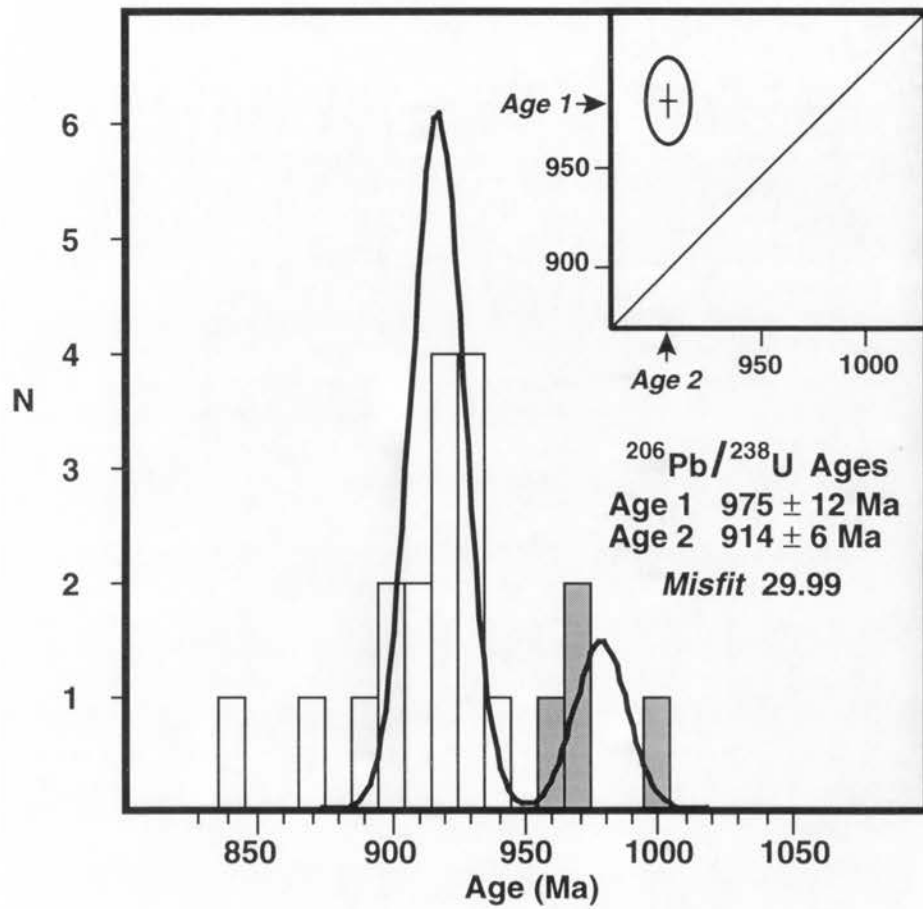
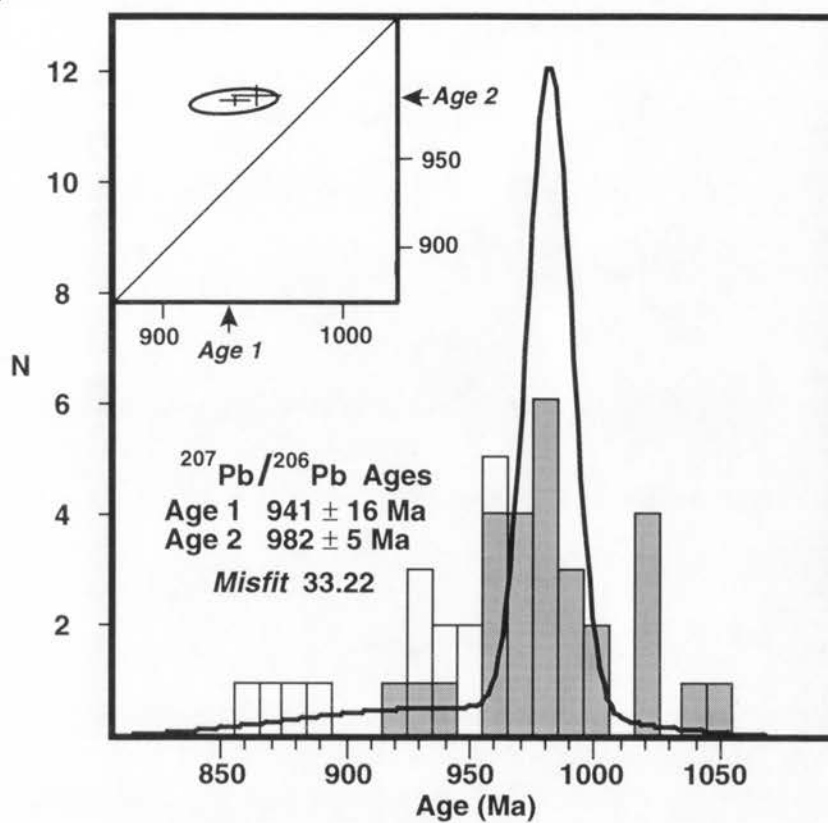


Fig. 4.16. Histograms of the mean values of $^{207}\text{Pb}/^{206}\text{Pb}$ age data for sample C, *charnockitic orthogneiss*.

a) Mean age histogram with results of mixture modelling. The curve represents the distribution of best-fit model age components obtained using the MIX program of Sambridge & Gallagher (1995). Model ages are quoted with 2σ errors. The inset compares the ages of the model components. A 95% confidence ellipse is plotted for the model ages, to show that the two ages lie well outside error of each other.

b) Mean age histogram showing the distribution of zircon analyses grouped by morphology and cathode luminescent behaviour.

a) Sample C *Charnockitic orthogneiss*



b)

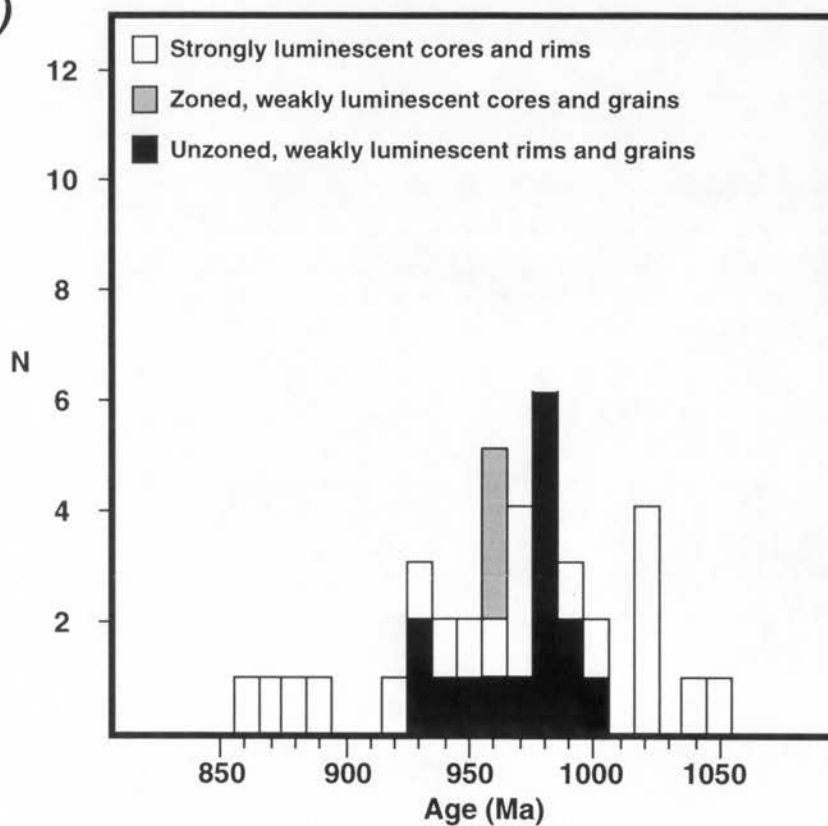
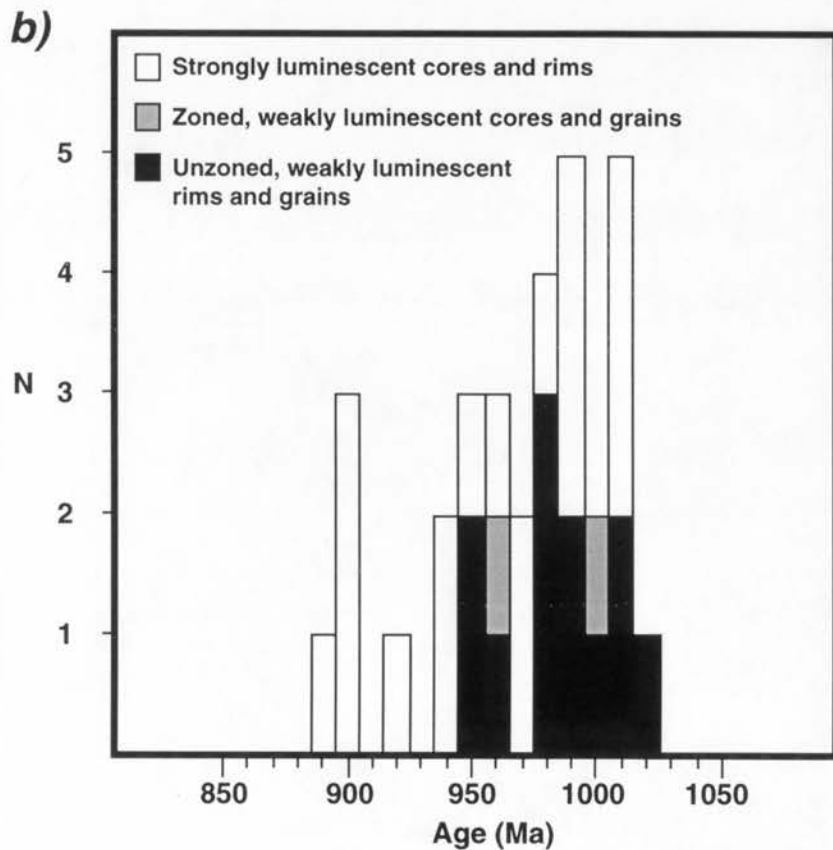
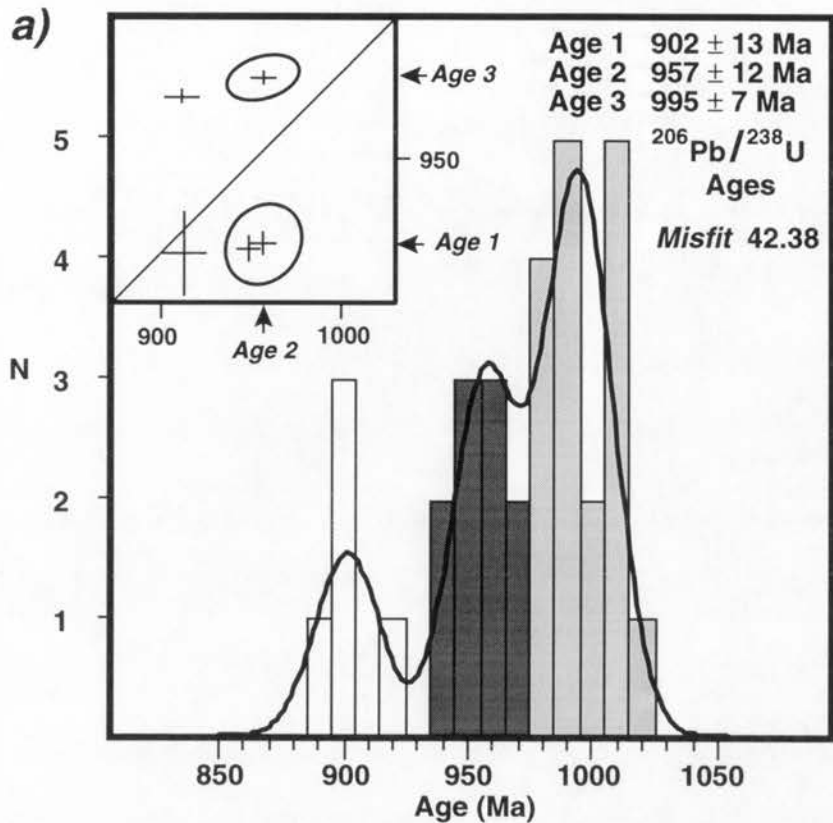


Fig. 4.17. Histograms of the mean values of $^{206}\text{Pb}/^{238}\text{U}$ age data for sample C, *charnockitic orthogneiss*.

a) Mean age histogram with results of mixture modelling. The curve represents the distribution of best-fit model age components obtained using the MIX program of Sambridge & Gallagher (1995). Model ages are quoted with 2σ errors. The inset compares the ages of the model components. Error ellipses at 95% confidence are plotted for the model ages to show that the three ages lie well outside error of each other.

b) Mean age histogram showing the distribution of zircon analyses grouped by morphology and cathode luminescent behaviour.

Sample C *Charnockitic orthogneiss*



zircon. This age consequently represents a minimum age only for the crystallization of the charnockite.

The significance of the *c.* 957 Ma population is unclear, as this population may contain data that were affected by partial radiogenic Pb loss during the *c.* 902 Ma event. As the *c.* 957 Ma population lies within 1σ error of the *c.* 975 Ma population of metamorphic grains from the *paragneiss* sample, both populations may be the result of one metamorphic event.

Red granitic orthogneiss (sample R)

Modelling of the variation in thirty concordant $^{207}\text{Pb}/^{206}\text{Pb}$ age data and thirty $^{206}\text{Pb}/^{238}\text{U}$ age data resolved two age components in both cases (Table. 4.9). The $^{207}\text{Pb}/^{206}\text{Pb}$ age data were divided into populations at 981 ± 5 Ma and 924 ± 14 Ma (Fig. 4.18a). $^{206}\text{Pb}/^{238}\text{U}$ age data yielded populations at 989 ± 5 Ma and 915 ± 9 Ma (Fig. 4.19a). In both cases, most of the data were assigned to the earlier event.

In both the $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ datasets, the older population is composed of analyses from zoned zircon, which has been interpreted to be igneous, and unzoned, weakly luminescent zircon, interpreted to be metamorphic in origin (Figs 4.18b & 4.19b). The *c.* 989 Ma population is therefore interpreted to define a metamorphic episode, that occurred at the same time as or later than the crystallization of the protolith. This estimate is within 1σ error of the *c.* 995 Ma population of analyses from the *charnockitic orthogneiss*.

The younger population is entirely composed of zoned zircon, which is inferred to have an igneous origin. If this is the case, the population may represent the time at which the isotopic systems in the zircon were completely reset, possibly during a metamorphic event. The *c.* 915 Ma population for $^{206}\text{Pb}/^{238}\text{U}$ age data lies within 2σ error of the *c.* 902 Ma population from the *charnockitic orthogneiss*, and the *c.* 914 Ma population from the *paragneiss*.

Fig. 4.18. Histograms of the mean values of $^{207}\text{Pb}/^{206}\text{Pb}$ age data for sample R, *red granitic orthogneiss*.

a) Mean age histogram with results of mixture modelling. The curve represents the distribution of best-fit model age components obtained using the MIX program of Sambridge & Gallagher (1995). Model ages are quoted with 2σ errors. The inset compares the ages of the model components. A 95% confidence ellipse is plotted for the model ages, to show that the two ages lie well outside error of each other.

b) Mean age histogram showing the distribution of zircon analyses grouped by morphology and cathode luminescent behaviour.

Sample R Red granitic orthogneiss

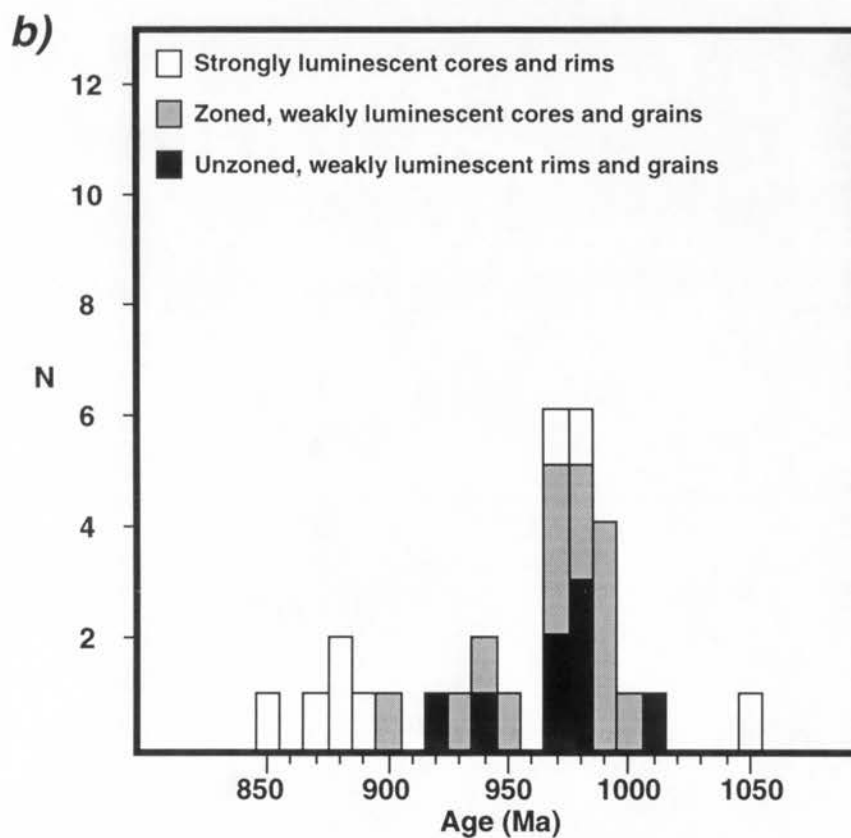
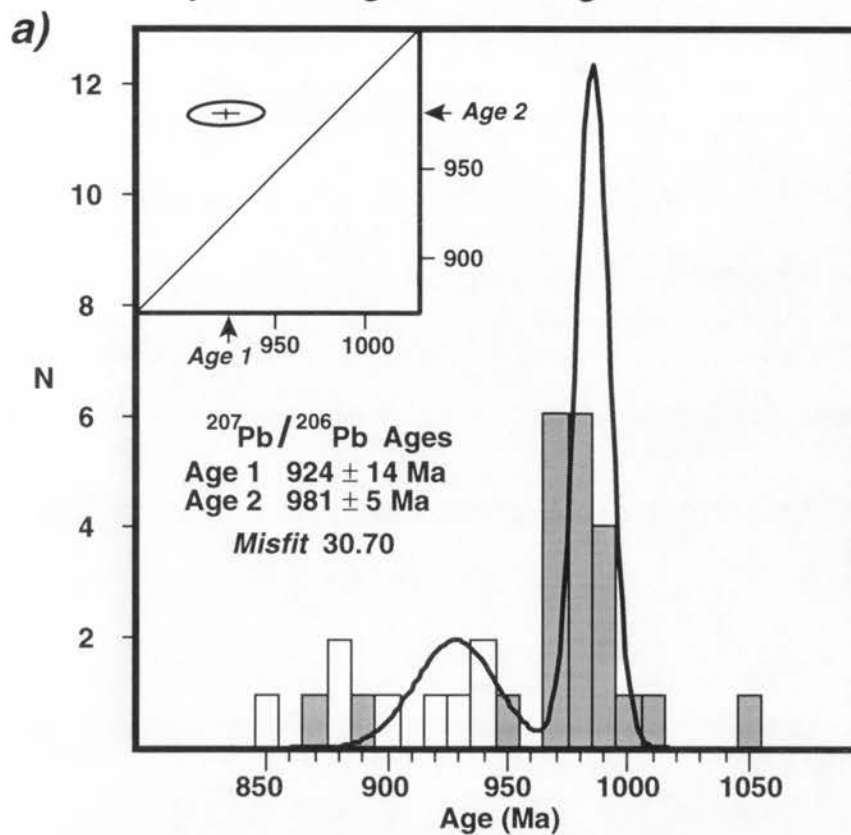
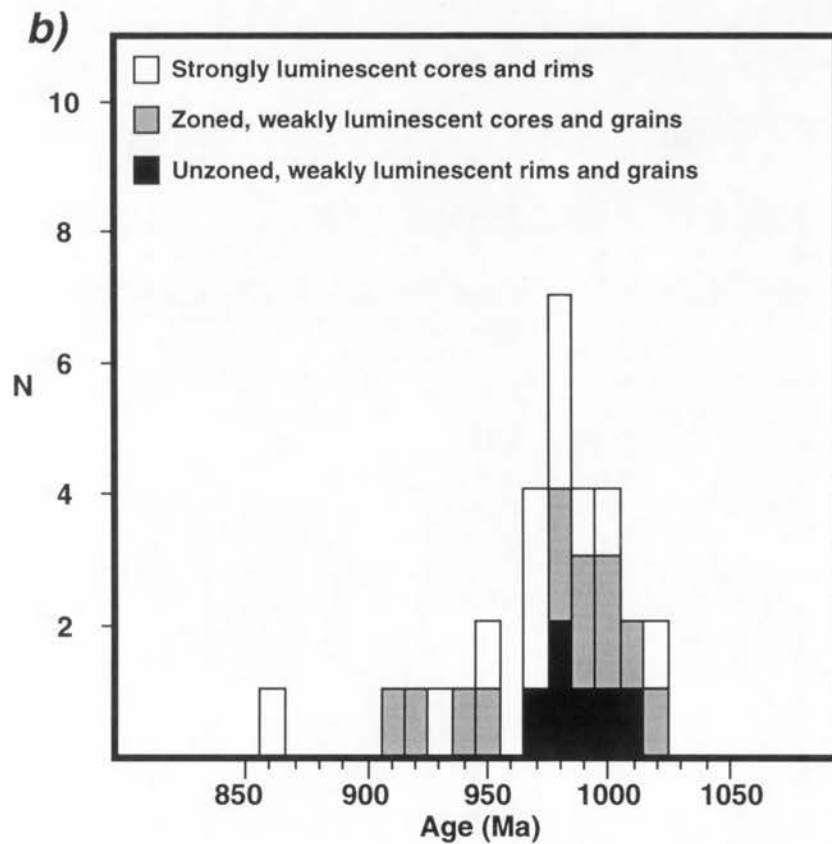
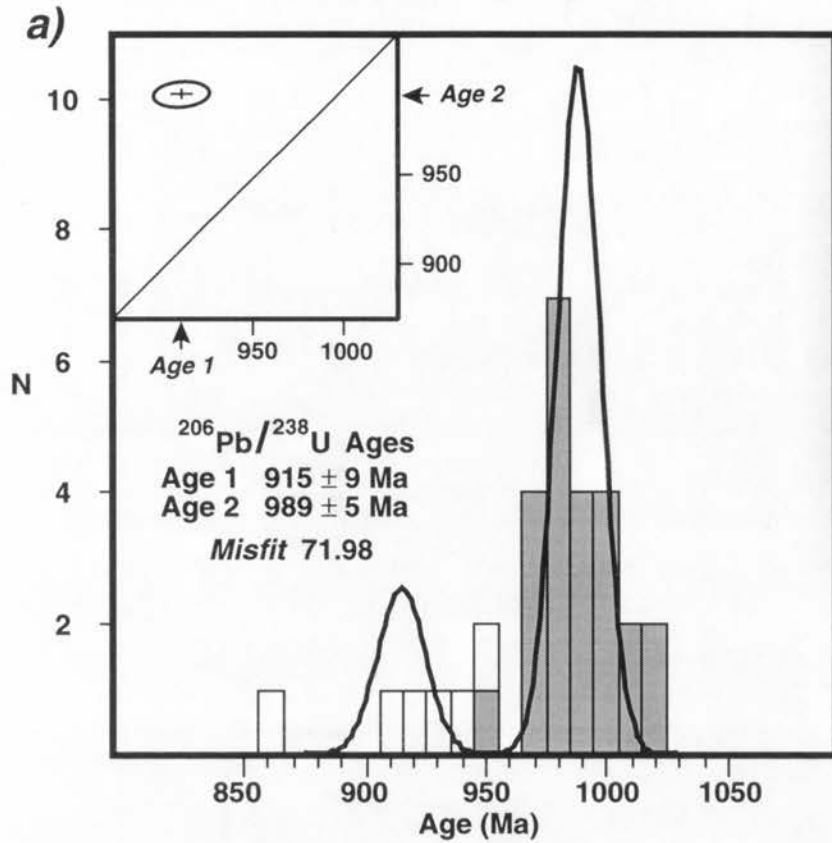


Fig. 4.19. Histograms of the mean values of $^{206}\text{Pb}/^{238}\text{U}$ age data for sample R, *red granitic orthogneiss*.

a) Mean age histogram with results of mixture modelling. The curve represents the distribution of best-fit model age components obtained using the MIX program of Sambridge & Gallagher (1995). Model ages are quoted with 2σ errors. The inset compares the ages of the model components. A 95% confidence ellipse is plotted for the model ages, to show that the two ages lie well outside error of each other.

b) Mean age histogram showing the distribution of zircon analyses grouped by morphology and cathode luminescent behaviour.

Sample R Red granitic orthogneiss



Pre-D₄ felsic pegmatite (sample D)

Twenty three concordant data were modelled for $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ages (Table 4.10). Due to large errors for $^{207}\text{Pb}/^{206}\text{Pb}$ age estimates, only one age of 979 ± 18 Ma was inferred from the modelling. For the $^{206}\text{Pb}/^{238}\text{U}$ age data, two age populations at 963 ± 6 Ma and 909 ± 7 Ma, were resolved (Fig. 4.20a). The *c.* 909 Ma population is composed entirely of zoned, strongly luminescent zircon, whereas the *c.* 963 Ma population is a mixture of strongly and weakly luminescent zircon (Fig. 4.20b). The *c.* 963 Ma population is interpreted as the crystallization age of the felsic dyke. The *c.* 909 Ma age is inferred to be the result of the isotopic resetting of igneous zircon by Pb loss or recrystallization. This younger population is within 1σ error of the younger populations derived from $^{206}\text{Pb}/^{238}\text{U}$ age data from the *paragneiss*, *charnockitic orthogneiss*, and *red granitic orthogneiss*.

4.8 Discussion

4.8.1 Cathode luminescence, zircon structure and age estimation

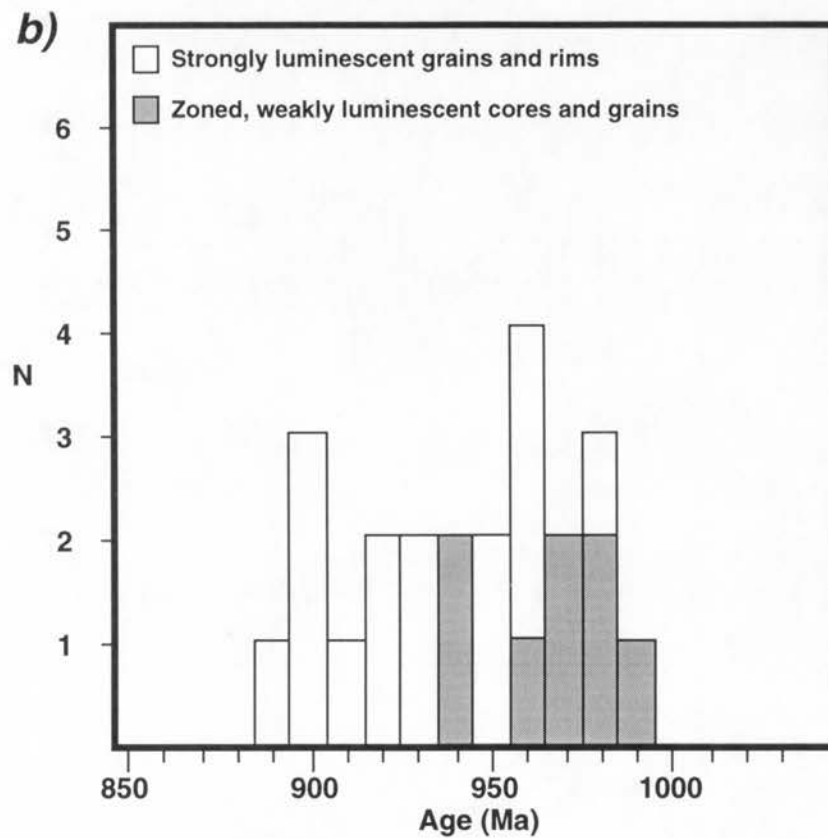
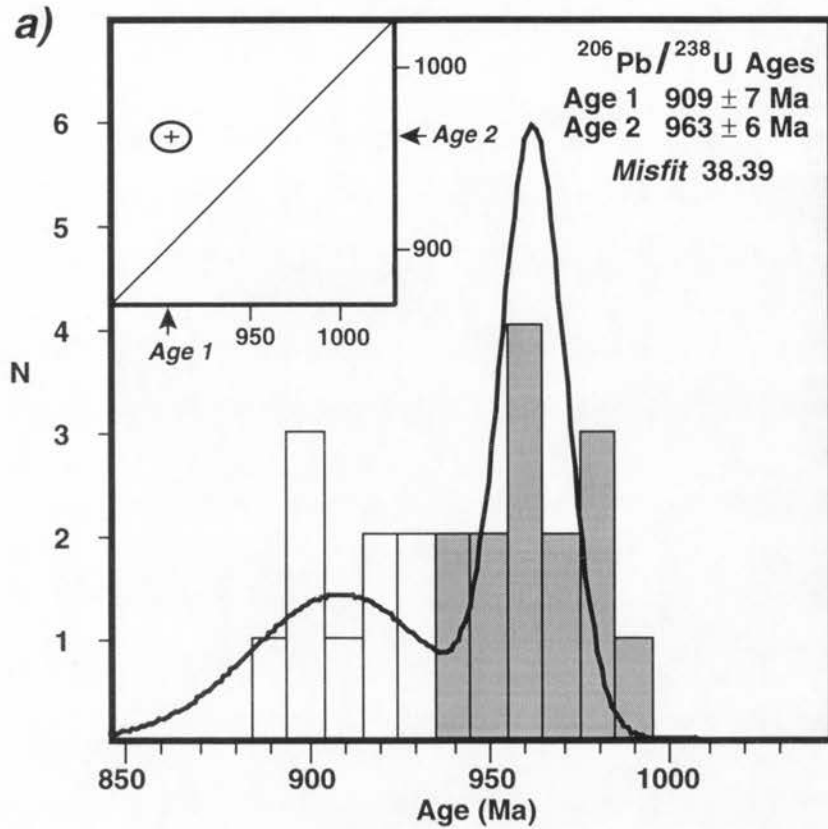
Internal structures in zircon grains, such as anhedral cores and growth zones, were observed by both cathodoluminescence imaging and transmitted light. Each type of CL response had a characteristic set of U-Th content and isotopic values. Strongly luminescent zircon had lower U and Pb content than weakly luminescent zircon, and thus yielded less precise age estimates. Zircon with more than 1000 ppm U had no CL response and commonly yielded reverse discordant ages. McLaren *et al.* (1994) attribute reverse discordance to Pb ‘sputtering’ (*sic*), that occurs during analysis of zircon damaged by radioactive decay. Radiation destruction also accounts for the absence of CL emission seen in high-U zircon (Koschek, 1993). Cathodoluminescence imaging can therefore be used to recognise the state of preservation of a zircon crystal, as well as the different stages of zircon growth.

Fig. 4.20. Histograms of the mean values of $^{206}\text{Pb}/^{238}\text{U}$ age data for sample D, a *pre-D₄ felsic pegmatite*.

a) Mean age histogram with results of mixture modelling. The curve represents the distribution of best-fit model age components obtained using the MIX program of Sambridge & Gallagher (1995). Model ages are quoted with 2σ errors. The inset compares the ages of the model components. A 95% confidence ellipse is plotted for the model ages to show that the two ages lie well outside error of each other.

b) Mean age histogram showing the distribution of zircon analyses grouped by morphology and cathode luminescent behaviour.

Sample D Pre-D₄ felsic pegmatite



However, relationships between structures that are visible with CL imaging, and their measured U-Pb ages, may be ambiguous. For example, grain R3 from the *red granitic orthogneiss* (Plate 4.3b) contains a round non-luminescent core, surrounded by zircon with a high CL response and oscillatory growth zones. The growth-zoned zircon is truncated at one end and enclosed within unzoned weakly luminescent zircon. These structures may be interpreted as a series of growth stages; *i.e.* zoned zircon crystallized from the granitic melt around an inherited zircon core, and was subsequently corroded and overgrown by U-rich zircon during metamorphism. Analysis R3.1 of the weakly luminescent core yields a discordant $^{207}\text{Pb}/^{206}\text{Pb}$ age significantly older than the rest of the grain, and thus the core is a xenocryst. Analysis R3.2 of the strongly luminescent zircon has significantly younger $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ ages than analysis R3.4, which is of the weakly luminescent rim. Both ages are concordant. The age of analysis R3.2 is unlikely to be the original crystallization age of the strongly luminescent zircon. Several such zircon grains, with cores that yielded younger ages than their rims, were found in various samples.

Due to such complexities, cathodoluminescence was not used in this study as a criteria for separating age populations. CL imaging is nonetheless important for understanding the meaning of the age estimates, especially where multiple ages are determined from a single data set.

Discrepancies between the order of stages of growth in a zircon crystal, and the dates of those stages, can often be attributed to isotopic disturbance after the formation of a zircon grain. Such disturbance is usually the product of Pb diffusion and/or recrystallization in the zircon crystal. The efficiency of Pb diffusion in zircon is poorly understood. As age information is preserved in zircons that have undergone high-temperature metamorphism, Mezger & Krogstad (1997) argue that the rate of Pb diffusion in pristine zircon is insignificant at temperatures below 1000°C. A zircon crystal that has been damaged by the decay of U and Th radioisotopes is highly susceptible to Pb leaching and/or diffusion at almost any temperature. This is

commonly the case in zircon of considerable age and high U and Th content. Recrystallization, with partial Pb loss, is unlikely to occur in zircon unless the crystal was damaged by radioactive decay and subsequently heated to temperatures above $\approx 600^{\circ}\text{C}$ (Mezger & Krogstad, 1997). It is also argued that Pb diffusion will not completely reset any isotopic system, and that concordant ages that are significantly younger than the original crystallization age of a zircon grain can only be produced by the dissolution and re-precipitation of zircon (Mezger & Krogstad, 1997).

Despite this, there have been cases described where U-Pb isotopic systems in zircon cores have been completely reset by a later event (Kroner *et al.*, 1994; Collins & Williams, 1995). Bibikova & Williams (1990) describe zircons from Archaean charnockitic samples in which low U zircon cores yield ages that are younger than the enclosing rims. In one case, the core age matches that of metamorphic overgrowths on much older igneous grains. Similar behaviour is observed in the samples dated from Cape Bruce; for example, zircon grain R26 from the *red granitic orthogneiss* contains an inherited core with a concordant $^{207}\text{Pb}/^{206}\text{Pb}$ age of 921 ± 20 Ma (R26.1). The core is enclosed in zoned igneous zircon, with a concordant $^{207}\text{Pb}/^{206}\text{Pb}$ age of 974 ± 7 Ma (R26.2). The inherited core is unstructured and contains high U, and the *c.* 921 Ma age probably records the annealing of radiation-damaged zircon during M_3 metamorphism. Since it is unlikely that only the core of a zircon crystal would be dissolved and re-precipitated, cores that are younger than rims probably formed by the solid state recrystallization of older zircon (Pigeon, 1992).

In the case of zircon grain R3 described above, Pb appears to have been lost from the grain without the destruction of growth zones. Radiation damage and recrystallization can occur within individual growth zones with high U contents, without damaging adjacent zones (McLaren *et al.*, 1994). If this is the case, analysis R3.2 represents a mixture of isotopic values from recrystallized and preserved growth zones. If, as interpreted above, the grain was crystallised at *c.* 980 Ma and isotopically

reset at *c.* 910 Ma, then the amount of radiation damage in the grain was likely to be minor. This implies that Pb loss occurred in zircon that was comparatively pristine.

Many of the analyses from orthogneisses that lie within the *c.* 910 Ma populations are of zircon grains and cores that have strong growth zoning visible in CL. These grains are interpreted as zircon that crystallised with the intrusion of each orthogneiss, and subsequently suffered Pb loss. Intrusion of the orthogneisses is inferred to have occurred immediately before or during metamorphism (Table 4.11). If the *c.* 910 Ma age represents a real metamorphic event that involved zircon growth and recrystallization, then complete isotopic resetting must have also occurred in many of the igneous zircon grains. These grains would not have had time to be damaged by radioactive decay.

An alternative interpretation is that isotopic disturbance in the zircon occurred much later than *c.* 910 Ma. If this is the case, then only a small amount of Pb was lost, so that concordance was maintained between most of the $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ ages. However, there are no zircons ages younger than *c.* 850 Ma, and therefore new zircon growth did not accompany the Pb loss. Arguing for late isotopic disturbance also assumes that zircons from four of the samples lost just enough Pb to form age populations at *c.* 910 Ma. Since these zircons come from unrelated lithologies with different ages of formation, this is unlikely to be the case.

4.8.2 Age constraints on field relationships

The sequence of geological events inferred in Chapters 2 and 3 can now be constrained using the age estimates discussed in this Chapter (Table 4.11). As the *paragneiss* and the *charnockitic orthogneiss* are the only dated units to have been affected by M_1 granulite facies metamorphism, the *c.* 995 Ma zircon population from

Table 4.11. Age constraints on Cape Bruce geology derived by this study. All ages are quoted with 1σ errors.

.

Table 4.11	Dated Lithology					
Sequence of events	<i>Paragneiss</i>	<i>Charnockitic orthogneiss</i>	<i>Biotite-pyroxene orthogneiss</i>	<i>Red granitic orthogneiss</i>	<i>Pre-D₃ felsic orthogneiss</i>	<i>Pre-D₄ felsic dykes</i>
Deposition of sediments	<1200?					
Charnockitic orthogneiss		≥995±7				
<i>D₁/M₁ granulite-facies metamorphism</i>		995±7				
Biotite-pyroxene orthogneiss			992±10			
White granitic orthogneiss						
Red granitic orthogneiss				≈995-989		
<i>D₂/M₂ retrograde metamorphism</i>	975±12	957±12		989±5		
Pre-D ₃ felsic orthogneiss dykes					944±10	
D ₃ folding						
Pre-D ₄ felsic dykes						963±6
D ₄ folding						
<i>M₃ metamorphism</i>	914±6	902±13		915±9		909±7
D ₅ mylonitic shear zones						

the orthogneiss most probably reflects the M_1 event. The *biotite-pyroxene orthogneiss* dyke intruded after D_1 , and places a lower constraint on the age of M_1 .

The *c.* 989 Ma zircon population from the *red granitic orthogneiss* is inferred to contain a metamorphic component, and therefore places a minimum time constraint on the crystallization of this orthogneiss. All of the lithologies described above were affected by M_2 metamorphism, which involved the hydration and retrogression of S_1 assemblages (see Chapter 3). Zircon populations from the *paragneiss*, the *charnockitic orthogneiss* and the *red granitic orthogneiss* are inferred to represent the age of M_2 . The M_2 event is further constrained by crystallization ages inferred for the *pre- D_3* and *pre- D_4 felsic dykes*, which were not affected by this event. M_2 is therefore constrained to the period *c.* 960-995 Ma. The upper limit of this range, however, overlaps with estimates for the M_1 event. These two events cannot therefore be reliably resolved, despite the modelling of three age populations from the charnockitic orthogneiss. This lack of separation between M_1 and M_2 is consistent with the possibility that they form a continuous period of metamorphism, involving cooling and rehydration after granulite facies peak metamorphism (see Chapter 3).

The ages of crystallization estimated for the *pre- D_3* and *pre- D_4 felsic dykes* are inverted relative to the order of emplacement observed in the field. It is therefore possible that both sets of dykes belong to a common suite, that intruded over a period of time during which D_3 folding also occurred. The timing of D_3 fits well with the interpretation of D_3 as a waning stage of D_1 - D_2 tectonism (see Chapter 2).

Metamorphic minerals in *pre- D_3* and *pre- D_4 felsic dykes* provide evidence that elevated conditions were present after M_2 . Four of the dated lithologies contain zircon with age populations of *c.* 910 Ma. The reproducibility of this age population between samples strongly supports the possibility of a distinct period of zircon growth, recrystallization, and/or isotopic resetting. As this is most likely to have occurred under metamorphic conditions, the *c.* 910 Ma population is correlated with

M₃ assemblages observed in all of the lithologies dated (see Chapter 3). M₃ is therefore interpreted as an event that is distinct from an earlier, continuous M₁-M₂ event.

4.8.3 A new geochronology of the Mawson Coast

Previous interpretations of the geology of the Mawson Coast (Clarke, 1988; Grew *et al.*, 1988; White & Clarke, 1993) have involved the tentative correlation of events between widely separated localities. Geochronological constraints on such correlations were limited. The data obtained by this study necessitates a re-interpretation of the history of parts of the Mawson Coast.

Archaean crust exposed along the Mawson Coast has only been recognised in the vicinity of the Stillwell Hills, where an age of *c.* 2700 Ma was interpreted by Clarke (1987) to reflect granulite facies metamorphism. This event was tentatively correlated with D₁/M₁ deformation and granulite facies metamorphism in the Colbeck Archipelago by White & Clarke (1993). However, the latter event is correlated in this study with the D₁/M₁ event at Cape Bruce, which occurred *c.* 995 Ma (see Chapter 2). This event represents the peak of tectonic and metamorphic activity during the Rayner Structural Episode, and corresponds to the D₁ event of Clarke (1988) that deformed gneisses in the Stillwell Hills.

At Cape Bruce and the Colbeck Archipelago, the D₁/M₁ event deforms and metamorphoses sediments and orthogneisses of unknown age. White & Clarke (1993) correlated these lithologies with similar gneisses in the Stillwell Hills, which were inferred by Clarke (1987) to have been metamorphosed during the Archaean. However, Rb-Sr whole rock ages for paragneiss from the Stillwell Hills merely constrain the age of these gneisses to ≥ 1250 Ma (Clarke, 1987). It is therefore possible that the deposition of sedimentary protoliths of paragneiss lithologies at Cape Bruce

occurred during the Proterozoic. This possibility is supported by the common occurrence of calc-silicates at Fold Is (James *et al.*, 1991), Cape Bruce, and the Framnes Mountains, whereas calc-silicates have only rarely been found in the Archaean Napier Complex (Sheraton *et al.*, 1987). There is evidence from other localities in the Rayner Complex that sedimentation occurred during the Mesoproterozoic (see Chapter 6).

The presence of Archaean crust beneath the rocks of the Mawson Coast is indicated by the ages of xenocrystic zircon in orthogneisses from Cape Bruce and (for one analysis) Mawson station (Young & Black, 1991). However, in these localities, most age estimates from the inherited zircon lie within the range 2000-1300 Ma. As the orthogneisses were probably derived from the anatexis of lower crustal sources (see Chapter 2), much of the crust underlying the Mawson Coast is likely to be Paleoproterozoic to early Mesoproterozoic in age. There are no data, however, to indicate if either the Paleoproterozoic or Archaean lithologies formed a basement onto which sediments were later deposited.

Zircon age data from this study indicate that the Rayner Structural Episode occurred over the period 1000-900 Ma. This is a shorter period than that proposed by White & Clarke (1993), who suggested that Rb-Sr ages of *c.* 1200 Ma from gneisses in the Framnes Mountains indicate the start of granulite facies metamorphism. The *c.* 1200 Ma dates, however, may be protolith ages, or may be partially reset by *c.* 1000 Ma metamorphism. The 1000-900 Ma period matches estimates for the timing of the Rayner Structural Episode in other parts of the Rayner Complex, such as the northern Prince Charles Mountains and Enderby Land (this is discussed in detail in Chapter 6). The maintenance of near-constant pressures during the Rayner Structural Episode (see Chapter 3) is consistent with the short-lived nature of the event.

With the possible exception of the pre-D₁ *charnockitic orthogneiss* and parts of the *proclamation gneiss*, all intrusive activity recorded at Cape Bruce and the Colbeck

Archipelago occurred during the Rayner Structural Episode. Igneous units that were emplaced between structural events have comparable ages where the events can be correlated between different localities. At Cape Bruce, felsic intrusions that were emplaced after D₂ and before or during D₄ range in age from *c.* 969 to *c.* 923 Ma. These correspond in age to pegmatites at Fold Island, that were emplaced at 940±80 Ma (Grew *et al.*, 1988). The Fold Island pegmatites were intruded during the D₃ event of James *et al.* (1991), which may correspond to the D₃ event of this study (see Chapter 2).

The Mawson Charnockite was emplaced at Chapman Ridge between the D₂ and D₄ events of this study. The new ages fit well with the 954±12 Ma estimate for the emplacement of the Mawson Charnockite (Young & Black, 1991). However there is no clear evidence that the post-D₂ felsic intrusions at Cape Bruce and Fold Island are co-magmatic with the Mawson Charnockite. As large amounts of migmatite were generated at Cape Bruce during D₁ and D₂, it is possible that the felsic intrusions were derived from the pooling of partial melts from rocks in the local vicinity. Pyroxene-bearing felsic intrusions, such as the *charnockitic orthogneiss*, the *biotite-pyroxene orthogneiss* and the Mawson Charnockite, were derived from magmas with drier sources of melting, and probably originated with the anatexis of lower crust. The presence of abundant biotite in these rocks, especially in the *biotite-pyroxene orthogneiss*, may be the result of either the mixing of locally derived anatectic melts with magma, or of the hydrous retrogression of solid intrusions after the solidification of the partial melts.

The polyphase nature of *c.* 1000 Ma metamorphism along the Mawson Coast, as suggested by White (1993), is supported by the new age data. Distinct periods of zircon growth at *c.* 995-945 and *c.* 910 Ma indicate two periods of elevated P-T conditions (M₁-M₂ and M₃), separated by a hiatus. These metamorphic events correspond to the M₂ and M₃ events of White (1993), that occurred before and after the emplacement of the Mawson Charnockite at Chapman ridge. The M₁ granulite

facies of White (1993) is now interpreted as the peak of a continuous M_1 - M_2 event. Granulite facies metamorphism before and after the emplacement of the Mawson Charnockite is also recognised in the Framnes Mountains (Clarke & Norman, 1993). Rb-Sr ages from the Mawson Charnockite (Black *et al.*, 1987; Young & Black, 1991), and a U-Pb zircon age for metamorphic zircon in an orthogneiss from Mawson base (Young & Black, 1991), are within error of the *c.* 910 Ma estimate for M_3 .

The D_5 event, which produced shear zones containing mylonite and pseudotachylite along the Mawson Coast, has not been temporally constrained. Clarke (1987) obtained a Rb-Sr age of 718 ± 10 Ma for biotite in an undeformed pegmatite which is displaced by a D_5 shear zone. Since Sr in biotite can be re-equilibrated above 200°C, the age may have been disturbed after the emplacement of the pegmatite, and therefore cannot be used to constrain the age of D_5 . Mineral growth adjacent to D_5 mylonites at Cape Bruce and the Colbeck Archipelago was shown by White & Clarke (1994) to have occurred at similar pressures to those present during metamorphism in the Rayner Structural Episode. Although the metamorphic conditions within D_5 shear zones may not reflect those in the host rocks (White & Clarke, 1994), the similarity of pressures is consistent with the final stage of 1000-900 Ma tectonism. Alternatively, D_5 may constitute the only evidence of tectonism at granulite facies conditions, that postdates the Rayner Structural Episode.

4.9 Conclusion

New chronological data from Cape Bruce provide insights into the geological evolution of the Mawson Coast. The application of the mixture modelling technique of Sambridge & Compston (1994) to structurally constrained samples has enabled individual events to be resolved within the *c.* 1000-900 Ma Rayner Structural Episode. At Cape Bruce, sedimentary and igneous protoliths were intruded by charnockitic, basic and granitic magmas, before and during deformation (D_1 - D_2) and granulite

facies metamorphism (M_1 - M_2), during the period *c.* 1000-945 Ma. Felsic pegmatites were intruded at 944 ± 10 Ma, and were subsequently deformed (D_3 - D_4), and then metamorphosed at granulite facies conditions (M_3) at *c.* 910 Ma. D_5 mylonitic shear zones were developed under granulite facies conditions (M_4) at some stage after M_3 . At least two distinct periods of metamorphism occurred during the Rayner Structural Episode at Cape Bruce.

Granulites at Cape Bruce and the Colbeck Archipelago were not metamorphosed at the same time as *c.* 2700 Ma granulites at Stillwell Hills, as suggested by White & Clarke (1993). Exposures east of the Stillwell Hills were metamorphosed and intruded into during the *c.* 1000 Ma Rayner Structural Episode. Age inheritance in metasediments and orthogneisses indicate that they were underlain by Archaean and Paleoproterozoic lithologies. Archaean crust exposed in the Stillwell Hills and underlying Proterozoic crust further east may represent an eastward extension of the Napier Complex, which was reworked by *c.* 1000 Ma tectonism. This possibility is discussed further in Chapter 6.

Chapter 5

Diffusion metasomatism in silica-undersaturated granulites



He told me one day that he was being followed by a lurking suspicion. p.27

-W.E. Bowman, The Ascent of Rum Doodle

5. Diffusion metasomatism in silica-undersaturated granulites

Abstract

At Rumdoodle Peak, near Mawson Base, east Antarctica, a xenolith of metasedimentary granulite is enclosed in Mawson Charnockite, an extensive *c.* 960 Ma intrusion. The xenolith contains a disrupted layer of black spinel-orthopyroxene-phlogopite gneiss, that was truncated by quartz-rich pegmatite. A reaction band that developed between these units is composed of a sequence of mineral zones that contain spinel, orthopyroxene, sapphirine, cordierite and plagioclase. The sequence of mineral zones approximately matches that predicted by a model of closed system diffusion metasomatism, involving the exchange of Si for Fe and Mg. The sequence differs from that predicted by the model in the presence of micro-scale disequilibrium textures, that include 'double coronas' composed of cordierite surrounding sapphirine and sapphirine surrounding spinel. The growth of the reaction band was controlled by diffusion along intergranular pathways, with local equilibrium maintained adjacent to grain boundaries. The presence of corona textures is a result of slow reaction rates, due to limited diffusive exchange of Si and Al across mineral grains.

Introduction

Several authors have produced simple models for mass transfer via diffusion between two reactant substances (Fletcher & Hofmann, 1974; Brady, 1977; Joesten, 1977; Frantz & Mao, 1979; Brady, 1983). Such models usually involve a simple three-component system with one to three mobile components. A metasomatic reaction band is inferred to develop as a result of steady state diffusion, where continuous

monotonic gradients in the chemical potential of the diffusing components are generated between the reactants (Brady, 1977). The model reaction band consists of a series of well defined zones of unique modal composition, with boundaries parallel to the initial contact between the reactants. The models assume that certain ideal conditions are required for diffusion to take place. These include the maintenance of local equilibrium at all points between the two reactants, and fixed mineral compositions (Brady, 1977).

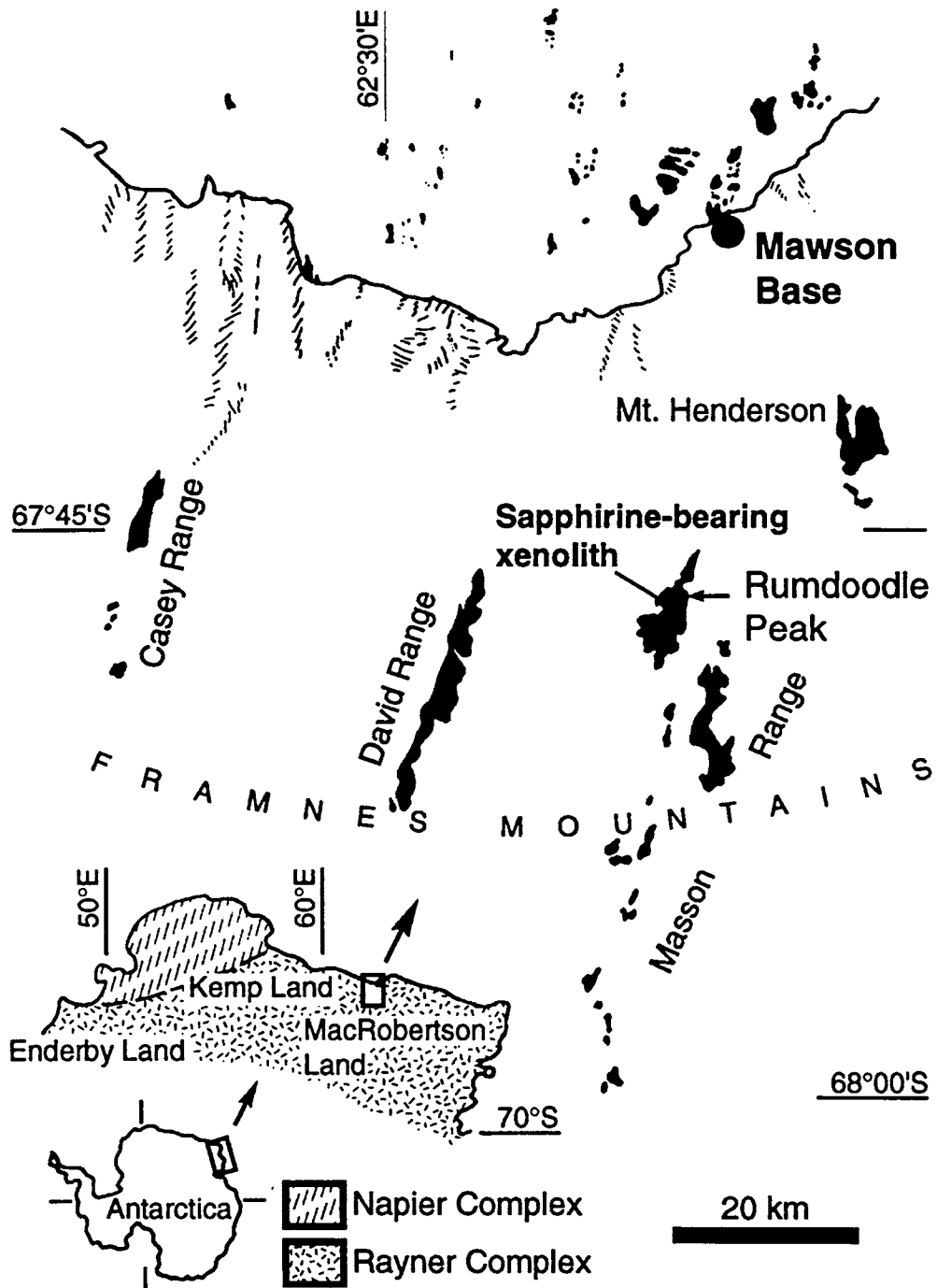
Whereas such models have been successful in explaining the occurrence of chemical zoning in metamorphic rocks, diffusive behaviour in real rocks is complicated by many variables, which may include variable reaction rates, ambient temperatures, water activities, and cationic exchange between co-existing minerals. Under non-ideal conditions diffusion can still take place, but the resulting reaction band may possess features not predicted by the simple models. Such features include irregular, inhomogenous, unpredicted or absent zones, incomplete reaction textures, and variations in mineral composition.

In this chapter, reaction bands that separate quartz-rich pegmatite and aluminous granulite are described. The morphology of the reaction bands broadly corresponds to that predicted by a simple model involving diffusion metasomatism. The reaction bands also have features that are inconsistent with the model, including reaction coronas and zones containing more than two minerals. Such disparities can be explained by considering the different behaviour of diffusing components within and between mineral grains in the reaction bands.

5.1 Geological Setting

The rocks of the MacRobertson Land coast (Fig. 5.1) are a part of the regionally extensive Rayner Complex (Sheraton, 1980), which comprises upper amphibolite to

Fig. 5.1. Locality map of the sapphirine-bearing xenolith near Rumdoodle Peak, Framnes Mountains, east Antarctica.



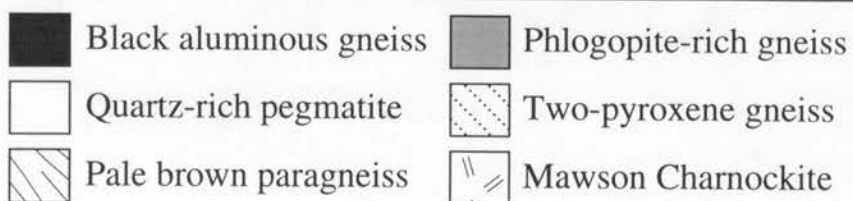
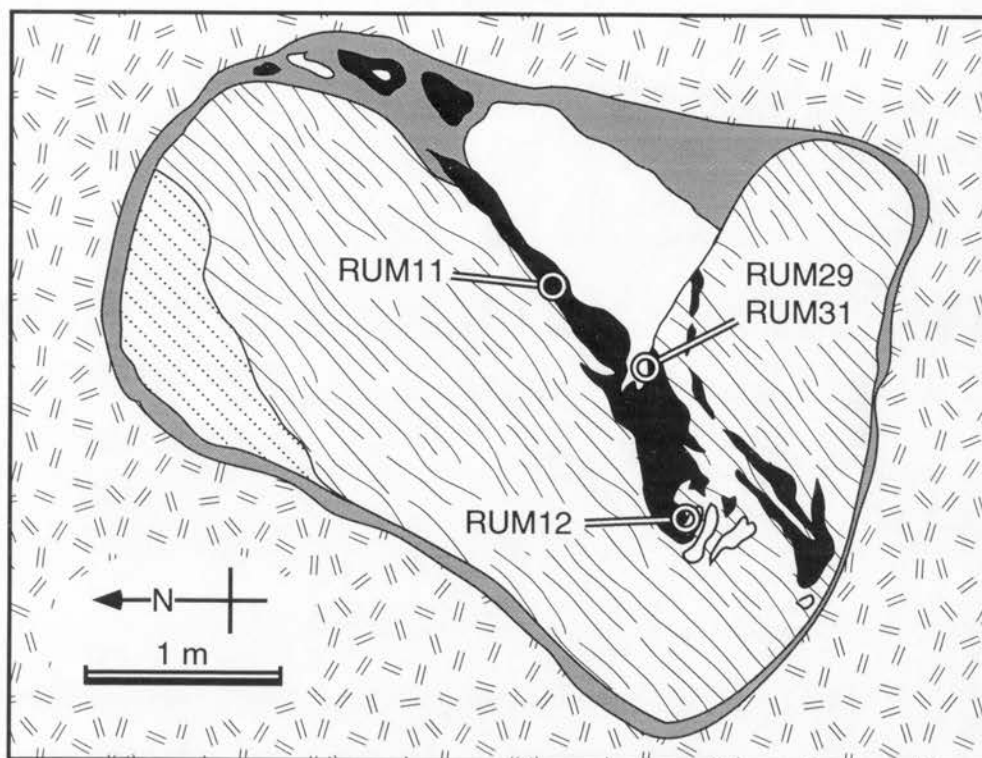
granulite facies gneisses deformed during the Neoproterozoic Rayner Structural Episode (Sandiford & Wilson, 1984). Four episodes of pervasive deformation (D_{1-4}), and the formation of discrete shear zones (D_5), are recognised in the area around Mawson Base, including the coastline and the Framnes Mountains to the south (Clarke, 1988; White & Clarke, 1993). This area of outcrop is dominated by the Mawson Charnockite (Trail, 1970). U-Pb dating of zircons by Young & Black (1991) place the age of crystallisation of this intrusion at *c.* 960 Ma. The charnockite was emplaced after D_2 , and was deformed by the D_{3-5} events (Clarke & Norman, 1993).

In the Framnes Mountains (Fig. 5.1), the Mawson Charnockite contains xenoliths that are centimetres to kilometres in size, which are grouped together as Painted Gneiss (Trail, 1970). This lithology includes paragneiss and orthogneiss units metamorphosed at granulite facies during or before a *c.* 1000 Ma D_1 event (Clarke & Norman, 1993). Retrograde S_2 and S_3 foliations in the Painted Gneiss are defined by aligned grains of biotite and hornblende.

5.2 The Rumdoodle xenolith

A small xenolith of Painted Gneiss, enclosed in Mawson Charnockite, is located approximately 1km south of Rumdoodle Peak in the North Masson Range (Fig. 5.1). The xenolith is composed of layers of aluminous paragneiss, two-pyroxene orthogneiss and felsic pegmatite (Fig. 5.2). Most of the xenolith is composed of light brown paragneiss, that consists of alternating layers cordierite-rich and plagioclase-rich gneiss. The cordierite-rich gneiss is composed of orthopyroxene, cordierite and phlogopite, with minor amounts of rutile and monazite. The plagioclase-rich gneiss contains orthopyroxene and antiperthite, with small amounts of phlogopite. It is typically coarser-grained than the cordierite-rich gneiss. Phlogopite grains in the paragneiss commonly define an S_2 foliation that lies at a high angle to the S_1 gneissic layering.

Fig. 5.2. Outcrop map of the Rumdoodle xenolith.



The xenolith also contains a narrow, dismembered layer of black aluminous gneiss, that lies sub-parallel to the S_1 gneissosity in the adjacent paragneiss. The black aluminous gneiss is unfoliated and is composed of pale brown orthopyroxene (60-70 modal %) and green spinel (30-35 modal %). Orthopyroxene grains have polygonal anhedral forms, and are larger (0.3-0.6mm) than spinel grains (0.1-0.4mm). Some samples (*e.g.* RUM11) contain up to 2 modal % rutile, with most grains containing rims and patches of secondary ilmenite. Minor amounts of phlogopite (less than 2 modal %) are present as randomly oriented grains.

The layer of black aluminous gneiss is cut by white pegmatite, which occurs as a series of irregular blobs aligned sub-parallel to S_1 in the paragneiss. The pegmatite is mostly composed of quartz, with 3-5 modal % of antiperthite. Minor amounts of orthopyroxene and phlogopite are present in a few samples. The entire xenolith is separated from the Mawson Charnockite by a sheath of biotite-rich gneiss 1-15cm thick.

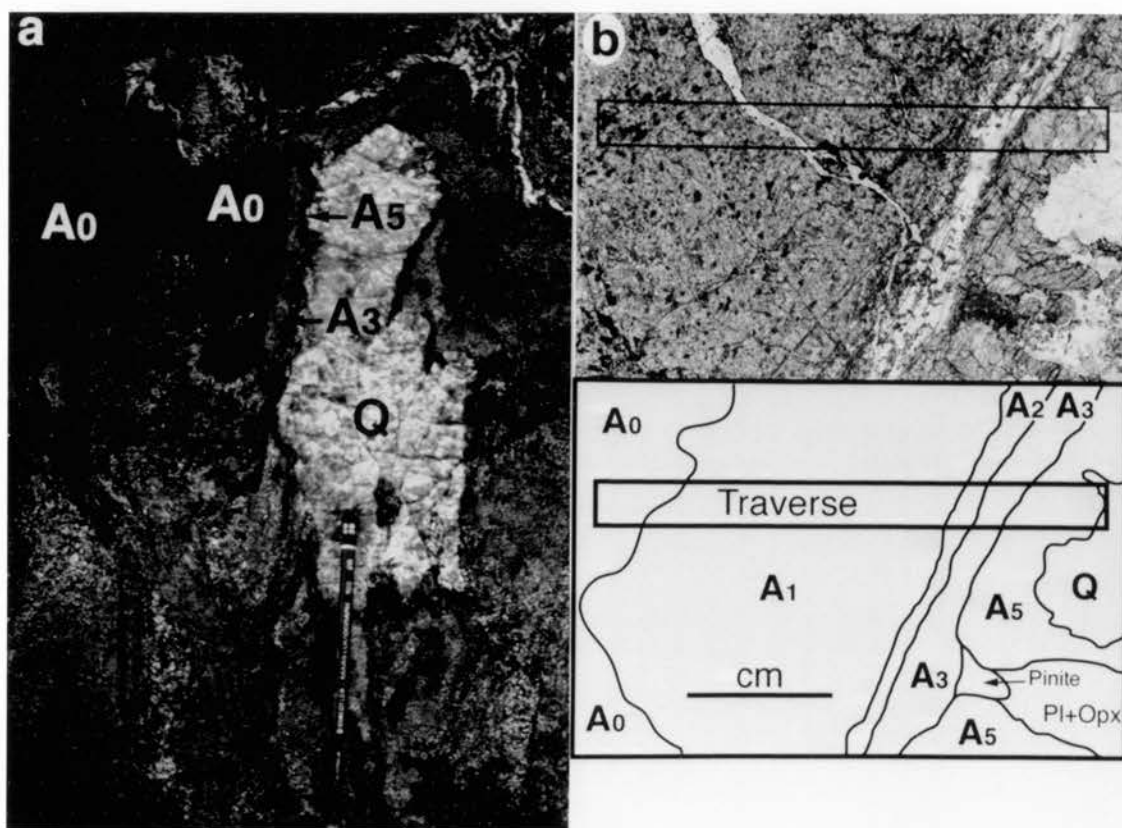
A north-west-trending S_1 in the xenolith is defined by gneissic layers 0.5-10cm thick. Aligned phlogopite grains define a sub-vertical north-east-trending S_2 . S_1 and S_2 are deformed into small open F_3 folds. These folds have upright north-west-trending axial planes, which are parallel to an S_3 foliation in the Mawson Charnockite defined by aligned megacrysts of alkali feldspar. A faint, vertical, north-trending S_4 foliation, defined by aligned grains of phlogopite, crosses earlier foliations in the xenolith and the Mawson Charnockite.

The contacts between quartz pegmatite and black aluminous gneiss are complex. Patches of quartz pegmatite occur in aluminous gneiss (Plate 5.1a) and pods of aluminous gneiss occur in quartz pegmatite. Where black aluminous gneiss lies adjacent to quartz-rich pegmatite, the two are separated by bands of light-coloured aluminous gneiss (Plate 5.1a). These 'reaction bands' have a complex structure and composition, which is described below.

Plate 5.1.

a) A pod of quartz-rich pegmatite (Q) enclosed in aluminous gneiss (A₀) and surrounded by zoned reaction bands. Mineral assemblages in zones A₀, A₃ and A₅ are described in the text.

b) Thin section of the reaction band from sample RUM12. The sketch shows the zones present and the location of the traverse of mineral analyses described in the text.



Petrography of the reaction bands

The reaction bands are 7-30mm wide, and typically have a uniform thickness. A single band may contain up to five mineralogically distinct zones. On planar contacts between the aluminous gneiss and quartz pegmatite, these zones occur as parallel layers (sample RUM12; Plate 5.1b). Pods of aluminous gneiss in quartz pegmatite are enclosed by a set of concentric zones (samples RUM29 and RUM31; Plate 5.2a). Zones in reaction bands that surround pods of aluminous gneiss are wider than those developed on planar contacts.

Mineral assemblages and textures in the zones of a typical reaction band are summarised in Fig. 5.3. To simplify the description of the reaction bands, the constituent zones have been labelled from A₁ to A₅. The sequence of zones runs from zone A₁, which is adjacent to the black aluminous gneiss (labelled A₀), to zone A₅, which lies adjacent to the quartz pegmatite (labelled Q). In small pods of aluminous gneiss that are enclosed in quartz pegmatite, A₀ may be absent from the pod's core, which may consist of zone A₁ or A₂ (Plate 5.2a).

In black aluminous gneiss (A₀) adjacent to a reaction band, small amounts of sapphirine may be present in patches and web-like networks (Plate 5.2b). Within 1-2cm of the contact with quartz pegmatite, these patches merge to form a distinct zone (A₁, Fig. 5.3), in which all grains of spinel are separated from orthopyroxene by coronas composed of single grains of sapphirine. Grains of sapphirine that pseudomorph spinel are also present (20-30 modal %). The amount of orthopyroxene in zone A₁ is similar to that in the adjacent A₀ gneiss (60-70 modal %).

There is a sharp compositional boundary between zone A₁ and zone A₂. Zone A₂ contains cordierite as single-grain coronas around sapphirine. Sapphirine grains commonly contain cores of spinel (Plate 5.2c). The A₂ zone contains less orthopyroxene (45-60 modal %) than the A₁ zone. The amount of cordierite in zone A₂ increases from approximately 15 modal % to 35 modal % with distance towards

Plate 5.2.

a) Thin section from sample RUM31, showing a concentric reaction band developed around a pod of aluminous gneiss, which is enclosed in quartz-rich pegmatite.

b) Aluminous gneiss (zone A₀), with the incipient development of Spr coronas around Spl in the top half of the photo. Sample RUM11.

c) Zone A₂, showing Crd coronas around Spr grains, that contain cores of Spl. Sample RUM31.

d) Opx grain in zone A₅. The grain extends across the width of the zone. Chlorite has replaced Opx adjacent to the quartz pegmatite (Q). Sample RUM29.

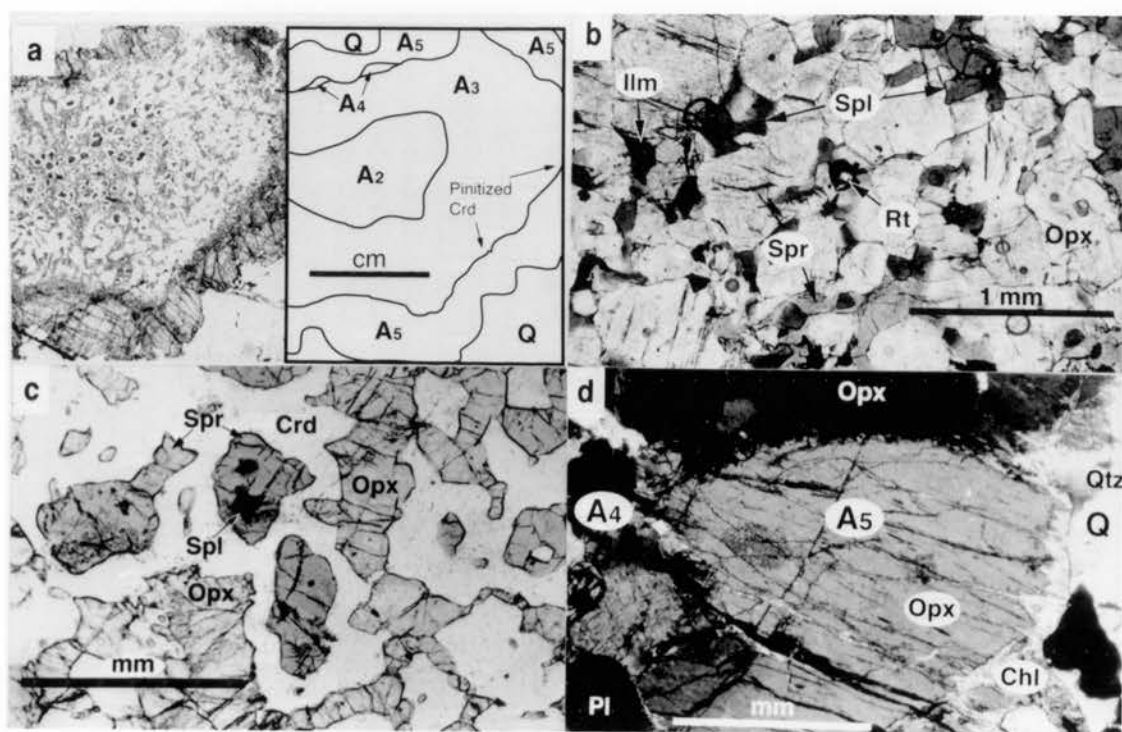
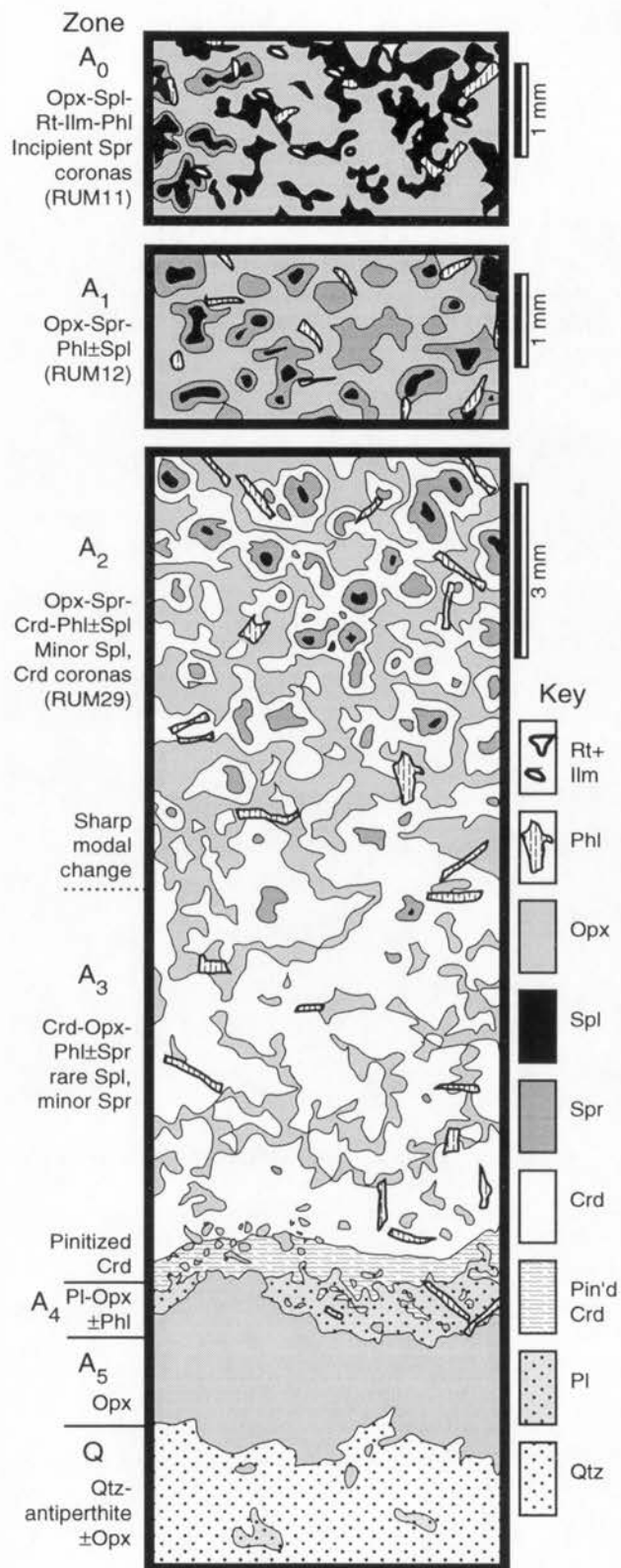


Fig. 5.3. Diagram of an idealised complete reaction band between aluminous gneiss (A₀) and quartz pegmatite (Q). Textures are based on photomicrographs of samples RUM11, RUM29 and RUM31.



the quartz pegmatite, at the expense of sapphirine (10-25 modal %) and spinel (less than 2%). In samples with a planar reaction band (Plate 5.1b), zone A₂ is always less than 2mm thick. In small aluminous pods (Plate 5.2a) it may form a shell or core up to 10mm across.

The A₃ zone has a sharp boundary with the A₂ zone, and is distinguished from the latter by a greater proportion of cordierite (60-70 modal %) and a smaller proportion of orthopyroxene (20-30 modal %). Phlogopite and sapphirine are present in minor amounts (less than 5%), and spinel is rarely present as cores in sapphirine grains. The A₃ zone is wider than the A₂ zone in all samples. It may be up to 3mm wide in samples with a planar reaction band, and up to 20mm wide in aluminous pods. In some samples (*e.g.* RUM12), a slight modal gradient is present in zone A₃, with cordierite increasing towards the quartz pegmatite. In the part of zone A₃ nearest to the quartz pegmatite, cordierite is always replaced by yellow pinite, which forms a continuous zone 0.3-1mm wide (Plates 5.1b, 5.2a).

Zone A₄ is present only in some samples of aluminous pods. It consists of albite or oligoclase, intergrown with small amounts of fine-grained orthopyroxene (2-15 modal %) and pinitized cordierite (2-10 modal %). Symplectic intergrowths of orthopyroxene, pinitized cordierite and/or plagioclase are common. Where present in a reaction band, zone A₄ is less than 1mm thick, and is discontinuous along the length of the reaction band, as illustrated in Fig. 5.3.

Quartz pegmatite is always separated from the aluminous reaction band by a well defined zone of orthopyroxene, 3 to 7mm thick (zone A₅). Individual crystals are large and commonly span the width of the A₅ zone (Plate 5.2d). Minor amounts of plagioclase, phlogopite and chlorite may be present in gaps between orthopyroxene grains.

5.3 Analytical methods

Analyses of the bulk composition of reaction band zones were conducted at the University of Edinburgh, Scotland. Samples were used containing reaction bands that were planar (RUM11, RUM12) and pod-shaped (RUM29, RUM31). The samples were processed by a method based on that described by Nicholls (1974). The reaction bands were cut into slabs 0.5cm thick. Pellets with a diameter of 3mm were obtained from each zone using a hollow-core drill. The A₄ zone, where present in the samples, was too narrow to obtain pellets. The pellets were powdered and melted on a molybdenum strip heater. Glass shards were then checked for unmelted patches and oxidation, before mounting and polishing. They were analysed using the Cameca Electron Microprobe at the University of Edinburgh, operating at 20kV and 5nA. The electron beam was rastered over an area of 5µm by 5µm. Minerals were analysed with the Cameca Camebax Electron Microprobe at the University of New South Wales, Australia, using an operating voltage of 15kV at 20nA, and applying PAP data reduction as described by Pouchou & Pichoir (1984). Total Fe was measured as FeO.

5.4 Chemistry of the reaction bands

5.4.1 Bulk rock chemistry

The compositions of zones in various samples are listed in Table 5.1. Variation trends for major elements from zone A₀ to A₃ are shown on SFMA (SiO₂ - FeO + MgO - AlO_{3/2}) and AFM (AlO_{3/2} - FeO - MgO) plots. Compositions are plotted as molar proportions. On the SFMA plot (Fig. 5.4a), a linear trend projects from the spinel-orthopyroxene tie-line (A₀ zone) towards the cordierite-orthopyroxene tie-line (A₃ zone), with SiO₂ increasing at the expense of FeO + MgO. There is also a slight

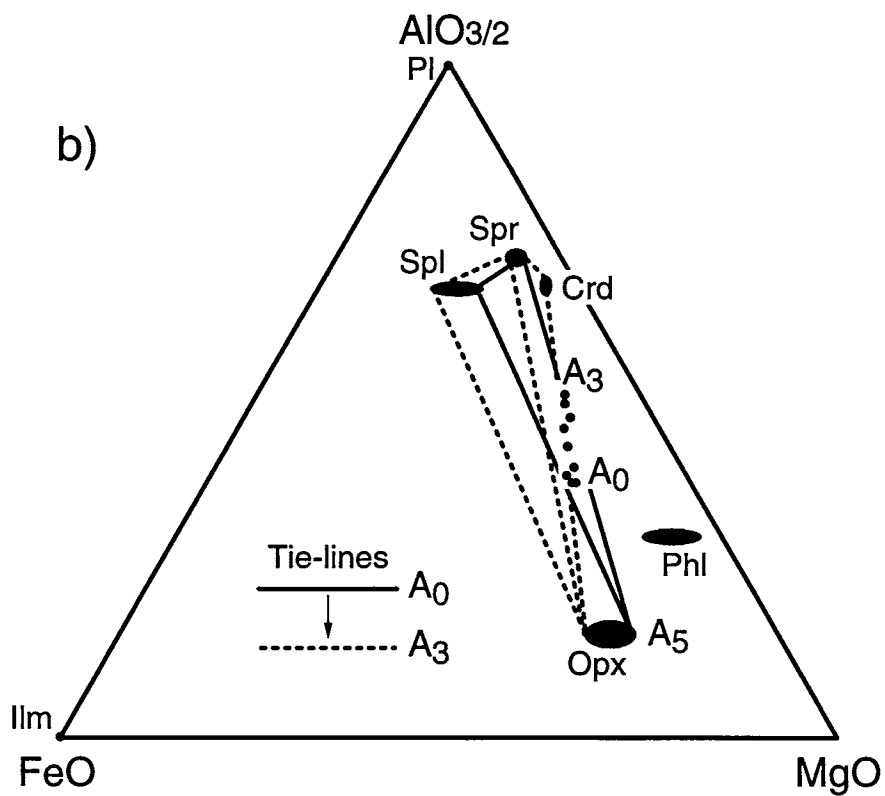
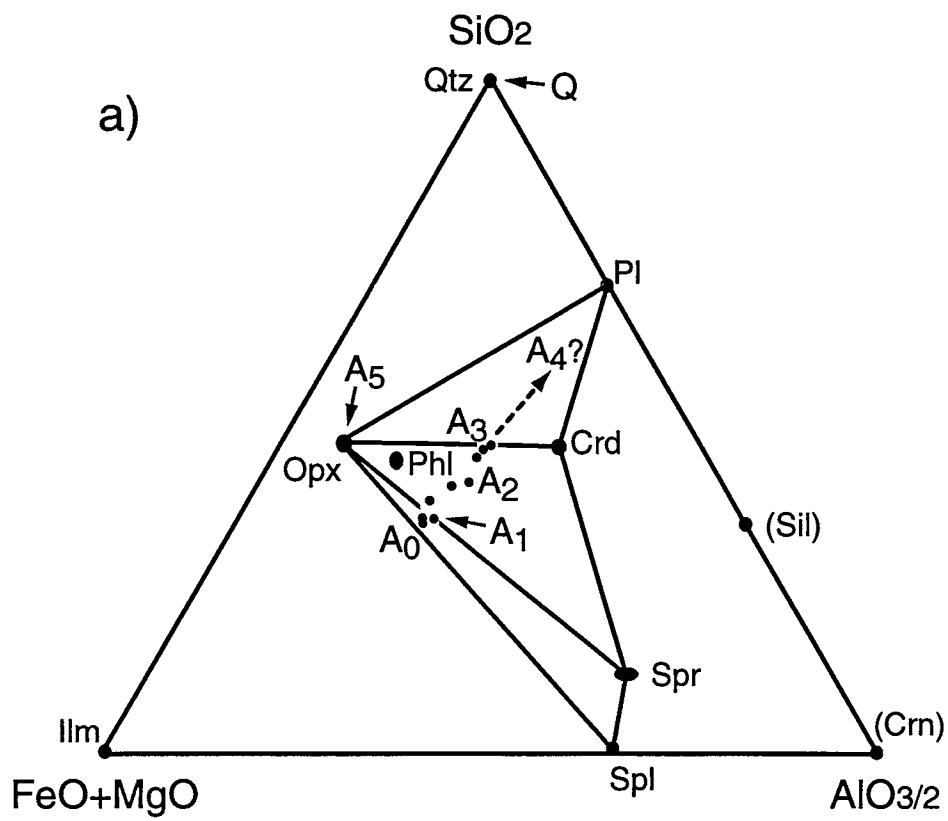
Table 5.1. Bulk rock composition of zones in reaction bands, expressed as weight % of oxides. Suffixes to zone labels indicate the number of analyses averaged for each zone (*e.g.* A0 3 is an average of 3 analyses). X_{Fe} = molar $\text{Fe}/(\text{Fe} + \text{Mg})$.

Sample Zone	RUM11	RUM12					RUM29			RUM31		
	A ₀ 3	A ₀ 4	A ₁ 10	A ₃ 4	A ₅ 2		A ₁ 2	A ₂ 2	A ₃ 4	A ₂ 4	A ₃ 6	A ₅ 2
SiO ₂	37.12	38.17	38.05	47.91	49.96		40.75	44.10	48.66	43.35	49.19	51.62
TiO ₂	1.21	0.09	0.12	0.08	0.18		0.17	0.12	0.07	0.17	0.09	0.12
Al ₂ O ₃	22.37	22.64	23.66	24.22	8.71		22.17	25.17	24.87	23.51	25.32	8.64
Cr ₂ O ₃	0.01	0.00	0.00	0.00	0.00		0.00	0.00	0.00	0.00	0.00	0.00
FeO	13.35	13.06	12.53	7.94	13.59		12.84	9.85	7.47	10.76	7.05	13.35
MnO	0.07	0.05	0.09	0.06	0.12		0.08	0.02	0.07	0.06	0.05	0.13
MgO	23.11	23.56	23.21	17.73	24.32		21.80	18.84	16.64	19.82	16.38	22.94
CaO	0.23	0.09	0.08	0.11	0.27		0.10	0.09	0.11	0.09	0.08	0.39
NiO	0.00	0.03	0.03	0.01	0.00		0.02	0.02	0.04	0.02	0.02	0.01
Na ₂ O	0.05	0.03	0.05	0.08	0.37		0.05	0.04	0.05	0.04	0.05	0.35
K ₂ O	0.72	0.30	0.36	0.31	0.35		0.30	0.29	0.31	0.63	0.37	0.22
SO ₃	0.00	0.00	0.01	0.00	0.00		0.00	0.00	0.00	0.00	0.00	0.00
Total	98.25	98.02	98.17	98.44	97.85		98.29	98.52	98.32	98.46	98.61	97.78
X _{Fe}	0.245	0.237	0.232	0.201	0.239		0.248	0.227	0.201	0.234	0.195	0.246

Fig. 5.4. Compatibility diagrams for samples in the SFMA system. Compositions are plotted as molar proportions. Minerals in parentheses were not observed in the samples.

a) Ternary plot for S(FM)A (SiO_2 - $\text{FeO}+\text{MgO}$ - $\text{AlO}_3/2$). Bulk rock composition varies in a linear fashion from zone A₀ to zone A₃. This trend can be extrapolated towards the A₄ zone ($\text{Opx} + \text{Pl} \pm \text{Phl}$).

b) Ternary plot for AFM (Al-Fe-Mg). A linear trend in bulk rock compositions is shown from zones A₀ to A₃. Orthopyroxene, sapphirine and spinel become more Fe-enriched from zone A₀ to zone A₃. The shift in mineral tie-lines from zone A₀ to A₃ is shown.



increase in $\text{AlO}_{3/2}$ in the trend. In the AFM plot (Fig. 5.4b), a linear trend from zone A_0 to A_3 is directed towards the $\text{AlO}_{3/2}$ - MgO join; *i.e.* with increasing $\text{AlO}_{3/2}$, FeO decreases to a greater degree than MgO . This relation is also expressed in Fig. 5.5 as a decrease in X_{Fe} (= molar $\text{Fe}/(\text{Fe}+\text{Mg})$) from zone A_0 to A_3 , within each of the samples analysed. Between samples, X_{Fe} values for A_0 are similar in samples RUM11 and RUM12, whereas X_{Fe} in each zone from the 'pod-shaped' samples (RUM29 and RUM 31) is higher than in the corresponding zone in sample RUM12. Zone A_5 (monomineralic orthopyroxene) does not follow the composition trends above. SiO_2 and FeO concentrations are higher in zone A_5 than in A_3 (and A_0), whereas Al is much lower.

With the exception of Na_2O and CaO in plagioclase in the A_4 zone, TiO_2 , CaO , Na_2O , K_2O and ZnO have low concentrations, that do not vary systematically across the reaction band. TiO_2 and K_2O vary sympathetically with modal proportions of rutile, ilmenite and phlogopite. For example, ilmenite and rutile are present in sample RUM11 ($\text{Ti} = 0.056$ cations per 10 oxygen atoms) , but not in the A_0 zone of sample RUM12 ($\text{Ti} = 0.004$ cations per 10 oxygens).

5.4.2 Mineral chemistry

Compositions of minerals in the reaction bands, averaged for each zone, are presented in Table 5.2. To examine variations in mineral chemistry across a representative reaction band, a thin section was cut from sample RUM12, perpendicular to the planar contact between quartz pegmatite and aluminous gneiss. A series of points were analysed along a 35mm by 3mm traverse, that lay at a high angle to the trend of the zones in the reaction band (Plate 5.1b). The site distribution of cations in orthopyroxene, spinel and sapphirine was calculated assuming perfect stoichiometry and ideal mixing, using the mineral formulas given by Deer *et al.* (1992, pp. 148, 217 & 559).

Fig. 5.5. X_{Fe} values for mineral and bulk rock compositions of zones in each sample. Analyses were averaged for each zone.

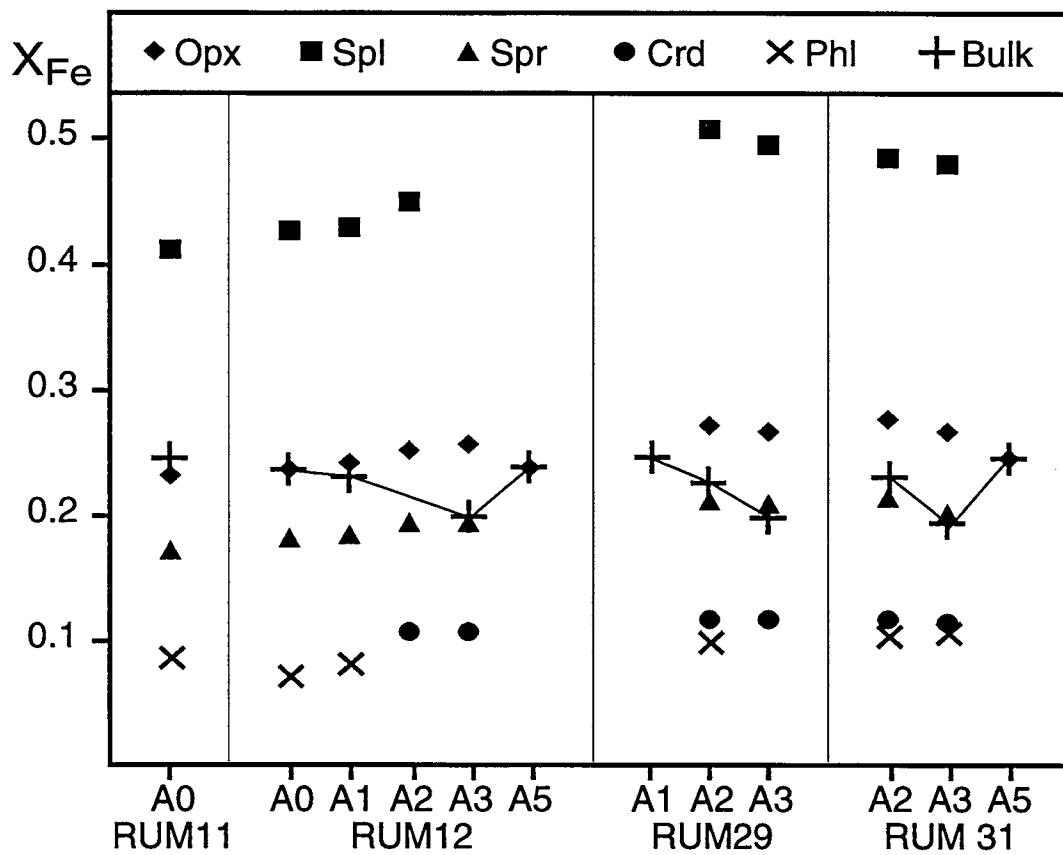


Table 5.2. Mineral composition averaged by zone in reaction bands, expressed weight % of oxides. Mineral abbreviations as *per* Kretz (1983). Suffixes to mineral labels indicate the number of analyses averaged for each zone (*e.g.* Opx 9 is an average of 9 analyses). X_{Fe} = molar $\text{Fe}/(\text{Fe} + \text{Mg})$.

Sample Zone Mineral	RUM11				RUM12							
	A ₀	A ₀	A ₀	A ₀	A ₀	A ₀	A ₀	A ₀	A ₁	A ₁	A ₁	A ₁
	Opx 9	Spl 12	Spr 12	Phl 6	Opx 52	Spl 12	Spr 30	Phl 2	Opx 52	Spl 18	Spr 59	Phl 3
SiO ₂	50.70	0.03	13.10	39.67	51.14	0.01	13.37	41.75	51.11	0.02	13.27	40.31
TiO ₂	0.14	0.01	0.08	2.94	0.04	0.01	0.02	1.24	0.06	0.01	0.03	1.67
Al ₂ O ₃	7.25	64.03	62.09	13.74	6.76	63.32	61.68	13.23	6.59	63.03	61.66	13.20
Cr ₂ O ₃	0.03	0.11	0.04	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.00
FeO	14.47	19.54	6.55	4.00	14.87	20.09	6.94	3.44	15.13	20.16	7.02	3.87
MnO	0.10	0.06	0.03	0.01	0.10	0.04	0.03	0.02	0.09	0.05	0.03	0.00
MgO	26.98	15.68	17.53	23.44	26.63	15.16	17.32	24.69	26.42	14.95	17.27	24.20
ZnO	0.01	0.34	0.04	0.03	0.04	0.58	0.04	0.00	0.03	0.54	0.02	0.04
CaO	0.07	0.01	0.01	0.01	0.08	0.01	0.01	0.00	0.08	0.00	0.01	0.02
Na ₂ O	0.01	0.01	0.01	0.37	0.01	0.02	0.01	0.41	0.01	0.01	0.01	0.42
K ₂ O	0.01	0.01	0.01	10.15	0.01	0.01	0.01	10.04	0.01	0.00	0.01	10.05
Total	99.77	99.83	99.49	94.35	99.71	99.26	99.45	94.84	99.53	98.81	99.34	93.80
X _{Fe}	0.231	0.412	0.173	0.087	0.239	0.427	0.184	0.072	0.243	0.431	0.186	0.082

Sample Zone Mineral	RUM12					RUM29				
	A ₂	A ₂	A ₂	A ₂	A ₃	A ₃	A ₃	A ₅	A ₂	A ₂
	Opx 3	Spl 3	Spr 2	Crd 3	Opx 5	Spr 2	Crd 8	Opx 15	Opx 8	Spl 2
SiO ₂	49.71	0.02	13.78	49.84	50.20	13.10	49.73	50.11	50.43	0.02
TiO ₂	0.08	0.01	0.04	0.00	0.09	0.08	0.01	0.12	0.06	0.00
Al ₂ O ₃	8.25	62.48	60.45	33.88	7.77	61.73	33.91	7.84	7.01	61.95
Cr ₂ O ₃	0.00	0.05	0.04	0.02	0.00	0.08	0.02	0.01	0.01	0.04
FeO	15.54	20.94	7.43	2.66	15.80	7.34	2.66	15.48	16.74	23.03
MnO	0.10	0.04	0.01	0.02	0.12	0.03	0.03	0.10	0.13	0.07
MgO	25.70	14.30	17.22	12.30	25.53	16.92	12.35	25.71	25.13	12.55
ZnO	0.02	0.73	0.08	0.03	0.01	0.03	0.01	0.04	0.01	1.31
CaO	0.04	0.01	0.01	0.01	0.06	0.00	0.01	0.04	0.05	0.02
Na ₂ O	0.02	0.02	0.01	0.02	0.01	0.02	0.02	0.01	0.00	0.04
K ₂ O	0.00	0.01	0.00	0.01	0.00	0.01	0.01	0.01	0.01	0.01
Total	99.47	98.62	99.06	98.78	99.59	99.33	98.75	99.48	99.59	99.03
X _{Fe}	0.253	0.451	0.195	0.108	0.258	0.196	0.108	0.253	0.272	0.507

Sample Zone Mineral	RUM29				RUM31							
	A ₃	A ₃	A ₃	A ₃	A ₂	A ₂	A ₂	A ₂	A ₂	A ₃	A ₃	A ₃
	Opx 10	Spl 6	Spr 6	Crd 13	Opx 8	Spl 11	Spr 13	Crd 9	Phl 6	Opx 12	Spl 5	Spr 14
SiO ₂	50.29	0.04	13.05	49.78	49.58	0.02	13.15	49.92	40.07	50.84	0.06	12.98
TiO ₂	0.09	0.02	0.05	0.02	0.07	0.02	0.03	0.00	2.18	0.04	0.01	0.04
Al ₂ O ₃	6.81	61.77	61.67	33.97	8.16	62.26	61.41	33.92	13.64	6.53	62.14	62.01
Cr ₂ O ₃	0.02	0.03	0.02	0.01	0.01	0.02	0.01	0.00	0.01	0.02	0.06	0.07
FeO	16.49	22.13	7.82	2.89	16.75	21.80	8.03	2.89	4.88	16.70	21.24	7.47
MnO	0.13	0.05	0.03	0.02	0.11	0.05	0.04	0.03	0.02	0.11	0.05	0.03
MgO	25.34	12.69	16.60	12.25	24.58	12.99	16.41	12.17	23.29	25.51	12.88	16.55
ZnO	0.02	1.91	0.05	0.05	0.05	1.88	0.04	0.03	0.07	0.06	2.70	0.04
CaO	0.05	0.01	0.01	0.01	0.06	0.01	0.02	0.01	0.00	0.05	0.07	0.01
Na ₂ O	0.02	0.06	0.01	0.02	0.01	0.05	0.01	0.01	0.21	0.01	0.09	0.01
K ₂ O	0.01	0.01	0.01	0.02	0.01	0.00	0.01	0.01	10.29	0.01	0.01	0.01
Total	99.26	98.72	99.31	99.03	99.39	99.11	99.16	98.99	94.67	99.87	99.30	99.20
X _{Fe}	0.267	0.495	0.209	0.117	0.277	0.485	0.215	0.117	0.105	0.269	0.481	0.202

The compositions of orthopyroxene, spinel and sapphirine vary between zones in the reaction band. These variations are discussed in detail below.

Orthopyroxene

Most of the Al in orthopyroxene was assigned to Mg and Fe- Tschermak components (Table 5.3). Fe^{3+} content was estimated at 4-15% of total molar Fe, with no systematic variation between different zones in the reaction bands.

Orthopyroxene in the RUM12 traverse shows some loosely-defined but systematic composition trends (Fig. 5.6a). Al content is variable within each zone, but is generally higher in zones A₂ to A₅ than in zone A₁. X_{Fe} increases gradually from zone A₁ to zone A₃. The X_{Fe} of orthopyroxene in the A₅ zone is comparable to that in zones A₂ and A₃.

X_{Fe} also increases slightly from zone A₀ (around 0.23) to zone A₃ (around 0.27) in averaged analyses from samples RUM11 and RUM12 (Fig. 5.5). This trend is antithetic to the X_{Fe} trend for the bulk rock composition of zones A₀ to A₃. In samples RUM31 and RUM29, the X_{Fe} of orthopyroxene from zones A₂ to A₃ decreases sympathetically with, although to a smaller degree to, the decreasing X_{Fe} of the bulk rock.

Other elements in orthopyroxene (Mn, Ca, Ti, Na) occur in trace amounts and show no systematic variation in concentration across the zones.

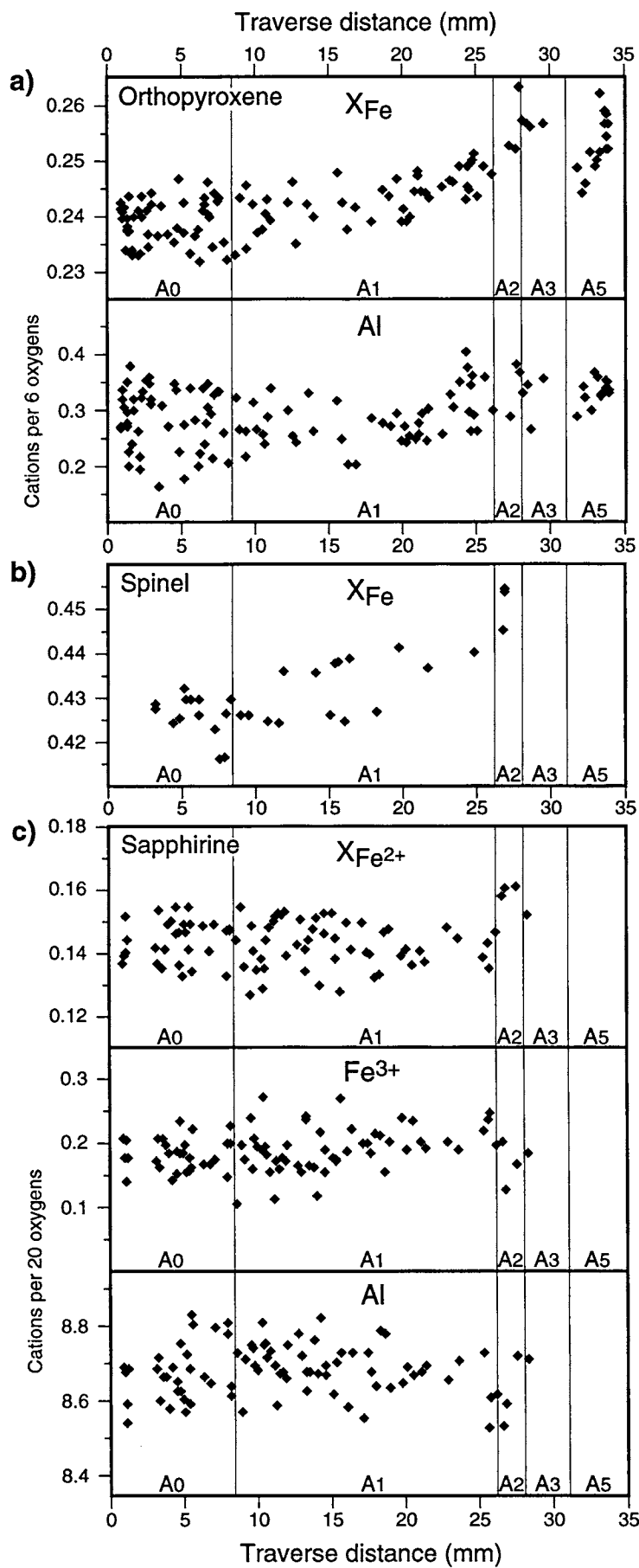
Spinel

Analysed spinel contains only minor amounts of magnetite in solid solution, with Fe^{3+} estimated at 6-10% of the total molar Fe measured (Table 5.3). Fe^{3+} does not vary systematically along the RUM12 traverse. X_{Fe} shows a poorly defined but gradually increasing trend from zone A₀ (around 0.43) to zone A₂ (around 0.45; Fig. 5.6b). The X_{Fe} of analyses averaged in zones increases from zone A₀ to zone A₂ for sample RUM12, and decreases slightly from zone A₂ to zone A₃ in samples RUM29

Table 5.3. Site distribution of cations in orthopyroxene and sapphirine, calculated assuming ideal mixing and perfect stoichiometry. Formulas used: orthopyroxene = (M2)(M1)[T₂O₆], and sapphirine = (M1)₇(M2)₂[T₆O₁₈], where M = octahedral sites, T = tetrahedral sites (after Deer *et al.*, 1992). $X_{Fe^{2+}} = Fe^{2+}/(Fe^{2+} + Mg)$.

Sample Zone Mineral	RUM12			RUM31			RUM12		RUM31			RUM12		RUM31		
	A ₀	A ₃	A ₅	A ₂	A ₃		A ₀	A ₃	A ₂	A ₃		A ₀	A ₂	A ₃		
	Op _x 52	Op _x 5	Op _x 15	Op _x 8	Op _x 12		Spr 30	Spr 2	Spr 13	Spr 14		Spl 12	Spl 11	Spl 5		
T site	Si	1.834	1.811	1.807	1.801	1.836	T site	Si	1.589	1.561	1.576	1.551	Si	0.000	0.001	0.002
	Ti	0.001	0.003	0.003	0.002	0.001	Ti	0.002	0.008	0.003	0.004	Ti	0.000	0.000	0.000	
	Cr	0.000	0.000	0.000	0.000	0.001	Cr	0.001	0.007	0.001	0.007	Cr	0.000	0.000	0.001	
	Al	0.164	0.187	0.189	0.197	0.163	Al	4.408	4.424	4.420	4.439	Al	1.954	1.956	1.952	
M site	Al	0.122	0.144	0.144	0.153	0.115	M site	Al	4.232	4.252	4.253	4.297				
	Fe ³⁺	0.044	0.044	0.047	0.045	0.049	Fe ³⁺	0.179	0.178	0.171	0.146	Fe ³⁺	0.045	0.043	0.047	
	Fe ²⁺	0.403	0.433	0.420	0.464	0.455	Fe ²⁺	0.511	0.554	0.633	0.601	Fe ²⁺	0.395	0.443	0.426	
	Mn	0.003	0.004	0.003	0.004	0.003	Mn	0.003	0.003	0.004	0.003	Mn	0.001	0.001	0.001	
	Mg	1.424	1.372	1.382	1.330	1.373	Mg	3.067	3.005	2.930	2.947	Mg	0.592	0.516	0.511	
	Zn	0.001	0.000	0.001	0.001	0.002	Zn	0.004	0.003	0.004	0.003	Zn	0.011	0.037	0.053	
	Ca	0.003	0.002	0.002	0.002	0.002	Ca	0.001	0.000	0.002	0.001	Ca	0.000	0.000	0.002	
	Na	0.001	0.001	0.001	0.001	0.001	Na	0.002	0.004	0.002	0.002	Na	0.001	0.003	0.005	
	K	0.001	0.000	0.000	0.000	0.000	K	0.001	0.002	0.001	0.001	K	0.000	0.000	0.000	
X _{Fe²⁺}	0.220	0.240	0.233	0.258	0.249	X _{Fe²⁺}	0.143	0.156	0.178	0.169	X _{Fe²⁺}	0.400	0.462	0.455		

Fig. 5.6. Mineral compositions along a traverse in sample RUM12 (see Fig. 5.3b). Zone A₄ is absent from this traverse. a) X_{Fe} and Al values in orthopyroxene. b) X_{Fe} values in spinel. c) Fe^{3+} , $X_{Fe^{2+}}$ and Al values in sapphirine, calculated from analyses assuming ideal 2-site mixing and perfect stoichiometry. $X_{Fe^{2+}} = \text{calculated } Fe^{2+} / (Fe^{2+} + Mg)$.



and RUM31 (Fig. 5.5). Other elements (Ti, Cr, Mn and Zn) have low concentrations without systematic variation.

Sapphirine

Fe^{3+} content was estimated at 17-34% of total molar Fe (Table 5.3). Along the RUM12 traverse, Fe^{3+} content is variable but slightly higher in zone A₁ than in zone A₀ (Fig. 5.6c). $X_{\text{Fe}^{2+}}$ (= calculated $\text{Fe}^{2+}/\text{Fe}^{2+}+\text{Mg}$) is similar in zones A₀ and A₁ (around 0.130-0.155), but is slightly higher in zones A₂ and A₃ (around 0.152-0.165). Al content is variable, with no systematic trend. The X_{Fe} of analyses averaged in zones increases from zones A₀ to A₃ in sample RUM12, and decreases slightly from zones A₂ to A₃ in samples RUM29 and RUM31 (Fig. 5.5). Minor elements (Mn, Ti, Zn) do not show any systematic variation in concentration across the zones.

Other Minerals

Cordierite has an X_{Fe} of approximately 0.11. There are no detectable differences in composition between cordierite from zone A₂ and zone A₃, either in the RUM12 traverse or in the analyses averaged by zone (Fig. 5.5). The X_{Fe} of phlogopite varies from approximately 0.08 in the A₀ and A₁ zones of samples RUM11 and RUM12, to approximately 0.11 in the A₂ and A₃ zones of samples RUM29 and RUM31.

Variations within and between zones

Traverses across corona textures in various samples were undertaken to determine if there was any variation in composition within individual mineral grains. No systematic trends were observed within mineral grains, and the degree of variation was negligible compared to that between different grains in the same zone.

Mineral compositions were also compared between equivalent zones in the reaction bands of different samples. In the A₃ zone, the averaged X_{Fe} for orthopyroxene, sapphirine and cordierite is slightly higher in the podiform samples RUM29 and RUM31, than in RUM12, which contains a planar reaction band. The X_{Fe} of bulk

rock analyses of the same A₃ zones is around 0.20 in all three samples. This reflects the higher modal proportion of cordierite in the A₃ zones of the podiform samples.

5.5 Development of the reaction bands

5.5.1 Evidence for diffusion metasomatism

The reaction assemblages developed between the aluminous gneiss and the quartz pegmatite possess several features which are characteristic of diffusion metasomatism:

1. The chemical composition of the reaction band is intermediate to the compositions of the reactants (Fig. 5.4). Furthermore, the composition of each zone within the reaction band lies between that of its neighbours, especially with respect to Si. Each zone can therefore be considered as the product of non-reversible reactions, that served to reduce the difference in chemical potential between adjacent zones (Korzhinskii, 1970, p. 115; Brady, 1977). Similarly, the entire reaction band was formed to reduce the chemical potential difference between the aluminous gneiss and the quartz pegmatite.

2. The reaction bands are only a few centimetres in width. Diffusion on such a limited scale over large periods of time ($>10^6$ years) is consistent with the measured and inferred diffusivities of cations in silicate mediums (Fletcher & Hofmann, 1974; Joesten, 1991).

3. The assemblages are nearly volatile-free, containing no carbonates and only minor amounts of hydrous minerals, such as biotite and cordierite. It is therefore unlikely that fluid infiltration was involved in mass transfer. This is supported by the limited extent of metasomatism, and by the presence of monotonic composition gradients across the reaction bands (Korzhinskii, 1970, p. 115).

The granoblastic texture of the unreacted aluminous gneiss was probably formed under granulite-grade conditions, which were present during D₁ deformation in the Framnes Mountains (Clarke & Norman, 1993). The quartz pegmatite lies sub-parallel to the S₁ gneissosity, and therefore was intruded before or during D₁. The reaction band must have formed during or after peak metamorphism, at temperatures sufficiently high to maintain large flux values for the diffusing components (Joesten, 1983).

5.5.2 A simple model for the growth of the reaction bands

The composition of the reaction band can be modelled in terms of the major components present, *i.e.* Si - Fe - Mg - Al - O (SFMA). The use of this model system presumes that the formation of plagioclase in zone A₄ is not strongly dependant on Na and Ca availability, *i.e.* the chemical potential of Na (μ_{Na}) does not have a controlling affect on zone formation. Initially, solid solution variation of Fe and Mg within minerals is also ignored, as it should not significantly affect the morphology of the predicted reaction band (Brady, 1977).

In the typical reaction band, Si is the only component which varies monotonically in concentration from the aluminous reactant (A₀) to the quartz pegmatite (Q). Assuming that mineral reactions took place in a closed system, the reaction band was formed in order to reduce the difference in the chemical potential of Si between the reactants. Zones A₁ to A₄ were formed from the A₀ reactant, by the exchange of Fe and Mg for Si. Al and O concentrations remained nearly constant in these zones. This exchange was balanced by that in zone A₅, which was formed from the quartz pegmatite by the exchange of Si for Fe, Mg and some Al. The reaction band grew as Si diffused into the aluminous gneiss while Fe and Mg diffused into the quartz pegmatite. Fe was exchanged for Si to a greater degree than Mg, so that X_{Fe} is lower in zones A₁ to A₃ and higher in zone A₅ than X_{Fe} in the A₀ reactant.

The growth of the reaction band can be modelled by treating Si as the only diffusing component (Brady, 1977). Reactions proceeded on either side of the initial boundary between the reactants, and a system of steady-state diffusion was generated that involved a continuous monotonic gradient in μ_{Si} across the reaction band. This system produced a series of bimineralic zones, that correspond to the divariant fields in a $\mu_{\text{Si}}\text{-X}_{\text{Al}}$ section (Fig. 5.7). Zones A_1 , A_3 and A_4 were formed as μ_{Si} was increased in the A_0 (Opx + Spl) reactant. The A_5 zone grew as μ_{Si} was reduced in the Q (Qtz) reactant. The growth of each zone was controlled by exchange reactions occurring at one or both of the zone's boundaries (Joesten, 1977). The system would have produced the following sequence of zones:

Qtz [Opx § Opx-Pl | Opx-Crd | Opx-Spr] Opx-Spl

where § marks the position of the initial reactant boundary. The zones in the real reaction bands that correspond to the bimineralic zones predicted by this model are:

Q [A_5 § A_4 | A_3 | A_1] A_0

The same sequence is predicted by considering Fe + Mg as a single diffusing component in the SFMA system, as this is complimentary to Si diffusion if Al is fixed (*i.e.* Fe + Mg is exchanged for Si).

The relative control of each diffusing component (Si, Fe and Mg) over the zone sequence can also be considered in the chemical potential space $\mu_{\text{Si}}\text{-}\mu_{\text{Fe}}\text{-}\mu_{\text{Mg}}$ (Fig. 5.8). Single-phase volumes in this chemical potential space are bound by bimineralic planes. Most of these planes are oriented at a low angle to the $\mu_{\text{Fe}}\text{-}\mu_{\text{Mg}}$ axial plane. Therefore bimineralic assemblages are more sensitive to changes in μ_{Si} than μ_{Fe} or μ_{Mg} . Since μ_{Si} controls diffusive behaviour, the path generated between the two reactants (from Qtz to Opx + Spl) will proceed along bimineralic planes, as in the case of single component diffusion (Brady, 1977). From the A_0 reactant (Opx + Spl) to zone A_4 in the reaction band (Opx + Pl), μ_{Si} increases along the path with the

Fig. 5.7. μ_{Si} -XAl section for the system Si-Fe-Mg-Al, constructed from Fig. 5.5 (after Brady, 1977). The approximate positions of the two reactants (A_0 and Q) in μ -X space are shown. Arrows show the changes in the μ_{Si} of the reactants that accompanied the growth of a reaction band. Bimineralic fields that correspond to observed zones are labelled accordingly. The dotted line represents the value of μ_{Si} at the original boundary between the reactants.

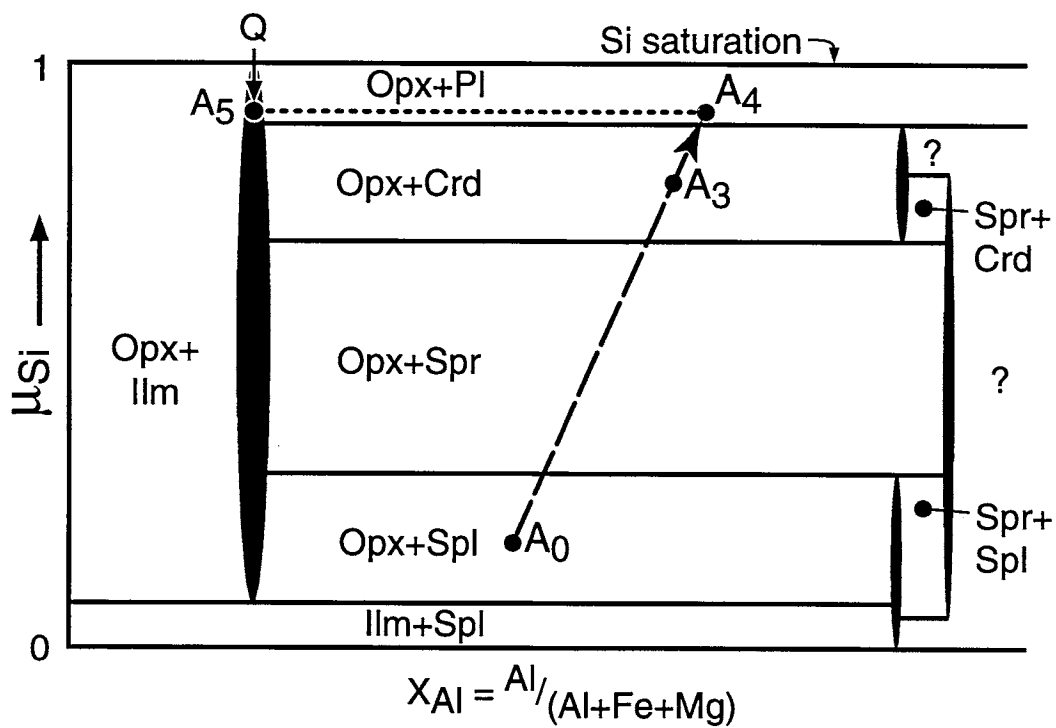
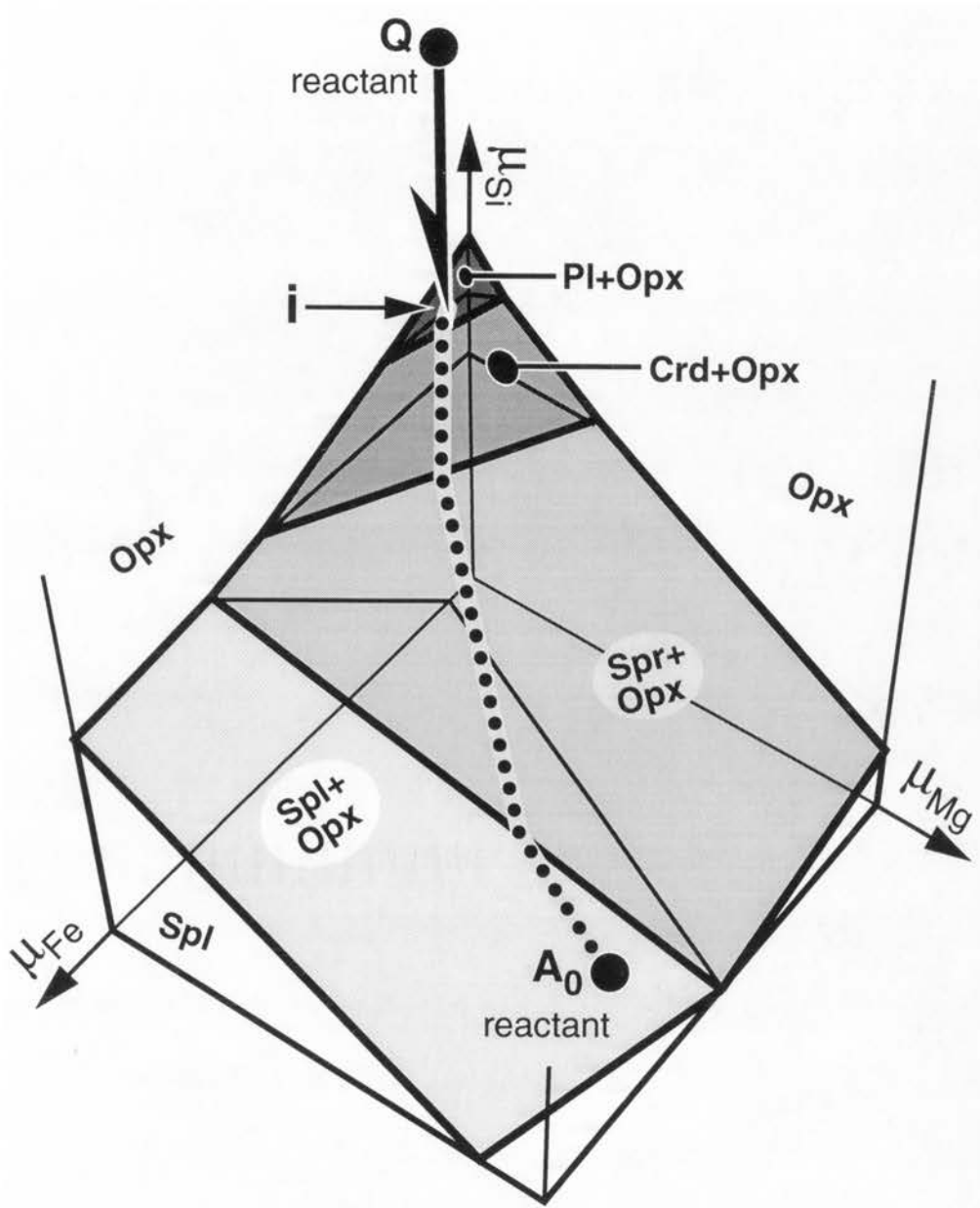


Fig. 5.8. Chemical potential diagram for the system Si-Fe-Mg-Al, showing the relationships between μ_{Si} , μ_{Fe} and μ_{Mg} . Phase field boundaries were constructed by the method of Korzhinskii (1959, p. 92), using the averaged compositions of minerals from zones A2 and A4 in sample RUM29. The approximate positions of the reactants (A0 and Q) are shown. The dotted path represents the shift in chemical potential from zone A0 to zone A4. Along this path, an increase in μ_{Si} involves a reduction in μ_{Fe} and μ_{Mg} . The arrow marks the shift in chemical potential from the quartz pegmatite (Q) to zone A5. Point 'i' represents $\mu_{\text{Si}}-\mu_{\text{Fe}}-\mu_{\text{Mg}}$ values at the initial boundary between the reactants.



decreasing μ_{Fe} and μ_{Mg} . From the quartz pegmatite (Q) to zone A₅, μ_{Si} decreases from unity along a gradient path through the trivariant Opx field. The point where the two gradient paths meet ('i' on Fig. 5.8) represents the values of μ_{Si} , μ_{Fe} and μ_{Mg} at the initial boundary between the two reactants.

5.5.3 Modifications to the model

The structure and composition of the typical reaction band is broadly consistent with the metasomatic model constructed above, but there are significant disparities.

1. Most importantly, zones in the real reaction bands are not bimineralic, but may contain up to four phases. In 3-dimensional chemical potential space (Fig. 5.8), four-phase assemblages are invariant. Four-phase zones should not be present, because they do not maintain a continuous chemical potential gradient across the reaction band.

Zone A₂, which contains spinel, sapphirine, cordierite and orthopyroxene, is not predicted by the model. The zone can be accounted for, however, if it is treated as a three-phase assemblage (Opx + Spr + Crd). This assemblage is univariant in Fig. 5.8. If μ_{Si} increased along a path that followed this invariant, then the predicted sequence would be:

Qtz [Opx § Opx-Pl | Opx-Crd | Opx-Crd-Spr | Opx-Spr] Opx-Spl

and the zone sequence that corresponds to the predicted sequence would be:

Q [A₅ § A₄ | A₃ | A₂ | A₁] A₀

2. The zones preserve clear evidence of textural disequilibrium, in the form of reaction coronas around individual grains (sapphirine around spinel, cordierite around sapphirine). To have maintained the continuous chemical gradients that drive diffusion, the diffusing components must have been in a state of local equilibrium at

each point along the reaction band (Thompson, 1959; Joesten, 1977). The preservation of reaction textures in several zones shows that this was not the case.

Despite the presence of four-mineral assemblages and reaction textures, the zoned structure of the reaction band is a typical product of diffusion metasomatism. These disparities may be resolved by a consideration of intergranular *versus* intragranular mediums of diffusion. Diffusion of mobile components through a polycrystalline substance will proceed more freely along grain boundaries than through the constituent grains (Vernon, 1976, p. 81; Brady, 1983). If local equilibrium is maintained for diffusing components within an 'exchange medium' that occupies intergranular space, steady state diffusion over large distances can still occur (Joesten, 1977). In the Rumdoodle reaction bands, diffusion along intergranular pathways is indicated by the growth of sapphirine in the A_0 reactant in a continuous network of patches. These patches penetrate deeply into the A_0 reactant, as the formation of sapphirine from spinel requires only a small increase in Si (Fig. 5.4a).

The growth of coronas of sapphirine around spinel grains and of cordierite around sapphirine grains involved the exchange of components between the reacting crystals at their grain boundaries. The complete reaction of spinel in zone A_1 and sapphirine in zone A_3 may have been prevented by two factors: a) a slow rate of reaction and growth at the grain interface (Christian, 1975), and/or b) the isolation of the spinel or sapphirine reactant from diffusing components by the formation of coronas. There is textural evidence for the latter case, since in most cases the sapphirine and cordierite coronas consist of single grains which completely enclose the core mineral. The formation of sapphirine from spinel and of cordierite from sapphirine involved the exchange of Si for Al as well as for Fe and Mg. The low diffusivity of Al away from the reaction site, especially through a solid mineral corona, would have caused the reaction rate to decrease rapidly as the corona widened.

3. If all reactions had occurred at the boundaries between zones, then each zone should have a uniform composition (Brady, 1977). The presence of modal gradients, especially in zones A₁ and A₂, indicates that growth also occurred within these zones during metasomatism (Frantz & Mao, 1979). These gradients occurred because the degree of replacement of reactants by products is a function of time (Fletcher & Hofmann, 1974). For example, zone A₁ grew by replacing the A₀ reactant along an advancing front, and was simultaneously consumed by the advancing front of zone A₂. The mineral assemblage near the trailing (high μ_{Si}) edge of zone A₁ had existed longer than the assemblage near the leading (low μ_{Si}) edge. Since the amount of sapphirine grown in zone A₁ was dependant on the time taken for the reaction, the assemblage on the high μ_{Si} side of the zone should contain more sapphirine (*i.e.* is more siliceous) than that on the low μ_{Si} side.

4. Since Fe was exchanged for Si at a higher rate than Mg, there is a decreasing trend in X_{Fe} from zones A₀ to A₃. A contrary trend, however, is seen between each of the constituent minerals of the zones. In RUM12, X_{Fe} in spinel, orthopyroxene and sapphirine increases from zones A₀ to A₃ (Figs 5.5 and 5.6). In samples RUM29 and RUM31, X_{Fe} decreases in orthopyroxene and sapphirine from zones A₂ to A₃, but to a smaller degree than the bulk rock X_{Fe} . These trends can be understood by considering the diffusive behaviour of Fe and Mg in the mineral assemblages. Fe and Mg have constant concentrations within individual mineral grains and coronas, indicating that these elements had rates of diffusion high enough to re-equilibrate between minerals in each assemblage. Equilibration between Fe and Mg may have occurred during or after metasomatism. If the former is the case, the exchange of Mg for Fe in orthopyroxene and spinel may have occurred to assist the appearance of more Mg-rich phases in the metasomatised assemblage. For example, consider the shift in the Opx-Spr tie-line between zones A₀ and A₃ on the AFM diagram (Fig. 5.4b). The growth of zone A₂ in the reaction band involved the growth of cordierite, as a result of the bulk rock composition having crossed the Opx-Spr tie-line from zone

A₁ to zone A₂. The growth of zone A₂ also involved an increase in the X_{Fe} of orthopyroxene and sapphirine, which shifted the Opx-Spr tie-line towards the FeO corner of the AFM diagram (Fig. 5.4b). This shift reduced the amount of Fe and Mg that needed to be removed for the bulk rock composition to cross the Opx-Spr tie-line. Solid solution exchange within minerals has therefore assisted in the growth of the reaction band.

The alternative possibility is that re-equilibration of Fe and Mg between minerals has occurred after metasomatism. A period of near-isobaric cooling has been inferred for the post-peak metamorphic history of the Painted Gneiss (Clarke *et al.*, 1989). In the A₂ and A₃ zones, cooling after metasomatism may have shifted cordierite to more magnesian compositions, and other minerals to more Fe-rich compositions.

5. The model assumes that the diffusivities of Na and Ca do not control the formation of the orthopyroxene-plagioclase (A₄) zone. These elements may have been derived from antiperthite grains in the quartz pegmatite. The poor development or total absence of the A₄ zone in many samples may be attributed to a lack of Na and Ca availability in the metasomatic system.

6. The presence of hydrous phases (phlogopite, chlorite and pinite) is inconsistent with the anhydrous metasomatic model postulated above. These minerals are considered to have been produced by slight rehydration of the gneiss at temperatures lower than those present during metasomatism. Phlogopite may have been present during metasomatism, but it has an even modal distribution throughout the reaction bands, and has no effect on composition trends. The interstitial habit of phlogopite grains is consistent with later growth. The growth of pinite and chlorite is restricted to lower temperatures (<400°C; Deer *et al.*, 1992, p. 340) than those under which diffusion metasomatism is likely to have occurred.

5.6 Conclusion

The textures and mineralogy of the reaction bands developed between aluminous gneiss and quartz pegmatite are consistent with a modified model of diffusion metasomatism under closed system, steady state conditions. Reaction bands were developed as a result of the diffusive exchange of Si for Fe and Mg, which served to decrease the difference in chemical potential between the two reactants. The preservation of corona textures is due to the limited diffusive exchange of Si and Al across mineral grains. Despite complexities in the structure of the reaction bands, their broad morphology could be successfully modelled in terms of a single diffusing component (Si).

Disparities between the model and real reaction bands can be partially attributed to the nature of diffusive transport in polycrystalline materials. Rates of diffusion are much greater along grains boundaries than through the constituent grains of a rock. This is particularly true for rocks composed of silicate minerals, as the diffusivities of Si and Al in these minerals are low. Diffusive exchange can still occur on a much larger scale, however, along a continuous network of intergranular pathways. Local equilibrium was maintained between minerals immediately adjacent to grain boundaries. The preservation of corona textures within the zones did not prevent the larger scale growth of the reaction band.

Chapter 6

Synthesis

6. Synthesis

Introduction

The new petrological and geochronological work presented in this thesis on the Cape Bruce granulites forms a small part of a series of detailed studies on the Rayner Complex, that have been produced over the past 15 years. The new work needs to be placed in context with previous work, so that an understanding of the large scale processes that affected rocks in the region can be approached. In this chapter, comparisons are made between the geochronological and structural histories of various provinces in MacRobertson Land. These comparisons are drawn upon to propose some simple tectonic models for the evolution of the Rayner Complex in MacRobertson and Kemp Land. Some possible mechanisms for the production of high-grade gneisses in the region are also considered.

6.1. Geochronological constraints on the history of the Rayner Complex

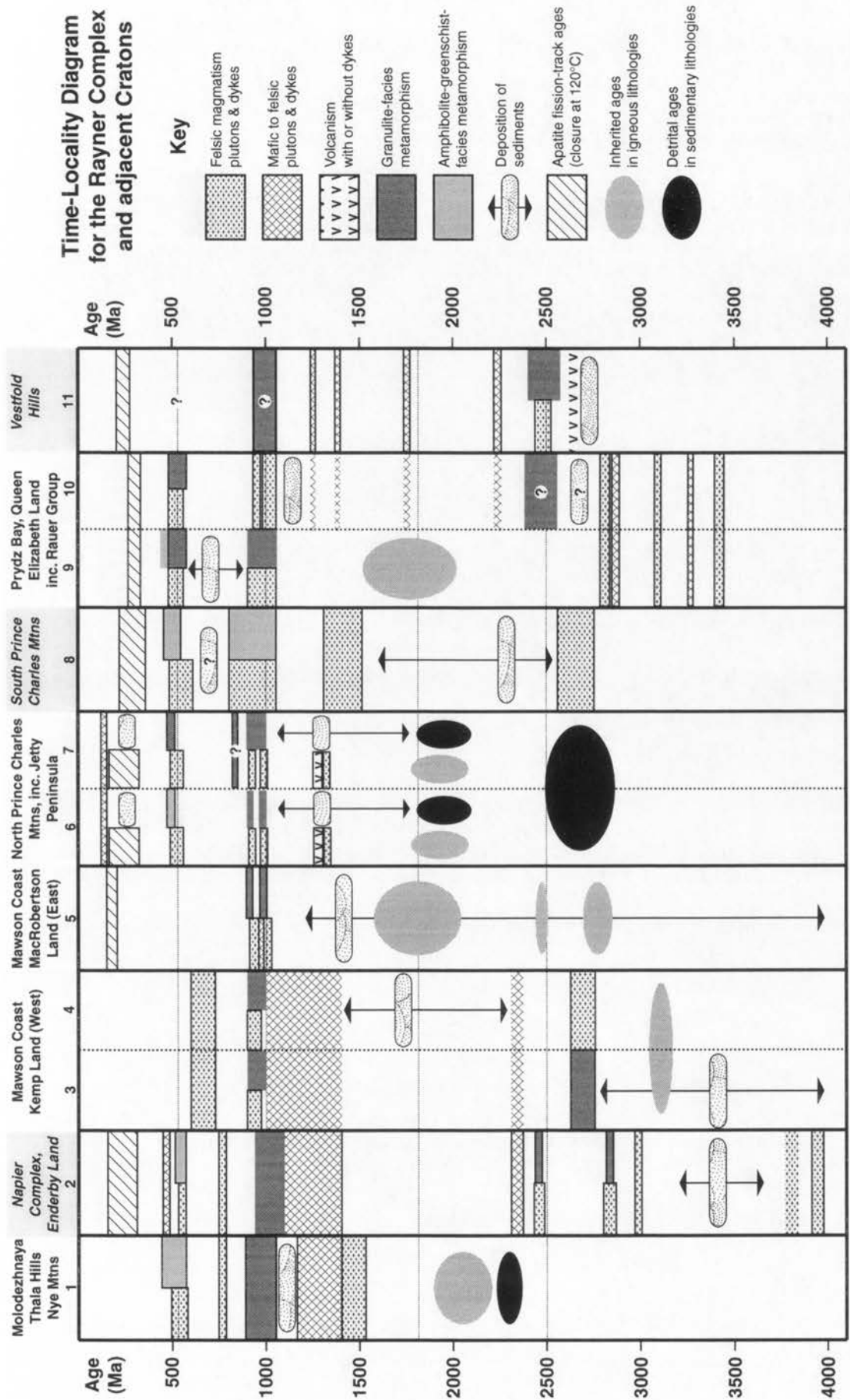
The geochronology of the Mawson Coast can now be placed in context with the development of the Rayner Complex. To do this, comparisons need to be made with the geochronological work detailed in Chapters 1 and 4. Geochronologically constrained events in various provinces have been schematised in a Time *vs.* Locality diagram (Fig. 6.1). The correlation on the regional scale of particular lithologies, such as igneous intrusions, is not the intention of this diagram; rather, it is intended to demonstrate the regional extent of processes during which different lithologies were formed.

Fig. 6.1. Time *versus* locality diagram for the Rayner Complex and adjacent Archaean terranes (shaded). The eight provinces compared are shown in Fig. 1.3 (Chapter 1). Provinces with alternate interpretations are represented by two columns separated by a dotted line. The height of an 'event block' indicates the amount of error on the known age of the event, rather than the duration of any given event. Events that have been dated with measured errors are shown as blocks with black borders; ages that are approximate, or that are only constrained by their relation to other dated events, have blocks without borders. Question marks indicate uncertain interpretations.

References for age data and geological interpretations of each province:

- 1: Sheraton *et al.* (1987), Black *et al.* (1987), Sheraton & Black (1981).
- 2: Harley & Black (1997), Sheraton & Black (1981), Black *et al.* (1984), Sheraton *et al.* (1987).
- 3: Clarke (1987), Grew *et al.* (1988), Sheraton & Black (1981).
- 4: this study, Clarke (1987), Grew *et al.* (1988).
- 5: this study, White & Clarke (1993), Young & Black (1991), Black *et al.* (1987).
- 6: Kinny *et al.* (1997), Hand *et al.* (1994), Manton *et al.* (1992), Beliatsky *et al.* (1994), Fitzsimons & Thost (1991), Tingey (1991).
- 7: Kinny *et al.* (1997), Hensen *et al.* (1997), Zhou & Hensen (1995), Carson *et al.* (1996), Manton *et al.* (1992), Beliatsky *et al.* (1994), Fitzsimons & Thost (1992), Tingey (1991).
- 8: Tingey (1991), Grikurov & Soloviev (1974), Halpern & Grikurov (1975), Carson *et al.* (1996).
- 9: Hensen & Zhou (1995), Kinny *et al.* (1993), Zhao *et al.* (1992), Zhao *et al.* (1995), Sheraton *et al.* (1984).
- 10: Carson *et al.* (1996), Kinny *et al.* (1993), Sims *et al.* (1994), Zhao *et al.* (1992).
- 11: Snape & Harley (1996), Lanyon *et al.* (1993), Black *et al.* (1991), Dirks *et al.* (1994).

Time-Locality Diagram for the Rayner Complex and adjacent Cratons



6.1.1 Relationships between Archaean and Proterozoic crust

Parts of the Rayner Complex have gradational boundaries with Archaean cratons. Outcrops in the Edward VIII Gulf, Kemp Land, have been interpreted as a section of the Archaean Napier Complex, that was reworked by the *c.* 1000 Ma Rayner Structural Episode (Sheraton *et al.*, 1987). Lithologies with compositional and structural similarities to the Edward VIII Gulf gneisses extend eastwards to the Stillwell Hills (Trail *et al.*, 1967; N. Kelly & C. Carson, *pers. comm.*, 1997). The *c.* 2700 Ma age of metamorphism in the Stillwell Hills, as well as ages for inherited zircon from Cape Bruce and the Framnes Mountains, fall within the *c.* 3100-2450 Ma window (Fig. 6.1) of high-grade metamorphism, magmatism and tectonism that occurred in the Napier Complex (Sheraton *et al.*, 1987). Therefore, although there are no direct correlations between the Archaean ages of particular events (Fig. 6.1), it is likely that Archaean lithologies that outcrop west of the Stillwell Hills represent an eastward extension of the Napier Complex.

The gradational boundary between the Napier and Rayner Complexes in Kemp Land contrasts sharply with that in the vicinity of the Rayner Glacier in Enderby Land (Fig. 1.3), where there is no evidence for crust in the Rayner Complex older than *c.* 2200 Ma (Black *et al.*, 1987). This boundary represents a tectonic juxtaposition of the two terranes caused by thrusting during the Rayner Structural Episode, which may also have been reworked during the *c.* 500 Ma event (Harley, 1989). The correlation of 1190 ± 200 Ma dykes between the Rayner and Napier Complexes in Enderby Land confirms that the two terranes were adjacent prior to the Rayner Structural Episode (Fig. 6.1).

The Rayner Complex in Enderby Land had a pre-1200 Ma evolution distinct to that of the complex further east. Evidence for a major episode of felsic magmatism at *c.* 1450 Ma in Enderby Land has not been recognised elsewhere, and there is no conclusive evidence for the *c.* 1000 Ma magmatism that occurred elsewhere in the

Rayner Complex. Thus the relationship of Proterozoic crust in Enderby Land to the rest of the Rayner Complex before the Rayner Structural Episode is uncertain.

In the Prince Charles Mountains, the boundary between Archaean and Proterozoic crust is fairly well established (see Fig. 1.3 in Chapter 1). There is no Archaean inheritance in the rocks of the northern Prince Charles Mountains, with the oldest inherited zircon dated at *c.* 1900 Ma (Kinny *et al.*, 1997). The paragneiss dated by Kinny *et al.* (1997) contains detrital zircon with ages in the ranges 2800-2500 Ma and 2100-1800 Ma. The former range of data corresponds to age estimates for felsic granulites in the southern Prince Charles Mountains (Fig. 6.1), and the latter probably derives from the erosion of Paleoproterozoic crust in the northern Prince Charles Mountains. This provides evidence that the northern and southern terranes were adjacent during erosion and sedimentation, at some time between *c.* 1800 and *c.* 1000 Ma.

6.1.2 Pre-metamorphic age correlations and the timing of sedimentation

Age estimates of *c.* 2700 Ma (Fig. 6.1) have been obtained from felsic orthogneisses in the Stillwell Hills (Clarke, 1987) and the Southern Prince Charles Mountains (Tingey, 1991). These ages correspond to dates obtained from inherited zircon at Cape Bruce (this study) and detrital zircon from the northern Prince Charles Mountains (Kinny *et al.*, 1997). It is therefore possible that all these terranes are based on a single block of Archaean crust. However, the data are spread over large area, and more data are required before this possibility can be fairly assessed.

Although data for the evolution of the Rayner Complex in the Paleoproterozoic and Mesoproterozoic are scant, correlations are possible between several provinces. Model

(T_{CHUR}) ages for Sm-Nd correspond at *c.* 2000-1800 Ma for felsic orthogneisses from Enderby Land and Mawson Station (Black *et al.*, 1987). Inherited and detrital zircon with ages between and *c.* 2000 Ma and *c.* 1700 Ma have been obtained from gneisses at Mawson Station (Young & Black, 1991) and the northern Prince Charles Mountains (Kinny *et al.*, 1997). Due to their close proximity, the Framnes Mountains and the northern Prince Charles Mountains are the most likely provinces to have a common Paleoproterozoic history.

There is no geochronological evidence for the formation of Paleoproterozoic crust in the Stillwell Hills (Clarke, 1987) or Cape Bruce (this study). This fact, along with the evidence for the involvement of Archaean crust in these localities (as opposed to the northern Prince Charles Mountains), may indicate that this part of the Mawson Coast was unrelated to crust further east before the Mesoproterozoic. There is also a lack of evidence for Paleoproterozoic crust in the Prydz Bay province.

There is some evidence for sedimentation having occurred in parts of the Rayner Complex during the Mesoproterozoic (Fig. 6.1). All of these sedimentary sequences, with the possible exception of those at Fisher Massif in the northern Prince Charles Mountains, were metamorphosed during the Rayner Structural Episode. In Enderby Land, sediments post-date the emplacement of *c.* 1200 Ma mafic dykes (Sheraton *et al.*, 1987). The minimum U-Pb ages of detrital zircon constrains the age of sedimentation to less than *c.* 1800 Ma for the northern Prince Charles Mountains (Kinny *et al.*, 1997). Volcanism occurred at Fisher Massif at *c.* 1300 Ma (Beliatsky *et al.*, 1994; Kinny *et al.*, 1997), and may have been contemporaneous with sedimentation in the province. Metasediments in the Mawson Coast, Framnes Mountains and the Prydz Bay region are not well constrained, but may be equivalent with Mesoproterozoic sediments in nearby provinces. Further geochronology is required to test this hypothesis.

6.1.3 Extent and duration of the Rayner Structural Episode

Granulite facies metamorphism, deformation and magmatism have been constrained to the *c.* 1000 Ma Rayner Structural Episode in all parts of the Rayner Complex, with the possible exception of the Prydz Bay province (Fig. 6.1). Activity of equivalent age has also been dated, albeit with low precision and at lower metamorphic grade, in the Napier Complex and the southern Prince Charles Mountains (Grew *et al.*, 1982; Halpern & Grikurov, 1975; Tingey 1991).

The effects of *c.* 500 Ma tectonism have obscured the *c.* 1000 Ma history of the Prydz Bay region. The formation of most of the gneisses in the province has now been controversially attributed to *c.* 500 Ma metamorphism (Fig. 6.1). However, there is consensus on the occurrence of magmatism at *c.* 1000 Ma (*e.g.* Kinny *et al.*, 1993; Hensen & Zhou, 1995; Carson *et al.*, 1996). Hensen & Zhou (1995) have suggested that *c.* 1000 Ma metamorphism did occur, but that it only affected mafic and felsic orthogneisses which have been interpreted as forming a basement to supercrustal lithologies (*e.g.* Fitzsimons & Harley, 1991; Carson *et al.*, 1995). Other authors suggest that activity at *c.* 1000 Ma was restricted to felsic magmatism (*e.g.* Carson *et al.*, 1996). In either case, *c.* 1000 Ma ages from the Prydz Bay province provide evidence that the area was a part of the Rayner Complex, perhaps as the periphery of a large continental block which was extensively reworked by *c.* 500 Ma tectonism.

Although there is some variance in age estimates of the Rayner Structural Episode between different provinces, the data constrains the event to a period of *c.* 1030-900 Ma. Individual events during this period have been resolved, and in some cases can be correlated between provinces. Dating in the northern Prince Charles Mountains (Kinny *et al.*, 1997) and Jetty Peninsula (Manton *et al.*, 1992) shows that, like the Cape Bruce area, magmatism occurred in the early (*c.* 1000 Ma) and later (*c.* 980-900 Ma) stages of the Rayner Structural Episode. The *c.* 980 Ma syn-tectonic charnockites in the northern Prince Charles Mountains correspond in age, composition and

tectonic framework to the Mawson Charnockite, and were probably generated by the same processes (Kinny *et al.*, 1997). Both were emplaced after peak metamorphism (White & Clarke, 1993; Fitzsimons & Thost, 1992). Felsic gneisses and pegmatites that immediately post-date peak metamorphism have been dated from the Stillwell Hills (Grew *et al.*, 1988) and Cape Bruce (this study). For the Mawson Coast and the Prince Charles Mountains, peak metamorphism and the early stages of deformation are restricted to the period *c.* 1030-950 Ma.

The new dating at Cape Bruce provides strong evidence that deformation and metamorphism continued shortly after the emplacement of *c.* 980-920 Ma intrusions. Post-peak deformation and granulite facies metamorphism in the northern Prince Charles Mountains, although poorly constrained in age, have a similar structural orientation and metamorphic grade to the *c.* 910 Ma D₄/M₃ event at Cape Bruce. Hand *et al.* (1994a) suggested that decompressive textures observed in the granulites of Jetty Peninsula were generated by a thermal pulse in the late stages of the Rayner Structural Episode, a model which corresponds to that provided by this study for M₃ metamorphism at Cape Bruce (see below). However, decompressive reaction textures observed at Jetty Peninsula and Depot Peak (Hand *et al.*, 1994a; Stüwe & Hand, 1992) are not present in any gneisses on the Mawson Coast. In the Prydz Bay province, decompression textures have been interpreted as the products of *c.* 500 Ma metamorphism (Hensen & Zhou, 1995; Fitzsimons, 1996; Carson *et al.*, 1996). Carson *et al.* (1996) re-interpreted the textures described by Hand *et al.* (1994a) as having been produced by a *c.* 500 Ma overprint of *c.* 1000 Ma granulites. Due to differences in the metamorphic evolution of rocks from the Mawson Coast and Jetty Peninsula, this author supports the interpretation of Carson *et al.* (1996).

Data from other areas support the evidence from Cape Bruce that high-grade metamorphism occurred in the latter stages of the Rayner Structural Episode. Syn-tectonic pegmatites containing garnet from the Stillwell Hills (Grew *et al.*, 1988) and an anatectic leucogneiss from the Jetty Peninsula (Manton *et al.*, 1992), crystallised at

940 $\pm$ 80 Ma and 940 $\pm$ 24 Ma, respectively. Zircons from gneisses in the Nye Mountains (Black *et al.*, 1987) and Mawson Base (Young & Black, 1991), that have an inferred metamorphic origin, grew in the range *c.* 940-902 Ma. Rb-Sr whole rock data from Mawson Station (Black *et al.*, 1987) and the Framnes Mountains (Young & Black, 1991) indicate that the isotopic system closed in these gneisses in the period *c.* 934-838 Ma. Therefore, for the Mawson Coast province at least, the prolonged nature of metamorphism during the Rayner Structural Episode is well established. However, the new data from Cape Bruce provides the only geochronological evidence that this metamorphism occurred as two or more discrete events. Further precise work is required to assess the significance of the data.

6.1.4 After the Rayner Structural Episode

There are only few data available to constrain the timing of geological activity between the *c.* 1000 Ma and the *c.* 500 Ma events (Fig. 6.1). A notable exception occurs in the northern Prince Charles Mountains, where Hensen *et al.* (1998) obtained garnet-whole rock Sm-Nd pseudo-isochrons with ages of 830-786 Ma and 625-541 Ma. Hensen *et al.* (1998) interpret these ages as those of garnet growth, and consequently of the formation of the granulites dated. No garnet-whole rock pseudo-isochrons in the study yielded *c.* 1000 Ma ages, a fact that Hensen *et al.* (1998) used to infer that high-pressure metamorphism occurred at *c.* 800 Ma, with a lower pressure event having occurred at *c.* 1000 Ma. However, Mezger *et al.* (1992) have suggested that in slowly cooled terranes, only garnets with radii larger than 5cm are likely to record Sm-Nd ages that were set before the closure of Sm-Nd diffusion, at around 600°C. As the *c.* 800 Ma garnets used by Hensen *et al.* (1998) were less than 2cm in diameter, it is possible that the Sm-Nd isotopic system was reset by diffusion at *c.* 800 Ma. However, to produce resetting, a period of heating above 600°C must still have occurred just before *c.* 800 Ma.

The *c.* 500 Ma tectonic episode is expressed in most parts of the Rayner Complex as discrete shear zones and felsic pegmatite swarms (Harley & Hensen, 1990). East-trending mylonitic shear zones that occur along the Mawson Coast (D₅, this study) may correlate with similar E-trending structures that have been related to *c.* 500 Ma thrusting in Enderby Land (Sheraton *et al.*, 1987; Clarke, 1988). Pressures estimated from syn-D₅ mineral growth in the Colbeck Archipelago (White & Clarke, 1994) are comparable to those present during the Rayner Structural Episode. There is no evidence along the Mawson Coast for exhumation or reburial during tectonism after the Rayner Structural Episode. In the northern Prince Charles Mountains, there is also a lack of evidence for decompression and uplift after the Rayner Structural Episode (Fitzsimons & Thost, 1992; Carson *et al.*, 1996).

In contrast, the Jetty Peninsula and provinces east of the Lambert Glacier (Fig. 1.3) preserve evidence for a major episode of tectonism, magmatism and amphibolite to granulite facies metamorphism at *c.* 500 Ma (Fig. 6.1; Manton *et al.*, 1992; Hensen & Zhou, 1995; Carson *et al.*, 1996; Fitzsimons, 1996). In the Prydz Bay province, this has been established as the main tectonothermal event that produced granulite facies gneisses in the area. The *c.* 500 Ma event extensively reworked *c.* 1000 Ma gneisses in the province, as well as introducing new intrusive material. The Prydz Bay province has been interpreted as the result of continent-continent collision, which involved the reworking and/or amalgamation of the margin of the Rayner Complex with the Vestfold Hills (Snape & Harley, 1996) and the Southern Prince Charles Mountains (Carson *et al.*, 1996).

Tectonic activity at *c.* 500 Ma has also occurred in Enderby Land (Fig. 6.1), where the Rayner Complex lies adjacent to the *c.* 500 Ma gneisses of the Lutzow-Holm Complex (Harley & Hensen 1990; Shiraishi *et al.*, 1994). Parts of the Rayner Complex in this area have been recently re-interpreted as *c.* 500 Ma crust (Shiraishi *et al.*, 1994). As with the Prydz Bay province, continent-continent collision at *c.* 500 Ma is inferred. This collision involved the reworking of the margin of the Rayner

Complex (Harley & Hensen, 1990), that may have occurred as a result of the amalgamation of the complex with adjacent terranes (Shiraishi *et al.*, 1994).

6.2 Structural relationships in Kemp and MacRobertson Land

Structural correlations on a regional scale are difficult to make in the Rayner Complex, as outcrops are sparse and early tectonic fabrics have been re-oriented by later deformations. However, some fabrics with similar orientations and structural styles can be observed in several localities. Geochronology and metamorphic histories can be used to constrain contemporaneous events. There are strong similarities between the deformation histories of the Mawson Coast and outcrops to the south-east, including the Framnes Mountains, Depot Peak and the northern Prince Charles Mountains. Correlations are suggested here in order to develop a model for tectonism during the Rayner Structural Episode in MacRobertson Land. Possible structural correlations are shown in Table 6.1.

In all provinces between Kemp Land and the northern Prince Charles Mountains, the earliest tectonic fabrics are defined by a high-grade gneissosity that was interpreted to have formed during the peak of metamorphism (White & Clarke, 1993, Fitzsimons & Thost, 1992, Hand *et al.*, 1994b, this study). In the Cape Bruce, northern Prince Charles Mountains and Jetty Peninsula localities, geochronology has constrained the formation of these gneisses to about *c.* 1000-980 Ma (this study; Kinny *et al.*, 1997; Manton *et al.*, 1992). Therefore, it is possible that these high-strain fabrics were developed in the early stages of the Rayner Structural Episode (Table 6.1).

The correlation of these early fabrics between provinces is suspect, as the attitude of gneissic foliations and intrafolial folds are highly variable (Table 6.1). Sub-horizontal gneissic foliations, associated with reclined to recumbent isoclinal folds, have been observed in the Stillwell Hills (D₁-D₂; Clarke, 1988), Depot Peak (D₁; Stüwe &

Table 6.1 Structural correlations in Kemp Land & MacRobertson Land.

Fold Is.	Stillwell Hills	Cape Bruce	Colbeck Archipelago
(James et al., 1991; Grew et al., 1988)	(Clarke, 1988)	(this study)	(White & Clarke, 1993; White, 1993)
Metasediments, granitoids & charnockite	Metasediments, granitoids, c.2700Ma pre-syn-Dc charnockite		
D1-D2 S1 dominant gneissosity. Intrafolial F1. NW trending horizontal L1/F1 axes. minor isoclinal F2	Dc Sc dominant gneissosity, partial melting. Intrafolial Fc		
?c. 1200Ma mafic dykes,	?c. 1200Ma mafic dykes		
		Metasediments, mafic orthogneiss, ≥1000Ma pre-syn-D1 charnockite	Metasediments, mafic orthogneiss, charnockite
?	D1 S1 dominant foliation // Sc. N-NE trending recumbent isoclinal F1. W over E shear sense. Sub-horizontal W trending L1, normal to parallel with F1 axes	D1 S1 dominant gneissosity, partial melting. Intrafolial F1	D1 S1 dominant gneissosity, partial melting. Intrafolial F1
		c. 990Ma pre-syn-D2 biotite-pyroxene orthogneiss & intermediate to mafic dykes. c.980Ma granitoids	Biotite-pyroxene orthogneiss & intermediate to mafic dykes. Granitoids
?D3 minor isoclinal F3.		D2 S2 dominant foliation // S1, partial melting. S2 with moderate N-NW to vertical NW-SW dip. N and S plunging isoclinal F2. L2 lineation & stretching // F2 axes	D2 S2 dominant foliation, partial melting. Macro-mesoscopic isoclinal F2 with steep E dipping axial planes in areas of low D3 strain. Steep NE-E plunging F2 axes. L2//F2 axes
		c. 950Ma late-post-D2 granitoid dykes & pegmatites	
		D3 Open-tight F3 with vertical N-trending axial planes and N plunging axes. F3 coaxial with F2	
c. 940Ma syn-post D3 pegmatites	Pegmatites	c. 950Ma pegmatites	c. 980-950Ma Mawson Charnockite, pegmatites
D4-D5 macro-mesoscopic isoclinal F4 with moderate NE dip on axial plane. Controls NW trend of S1. Macroscopic tight F5 with NW trend & shallow SE plunging axis	D2 Low strain. Vertical W trending S2. Open-isoclinal upright F2 with horizontal W trending axes and stretching direction. Hinge area of tight macroscopic F2? c. 920 Ma metamorphism	D4 Macro-mesoscopic open F4 with vertical E-trending axial planes and shallow to steep W plunging axes. Controls trend of S2/S1.	D3 Macro-mesoscopic open-isoclinal F3 with steep S dipping axial planar S3. F3 axes plunge steeply E. F3 possibly coaxial with F2. Controls NE trend of S2/S1 in high D3-strain zone.
	Pegmatites		
	D3 Steep E dipping S3. Open-isoclinal upright F3 with sub-horizontal N trending axes. E plunging L3		D4 ?Weak vertical N-trending S4 in Mawson charnockite
	≥ 770Ma? pegmatites		
MPS shear zones with moderate NE dip & N plunging lineation	D5 MPS shear zones with S dip and S over N movement. SW-SE plunging L5	D5 MPS shear zones with moderate NE dips and S over N movement. NE plunging L5.	D5 MPS shear zones with steep N-NW dip and S over N movement. N plunging L5.

Table 6.1 MPS = mylonite-pseudotachylite, // = parallel to

Framnes Mountains (*Clarke 1988, Trail 1970, unpublished data)	Depot Peak (Stüwe & Hand 1991)	N Prince Charles Mtns (Fitzsimons & Thost 1991, Kinny et al. 1997)	Jetty Peninsula (Hand et al. 1994a,b, Manton et al., 1992)
Metasediments, mafic orthogneiss & dykes, granitoids	Metasediments, granitoids	Metasediments. c.1300Ma volcanics & diorite. Granitoids, mafic dykes. ≥1020Ma granitoid	Metasediments, mafic orthogneiss
= D1 S1 dominant gneissosity, partial melting. Intrafolial F1 = Pre-syn-D2 granitoids, mafic dykes = D2 (D1*) S2 dominant foliation, partial melting. Mostly moderate-steep E dip Tight-isoclinal E plunging F2. L2//F2 axes. <i>In areas of low D4 strain?</i>: moderate-steep S-SE dipping S2. F2 axes plunge SW // L2 Mafic dykes, pegmatites, ?granitoid = D3 <i>In areas of low D4 strain?</i>: Open F3 with SE-E dipping axial planes and S plunging axes. F3 coaxial with F2?	= D1 S1 sub-horizontal dominant gneissosity & foliation, partial melting. Recumbent F1. Shallow N plunging L1 stretching ? D2 macroscopic recumbent F2 with E? trending axis. S2 shear zones with shallow N dip & S over N movement. N plunging L2 in shear zones // L1	= D1-D2 S1 gneissosity & foliation. F2 folding. Preserved only in mafic pods. Possibly early D3 = D3-D4 S3 dominant gneissosity & foliation, sub-horizontal to shallow E dip areas of low D6 strain. Recumbent isoclinal F4. Syn-post D3-4 partial melting	= D1 S1 dominant gneissosity, partial melting. c. 1000Ma syn-D1 granitoids
c. 980-950Ma Mawson Charnockite.		c. 980Ma charnockite. Mafic dykes	
Gentle open warping? (Clarke, 1988)		= D5 upright E-trending F5 with shallow-moderate E plunging axes. L5//F5 axes = D6 Vertical E-trending S6 // S3 in high D6-strain zones. Macro-mesoscopic tight-isoclinal F6 in low D6-strain zones. Shallow E-plunging L6//F6 axes. Partial melting. Controls E trend of S3	= D2 S2 dominant foliation in high D2-strain zones. Vertical E trending S2 with S over N shear sense. Macro-mesoscopic tight-isoclinal F2 with steep E plunging axes //L2.. c. 940Ma syn-D2 granitoids, partial melting
Pre-syn-D4 mafic dykes?	pegmatites	Granitoid dykes & pegmatites	Pre-syn-D3 granitoid dykes & pegmatites
D4 (D3*) S4 dominant foliation in Mawson Charnockite. Vertical to steep E dipping S4. Minor open-isoclinal F4 with sub-horizontal N-trending axes. Controls N trend of S2/S1 in most areas.	< Not correlated >	D7 minor shear zones	D3 minor hear zones with vertical N trending S3. Partial melting. c.500Ma granitoid dykes
D5 (D4*) MPS shear zones with moderate to steep SW-SE dips and S over N movement. S-SW plunging L5	= ??? =	D8 MPS shear zones with sub-horizontal to steep N trending S8.	D4 MPS shear zones with vertical NE & NW trending orientation & NW over NE movement

Hand, 1992) and the northern Prince Charles Mountains (D₃-D₄; Fitzsimons & Thost, 1992). In both the Stillwell Hills and the northern Prince Charles Mountains, these shallowly dipping structures are found in the low-strain domains of later deformations (Trail *et al.*, 1967; Fitzsimons & Thost, 1992). In the Cape Bruce, Colbeck Archipelago, Framnes Mountains and Jetty Peninsula localities, the attitudes of gneissic fabrics are highly variable, with a predominance of moderately to steeply dipping foliations and axial planes (D₁-D₂; this study; D₂, White & Clarke 1993; D₁, Clarke 1988; D₁, Hand *et al.* 1994b). This variability can be partially attributed to later deformation in all of these localities; however, it cannot be determined if the fabrics had the same orientation as the low-angle structures elsewhere.

Evidence published for east over west thrusting during the Rayner Structural Episode on the Mawson Coast, as postulated by Clarke (1988), is present in the Stillwell Hills only (see Chapter 2). Elsewhere, the absence of clear kinematic indicators makes it impossible to determine the strain components of early Rayner Structural Episode deformation. However, in the Stillwell Hills (Clarke, 1988) and the Framnes Mountains (unpublished observations by the author), the dominant gneissosity has a northerly trend. In areas of low strain for subsequent structures, north-trending foliations and axial planes are seen in the Colbeck Archipelago (White & Clarke, 1993) and the northern Prince Charles Mountains (Fitzsimons & Thost, 1992). Therefore, it is likely that the early high-strain stages of the Rayner Structural Episode had a significant component of east-west compression.

The formation of gneissic fabrics was followed by the emplacement of large bodies of charnockite on the Mawson Coast and in the northern Prince Charles Mountains, and by the intrusion of felsic pegmatites west of the outcrop area of the Mawson Charnockite (Table. 6.1). Geochronology has constrained the timing of intrusion to a period of *c.* 990-920 Ma (Grew *et al.*, 1988; this study; Young & Black, 1991; Kinny *et al.*, 1997). Intrusive activity pre-dated further deformation in all areas between Kemp Land and the northern Prince Charles Mountains (Table 6.1). Upright east-

trending folds deform the early gneissosity in the Stillwell Hills (D₂, Clarke 1988), Cape Bruce (D₄, this study), and the Colbeck Archipelago (D₃, White & Clarke, 1993). In the northern Prince Charles Mountains and at Jetty Peninsula, high-strain shear zones with an east-trending sub-vertical fabric completely reorient the earlier gneissosity (D₅-D₆; Fitzsimons & Thost, 1992; D₂, Hand *et al.*, 1994b). Macroscopic folds and shear zones at Depot Peak are also east-trending (Stüwe & Hand, 1992), but unlike other fabrics have axial planes and foliations that dip shallowly to the north. At Fold Island, tight D₄-D₅ macroscopic folds with north-east dipping axial planes deform pegmatites (James *et al.*, 1991; Trail *et al.*, 1967), which are correlated with pegmatites in the Stillwell Hills that were deformed by east-trending D₂ folds (Clarke, 1987; Clarke, 1988). Therefore, these folding events may have a similar timing.

There are further similarities between the east-trending fabrics on the Mawson Coast and the northern Prince Charles Mountains. In the Stillwell Hills, Cape Bruce and the northern Prince Charles Mountains, the deformation is associated with an east-trending mineral lineation that mostly lies parallel to fold axes (Clarke, 1988; this study; Fitzsimons & Thost, 1992). In the Cape Bruce and the northern Prince Charles Mountains areas, east-trending foliations are associated with high-grade metamorphism and partial melting (Table 6.1). Geochronology also provides evidence that the east-trending fabrics were produced contemporaneously. At Cape Bruce, *c.* 920 Ma metamorphism is correlated with the D₄ event (see Chapter 4), and syn- to post-D₂ felsic intrusions at the Jetty Peninsula have been dated at *c.* 939 Ma and 940±20 Ma (Hand *et al.*, 1994a; Manton *et al.*, 1992). Granitic dykes and pegmatites intrude across east-trending structures in the Stillwell Hills, at Depot Peak and in the northern Prince Charles Mountains (Table 6.1).

Therefore, it is likely that east-trending fold structures along the Mawson Coast and east-trending shear zones in the northern Prince Charles Mountains were produced by the same tectonic event, which occurred in the latter stages of the Rayner Structural

Episode, *i.e.* before *c.* 900 Ma. This event would have involved a component of north-south compression. A south over north sense of movement has been inferred for shear zones at Depot Peak (Stüwe & Hand, 1992) and Jetty Peninsula (Hand *et al.*, 1994b). However, this is orthogonal to the east-trending mineral lineation observed in several localities, which was inferred to represent a direction of maximum extension by Clarke (1988). If the deformation involved a large component of simple shear, east-trending lineations can be used to infer a strike-slip movement in the shear zones of the northern Prince Charles Mountains. This has important implications for the tectonic modelling of the Rayner Complex (see below).

Deformation continued in some localities with the development of north-trending fabrics. Upright F_3 folds in the Stillwell Hills, with a steep east-dipping axial planar foliation, were correlated by Clarke (1988) with vertical north-trending S_3 foliations and associated folds in the Framnes Mountains (Table 6.1). As with earlier structures in the Stillwell Hills, F_3/S_3 is associated with an east-trending mineral lineation. Similar structures have not been reported from the northern Prince Charles Mountains.

Semi-ductile shear zones containing mylonite and/or pseudotachylite transect earlier structures in all localities between the Mawson Coast and the northern Prince Charles Mountains (Table 6.1). On the Mawson Coast, these shear zones vary from north to south dipping, with a down dip stretching lineation. South over north displacement is inferred for most localities (Clarke, 1988; White & Clarke, 1993; this study; unpublished observations by the author in the Framnes Mountains). In the northern Prince Charles Mountains, Fitzsimons & Thost (1992) reported south-dipping D_8 mylonitic shear zones with south over north movement and down-dip lineations. In the vicinity of the Jetty Peninsula, sub-vertical north-east trending mylonite zones (MY2, Nichols, 1995) and north-west trending shear zones (D_4 , Hand *et al.*, 1994b) are present. Nichol (1995) has inferred north-west over south-east movement for the MY2 zones. Therefore, while the D_8 mylonite zones in the western parts of the

northern Prince Charles Mountains may be related to similar structures along the Mawson Coast, their relationship to shear zones in the Jetty Peninsula area is uncertain.

6.3 Possible tectonic models

Similarities between the deformation histories of the localities discussed above can be used to infer a common tectonic mechanism for the formation of the Rayner Complex in Kemp Land and MacRobertson Land. As there is a lack of clear kinematic information from gneisses in most localities, it is impossible to develop a conclusive model for tectonism during the Rayner Structural Episode. However, based on structural and geochronological evidence, a few alternatives can be proposed. It is recognised by the author that the proposed models are, at best, tentative.

6.3.1 Structural controls on tectonic modelling

The earliest stage of the Rayner Structural Episode produced an intense pervasive gneissosity under granulite facies conditions. Although the orientation of the gneissic fabric varies between different localities, the predominance of north-trending structures in Kemp Land and MacRobertson Land can be attributed to a strain regime with a large component of east-west compression. Low-angle structures may have been produced by thrusting, whereas high-angle structures may represent either deformation orthogonal to the direction of maximum compression, or thrust structures that were re-oriented during subsequent deformation. Simple shear was also involved in the deformation, as indicated by the presence of asymmetrical structures, mineral lineations, S-C fabrics and/or fold vergence in all localities. However, stretching directions, as indicated by mineral lineations, are highly variable. Although this variance can be partially attributed to re-orientation by later events, it is possible

that strain partitioning also played a part (Tikoff & Greene, 1997). Sub-horizontal east-trending stretching lineations, associated with shallowly dipping foliations, may have been the result of thrust-type movement in simple-shear dominated high-strain zones. In the Stillwell Hills, this thrusting appears to have had a west over east sense of movement (Clarke, 1988). In localities with steeply dipping foliations and down-dip lineations, such as the Framnes Mountains, a larger component of pure shear may have been present, with stretching having occurred orthogonally to the direction of maximum contraction (Tikoff & Greene, 1997).

At Cape Bruce and in the Framnes Mountains (unpublished observations by the author), the formation of high-strain gneissic fabrics was followed by upright open folding, as a product of east-west compression. The folding in these localities is possibly a reflection of decreasing strain rates after the early high-strain stages of the Rayner Structural Episode. Decreasing strain may have been the result of decreasing stresses in a compression setting, or of a spreading out of stresses across a cross-section of over-thickened crust.

There was a sudden and significant change in the style and direction of deformation between the early and late stages of the Rayner Structural Episode. A transition from east-west contraction to north-south contraction is seen in all localities, along with a shift from low-angle to steeply inclined structures. Deformation has produced discrete vertical high-strain zones in the northern Prince Charles Mountains and open upright folding along the Mawson Coast. Between these two structural styles, there is a consistent development of sub-horizontal east-trending mineral lineations. These lineations can be used to infer that, despite the change in the orientation of planar structures, the direction of transport did not change significantly between the early and late stages of the Rayner Structural Episode.

There are two extreme scenarios that can be proposed to explain the widespread change in deformation styles during the Rayner Structural Episode. Firstly, the

transition may have been the result of a change in strain as a consequence of progressive deformation, as in the case of a transition from contraction-dominant to transcurrent-dominant convergence in an oblique-collision setting (*e.g.* Jacobs & Thomas, 1994; Tikoff & Saint Blanquat, 1997). The second scenario infers that the transition reflects a change in the forces driving compression, such as a change in the motion of colliding crustal blocks, or the involvement of more than two crustal blocks in tectonic collision. These two scenarios are considered below.

The development of east-trending mylonitic shear zones postdated the ductile events of the Rayner Structural Episode in MacRobertson Land and Kemp Land. The sense of movement varies greatly between localities. However, in MacRobertson Land there is a dominance of south over north reverse displacement. Therefore, the shear zones may have developed in response to a further stage of north-south compression. However, down-dip stretching lineations in the shear zones indicate that, unlike the structures developed in the late stages of the Rayner Structural Episode, deformation proceeded without a significant component of strike-slip displacement. As the timing of the shear zones is only loosely constrained to the period *c.* 900-500 Ma, it is possible that there was a prolonged hiatus between the ductile and semi-ductile events.

6.3.2 Age-defined terranes in Kemp and Mac Robertson Land

Geochronological studies of the gneisses of Kemp and MacRobertson Land indicate that the Rayner Complex is a composite of crustal units with a wide variety of ages of formation. Distinct periods of crustal growth and reworking can be recognised from protolith and metamorphic ages, as well as from age data deriving from detritus in metasediments and material inherited from the source rocks of orthogneisses (see section 6.1). However, the structural relationships between these crustal units are obscured by the effects of tectonism and metamorphism. The relationships between

lithologies and the sources of detrital or melted materials are unknown, since in most cases the inferred source rocks are not exposed. Age data can only be used to define 'terranes', or regions within the crust that have distinct ages of formation. The relationships between these terranes are generally unknown.

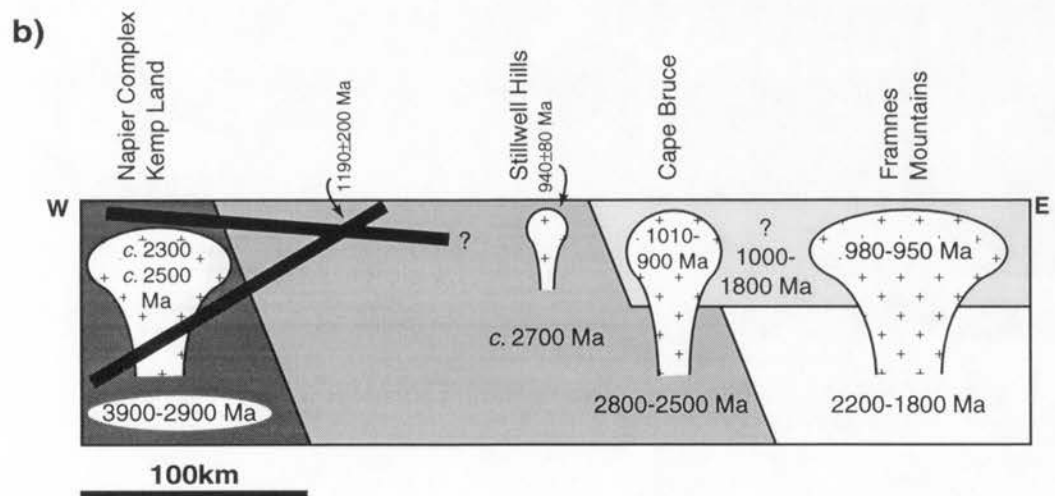
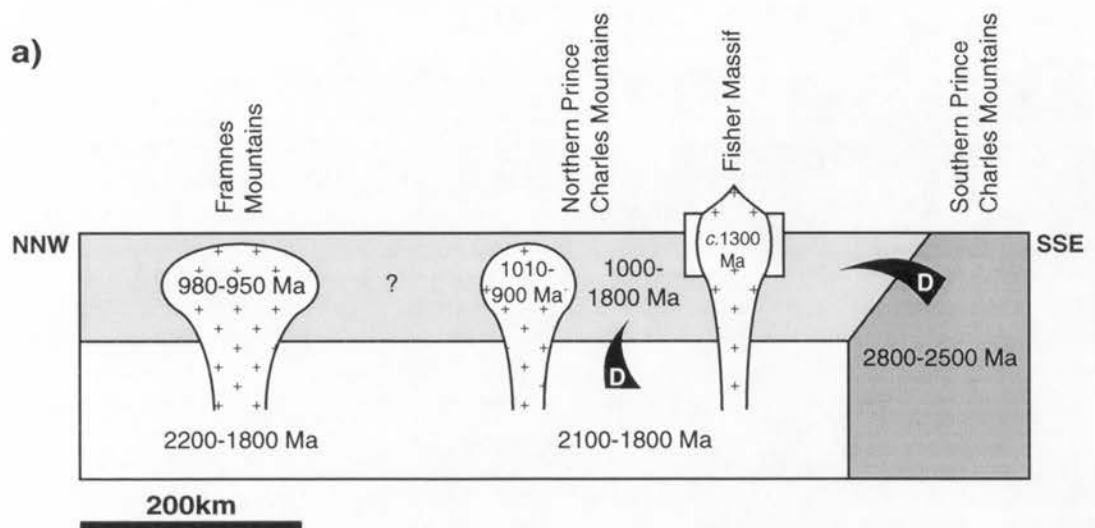
The best constrained ages for the formation of continental crust come from the northern Prince Charles Mountains. Granitoids and charnockites intruded metamorphosed sedimentary and igneous lithologies at *c.* 1000-900Ma. Dates from a sample of metasediment (Kinny *et al.*, 1997) suggest that the sediments included detritus that was derived from exposed crust with ages of *c.* 1800-2100 Ma and *c.* 2500-2800 Ma. The latter age range corresponds to the ages of lithologies in the southern Prince Charles Mountains, providing evidence for the two terranes having been adjacent at some time between *c.* 1800 and 1000 Ma. A few dated orthogneisses have inherited zircons with ages of *c.* 1900 Ma (Kinny *et al.*, 1997). This zircon may be derived from country rocks that are not exposed in the northern Prince Charles Mountains, or from a lower crustal source that has been inferred for most of the intrusions (Kinny *et al.*, 1997). There is no evidence of Archaean inheritance in any of the dated orthogneisses. The Fisher Massif, which appears to be an allochthonous block, contains volcanic and intrusive units with island-arc and continental margin geochemical affinities (Mikhalsky *et al.*, 1996). These igneous rocks have been dated at *c.* 1300 Ma. The relationships between age-defined terranes in the northern Prince Charles Mountains are summarised in Fig. 6.2a.

In the Mawson Base / Framnes Mountains province, inherited zircon from a pre-1000 Ma orthogneiss (Young & Black, 1991), and Sm-Nd T_{DM} model ages from the *c.* 980-950 Ma Mawson Charnockite (Young *et al.*, 1997), yield ages in the range 2200-1800 Ma. These orthogneisses probably derived from lower crustal sources, and the ages correspond to inherited ages in the northern Prince Charles Mountains (Fig. 6.2a). Metasediments in the Framnes Mountains were deposited at an unknown time

Fig. 6.2. Schematic relationships between age-defined terranes in Kemp Land and MacRobertson Land. Ages indicate the time of formation of the various lithologies. Boundary relationships between adjacent terranes are inferred. Black bars represent mafic dykes; arrows marked “D” represent the transfer of detrital material from the source terrane to the depositional terrane. As the relationships between the terranes are based on age inheritance, the cross sections do not necessarily reflect the present day structure of the Rayner Complex.

a) Schematic NNW-SSE cross section, extending from the Framnes Mountains on the Mawson Coast to the southern Prince Charles Mountains.

b) Schematic W-E cross section, extending from the Framnes Mountains on the Mawson Coast to the eastern margin of the Napier Complex in Kemp Land.



before *c.* 1000 Ma metamorphism. However, deposition may have occurred in a similar time-frame to that in the northern Prince Charles Mountains (Fig. 6.2a).

The new dating of the gneisses at Cape Bruce suggests a history distinct to that of the provinces further east (Fig. 6.2b). Intrusive activity, metamorphism and deformation occurred at Cape Bruce in the period *c.* 1000-900 Ma. The dated orthogneisses contained inherited zircon with ages of *c.* 1300-1000 Ma and *c.* 2800-2500 Ma. The age of the metasediments is unknown. However, there is no evidence for metamorphism having occurred at Cape Bruce before *c.* 1000 Ma (*cf.* the Stillwell Hills), and the lack of prehistory in the area may be an indicator of younger crustal ages.

In the Stillwell Hills, Archaean crust is exposed, as indicated by a *c.* 2700 Ma age of metamorphism and/or emplacement of charnockitic orthogneiss (Clarke, 1987). Pegmatites that intruded at *c.* 940 Ma also contain inherited Archaean zircon (Grew *et al.*, 1988). The *c.* 2700 Ma age does not correspond to Archaean intrusive or metamorphic events that have been dated in the adjacent Napier Complex (*e.g.* Harley & Black, 1997). The relationship of Archaean crust in the Stillwell Hills to the Napier Complex is therefore unclear. However, the correlation of mafic orthogneisses in the Stillwell Hills with 1190 ± 200 Ma dykes in the Napier Complex (Fig. 6.2b), and the occurrence of *c.* 1000 Ma metamorphism in the Stillwell Hills and further west, provides evidence that the two Archaean provinces were adjacent at the start of the Rayner Structural Episode.

As relationships between the age-defined terranes are poorly known, there is little information of the timing of the amalgamation of the Rayner Complex. In the northern Prince Charles Mountains, Archaean, Paleoproterozoic and Mesoproterozoic terranes may have been adjacent before the *c.* 1300 Ma magmatic event in the Fisher Massif; however, the relationship between Fisher Massif and the surrounding granulite

province is unclear (Beliatsky *et al.*, 1994; Mikhalsky *et al.*, 1996). In other provinces, the amalgamation of age-defined terranes may have occurred before or at the start of the Rayner Structural Episode.

6.3.3 Settings for the growth of Proterozoic crust

Gneisses and felsic intrusions that are exposed in MacRobertson Land have isotopic characteristics that preserve evidence for the reworking of Proterozoic crust in the middle and lower levels of the crust. The data available can be used to infer that the majority of the Rayner Complex in this area is based on extensive sections of post-Archaean crust. These sections may represent a large fragment of Proterozoic continental crust, that was formed *in situ* or tectonically juxtaposed with adjacent Archaean cratons. Alternatively, the Rayner Complex may be an amalgamation of Proterozoic terranes, that were originally developed on the margins of Archaean continental blocks. In the latter case, crustal growth would have proceeded by magmatic activity in island arcs and on active continental margins. Amalgamation of the terranes may have occurred during plate convergence, in Andean or Tibetan-type settings.

While there are no gneisses in MacRobertson Land that can be clearly related to oceanic crust, there is some geochemical evidence that some of the terranes contain lithologies produced on an active continental margin. Magmatic rocks at Fisher Massif have geochemical signatures that are consistent with a mixture of island arc and back-arc magmatism (Mikhalsky *et al.*, 1996; Kinny *et al.*, 1997). The *c.* 1000 Ma charnockites in MacRobertson Land have compositions that are consistent with a derivation by the partial melting of lower crust, which contained igneous material with an intermediate calc-alkaline composition (Young *et al.*, 1997; Zhao *et al.*, 1997). The presence of calc-alkaline geochemical signatures provide evidence for the development of Proterozoic crust in MacRobertson Land in an active continental margin setting.

6.3.4 Tectonic scenarios for the Rayner Structural Episode

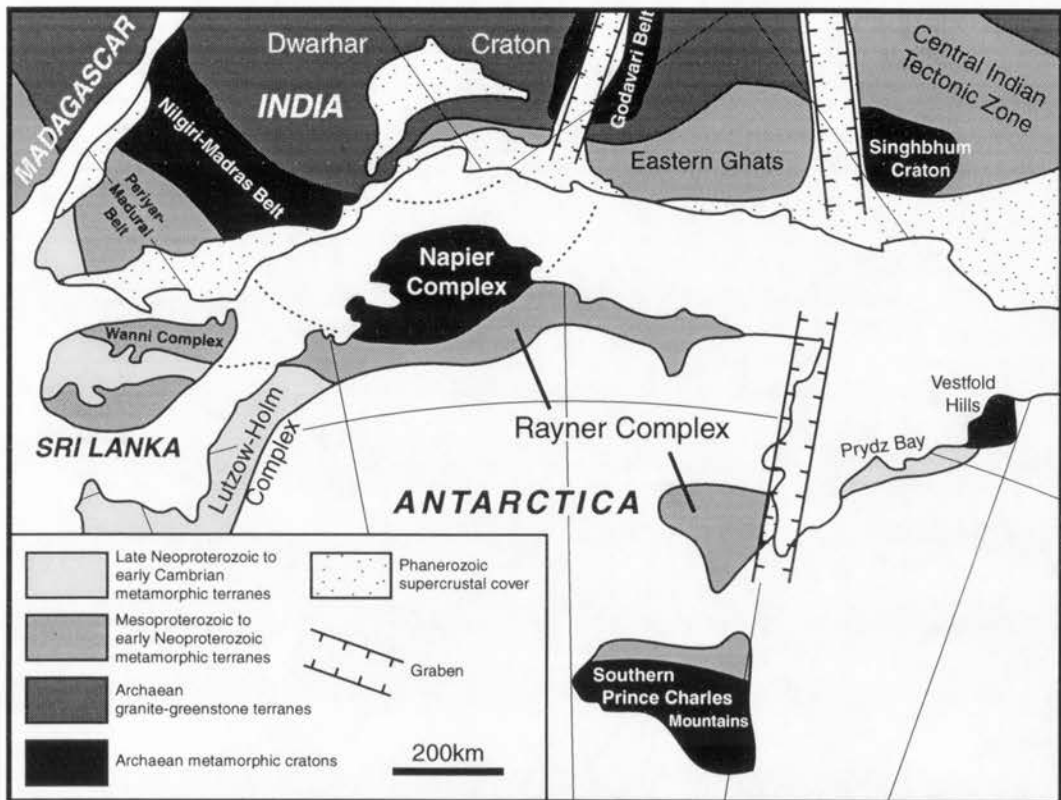
The Rayner Structural Episode in Kemp and MacRobertson Land involved the compression of Proterozoic crust against the Archaean craton of the Napier Complex. According to reconstructions of east Gondwana (*e.g.* Yoshida *et al.*, 1992), the mobile belt of the Rayner Complex extended across into the Eastern Ghats Granulite Belt, on the eastern edge of the Indian peninsula (Fig. 6.3; Grew & Manton, 1986; Yoshida *et al.*, 1996). The fabric of rocks in the Eastern Ghats Mobile Belt is dominated by an intense NE-trending foliation, which has been attributed to the compression of the province in a NW direction towards Archaean cratonic nuclei and greenstone belts (Bhattacharya, 1996). The Napier Complex has been interpreted by Yoshida *et al.* (1996) as a segment of an Archaean granulite belt, that extended from the Napier Complex into the southern part of the Indian peninsula, and also into the west side of the Eastern Ghats Granulite Belt (Fig. 6.3).

As previously discussed, there is a major change in the nature of ductile deformation during the Rayner Structural Episode. Two simple scenarios for tectonism are suggested here, and the pros and cons of each are discussed.

Scenario One: Single phase collision

It is possible that the different structural regimes in the early and late stages of the Rayner Structural Episode are the product of progressive deformation without a significant change in the direction of convergence. The transition from thrust-type to transcurrent convergence has been observed in several orogens that evolved as a result of oblique collision (*e.g.* Brown & Talbot, 1989; Jacobs & Thomas, 1994; Tobisch *et al.*, 1995; Dias & Ribeiro, 1995; Gresse, 1995). Such transitions are seen in orogens involving magmatic arc accretion on a continental margin (*e.g.* the Sierra Nevada of California; Tobisch *et al.*, 1995), and in zones of continent-continent collision

Fig. 6.3. Geological relationships in East Gondwana, based on the reconstruction by Yoshida *et al.* (1992). The extent of metamorphic terranes is based on maps by Grew & Manton (1986), Yoshida *et al.* (1992), Shiraishi *et al.* (1994), Yoshida *et al.* (1996), and Shaw *et al.* (1997).

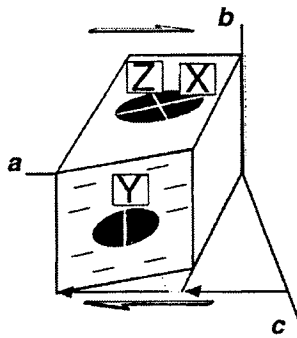


following the closure of an oceanic basin (*e.g.* the Variscan orogeny in western Europe; Dias & Ribeiro, 1995).

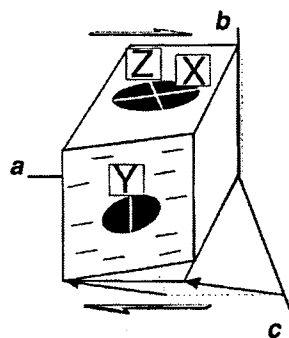
In the early stages of an oblique collision, crustal shortening is often accommodated by either folding dominated by pure shear, and/or thrusting parallel to the direction of maximum convergence (Gresse, 1995). Thrust structures typically contain a sub-horizontal stretching lineation, parallel to the transport direction, whereas steeply dipping structures produced with a large component of pure shear contain a sub-vertical stretching lineation (Fig. 6.4; Tikoff & Greene, 1997). Progressive deformation during convergence causes significant crustal thickening. At this stage, convergence can be further accommodated by the formation of transcurrent shear zones (Tobisch *et al.*, 1995). Transcurrent displacement is commonly associated with a high degree of strain partitioning, with the development of sub-vertical high strain zones that trend at a low angle to the continental margin. Movement in the shear zones is dominantly lateral, with the development of sub-horizontal stretching lineations (Tikoff & Greene, 1997).

A tectonic scenario involving oblique collision can be postulated for the Rayner Structural Episode in Kemp and MacRobertson Land. The early stage tectonism (*c.* 1000-950 Ma) was characterised by a combination of east over west thrusting and orthogonal compression (Fig. 6.5a). East-trending stretching lineations in thrust and sheath folds, and down-dip lineations in steep north-trending fabrics, are consistent with an westward direction of transport in a transpressional setting (Tikoff & Greene, 1997). The development of east-trending structures in the latter stages of the Rayner Structural Episode (*c.* 950-900 Ma) represent a shift towards transcurrent convergence (Fig. 6.5b). Shear movement in high-strain zones in the northern Prince Charles Mountains involved a large horizontal component, as indicated by stretching lineations, and a reverse vertical component, as indicated by south over north displacement (Hand *et al.*, 1994b; Fitzsimons & Thost, 1992). Upright E-trending folds developed in other provinces may represent a lower strain response to transpression. The

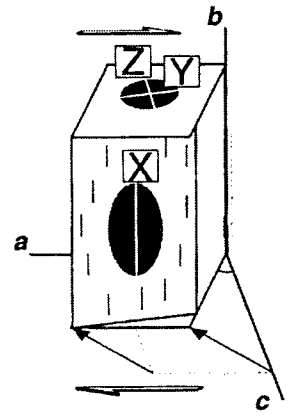
Fig. 6.4. The development of stretching lineations in a transpressional setting, from Tikoff & Greene (1997). Stretching lineations in a simple shear setting lie parallel to the direction of transport. In a transpressional setting, lineations may lie parallel or orthogonal to the simple shear component of strain. The stretching direction may be orthogonal to the direction of simple shear, if there is a large component of pure shear in transpression, or if the degree of strain is very high.



Simple shear



Wrench-dominated



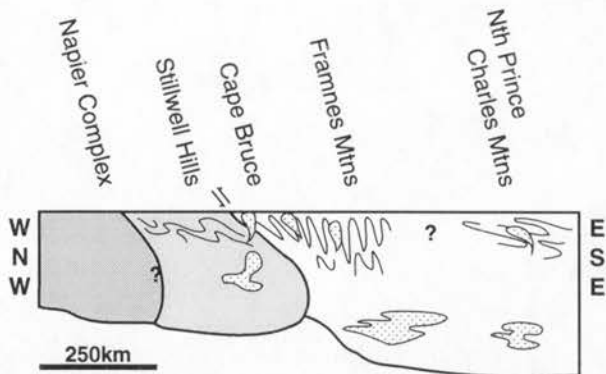
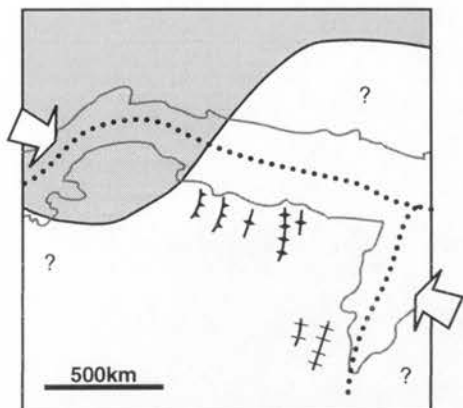
Wrench-dominated or
pure-shear dominated

Transpression

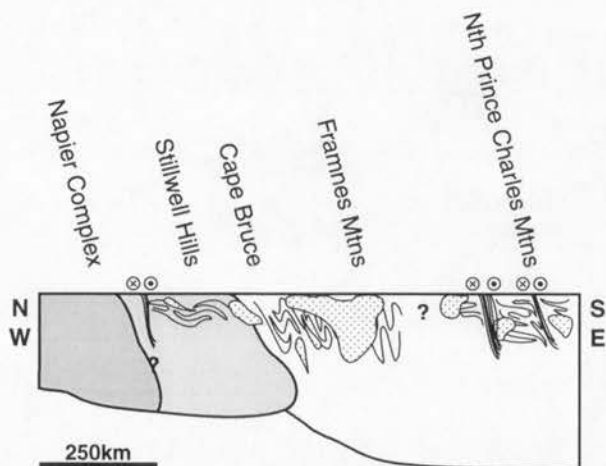
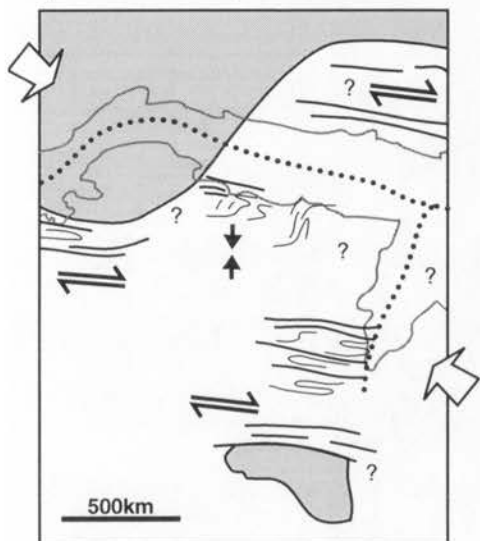
Fig. 6.5. Possible scenarios for the development of the Rayner Structural Episode in Kemp Land and MacRobertson Land. Block diagrams are based on surface relationships between exposed lithologies, and the possible locations of age-defined terranes.

- a) Oblique collision during the early stages of the Rayner Structural Episode. Crustal thickening and shortening is achieved by the westward transport of Proterozoic crust against Archaean cratons in Enderby Land and India.
- b) The development of transpressional structures in the late stages of the Rayner Structural Episode, as a result of progressive deformation during continued oblique convergence. This scenario does not require a change in the tectonic forces that produced the orogen.
- c) The development of transpressional structures in the late stages of the Rayner Structural Episode, as a result of a change in the direction of collision from westward to north westward. This scenario implies that the similarities in the stretching directions between the early and late stages of the Rayner Structural Episode are coincidental.

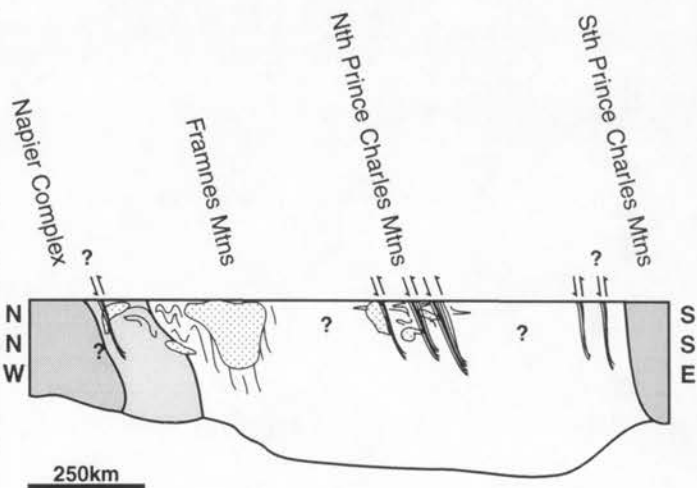
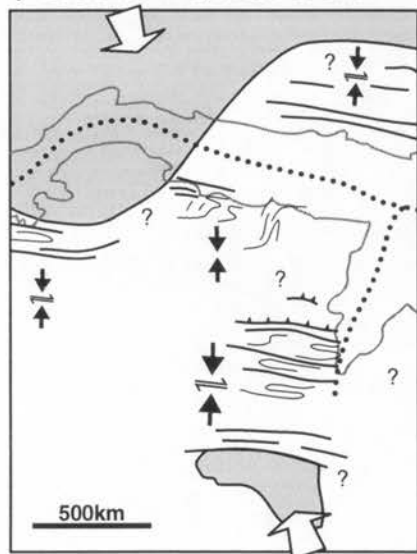
a) c.1000-980Ma



b) Scenario One, c.980-900Ma



c) Scenario Two, c.980-900Ma



development of these structures required strike-slip along a west-trending continental margin, and an approximately north-west direction of compression.

The main shortcoming of this model is the low angle between the transport direction inferred for the early stage of deformation, and the direction of strike-slip movement in the later events. This requires a low angle between the direction of convergence and the trend of the continental margin. If this is the case, it is perhaps unlikely that a large amount of thrusting and orthogonal contraction would have taken place. In the Natal Metamorphic Province of South Africa, Jacobs & Thomas (1994) found a similar relationship between earlier thrust structures and later transcurrent shear zones; however, in the Natal Province, the angle between the trends of earlier and later stretching lineations was approximately 45° . Another problem is that the later deformation in the Rayner Structural Episode requires convergence in a north-west direction, whereas there is no evidence for a northward vector in the earlier structures.

A possible answer to these problems may be found in an examination of the shape of the continental margin (Fig. 6.3). From the Napier Complex to the western side of the Eastern Ghats Mobile Belt, the boundary between Archaean crust and the mobile belt is distinctly S-shaped. Convergence and collision in a north-westerly direction (from an Antarctic frame of reference) is likely to have produced different effects on different parts of the continental margin. For converging crust immediately to the east of the Napier Complex, contraction may have occurred perpendicular to the margin of the craton, producing north-trending structures with a westward direction of transport (Fig. 6.5a). Elsewhere, north-west convergence may have resulted in transpression, that evolved into the development of east-trending shear zones which were sub-parallel to the east-trending sections of the continental margin (Fig. 6.5b).

Scenario Two: Two-stage collision

An alternate, and potentially simpler, explanation for the change in deformation style during the Rayner Structural Episode, proposes that the early and late stages of

deformation were produced by different convergent regimes. During the early stages of the Rayner Structural Episode, east-west convergence produced thrusting and upright contraction in the same fashion as described in Scenario One (Fig. 6.5a). In the later stages, there is a transition from westward convergence to northward or north-westward convergence. This change in tectonism may have been caused by the late addition of a new northward moving indenter, or by the rotation of the original indenter from a westward to northward direction of motion (*cf.* Dias & Ribeiro, 1995). Northward convergence produced upright east-trending folds in low strain zones (as found on the Mawson Coast) and discrete high-strain shear zones with a north-east to north-west direction of transport (Fig. 6.5c).

This second tectonic scenario implies that the similar orientations of stretching directions in the early and late stages of the Rayner Structural Episode are coincidental, and are not necessarily indicative of the direction of transport during collision. In the high-strain zones of the northern Prince Charles Mountains, the east-trending lineations may have been developed orthogonally to the direction of transport if there was extremely high strain, and/or a large component of pure shear (Tikoff & Greene, 1997). However, it is unlikely that such a large degree of stretching could be accommodated in a horizontal direction (Brown & Talbot, 1989; Bell, 1981). The high degree of strain partitioning in the structures of the northern Prince Charles Mountains is also suggestive of shear-strain dominated deformation in a transpressional setting. On the Jetty Peninsula, the presence of south-east plunging lineations and south-upward kinematic indicators implies a south-east-upward (dextral) direction of movement in the high-strain zones (Hand *et al.*, 1994b). This is consistent with oblique convergence in a north-west direction.

6.4 Causes of magmatism and metamorphism

The Rayner Structural Episode in Kemp Land and MacRobertson Land was associated with widespread granulite facies metamorphism and the emplacement of large amounts of felsic magma. The relationships between magmatism and metamorphism are not clearly established. At Cape Bruce and in the northern Prince Charles Mountains, charnockites and granitoids were emplaced just before and during the development of the dominant gneissosity, which is associated with peak metamorphism (this study; Kinny *et al.*, 1997). However, the major intrusive event occurred after the peak of metamorphism (*i.e.* at ≤ 980 Ma), with the emplacement of large volumes of charnockite along the Mawson Coast and in the northern Prince Charles Mountains (Young & Black, 1991; Kinny *et al.*, 1997). The generation of high temperatures in the middle crust during tectonism cannot, therefore, be simply attributed to magmatic activity.

Charnockites in MacRobertson Land have geochemical and isotopic characteristics that are consistent with their derivation by the partial melting of anhydrous granulitic lower crust (Young & Ellis, 1991; Zhao *et al.*, 1997). The sources of partial melting are inferred to have a composition corresponding to dehydrated calc-alkaline lithologies, that were melted at 950–1050°C to produce the charnockitic magmas (Zhao *et al.*, 1997). A mechanism is required for the production of such high temperatures in the lower crust.

It has been suggested that temperatures in the lower crust that are high enough to cause substantial partial melting could have been produced by thinning of the underlying mantle lithosphere (Sandiford & Powell, 1991; Zhao *et al.*, 1997). Such thinning would have involved the delamination of the lower section of the continental lithosphere. Delamination has been suggested as a cause of high heat flow during the extensional collapse of an orogen (Sandiford, 1989), or during compression where there is a large degree of strain partitioning between the upper and lower lithosphere

(Sandiford & Powell, 1991). In the latter case, a significant degree of melting in the lower crust can be achieved during compression only if thinning of the mantle lithosphere has accompanied crustal thickening.

Zhao *et al.* (1997) have suggested that the large volumes of charnockite in the northern Prince Charles Mountains were generated by decompression melting, as a result of the isostatic rebound of over-thickened crust after the waning of compressive strain. However, there are problems with this model. Post-orogenic isostatic rebound is typically associated with the extensional collapse of an orogen. In the northern Prince Charles Mountains and on the Mawson Coast, there is no structural evidence for extension having occurred before or during the emplacement of charnockitic intrusions. There is also ample evidence that compressional orogenesis continued shortly after the magmatic event. Thirdly, there is a lack of decompression textures in the metamorphic rocks of the region. Geochronological constraints on isobaric cooling paths imply that, at least for the Cape Bruce area, the gneisses were present at similar mid-crustal levels during the early and late stages of the Rayner Structural Episode.

In the lower crust, temperatures high enough to cause anhydrous partial melting could have been achieved by thinning of the mantle lithosphere (Sandiford & Powell, 1991). Such thinning may have occurred during continued collision, and need not have involved extensional failure of the orogen if the forces of lateral compression still exceeded buoyancy forces in the thinned lithosphere (Turner *et al.*, 1992). Although melting and magma emplacement may have caused crustal weakening, such weakening was probably transient (Sandiford *et al.*, 1991). After the cooling and solidification of magmas, convergent strain was renewed. Such a scenario has been suggested by Tobisch *et al.* (1995) for the tectonic history of the central Sierra Nevada magmatic arc. In their model, large scale plutonism occurred in a tectonically neutral setting, in-between an earlier episode of crustal contraction and the later development of strike-slip shear zones. Tobisch *et al.* (1995) attribute the sequence of events to a single progressive episode of oblique plate convergence. This episode

involved a decrease in the angle between the direction of convergence and the trend of the orogen, that occurred between the early and late stages of collision.

Although thinning of the mantle lithosphere could produce high temperatures in the lower crust, modelling by Sandiford & Powell (1991) indicates that heating of the middle crust by diffusive transfer could not exceed 700°C in any feasible scenario. The production of granulites and migmatites in most of the Rayner Complex required higher temperatures for the peak of metamorphism. As an alternative to diffusive heat transfer, heat transport by the ascent of magmas into the mid crust has been invoked by many authors (*e.g.* Sandiford & Powell, 1991; Stüwe *et al.*, 1993).

In MacRobertson Land, magmatic rocks that were emplaced immediately before or during the early stages of the Rayner Structural Episode are present in most localities. The emplacement of these intrusions were probably contemporaneous with the peak of metamorphism. However, it is not clear if the intrusions provided heat for metamorphism, as their total volume has not been measured. It is also likely that many of the granitic intrusions were derived by the partial melting of local lithologies, as indicated by the high degree of migmatization during peak metamorphism (White & Clarke, 1993; Fitzsimons & Thost, 1992; this study). Although there are sufficient volumes of charnockite emplaced to account for elevated temperatures in some localities, these intrusions post-dated the peak of metamorphism. For granulites in the Colbeck Archipelago that lie near the western edge of the Mawson Charnockite, White & Clarke (1993) suggested that the early stages of metamorphism were driven by the emplacement of large amounts of magma below the present level of exposure. Therefore, although advective heating is the most likely explanation for granulite facies metamorphism in MacRobertson Land, the source of heating is yet to be identified. An examination is required of the petrogenesis and volumetric extent of intrusions that were emplaced in the early stages of the Rayner Structural Episode, to determine their role in widespread metamorphism.

Summary



Summary

The Rayner Complex in MacRobertson Land is composed of granulite facies gneisses that were metamorphosed at mid-crustal levels during the *c.* 1000-900 Ma Rayner Structural Episode. New geological information presented in this study, derived from detailed investigations of the Cape Bruce area on the Mawson Coast, has important implications for our understanding of the origin and evolution of the Rayner Complex. By comparing this new information with published data from localities in MacRobertson Land and Kemp Land, I have attempted to shed light on the origin of lithologies in the Rayner Complex, the processes operating during the Rayner Structural Episode, and the tectonic regime under which the complex evolved.

Gneisses in the Cape Bruce area represent a supercrustal sequence that was deformed, metamorphosed and intruded by igneous bodies during the Rayner Structural Episode. Felsic and pelitic sedimentary rocks of probable Proterozoic age were buried at mid-crustal levels prior to the emplacement of charnockitic bodies before *c.* 1000 Ma. These lithologies were metamorphosed, migmatized and deformed into a gneissic fabric by a D₁-M₁ event, during the early stages of the Rayner Structural Episode. Geothermobarometry conducted on samples from Cape Bruce, that included garnet-bearing orthogneisses and paragneisses containing evidence for the stability of spinel with quartz, provide an estimate for P-T conditions during M₁ of 6-7 kbar and $\geq 800^{\circ}\text{C}$. Cooling after M₁ at near-constant pressures, that coincided with the crystallisation of leucosome, produced hydrous M₂ reaction assemblages in many of the gneiss samples studied. Minerals that grew during M₂, including garnet, biotite, hornblende and magnetite, were aligned with a strong foliation during a D₂ event, which also re-oriented the gneissosity produced during D₁. D₁ and D₂ may have occurred during a continuous period of deformation.

Between D₁ and D₂, granitic and biotite+pyroxene-bearing plutons and dykes were intruded into the gneisses. Felsic, garnet-bearing dykes and pegmatites were intruded after D₂. These felsic magmas may have derived from the partial melting of lithologies in the vicinity of Cape Bruce during D₁ and D₂. Lithologies were subsequently deformed into upright north-trending folds by a D₃ event. Similarities in the orientation of F₂ and F₃ fold axes provide evidence that D₃ was a stage of decreasing strain during a period of continuous deformation from D₁ to D₃. Felsic pegmatites were emplaced after D₃, and were subsequently deformed along with other lithologies into upright east-trending folds during a D₄ event. The growth of M₃ minerals in pre-D₃ and post-D₃ lithologies occurred before or during the D₄ event. Mineral assemblages are consistent with the M₃ event having occurred at lower temperatures ($\geq 550^{\circ}\text{C}$) than, and at similar pressures to, the M₁ and M₂ events. Narrow east-trending D₅ shear zones, containing mylonite and pseudotachylite, were formed after the ductile D₁-D₄ events. These shear zones contain M₄ mineral assemblages that grew under granulite facies conditions at approximately 6.5 kbar and 550-800°C (White & Clarke, 1994). Whereas D₁ to D₄ probably belonged to the *c.* 1000 Ma Rayner Structural Episode, the D₅ event may have occurred at a much later time.

SHRIMP analyses of U-Pb isotopes in zircons extracted from six lithologies provide age constraints on the timing of geological events. The mixture modelling technique of Sambridge & Compston (1994) was applied to four of the samples, in order to resolve multiple age populations in data from each sample. Orthogneisses that were intruded during the Rayner Structural Episode contain some zircon that was inherited from lower crustal sources or mid-crustal country rocks. Isotopic data obtained from inherited zircon provide evidence that both Archaean and Mesoproterozoic rocks were present in the region at the time of the Rayner Structural Episode. The early stages of the Rayner Structural Episode, which at Cape Bruce involved D₁-D₂ deformation and M₁-M₂ metamorphism, occurred during the period *c.* 1000-945 Ma. A sample of

biotite-pyroxene orthogneiss, that was emplaced just before or during the D₂ event, was dated at 995±7 Ma. A sample of pre-D₃ felsic orthogneiss was dated at 944±10 Ma. Four of the samples yielded isotopic data which included age populations at about 910 Ma. This age is interpreted as the timing of zircon growth, recrystallization and isotopic resetting during M₃. The M₃ event may have represented a brief period of granulite facies metamorphism, distinct from a progressive M₁-M₂ event that occurred at higher temperatures and similar pressures.

The geological evolution of the Cape Bruce gneisses possessed important differences to other localities on the Mawson Coast and in the northern Prince Charles Mountains. On the basis of published geochronological data, terranes with distinct ages of formation are proposed for Kemp and MacRobertson Land. Exposures of Archaean gneisses in the Rayner Complex are limited to the Mawson Coast in Kemp Land, including the Stillwell Hills. In MacRobertson Land, the protoliths of exposed gneisses were formed within the period 1800-1000 Ma. These protoliths were metamorphosed at mid-crustal levels during the 1000-900 Ma Rayner Structural Episode. Large volumes of charnockite, including the Mawson Charnockite batholith on the Mawson Coast, were emplaced at mid-crustal levels in MacRobertson Land during the Rayner Structural Episode. At Cape Bruce, Archaean ages from dated lithologies are interpreted as having been derived from crustal material that is not exposed in the region. In the northern Prince Charles Mountains and the Framnes Mountains, inherited isotopic data yielded ages in the range 2200-1800 Ma. In both cases, the inherited crustal material from which these isotopic data were obtained may have been partially derived from magmatic sources in the lower crust. If this is the case, then the section of Rayner Complex between the northern Prince Charles Mountains and the Framnes Mountains may have been underlain by Paleoproterozoic crust during the Rayner Structural Episode. By contrast, the Cape Bruce area was underlain in part by Archaean crust.

Similarities between the structural histories of localities in MacRobertson Land can be attributed to tectonic processes that operated on a regional scale during the Rayner Structural Episode. The early stages of the Rayner Structural Episode involved the generation of gneissic fabrics during high-strain deformation. This deformation was associated with a large component of east-west compression, which produced east-over-west thrust structures in the Stillwell Hills and upright north-trending structures in the Framnes Mountains. Charnockitic magmas were emplaced in the Rayner Complex during or after this period of compression. In the late stages of the Rayner Structural Episode, there was a regionally significant transition from a contraction-dominated tectonic regime to a transcurrent tectonic regime. The latter produced lateral movement along sub-vertical east-trending high-strain zones, and north-south compression in domains of lower strain. East-trending semi-brittle shear zones, that have a predominantly south-over-north sense of movement, were generated by north-south compression at some time after the Rayner Structural Episode.

A simple tectonic model is proposed to account for the various styles of deformation during the Rayner Structural Episode. The Rayner Complex in MacRobertson Land and Kemp Land was formed during the compression of Proterozoic crust against Archaean continental crust in Enderby Land and the Indian peninsula. The direction of maximum convergence was oblique to the boundary between Archaean and Proterozoic crust. Two scenarios are proposed for the transition from normal contraction to transcurrent contraction during the Rayner Structural Episode. In the first scenario, the transition is attributed to progressive deformation during east-west or southeast-northwest convergence, that occurred at an oblique angle to the continental margin. Similar transitions have been commonly observed in oblique-collisional orogens. In the second scenario, the transition is attributed to a change in the direction of tectonic convergence, from east-west convergence to north-south convergence. An oblique-collisional setting is also inferred for this scenario, to account

for the generation of east-trending transpressional shear zones during the late stages of the Rayner Structural Episode.

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Appendices

S1			S2		S3	S4
dir. dip	dir. dip	dir. dip	dir. dip	dir. dip	dir. dip	dir. dip
302 64	012 47	055 80	340 25	275 30	080 75	353 71
310 59	350 62	150 75	317 55	295 25	080 70	000 70
330 65	072 19	120 85	005 66	107 87	060 80	000 83
349 60	040 13	180 88	010 37	115 85	070 90	020 85
297 65	092 10	235 80	008 50	115 80	065 55	000 85
271 76	055 11	070 80	005 45	325 85	060 70	000 85
257 64	348 36	084 84	356 55	350 80	070 85	350 90
270 75	002 38	115 80	013 36	045 80	300 80	005 85
290 52	000 36	350 70	250 68	020 80	290 80	020 90
282 53	347 30	300 80	264 63	100 80	285 75	020 80
305 65	354 45	310 80	283 76	350 70	290 90	355 85
317 40	332 47	020 50	236 78	288 72	060 70	020 80
305 52	328 28	300 50	100 20	350 65	065 75	010 80
272 32	337 34	320 60	240 65	290 80	065 85	000 80
207 74	323 43	085 65	240 75	335 75	095 60	020 75
290 27	270 74	095 90	250 75	020 70	104 90	000 80
349 35	320 62	260 60	110 90	300 60	295 85	000 85
200 53	313 63	350 55	225 52	350 68	280 80	005 85
225 65	317 43	340 71	065 78	330 75	105 90	000 87
235 76	282 78	020 55	235 80	325 65	105 90	005 82
272 36	114 90	335 42	255 75	290 85	080 70	355 80
285 77	252 82	325 40	260 25	000 55	100 85	000 85
350 56	350 83	335 30	065 60	350 40		020 85
290 65	063 90	355 47	045 20	270 70		
346 30	052 79	346 54	345 15	250 32		
346 46	059 90	330 38	280 30	180 10		
328 37	266 52	335 37	275 43	090 60		
344 60	242 74	338 55	255 65	280 55		
346 55	060 40	010 35	025 25	285 50		
325 44	015 42	235 85	030 35	355 45		
338 42	019 25	340 53	045 40	300 65		
325 44	310 20	305 53	053 56	305 75		
355 45	290 70	345 40	230 34	305 70		
330 39	290 50	010 45	265 40	075 80		
226 31	285 45	330 72	265 80	040 15		
219 55	336 50	310 20	280 30	340 70		
216 60	240 75	045 8	075 82	350 72		
205 73	200 80	040 10	286 75	010 80		
246 45	300 75	330 75	355 62	354 82		
294 43	070 70	285 85	010 45	350 80		
255 79	070 72	005 50	000 50	350 74		
235 46	274 74	210 75	015 65	295 30		
280 58	200 80	260 70	270 81	275 25		
255 73	200 80	245 82	005 70	300 15		
200 90	230 80	245 80	270 85	275 15		
224 78	320 65	260 85	285 75	285 20		
225 75	252 64	070 82	295 70	330 30		
220 58	248 78	270 80	325 65	324 52		
239 46	240 90	310 80	285 75	320 58		
229 55	215 80	310 80	270 80	205 87		
300 51	290 75	280 60	310 80	325 75		
028 42	300 65		310 80	000 22		
013 53	055 80		280 60	020 25		
013 61	140 72		330 20	350 60		
002 52	056 82		025 20			

S5

dir. dip
095 40
025 33
030 66
050 51
010 80
075 35
005 50
047 38
025 40
046 55
047 47

F2	L2	F3	F4	L4	L5
dir. pl.	dir. pl.	dir. pl.	dir. pl.	dir. pl.	dir. pl.
010 16	356 26	187 74	094 68	084 55	045 47
031 37	044 42	340 30	105 65	078 65	020 80
070 70	010 37	340 10	285 65	082 55	070 56
070 30	005 5	335 8	110 82		020 40
060 45	234 53	020 55	280 20		031 32
340 65	357 28	020 45	285 15		
320 68	359 56	020 44	260 60		
335 65	010 35	025 50	270 60		
330 65	210 30	010 30			
323 73	285 75	015 35			
327 80	185 20	020 45			
180 5	200 50	005 35			
170 45	055 30	015 60			
170 65	345 50	005 40			
060 25		005 60			
045 50		000 55			
005 25		355 65			
180 15		030 15			
050 25		025 30			
055 25		019 35			
330 65		014 10			
152 48		350 80			
		355 70			
		010 75			
		000 68			
		005 75			
		355 30			
		000 40			
		180 10			
		180 5			
		025 40			
		000 45			
		005 50			
		010 45			
		020 55			
		195 75			
		010 50			
		005 50			
		005 55			
		180 59			
		020 20			
		030 50			
		230 80			
		210 75			

Appendix 3.1. Analyses for compositional profiles (Fig. 3.1).

Sample No. Trav um Analysis Mineral	O4 0	O4 13	O4 26	O4 40	O4 53	O4 66	O4 80	O4 93	O4 106	O4 120	O4 133	O4 159	O4 173	O4 186	O4 213	O4 226	O4 239	O4 253	O4 266	O4 279	O4 293	Sample No. Trav um Analysis Mineral
	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	b3opxtr opx	
SiO ₂	49.77	49.84	49.91	49.83	49.87	49.97	50.00	49.90	50.36	50.13	49.67	50.10	49.77	50.09	50.27	49.87	50.40	49.96	49.91	50.08	50.71	SiO ₂
TiO ₂	0.15	0.05	0.05	0.17	0.09	0.01	0.07	0.10	0.05	0.03	0.06	0.03	0.03	0.05	0.06	0.05	0.04	0.06	0.04	0.05	0.01	TiO ₂
Al ₂ O ₃	1.58	1.55	1.58	1.60	1.50	1.46	1.47	1.45	1.49	1.40	1.90	1.87	1.54	1.49	1.57	1.52	1.23	1.28	1.50	1.42	1.23	Al ₂ O ₃
Cr ₂ O ₃	0.00	0.00	0.00	0.00	0.03	0.00	0.00	0.00	0.00	0.02	0.04	0.00	0.01	0.02	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Cr ₂ O ₃
FeO	31.76	31.19	31.29	31.06	30.96	31.51	30.97	30.96	31.34	31.13	30.62	30.43	31.56	31.66	30.66	31.11	30.40	30.26	30.70	30.80	30.74	FeO
MnO	0.55	0.45	0.44	0.49	0.51	0.53	0.44	0.58	0.55	0.45	0.53	0.51	0.41	0.50	0.46	0.43	0.48	0.48	0.37	0.43	0.41	MnO
MgO	15.97	16.26	15.97	16.12	15.99	15.89	16.00	16.27	16.25	16.08	16.08	15.82	16.07	16.36	16.16	16.12	16.66	16.35	16.27	16.62	16.70	MgO
ZnO	0.00	0.08	0.00	0.02	0.09	0.04	0.17	0.01	0.00	0.08	0.24	0.07	0.01	0.12	0.03	0.14	0.05	0.04	0.00	0.15	0.14	ZnO
CaO	0.29	0.32	0.33	0.32	0.33	0.37	0.38	0.34	0.33	0.35	0.34	0.54	0.36	0.22	0.27	0.30	0.22	0.19	0.22	0.23	0.16	CaO
Na ₂ O	0.03	0.03	0.04	0.01	0.00	0.01	0.00	0.02	0.03	0.00	0.00	0.03	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.01	Na ₂ O
K ₂ O	0.01	0.02	0.00	0.02	0.01	0.02	0.01	0.00	0.03	0.00	0.06	0.04	0.00	0.00	0.01	0.00	0.00	0.03	0.00	0.00	0.01	K ₂ O
F																						F
Total Ox	100.11	99.80	99.61	99.62	99.38	99.79	99.52	99.63	100.43	99.67	99.54	99.24	99.76	100.53	99.49	99.53	99.48	98.64	99.00	99.78	100.10	Total Ox
No. Ox	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	No. Ox
Si	1.94	1.95	1.95	1.95	1.95	1.95	1.96	1.95	1.95	1.96	1.94	1.96	1.95	1.94	1.96	1.95	1.96	1.96	1.96	1.95	1.97	Si
Ti	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Ti
Al	0.07	0.07	0.07	0.07	0.07	0.07	0.07	0.07	0.07	0.06	0.09	0.08	0.07	0.07	0.07	0.07	0.06	0.06	0.07	0.07	0.06	Al
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Cr
Fe	1.04	1.02	1.02	1.02	1.01	1.03	1.01	1.01	1.02	1.02	1.00	1.00	1.03	1.03	1.00	1.02	0.99	1.00	1.01	1.00	1.00	Fe
Mn	0.02	0.02	0.01	0.02	0.02	0.02	0.01	0.02	0.02	0.02	0.02	0.02	0.01	0.02	0.01	0.02	0.02	0.02	0.01	0.01	0.01	Mn
Mg	0.93	0.95	0.93	0.94	0.93	0.93	0.93	0.95	0.94	0.94	0.94	0.92	0.94	0.95	0.94	0.94	0.97	0.96	0.95	0.97	0.96	Mg
Zn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Zn
Ca	0.01	0.01	0.01	0.01	0.01	0.02	0.02	0.01	0.01	0.01	0.01	0.02	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	Ca
Na	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Na
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	K
F	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	F
Total Cat	4.02	4.02	4.01	4.01	4.01	4.01	4.01	4.01	4.01	4.01	4.01	4.00	4.02	4.02	4.00	4.01	4.01	4.01	4.01	4.01	4.01	Total Cat
Fe/(Fe+Mg)	0.53	0.52	0.52	0.52	0.52	0.53	0.52	0.52	0.52	0.52	0.52	0.52	0.52	0.52	0.52	0.52	0.51	0.51	0.51	0.51	0.51	Fe/(Fe+Mg)

Appendix 3.1. Analyses for compositional profiles (Fig. 3.1).

Sample No. Trav um Analysis Mineral	O4 306 b3opxtr opx	O5 0 a2grb grt	O5 63 a2grb grt	O5 84 a2grb grt	O5 105 a2grb grt	O5 126 a2grb grt	O5 146 a2grb grt	O5 167 a2grb grt	O5 188 a2grb grt	O5 209 a2grb grt	O5 230 a2grb grt	O5 251 a2grb grt	O5 272 a2grb grt	O5 292 a2grb grt	O5 313 a2grb grt	O5 334 a2grb grt	O5 397 a2grb grt	O5 418 a2grb grt	O5 438 a2grb grt	O5 501 a2grb grt	Sample No. Trav um Analysis Mineral
SiO2	50.65	37.84	37.60	37.84	37.72	37.95	37.85	37.76	37.99	38.08	37.87	37.96	37.90	37.93	38.11	39.44	37.61	37.85	37.98	37.63	SiO2
TiO2	0.07	0.05	0.00	0.05	0.04	0.05	0.02	0.05	0.06	0.02	0.02	0.09	0.04	0.11	0.05	0.01	0.01	0.02	0.03	0.05	TiO2
Al2O3	1.16	21.54	21.60	21.46	21.43	21.52	21.49	21.56	21.46	21.45	21.47	21.46	21.56	21.45	21.75	22.46	21.62	21.62	21.46	21.62	Al2O3
Cr2O3	0.05	0.08	0.03	0.02	0.00	0.09	0.00	0.03	0.04	0.02	0.04	0.00	0.04	0.00	0.02	0.01	0.00	0.01	0.00	0.02	Cr2O3
FeO	30.40	30.81	30.54	30.68	30.82	30.90	30.69	30.73	30.15	30.77	30.80	30.74	30.34	30.54	30.61	31.53	30.69	30.35	30.53	30.61	FeO
MnO	0.47	1.23	1.21	1.26	1.15	1.14	1.44	1.24	1.25	1.31	1.23	1.22	1.20	1.14	1.34	1.11	1.17	1.07	1.31	1.11	MnO
MgO	16.82	6.11	6.16	6.11	5.97	5.96	6.05	6.04	5.95	5.94	5.89	6.01	6.06	5.86	5.99	5.73	6.07	5.97	5.94	5.89	MgO
ZnO	0.05	0.00	0.10	0.00	0.00	0.01	0.08	0.06	0.00	0.00	0.13	0.00	0.00	0.00	0.06	0.12	0.07	0.00	0.00	0.15	ZnO
CaO	0.18	2.71	2.63	2.65	2.53	2.60	2.65	2.59	2.64	2.63	2.70	2.60	2.67	2.60	2.69	2.65	2.69	2.86	2.84	2.90	CaO
Na2O	0.01	0.04	0.03	0.00	0.01	0.00	0.00	0.01	0.01	0.05	0.01	0.02	0.00	0.01	0.00	0.03	0.01	0.02	0.02	0.00	Na2O
K2O	0.00	0.05	0.00	0.01	0.00	0.00	0.00	0.00	0.02	0.01	0.01	0.01	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.01	K2O
F																					F
Total Ox	99.86	100.45	99.89	100.07	99.66	100.22	100.28	100.06	99.54	100.27	100.17	100.12	99.79	99.64	100.62	103.10	99.92	99.78	100.10	99.98	Total Ox
No. Ox	6	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12	12	No. Ox
Si	1.97	2.97	2.97	2.98	2.98	2.98	2.98	2.97	3.00	2.99	2.98	2.99	2.99	2.99	2.98	3.01	2.97	2.98	2.99	2.97	Si
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	Ti
Al	0.05	1.99	2.01	1.99	2.00	1.99	1.99	2.00	2.00	1.99	1.99	1.99	2.00	2.00	2.01	2.02	2.01	2.01	1.99	2.01	Al
Cr	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Cr
Fe	0.99	2.02	2.02	2.02	2.04	2.03	2.02	2.02	1.99	2.02	2.03	2.02	2.00	2.02	2.00	2.01	2.02	2.00	2.01	2.02	Fe
Mn	0.02	0.08	0.08	0.08	0.08	0.08	0.10	0.08	0.08	0.09	0.08	0.08	0.08	0.08	0.09	0.07	0.08	0.07	0.09	0.07	Mn
Mg	0.97	0.71	0.72	0.72	0.70	0.70	0.71	0.71	0.70	0.70	0.69	0.71	0.71	0.69	0.70	0.65	0.71	0.70	0.70	0.69	Mg
Zn	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.01	0.01	Zn
Ca	0.01	0.23	0.22	0.22	0.21	0.22	0.22	0.22	0.22	0.22	0.23	0.22	0.22	0.22	0.23	0.22	0.23	0.24	0.24	0.25	Ca
Na	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Na
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	K
F	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	F
Total Cat	4.01	8.03	8.03	8.02	8.02	8.01	8.02	8.02	8.00	8.02	8.02	8.02	8.01	8.00	8.01	7.99	8.03	8.01	8.02	8.02	Total Cat
Fe/(Fe+Mg)	0.50	0.74	0.74	0.74	0.74	0.74	0.74	0.74	0.74	0.74	0.75	0.74	0.74	0.75	0.74	0.76	0.74	0.74	0.74	0.74	Fe/(Fe+Mg)

Appendix 3.1. Analyses for compositional profiles (Fig. 3.1).

Sample No. Trav um Analysis Mineral	O5 504 a2grt	O5 514 a2grt	O5 522 a2grt	O5 524 a2grt	O5 533 a2grt	O5 543 a2grt	O5 562 a2grt	O5 564 a2grt	O5 572 a2grt	O5 582 a2grt	O5 585 a2grt	O7 0 3.3trav opx	O7 47 3.3trav opx	O7 83 3.3trav opx	O7 121 3.3trav opx	O7 144 3.3trav opx	O7 166 3.3trav M2grt	O7 181 3.3trav M2grt	O7 196 3.3trav M2grt	O7 212 3.3trav M2grt	Sample No. Trav um Analysis Mineral
SiO2	37.55	38.27	38.00	37.99	38.29	38.18	38.11	37.98	37.95	38.51	38.18	48.52	48.24	48.49	48.65	48.78	37.18	37.16	37.12	37.12	SiO2
TiO2	0.00	0.00	0.01	0.00	0.00	0.01	0.00	0.02	0.02	0.05	0.01	0.05	0.00	0.06	0.06	0.04	0.04	0.00	0.02	0.06	TiO2
Al2O3	21.53	21.46	21.56	21.49	21.61	21.50	21.22	21.45	21.50	21.77	21.56	0.43	0.35	0.31	0.28	0.34	20.33	20.32	20.25	20.35	Al2O3
Cr2O3	0.00	0.00	0.00	0.00	0.01	0.02	0.00	0.00	0.00	0.01	0.04	0.00	0.00	0.03	0.00	0.00	0.01	0.06	0.00	0.01	Cr2O3
FeO	30.60	30.46	30.23	30.55	30.49	30.63	29.94	30.63	30.25	30.20	29.76	38.22	38.50	38.62	39.09	38.18	31.84	31.57	31.12	31.04	FeO
MnO	1.08	1.12	1.13	1.12	1.15	1.10	1.16	1.14	1.17	1.14	1.09	0.37	0.39	0.37	0.39	0.40	1.01	1.08	1.00	0.97	MnO
MgO	6.00	5.91	5.94	6.04	5.98	5.93	5.74	5.98	5.79	5.82	5.85	10.73	10.57	10.76	10.61	10.75	1.91	1.89	1.70	1.77	MgO
ZnO	0.10	0.07	0.06	0.00	0.00	0.00	0.00	0.05	0.03	0.00	0.05	0.63	0.68	0.67	0.68	0.55	7.61	7.50	7.87	7.81	ZnO
CaO	3.01	2.99	2.97	2.99	2.99	3.13	3.03	3.01	3.16	2.95	3.04	0.00	0.00	0.03	0.01	0.04	0.00	0.01	0.03	0.00	CaO
Na2O	0.00	0.00	0.05	0.00	0.04	0.02	0.03	0.00	0.00	0.03	0.02	0.01	0.00	0.00	0.02	0.01	0.00	0.00	0.00	0.01	Na2O
K2O	0.02	0.04	0.01	0.00	0.01	0.04	0.05	0.04	0.02	0.00	0.02	0.01	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.01	K2O
F																					F
Total Ox	99.90	100.31	100.02	100.18	100.56	100.57	99.27	100.29	99.90	100.48	99.62	98.96	98.75	99.33	99.77	99.08	99.92	99.58	99.11	99.12	Total Ox
No. Ox	12	12	12	12	12	12	12	12	12	12	12	6	6	6	6	6	12	12	12	12	No. Ox
Si	2.97	3.00	2.99	2.99	2.99	2.99	3.02	2.98	2.99	3.01	3.01	1.98	1.98	1.98	1.98	1.99	3.00	3.00	3.01	3.01	Si
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Ti
Al	2.00	1.98	2.00	1.99	1.99	1.99	1.98	1.99	2.00	2.00	2.00	0.02	0.02	0.01	0.01	0.02	1.93	1.94	1.94	1.94	Al
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Cr
Fe	2.02	2.00	1.99	2.01	1.99	2.01	1.98	2.01	1.99	1.97	1.96	1.31	1.32	1.32	1.33	1.30	2.15	2.13	2.11	2.10	Fe
Mn	0.07	0.07	0.08	0.07	0.08	0.07	0.08	0.08	0.08	0.08	0.07	0.01	0.01	0.01	0.01	0.01	0.07	0.07	0.07	0.07	Mn
Mg	0.71	0.69	0.70	0.71	0.70	0.69	0.68	0.70	0.68	0.68	0.69	0.65	0.65	0.65	0.64	0.65	0.23	0.21	0.21	0.21	Mg
Zn	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.03	0.03	0.03	0.03	0.02	0.66	0.65	0.68	0.68	Zn
Ca	0.25	0.25	0.25	0.25	0.25	0.26	0.26	0.25	0.27	0.25	0.26	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	Ca
Na	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Na
K	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	K
F	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	F
Total Cat	8.03	8.01	8.01	8.02	8.01	8.02	8.00	8.02	8.01	7.99	7.99	4.01	4.01	4.01	4.01	4.00	8.03	8.03	8.02	8.02	Total Cat
Fe/(Fe+Mg)	0.74	0.74	0.74	0.74	0.74	0.74	0.75	0.74	0.75	0.74	0.74	0.67	0.67	0.67	0.67	0.67	0.90	0.90	0.91	0.91	Fe/(Fe+Mg)

Appendix 3.1. Analyses for compositional profiles (Fig. 3.1).

Sample No. Trav um Analysis Mineral	O7 228 3.3trav M2grt	O7 247 3.3trav pl	O7 269 3.3trav pl	P13 0 a5sptr spl	P13 12 a5sptr spl	P13 24 a5sptr spl	P13 36 a5sptr spl	P13 47 a5sptr spl	P13 59 a5sptr spl	P13 71 a5sptr spl	P13 83 a5sptr spl	P13 94 a5sptr spl	P13 106 a5sptr spl	P13 118 a5sptr spl	P13 129 a5sptr spl	P13 153 a5sptr spl	P13 165 a5sptr spl	P13 176 a5sptr spl	P13 188 a5sptr spl	P13 200 a5sptr spl	P13 212 a5sptr spl	Sample No. Trav um Analysis Mineral
SiO ₂	36.96	46.20	48.01	0.02	0.00	0.04	0.02	0.00	0.01	0.01	0.00	0.02	0.00	0.03	0.00	0.00	0.01	0.01	0.00	0.01	0.00	SiO ₂
TiO ₂	0.03	0.00	0.03	0.03	0.00	0.05	0.05	0.01	0.00	0.01	0.03	0.00	0.00	0.00	0.01	0.00	0.02	0.03	0.02	0.01	0.02	TiO ₂
Al ₂ O ₃	20.55	33.76	32.39	57.03	57.19	57.58	57.10	56.81	56.45	56.54	50.39	57.22	57.21	57.71	56.81	56.74	56.66	57.08	56.10	54.93	56.13	Al ₂ O ₃
Cr ₂ O ₃	0.00	0.00	0.00	1.30	1.12	1.28	1.21	1.17	1.17	1.29	0.95	0.95	1.30	1.08	1.14	1.21	1.04	1.13	1.17	1.07	1.09	Cr ₂ O ₃
FeO	30.71	0.35	0.49	31.09	31.20	31.49	31.41	32.02	33.03	32.43	38.50	31.86	31.85	30.30	31.94	31.79	32.70	31.55	32.47	34.17	32.20	FeO
MnO	1.04	0.00	0.00	0.00	0.02	0.00	0.02	0.03	0.03	0.08	0.05	0.03	0.06	0.04	0.09	0.07	0.01	0.09	0.01	0.00	0.07	MnO
MgO	1.76	0.00	0.00	7.43	7.60	7.34	7.43	7.12	7.36	7.24	7.07	7.26	7.35	7.53	7.19	7.10	7.12	7.59	7.21	7.76	7.26	MgO
ZnO	8.03	17.42	15.50	1.43	1.39	1.18	1.30	1.30	1.64	1.28	1.27	1.24	1.42	1.41	1.30	1.40	1.30	1.20	1.31	1.33	1.20	ZnO
CaO	0.00	0.00	0.00	0.00	0.02	0.01	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.02	0.00	0.01	0.03	0.00	0.02	0.00	0.03	CaO
Na ₂ O	0.00	1.58	2.48	0.00	0.01	0.05	0.03	0.00	0.04	0.05	0.02	0.05	0.02	0.06	0.01	0.04	0.02	0.00	0.05	0.05	0.05	Na ₂ O
K ₂ O	0.00	0.03	0.06	0.01	0.00	0.01	0.00	0.01	0.00	0.00	0.01	0.02	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.00	K ₂ O
F																						F
Total Ox	99.09	99.34	98.96	98.33	98.55	99.03	98.56	98.47	99.73	98.95	98.31	98.64	99.21	98.17	98.49	98.34	98.91	98.69	98.35	99.33	98.04	Total Ox
No. Ox	12	8	8	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	4	No. Ox
Si	2.99	2.14	2.22	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Si
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Ti
Al	1.96	1.84	1.77	1.91	1.91	1.92	1.91	1.91	1.89	1.90	1.77	1.92	1.91	1.93	1.91	1.91	1.90	1.91	1.90	1.86	1.90	Al
Cr	0.00	0.00	0.00	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.02	0.02	0.03	0.02	0.03	0.03	0.02	0.03	0.02	0.02	0.02	Cr
Fe	2.08	0.01	0.02	0.74	0.74	0.74	0.75	0.76	0.78	0.77	0.96	0.76	0.75	0.72	0.76	0.78	0.78	0.75	0.78	0.82	0.77	Fe
Mn	0.07	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Mn
Mg	0.21	0.00	0.00	0.31	0.32	0.31	0.31	0.30	0.31	0.31	0.31	0.31	0.31	0.32	0.31	0.30	0.30	0.32	0.31	0.33	0.31	Mg
Zn				0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	Zn
Ca	0.70	0.86	0.77	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Ca
Na	0.00	0.14	0.22	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Na
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	K
F				0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	F
Total Cat	8.02	5.01	5.01	3.03	3.03	3.03	3.03	3.03	3.04	3.04	3.10	3.03	3.03	3.02	3.03	3.03	3.04	3.03	3.04	3.06	3.04	Total Cat
Fe/(Fe+Mg)	0.91			0.70	0.70	0.71	0.70	0.72	0.72	0.72	0.75	0.71	0.71	0.69	0.71	0.72	0.72	0.70	0.72	0.71	0.71	Fe/(Fe+Mg)

Appendix 3.1. Analyses for compositional profiles (Fig. 3.1).

Sample No.	P13	P13	P13	P13	P13	P13	P13	P13	P13	P13	P13	P13
Trav um	223	235	247	258	270	282	294	305	317			
Analysis	a5sptr	a5sptr	a5sptr	a5sptr	a5sptr	a5sptr	a5sptr	a5sptr	a5sptr			
Mineral	spl	spl	spl	spl	spl	spl	spl	spl	spl			
SiO2	0.00	0.01	0.02	0.03	0.03	0.04	0.00	0.00	0.00			
TiO2	0.05	0.03	0.02	0.00	0.03	0.03	0.11	0.13	0.28			
Al2O3	56.68	56.54	55.32	54.89	56.41	56.83	56.85	57.31	58.17			
Cr2O3	1.21	1.18	1.12	1.02	1.27	1.14	1.27	1.25	1.35			
FeO	32.07	32.23	33.66	33.51	32.43	31.92	31.15	30.51	29.08			
MnO	0.02	0.05	0.03	0.10	0.03	0.06	0.02	0.00	0.02			
MgO	7.32	7.44	7.25	7.30	7.42	7.50	7.68	7.93	8.52			
ZnO	1.19	1.22	1.24	1.35	1.33	1.46	1.33	1.27	1.41			
CaO	0.02	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.03			
Na2O	0.04	0.04	0.03	0.05	0.08	0.03	0.04	0.04	0.06			
K2O	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.01	0.00			
F												
Total Ox	98.60	98.75	98.70	98.25	99.01	99.00	98.43	98.46	98.92			
No. Ox	4	4	4	4	4	4	4	4	4			
Si	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00			
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01			
Al	1.90	1.90	1.87	1.87	1.89	1.90	1.91	1.91	1.92			
Cr	0.03	0.03	0.03	0.02	0.03	0.03	0.03	0.03	0.03			
Fe	0.76	0.77	0.81	0.81	0.77	0.76	0.74	0.72	0.68			
Mn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00			
Mg	0.31	0.32	0.31	0.31	0.31	0.32	0.33	0.33	0.36			
Zn	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.03			
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00			
Na	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00			
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00			
F	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00			
Total Cat	3.03	3.04	3.05	3.05	3.04	3.04	3.03	3.03	3.02			
Fe/(Fe+Mg)	0.71	0.71	0.72	0.72	0.71	0.70	0.69	0.68	0.66			

Appendix 3.2. Orthogneiss analyses used for geothermobarometry (Chapter 3)

Sample No.	O1	O1	O2	O2	O2	O2	O2	O3	O3	O3	O3	O3	O4	O4	O4	O4	O4	O4	O5	O5	Sample No.
Lithology	Bl-pxz o/g	Bl-pxz o/g	Chk o/g	Chk o/g	Chk o/g	Chk o/g	Chk o/g	Chk o/g	Chk o/g	Chk o/g	Chk o/g	Chk o/g	Chk o/g	Chk o/g	Chk o/g	Chk o/g	Chk o/g	Chk o/g	Chk o/g	Lithology	
Analysis	1a bt	1a opx	c11bt	c13cgt	c10opx	c10pl	c21lim	1.9.bbt	1.8gt	1.8opx	1.8pl	1.11lim	b2btb	b2gtc	b2opx	b2pl	b1lim	a3bt4	a2gtb	Analysis	
Mineral	M2gt	M2grt	bt	grt	opx	pl	lim	bt	grt	opx	pl	lim	bt	grt	opx	pl	lim	bt	grt	Mineral	
SiO2	35.03	49.75	36.34	38.07	50.23	55.84	0.04	35.63	37.77	49.88	56.28	0.02	36.32	37.26	50.24	53.27	0.03	35.62	37.72	SiO2	
TiO2	4.87	0.08	5.12	0.04	0.06	0.02	52.33	5.65	0.01	0.05	0.02	51.80	3.91	0.01	0.05	0.00	48.76	4.40	0.04	TiO2	
Al2O3	14.52	3.07	14.64	21.48	2.10	27.65	0.00	14.75	21.31	2.35	27.02	0.00	14.00	20.84	1.38	29.16	0.00	14.66	21.43	Al2O3	
Cr2O3	0.07	0.06	0.06	0.03	0.02	0.01	0.00	0.05	0.03	0.04	0.01	0.00	0.03	0.03	0.02	0.00	0.00	0.06	0.00	Cr2O3	
FeO	13.34	24.44	16.87	30.51	29.21	0.05	44.81	16.97	30.87	27.97	0.03	44.77	17.02	32.32	30.77	0.12	48.30	17.22	30.82	FeO	
MnO	0.07	0.96	0.01	1.17	0.36	0.03	0.30	0.03	1.08	0.32	0.01	0.40	0.04	1.30	0.46	0.00	0.27	0.05	1.15	MnO	
MgO	14.52	20.30	12.54	6.06	17.21	0.00	1.05	11.57	5.75	17.59	0.00	0.82	13.09	4.09	16.07	0.00	0.29	12.31	5.97	MgO	
ZnO			0.03	0.02	0.14	0.03	0.13						0.07	0.00	0.08	0.00	0.02	0.06	0.00	ZnO	
CaO	0.01	0.19	0.01	2.70	0.20	9.76	0.01	0.01	2.55	0.19	9.29	0.04	0.06	4.22	0.26	11.61	0.05	0.04	2.53	CaO	
Na2O	0.09	0.01	0.04	0.01	0.02	5.80	0.04	0.07	0.02	0.01	6.09	0.03	0.10	0.03	0.01	4.74	0.00	0.03	0.01	Na2O	
K2O	8.77	0.00	9.83	0.00	0.01	0.26	0.00	9.63	0.00	0.01	0.27	0.00	9.19	0.02	0.00	0.12	0.00	9.52	0.00	K2O	
F													1.56			0.10		1.14		F	
Total Ox	91.30	98.85	95.51	100.09	99.54	99.45	98.70	94.37	99.39	98.41	99.04	97.88	95.38	100.13	99.34	99.12	97.71	95.11	99.66	Total Ox	
No. Ox	24	6	24	12	6	8	3	24	12	6	8	3	24	12	6	8	3	24	12	No. Ox	
Si	5.93	1.90	5.99	2.99	1.94	2.52	0.00	5.96	3.00	1.94	2.55	0.00	5.99	2.98	1.96	2.43	0.00	5.91	2.98	Si	
Ti	0.62	0.00	0.64	0.00	0.00	0.00	1.00	0.71	0.00	0.00	0.00	1.00	0.48	0.00	0.00	0.00	0.96	0.55	0.00	Ti	
Al	2.90	0.14	2.85	1.99	0.10	1.47	0.00	2.91	1.99	0.11	1.44	0.00	2.72	1.96	0.06	0.00	0.00	2.87	2.00	Al	
Cr	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	Cr	
Fe	1.89	0.78	2.33	2.01	0.95	0.00	0.95	2.37	2.05	0.91	0.00	0.96	2.35	2.16	1.01	0.00	1.06	2.39	2.04	Fe	
Mn	0.01	0.03	0.00	0.08	0.01	0.00	0.01	0.00	0.07	0.01	0.00	0.01	0.01	0.09	0.02	0.00	0.01	0.01	0.08	Mn	
Mg	3.67	1.16	3.08	0.71	0.99	0.00	0.04	2.88	0.68	1.02	0.00	0.03	3.21	0.49	0.94	0.00	0.01	3.04	0.70	Mg	
Zn			0.00	0.00	0.00	0.00	0.00						0.01	0.00	0.00	0.00	0.00	0.01	0.00	Zn	
Ca	0.00	0.01	0.00	0.23	0.01	0.47	0.00	0.00	0.22	0.01	0.45	0.00	0.01	0.36	0.01	0.57	0.00	0.01	0.21	Ca	
Na	0.03	0.00	0.01	0.00	0.51	0.00	0.00	0.02	0.00	0.00	0.54	0.00	0.03	0.00	0.00	0.42	0.00	0.01	0.00	Na	
K	1.89	0.00	2.07	0.00	0.00	0.02	0.00	2.05	0.00	0.00	0.02	0.00	1.93	0.00	0.00	0.01	0.00	2.02	0.00	K	
F													0.81	0.00	0.00	0.01	0.00	0.60	0.00	F	
Total Cat	16.95	4.02	16.99	8.01	4.01	5.00	2.00	16.92	8.01	4.00	5.00	2.00	16.74	8.04	4.00	4.99	2.04	16.82	8.02	Total Cat	
Fe/(Fe+Mg)	0.34	0.40	0.43	0.74	0.49		0.96	0.45	0.75	0.47		0.97	0.42	0.82	0.52		0.99	0.44	0.74	Fe/(Fe+Mg)	

Appendix 3.2. Orthogneiss analyses used for geothermobarometry (Chapter 3)

Sample No.	O8 Pr m/ o/g	O8 bopx	O8 opx	O8 Pr m/ o/g	O8 bpl	O8 Pr m/ o/g	O8 bcpx	O8 Pr m/ o/g	O8 b10hbl	O9 Pr m/ o/g	O9 2.5qt	O9 grt	O9 Pr m/ o/g	O9 2.5opx	O9 opx	O9 Pr m/ o/g	O9 2.5pl	O9 pl	O9 Pr m/ o/g	O9 2.2ilm	O9 cpx	O10 Pr m/ o/g	O10 c7qt	O10 grt	O10 Pr m/ o/g	O10 c7opx	O10 opx	O10 Pr m/ o/g	O10 c7pl	O10 pl	O10 Pr m/ o/g	O10 c3ilq	O10 ilm	O10 Pr m/ o/g	O10 cpx	O10 hbl	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g	Pr m/ o/g

Appendix 3.3. Paragneiss analyses used for geothermobarometry (Chapter 3)

Sample No.	P1	P1	P1	P2	P2	P2	P2	P2	P2	P2	P3	P3	P3	P3	P4	P4	P4	P4	P4	Sample No.
Lithology	metapel	metapel	metapel	metapel	metapel	metapel	metapel	metapel	metapel	metapel	metapel	metapel	metapel	metapel	fels para	fels para	fels para	fels para	fels para	Lithology
Analysis	b3.1 b2	b3cd8	b3gt3	c5 cd7	c5 g8	c5 p17	a1 b14	a2 cd11	a2 g5	a2 p11	a1 b14	a2 cd11	a2 g5	a2 p11	b3.2 b3	b2 cd9	b3.c g12	b3.r g12	b3.r g12	Analysis
Mineral	bt	crd	grt	crd	grt	pl	bt	crd	grt	pl	bt	crd	grt	pl	bt	crd	grtcore	grttrm	grttrm	Mineral
SiO ₂	35.91	48.36	37.44	47.40	37.20	58.83	35.59	48.53	37.96	60.37	35.59	48.53	37.96	60.37	36.15	48.83	37.41	37.47	37.47	SiO ₂
TiO ₂	4.13	0.01	0.02	0.01	0.06	0.02	3.21	0.01	0.03	0.01	3.21	0.01	0.03	0.01	3.84	0.01	0.03	0.00	0.00	TiO ₂
Al ₂ O ₃	16.45	33.08	21.13	32.19	20.89	25.43	17.47	32.98	21.45	24.76	17.47	32.98	21.45	24.76	16.24	33.22	21.24	21.37	21.37	Al ₂ O ₃
Cr ₂ O ₃	0.16	0.02	0.00	0.01	0.01	0.01	0.06	0.01	0.01	0.02	0.06	0.01	0.01	0.02	0.10	0.02	0.03	0.05	0.05	Cr ₂ O ₃
FeO	14.97	5.92	32.58	27.42	31.88	0.05	13.80	5.29	30.46	0.06	13.80	5.29	30.46	0.06	13.70	5.59	32.47	32.10	32.10	FeO
MnO	0.02	0.03	0.74	0.06	0.02	0.06	0.02	0.05	0.06	0.01	0.05	0.06	0.06	0.01	0.02	0.03	0.69	0.61	0.61	MnO
MgO	13.22	9.83	6.07	9.85	5.81	0.00	13.65	10.21	7.27	0.00	13.65	10.21	7.27	0.00	13.84	10.24	6.33	6.34	6.34	MgO
ZnO	0.05	0.03	0.01	0.07	0.03	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.08	0.05	0.03	0.04	0.04	ZnO
CaO	0.00	0.00	0.93	0.01	0.02	7.04	0.00	0.01	0.97	6.25	0.00	0.01	0.97	6.25	0.00	0.01	0.81	0.77	0.77	CaO
Na ₂ O	0.12	0.04	0.01	0.07	0.01	7.33	0.04	0.04	0.01	7.93	0.04	0.04	0.01	7.93	0.11	0.04	0.01	0.01	0.01	Na ₂ O
K ₂ O	9.48	0.01	0.01	0.01	0.00	0.21	9.24	0.01	0.01	0.20	9.24	0.01	0.01	0.20	9.37	0.01	0.01	0.01	0.01	K ₂ O
F																				F
Total Ox	94.50	97.33	98.95	98.14	94.96	97.61	98.96	93.12	97.16	99.40	93.12	97.16	99.40	99.63	93.45	98.03	99.05	98.77	98.77	Total Ox
No. Ox	24	18	12	18	12	8	24	18	12	8	24	18	12	8	24	18	12	12	12	No. Ox
Si	5.91	4.98	2.99	4.99	3.01	2.65	5.90	4.99	2.99	2.70	5.90	4.99	2.99	2.70	5.98	4.98	2.98	2.99	2.99	Si
Ti	0.51	0.00	0.00	0.00	0.00	0.00	0.40	0.00	0.00	0.00	0.40	0.00	0.00	0.00	0.48	0.00	0.00	0.00	0.00	Ti
Al	3.19	4.01	1.99	3.99	1.99	1.35	3.41	4.00	1.99	1.30	3.41	4.00	1.99	1.30	3.16	3.99	2.00	2.01	2.01	Al
Cr	0.02	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	Cr
Fe	2.06	0.51	2.18	0.47	2.16	0.00	1.91	0.45	2.01	0.00	1.91	0.45	2.01	0.00	1.89	0.48	2.17	2.14	2.14	Fe
Mn	0.00	0.00	0.05	0.01	0.05	0.00	0.01	0.01	0.08	0.00	0.01	0.01	0.08	0.00	0.00	0.00	0.05	0.04	0.04	Mn
Mg	3.24	1.51	0.72	1.54	0.70	0.00	3.37	1.56	0.85	0.00	3.37	1.56	0.85	0.00	3.41	1.56	0.75	0.75	0.75	Mg
Zn	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	Zn
Ca	0.00	0.00	0.08	0.00	0.08	0.34	0.00	0.00	0.08	0.30	0.00	0.00	0.08	0.30	0.00	0.00	0.07	0.07	0.07	Ca
Na	0.04	0.01	0.00	0.03	0.01	0.64	0.01	0.01	0.00	0.69	0.01	0.01	0.00	0.69	0.04	0.01	0.00	0.00	0.00	Na
K	1.99	0.00	0.00	1.98	0.00	0.01	1.95	0.00	0.00	0.01	1.95	0.00	0.00	0.01	1.98	0.00	0.00	0.00	0.00	K
F																				F
Total Cat	16.98	11.02	8.01	11.02	7.99	5.00	16.98	11.02	8.01	5.00	16.98	11.02	8.01	5.00	16.96	11.03	8.02	8.01	8.01	Total Cat
Fe/(Fe+Mg)	0.39	0.25	0.75	0.43	0.76		0.36	0.23	0.70		0.36	0.23	0.70		0.36	0.23	0.74	0.74	0.74	Fe/(Fe+Mg)

Appendix 3.3. Paragneiss analyses used for geothermobarometry (Chapter 3)

[illegible]

Appendix 3.3. Paragneiss analyses used for geothermobarometry (Chapter 3)

Sample No. Lithology Analysis Mineral	P12 metapel b2spb spl	P13 metapel a7btb 7 bt	P13 metapel a9cd 3 crd	P13 metapel a3gb grt	P13 metapel afilm ilm	P13 metapel a3kf ksp	P13 metapel a3pl 7 pl	P13 metapel a5spa spl	P16 metapel c4 b2 bt	P16 metapel c4 cd5 crd	P16 metapel c1.1 gt2 grt	P16 metapel c4 ilm3 ilm	P16 metapel c1.1 pl4 pl	P16 metapel cf ghm5 spl	Sample No. Lithology Analysis Mineral
SiO ₂	0.03	36.62	48.79	38.23	0.00	63.94	61.56	0.00	35.06	46.35	37.12	0.00	63.37	0.33	SiO ₂
TiO ₂	0.00	4.50	0.00	0.00	48.06	0.04	0.02	0.28	4.19	0.01	0.02	52.07	0.01	0.01	TiO ₂
Al ₂ O ₃	54.75	15.52	32.99	21.71	0.00	18.91	24.02	58.17	16.24	31.51	20.87	0.00	22.13	53.42	Al ₂ O ₃
Cr ₂ O ₃	1.01	0.08	0.04	0.00	0.10	0.00	0.00	1.35	0.04	0.01	0.07	0.02	2.83	2.83	Cr ₂ O ₃
FeO	35.40	14.57	5.44	30.67	49.08	0.03	0.03	29.08	12.64	4.47	30.23	44.41	0.03	15.04	FeO
MnO	0.06	0.01	0.04	0.85	0.20	0.02	0.02	0.02	0.04	0.03	0.59	0.30	0.03	0.01	MnO
MgO	6.15	14.16	10.35	7.71	0.78	0.00	0.00	8.52	13.16	9.78	6.59	0.49	0.00	3.35	MgO
ZnO	2.21	0.02	0.04	0.00	0.00	0.00	0.02	1.41	0.05	0.02	0.05	0.09	0.00	17.95	ZnO
CaO	0.00	0.00	0.00	0.82	0.00	0.27	5.32	0.03	0.00	0.02	0.81	0.01	3.39	0.00	CaO
Na ₂ O	0.05	0.16	0.04	0.01	0.00	2.28	8.40	0.06	0.11	0.05	0.02	0.00	9.55	0.63	Na ₂ O
K ₂ O	0.02	9.51	0.00	0.01	0.00	13.20	0.28	0.00	8.78	0.01	0.01	0.01	0.15	0.02	K ₂ O
F		1.36	0.02			0.07	0.04								F
Total Ox	99.67	96.50	97.64	100.01	98.23	98.75	99.71	98.92	90.31	92.26	96.77	97.40	98.67	93.59	Total Ox
No. Ox	4	24	18	12	3	8	8	4	24	18	12	3	8	4	No. Ox
Si	0.00	5.89	4.99	2.99	0.00	2.97	2.74	0.00	5.96	5.00	3.01	0.00	2.83	0.01	Si
Ti	0.00	0.54	0.00	0.00	0.94	0.00	0.00	0.01	0.54	0.00	0.00	1.01	0.00	0.00	Ti
Al	1.86	2.94	3.97	2.00	0.00	1.03	1.26	1.92	3.25	4.01	1.99	0.00	1.17	1.94	Al
Cr	0.02	0.01	0.00	0.00	0.00	0.00	0.00	0.03	0.01	0.00	0.00	0.00	0.00	0.07	Cr
Fe	0.85	1.96	0.47	2.00	1.07	0.00	0.00	0.68	1.80	0.40	2.05	0.96	0.00	0.39	Fe
Mn	0.00	0.00	0.00	0.06	0.00	0.00	0.00	0.00	0.01	0.00	0.07	0.01	0.00	0.00	Mn
Mg	0.26	3.39	1.58	0.90	0.03	0.00	0.00	0.36	3.33	1.57	0.80	0.02	0.00	0.15	Mg
Zn	0.05	0.00	0.00	0.00	0.00	0.00	0.00	0.03	0.01	0.00	0.00	0.00	0.00	0.41	Zn
Ca	0.00	0.00	0.00	0.07	0.00	0.01	0.25	0.00	0.00	0.00	0.07	0.00	0.16	0.00	Ca
Na	0.00	0.05	0.01	0.00	0.00	0.20	0.72	0.00	0.04	0.01	0.00	0.00	0.83	0.04	Na
K	0.00	1.95	0.00	0.00	0.00	0.00	0.02	0.00	1.90	0.00	0.00	0.00	0.01	0.00	K
F	0.00	0.69	0.00	0.00	0.00	0.01	0.01	0.00							F
Total Cat	3.06	16.74	11.02	8.01	2.05	5.00	5.00	3.02	16.84	11.00	7.99	1.99	5.00	3.01	Total Cat
Fe/(Fe+Mg)	0.76	0.37	0.23	0.69	0.97			0.66	0.35	0.20	0.72	0.98		0.72	Fe/(Fe+Mg)

Appendix 3.4. THERMOCALC 2.3 sample output

Sample 05, Garnet-bearing charnockitic orthogneiss

Sample 05, Garnet-bearing charnockitic orthogneiss

an independent set of reactions has been calculated

average P_T (for a(H₂O) = 0.2)

a phl ann py gr alm en fs
sd(a)/a 0.0990 0.0354 0.0145 6.07e-4 0.277 0.283 0.184
0.28181 0.40614 0.51469 0.75405 0.15000 0.15427 0.19976

a mgt an q H₂O ksp
sd(a)/a 0.0182 0.597 1.00 0.200 1.00 0
0.54852 0.05000

reactions

- 1) 2phl + 3mgt + 6q = 3py + 2H₂O + 2ksp
- 2) 2phl + 3en + 6an = 4py + 2gr + 2H₂O + 2ksp
- 3) 2phl + 6an = py + 2gr + 3mgt + 2H₂O + 2ksp
- 4) 2ann + 3fs + 6an = 2gr + 4alm + 2H₂O + 2ksp
- 5) 2ann + 3an + 3q = gr + 2alm + 2H₂O + 2ksp

calculations
P(T) sd(P) a sd(a) b c ln_K sd(ln_K)

- 1 7.4 14.12 136.47 3.34 -0.17018 -1.286 3.935 2.326
- 2 4.7 1.66 191.57 3.57 0.02511 -12.510 -20.240 2.671
- 3 5.8 1.87 203.42 3.55 0.03625 -9.939 -23.338 2.378
- 4 4.7 1.10 42.44 3.97 0.06718 -13.955 -5.094 1.935
- 5 4.0 1.50 78.10 3.55 -0.04157 -6.359 -1.744 1.158

average P_T (for a(H₂O) = 0.2)

T = 700°C, sd = 47,
P = 4.9 kbars, sd = 1.4, cor = 0.440, f = 1.19

Sample 05, Garnet-bearing charnockitic orthogneiss

an independent set of reactions has been calculated

average pressures (for a(H₂O) = 0.5)
(suggested T = 800°C)

a phl ann py gr alm en fs
sd(a)/a 0.0990 0.0354 0.0145 6.07e-4 0.277 0.283 0.184
0.28181 0.40614 0.51469 0.75405 0.15000 0.15427 0.19976

a mgt an q H₂O ksp
sd(a)/a 0.0182 0.597 1.00 0.500 1.00 0
0.54852 0.05000

reactions

- 1) 2ann + 3fs + 6an = 2gr + 4alm + 2H₂O + 2ksp
- 2) 2ann + 2py + gr + 3q = 2phl + 3fs + 3an
- 3) 2phl + gr + 2alm + 3q = 2ann + 3en + 3an
- 4) 4ann + 3en + 6an = 2phl + 2gr + 4alm + 2H₂O + 2ksp

5) ann + gr + 3mgt + 3q = phl + alm + 3an

calculations

P(T) sd(P) a sd(a) b c ln_K sd(ln_K)

- 1 6.1 1.31 37.31 3.97 0.08451 -13.471 -5.094 1.935
- 2 6.6 2.27 -49.96 1.95 -0.08964 6.859 11.307 1.728
- 3 4.3 1.61 74.54 1.40 -0.12173 7.634 2.581 1.368
- 4 4.7 1.64 2.61 4.30 0.09322 -13.492 -4.326 2.428
- 5 4.8 4.33 -81.52 1.04 -0.08806 3.900 17.616 1.888

average pressures (for a(H₂O) = 0.5)

T°C 700 800 900
av P 4.6 6.0 7.6
sd 1.59 1.13 1.53
f 1.6 1.1 1.3

Sample 05, Garnet-bearing charnockitic orthogneiss

an independent set of reactions has been calculated

average temperatures (for a(H₂O) = 0.5)
(suggested P = 7.0 kbar)

a phl ann py gr alm en fs
sd(a)/a 0.0990 0.0354 0.0145 6.07e-4 0.277 0.283 0.184
0.28181 0.40614 0.51469 0.75405 0.15000 0.15427 0.19976

a mgt an q H₂O ksp
sd(a)/a 0.0182 0.597 1.00 0.500 1.00 0
0.54852 0.05000

reactions

- 1) 2phl + 6q = 3en + 2H₂O + 2ksp
- 2) 2phl + 3mgt + 6q = 3py + 2H₂O + 2ksp
- 3) 2ann + 6q = 3fs + 2H₂O + 2ksp
- 4) phl + alm = ann + py
- 5) 4phl + 3fs + 6an = 2ann + 4py + 2gr + 2H₂O + 2ksp

calculations

T(P) sd(T) a sd(a) b c ln_K sd(ln_K)

- 1 872 44 147.12 3.26 -0.14117 1.111 0.837 0.729
- 2 773 111 138.01 3.34 -0.15546 -1.446 3.935 2.326
- 3 732 62 112.64 3.77 -0.13242 1.047 1.606 1.009
- 4 789 409 46.21 0.60 -0.01303 0.375 -3.979 0.729
- 5 699 718 228.10 3.97 0.03065 -12.624 -21.009 2.982

average temperatures (for a(H₂O) = 0.5)

P 6.0 7.0 8.0
av T 802 820 837
sd 46 52 62
f 1.3 1.4 1.7

Sample 05, Garnet-bearing charnockitic orthogneiss

an independent set of reactions has been calculated

average PT (for a(H₂O) = 0.5)

a phl ann py gr alm en fs
sd(al)/a 0.0990 0.0354 0.0145 6.07e-4 0.277 0.283 0.184
 0.28181 0.40614 0.51469 0.75405 0.15000 0.15427 0.19976
a mgts an q H₂O ksp
sd(al)/a 0.0182 0.597 1.00 0.500 1.00 0
 0.54852 0.05000 0

reactions

- 1) 2phl + 3en + 6an = 4py + 2gr + 2H₂O + 2ksp
- 2) 2phl + 6an = py + 2gr + 3mgts + 2H₂O + 2ksp
- 3) 2phl + 9an = 3gr + 6mgts + 3q + 2H₂O + 2ksp
- 4) 2arm + 3fs = 2gr + 4alm + 2H₂O + 2ksp
- 5) 2arm + 3an + 3q = gr + 2alm + 2H₂O + 2ksp

calculations

	P(T)	sd(P)	a	sd(a)	b	c	ln_K	sd(ln_K)
1	5.9	1.66	191.57	3.57	0.04035	-12.510	-20.240	2.671
2	7.2	1.87	203.42	3.55	0.05149	-9.939	-23.338	2.378
3	6.7	2.20	236.90	3.94	0.15470	-14.265	-36.975	4.058
4	5.7	1.10	42.44	3.97	0.08242	-13.955	-5.094	1.935
5	6.4	1.50	78.10	3.55	-0.02634	-6.359	-1.744	1.158

average PT (for a(H₂O) = 0.5)

T = 806°C, sd = 59,

P = 5.9 kbars, sd = 1.6, cor = 0.486, f = 1.23

Sample 05, Garnet-bearing charnockitic orthogneiss

an independent set of reactions has been calculated

average pressures (for a(H₂O) = 0)

a py gr alm en fs mgts an q
sd(al)/a 0.0145 6.07e-4 0.277 0.283 0.184 0.0182 0.597 1.00
 0.51469 0.75405 0.15000 0.15427 0.19976 0.54852 0.05000

reactions

- 1) 2py + gr + 3q = 3en + 3an
- 2) gr + 3mgts + 3q = py + 3an
- 3) gr + 2alm + 3q = 3fs + 3an

calculations

	P(T)	sd(P)	a	sd(a)	b	c	ln_K	sd(ln_K)
1	3.9	1.77	-15.26	0.73	-0.09834	6.881	10.539	1.366
2	5.1	3.94	-36.62	0.85	-0.09975	4.277	13.637	1.888
3	6.8	1.21	39.84	1.54	-0.11302	7.612	3.350	1.020

average pressures

T°C 700 800 900
av P 5.3 6.2 7.1
sd 1.06 1.51 2.09
f 0.9 1.3 1.6

Sample 05, Garnet-bearing charnockitic orthogneiss

an independent set of reactions has been calculated

average temperatures (for a(H₂O) = 0)

a py gr alm en fs mgts an q
sd(al)/a 0.0145 6.07e-4 0.277 0.283 0.184 0.0182 0.597 1.00
 0.51469 0.75405 0.15000 0.15427 0.19976 0.54852 0.05000

reactions

- 1) gr + 2alm + 3q = 3fs + 3an
- 2) 2py + 3fs = 2alm + 3en
- 3) 3fs + 3mgts = py + 2alm

calculations

	T(P)	sd(T)	a	sd(a)	b	c	ln_K	sd(ln_K)
1	820	73	35.68	1.54	-0.10872	7.589	3.350	1.020
2	541	614	-57.21	1.45	0.01657	-0.696	7.189	1.313
3	736	2674	-69.00	1.59	0.00534	-3.263	10.287	1.850

average temperatures

P 7.0

av T 733

sd 124

f 1.5

Sample 05, Garnet-bearing charnockitic orthogneiss

an independent set of reactions has been calculated

average PT (for a(H₂O) = 0)

a py gr alm en fs mgts an q
sd(al)/a 0.0145 6.07e-4 0.277 0.283 0.184 0.0182 0.597 1.00
 0.51469 0.75405 0.15000 0.15427 0.19976 0.54852 0.05000

reactions

- 1) gr + 3mgts + 3q = py + 3an
- 2) gr + 2mgts + 3q = en + 3an
- 3) gr + 2alm + 3q = 3fs + 3an

calculations

	P(T)	sd(P)	a	sd(a)	b	c	ln_K	sd(ln_K)
1	5.7	3.16	-32.67	0.85	-0.10424	4.343	13.637	1.888
2	4.8	1.89	-29.59	0.74	-0.09954	5.199	12.604	1.348
3	4.6	0.99	34.65	1.54	-0.10765	7.600	3.350	1.020

average PT

T = 562°C, sd = 106,
P = 4.1 kbars, sd = 1.2, cor = 0.683, f = 0.63

Sample O8, Garnet-bearing mafic orthogneiss, Proclamation gneiss

Sample O8, Garnet-bearing mafic orthogneiss, Proclamation gneiss

an independent set of reactions has been calculated

average PT (for a(H₂O) = 0.1)

a di hed cats gr alm en fs
0.418 0.427 0.0150 0.0113 0.294 0.109 0.401
sd(a)/a 0.10153 0.10000 0.66467 0.54177 0.15000 0.23794 0.10000

a an ab parg q H₂O
0.839 0.293 0.00123 1.00 0.100
sd(a)/a 0.05000 0.14980 0.81787 0

reactions

- 1) 2parg + 6q = 2di + 2cats + 3en + 2ab + 2H₂O
- 2) 14di + 6an + 2parg = 8gr + 11en + 2ab + 2H₂O
- 3) 3gr + 2parg + 11q = 8di + 5an + 2ab + 2H₂O
- 4) 33hed + 18an + 6parg = 24di + 13gr + 11alm + 6ab + 6H₂O
- 5) 22hed + 6an + 2parg = 8di + 8gr + 11fs + 2ab + 2H₂O

calculations P(T) sd(P) a sd(a) b c ln_K sd(ln_K)

1 6.3 14.93 198.34 4.07 -0.16351 1.252 -5.818 2.255
2 5.4 3.25 105.75 6.18 0.26790 -13.897 -36.020 5.354
3 4.2 1.91 184.88 4.29 -0.37412 10.741 16.567 2.476
4 6.8 1.71 -36.68 18.86 0.60669 -47.244 -28.616 9.739
5 7.7 3.07 -74.26 8.91 0.37501 -14.348 -22.168 5.324

average PT (for a(H₂O) = 0.1)

T = 676°C, sd = 84,
P = 4.9 kbars, sd = 1.1, cor = 0.793, f = 0.88

Sample O8, Garnet-bearing mafic orthogneiss, Proclamation gneiss

an independent set of reactions has been calculated

average PT (for a(H₂O) = 0.5)

a di hed cats gr alm en fs
0.418 0.427 0.0150 0.0113 0.294 0.109 0.401
sd(a)/a 0.10153 0.10000 0.66467 0.54177 0.15000 0.23794 0.10000

a an ab parg q H₂O
0.839 0.293 0.00123 1.00 0.500
sd(a)/a 0.05000 0.14980 0.81787 0

reactions

- 1) 14di + 6an + 2parg = 8gr + 11en + 2ab + 2H₂O
- 2) 6an + 2parg = 7cats + gr + 4en + 2ab + 2H₂O
- 3) cats + 2fs = hed + alm
- 4) 9hed + 12an + 4parg = 16di + 13cats + 3alm + 4ab + 4H₂O
- 5) 9fs + 4an + 2parg + 2q = 8di + 6alm + 2ab + 2H₂O

calculations P(T) sd(P) a sd(a) b c ln_K sd(ln_K)

1 7.3 3.21 104.68 6.18 0.29679 -14.052 -36.020 5.524
2 5.1 5.10 277.04 4.66 0.01336 -8.092 -30.706 5.070
3 8.6 4.61 -24.09 0.82 0.00311 -1.272 3.948 0.717
4 6.4 4.16 433.02 9.77 0.01340 -18.552 -40.505 9.473
5 8.5 1.58 73.28 6.25 -0.01503 -12.231 5.557 2.255

average PT (for a(H₂O) = 0.5)

T = 800°C, sd = 160,
P = 6.1 kbars, sd = 2.1, cor = 0.831, f = 1.51

Sample O8, Garnet-bearing mafic orthogneiss, Proclamation gneiss

an independent set of reactions has been calculated

average pressures (suggested T = 800°C) (for a(H₂O) = 0)

a di hed gr alm en fs
0.418 0.427 0.0113 0.294 0.109 0.401
sd(a)/a 0.10153 0.10000 0.54177 0.15000 0.23794 0.10000

a an q
0.839 1.00
sd(a)/a 0.05000 0

reactions

- 1) gr + en + q = 2di + an
- 2) 2gr + alm + 3q = 3hed + 3an
- 3) gr + fs + q = 2hed + an

calculations P(T) sd(P) a sd(a) b c ln_K sd(ln_K)

1 5.7 2.51 7.36 0.57 -0.05855 2.243 4.781 0.628
2 7.1 1.43 56.15 1.54 -0.15960 7.237 7.120 1.144
3 7.9 2.32 24.16 0.82 -0.06871 2.285 3.523 0.588

average pressures

T°C 800
av P 6.4
sd 0.82
f 1.1

Sample O8, Garnet-bearing mafic orthogneiss, Proclamation gneiss

an independent set of reactions has been calculated

average temperatures (suggested P = 7.0 kbar) (for a(H₂O) = 0)

a di hed gr alm en fs
0.418 0.427 0.0113 0.294 0.109 0.401
sd(a)/a 0.10153 0.10000 0.54177 0.15000 0.23794 0.10000

a an q
0.839 1.00
sd(a)/a 0.05000 0

reactions

- 1) gr + 3fs = 3hed + alm
- 2) 2gr + alm + 3q = 3hed + 3an
- 3) 2di + fs = 2hed + en

calculations
T(P) sd(T) a sd(a) b c ln_K sd(ln_K)

1	506	118	17.19	1.36	-0.04743	-0.379	3.450	0.704
2	791	57	52.87	1.54	-0.15615	7.210	7.120	1.144
3	550	315	16.18	0.60	-0.00955	0.042	-1.257	0.385

average temperatures

P	6.0	7.0	8.0
av T	728	809	889
sd	60	81	103
f	0.9	1.4	1.8

Sample O8, Garnet-bearing mafic orthogneiss, Proclamation gneiss

an independent set of reactions has been calculated

average PT (for a(H2O) = 0)

a	0.418	bed	cats	py	gr	alm	en	fs
sd(a)/a	0.10153	0.10000	0.66467	0.75607	0.54177	0.15000	0.23794	0.10000

a	an	q
sd(a)/a	0.839	1.00
	0	0.05000

reactions

- 1) py + 2gr = 3di + 3cats
- 2) py + 2gr + 3q = 3di + 3an
- 3) gr + en + q = 2di + an
- 4) cats + 2fs = bed + alm
- 5) 2gr + alm + 3q = 3bed + 3an

calculations

P(T)	sd(P)	a	sd(a)	b	c	ln_K	sd(ln_K)
1	5.9	7.38	53.37	1.04	-0.08115	2.648	1.219
2	3.3	1.63	0.95	1.08	-0.13457	6.802	13.283
3	4.9	2.28	7.20	0.57	-0.05836	2.238	4.781
4	8.6	4.60	-24.12	0.82	0.00317	-1.275	3.948
5	5.8	1.30	53.75	1.54	-0.15715	7.219	7.120

average PT

T = 493°C, sd = 68.

P = 3.2 kbars, sd = 0.8, cor = 0.717, f = 0.45

Sample P5, Cordierite-free metapelite paragneiss

Sample P5, Cordierite-free metapelite paragneiss

an independent set of reactions has been calculated

average pressures (suggested T = 800°C) (for a(H2O) = 0.5)

a	phl	py	gr	an	ksp
sd(a)/a	0.0797	0.0290	3.50e-5	0.273	0.843
	0.30899	0.43210	0.91713	0.15868	0.05000

a	q	H2O	sill
sd(a)/a	1.00	0.500	1.00
	0		0

reactions

- 1) 2phl + 3an + 3q = 2py + gr + 2ksp + 2H2O
- 2) phl + 6an = py + 2gr + ksp + H2O + 3sill

calculations

F(T)	sd(P)	a	sd(a)	b	c	ln_K	sd(ln_K)
1	6.5	2.62	166.13	3.34	-0.05046	-5.221	-8.729
2	7.2	1.82	40.66	2.05	0.14825	-10.533	-13.908

average pressures (for a(H2O) = 0.5)

T°C	800
av P	7.4
sd	1.74
f	0.4

Sample P5, Cordierite-free metapelite paragneiss

an independent set of reactions has been calculated

average temperatures (suggested P = 7.0 kbar) (for a(H2O) = 0.5)

a	phl	arm	py	gr	alm	an	ksp
sd(a)/a	0.097	0.0323	0.0290	3.50e-5	0.249	0.273	0.843
	0.30899	0.41804	0.43210	0.91713	0.15000	0.15868	0.05000

a	q	H2O	sill
sd(a)/a	1.00	0.500	1.00
	0		0

reactions

- 1) phl + 2q + sill = py + ksp + H2O
- 2) arm + q + sill = alm + ksp + H2O
- 3) 2phl + 3an + 3q = 2py + gr + 2ksp + 2H2O

calculations

T(P)	sd(T)	a	sd(a)	b	c	ln_K	sd(ln_K)
1	775	51	98.61	1.65	-0.08259	-0.246	-1.184
2	617	53	52.40	1.76	-0.06956	-0.622	1.872
3	780	222	170.37	3.34	-0.05068	-5.768	-8.729

average temperatures (for a(H2O) = 0.5)

P	6.0	7.0	8.0
av T	729	736	742
sd	56	59	67
f	1.5	1.5	1.7

Sample P5, Cordierite-free metapelite paragneiss

an independent set of reactions has been calculated

average PT (for a(H2O) = 0.9)

a	phl	arm	py	gr	alm	an	ksp
sd(a)/a	0.0797	0.0323	0.0290	3.50e-5	0.249	0.273	0.843
	0.30899	0.41804	0.43210	0.91713	0.15000	0.15868	0.05000

```

a          q      H2O      sill
sd(a)/a    1.00  0.900  1.00  0

reactions
1) phl + 6an = py + 2gr + ksp + H2O + 3sill
2) 2ann + 3an + 3q = gr + 2alm + 2ksp + 2H2O
3) ann + 6an = 2gr + alm + ksp + H2O + 3sill

calculations
P(T)      sd(P)      a      sd(a)      b      c      ln_K      sd(ln_K)
1  7.5  1.74  43.13  2.05  0.15067 -10.513 -13.908  2.134
2  6.1  2.07  78.52  3.55 -0.01998 -15.870 -12.619  1.366
3  6.6  1.66 -2.19  2.13  0.16272 -10.862 -10.853  2.114

average PT (for a(H2O) = 0.9)
T = 799°C, sd = 66,
P = 7.0 kbars, sd = 2.4, cor = 0.236, f = 1.34
*****

Sample P5, Cordierite-free metapeliteic paragneiss
an independent set of reactions has been calculated
average PT (for a(H2O) = 0.5)
a          phl      ann      py      gr      alm      an      ksp
sd(a)/a    0.0797  0.0323  0.0254  3.5045  0.249  0.273  0.843
a          q      H2O      sill
sd(a)/a    1.00  0.500  1.00  0

reactions
1) phl + 6an = py + 2gr + ksp + H2O + 3sill
2) 2ann + 3an + 3q = gr + 2alm + 2ksp + 2H2O
3) ann + 6an = 2gr + alm + ksp + H2O + 3sill

calculations
P(T)      sd(P)      a      sd(a)      b      c      ln_K      sd(ln_K)
1  7.0  1.74  43.13  2.05  0.14578 -10.513 -13.908  2.134
2  4.4  2.07  78.52  3.55 -0.02976 -15.870 -12.619  1.366
3  6.2  1.66 -2.19  2.13  0.15783 -10.862 -10.853  2.114

average PT (for a(H2O) = 0.5)
T = 734°C, sd = 79,
P = 6.2 kbars, sd = 3.0, cor = 0.216, f = 1.81
*****

Sample P6, Cordierite-bearing felsic paragneiss
an independent set of reactions has been calculated
average pressures (suggested T = 800°C) (for a(H2O) = 0.5)
a          phl      ann      crd      fcrd      py      alm
sd(a)/a    0.146  0.0119  0.666  0.0338  0.0239  2.60e-5  0.277
a          an      ksp      q      H2O      sill
sd(a)/a    0.263  0.843  1.00  0.500  1.00  0

reactions
1) 4phl + 3crd + 3q = 6py + 4ksp + 4H2O
2) 5phl + 6crd = 9py + 5ksp + 5H2O + 3sill
3) 2ann + 9q + 6sill = 3fcrd + 2ksp + 2H2O
4) ann + 2q + sill = alm + ksp + H2O
5) 2phl + 3crd + gr = 4py + 3an + 2ksp + 2H2O

calculations
T(P)      sd(T)      a      sd(a)      b      c      ln_K      sd(ln_K)
1  662  106  428.90  6.87 -0.22045 -15.973 -14.179  2.898
2  637  176  563.16  9.01 -0.19321 -31.366 -22.411  4.281
3  768  52  159.70  3.68 -0.24824  15.052 -1.641  1.611
4  797  66  52.35  1.76 -0.06989 -0.571  2.978  0.559
5  660  102  259.61  3.82 -0.16990 -10.296 -3.663  2.199

```

average temperatures (for a(H2O) = 0.5)

P 6.0 7.0 8.0
av T 712 737 748
sd 20 19 22
f 0.7 0.7 1.2

Sample P6, Cordierite-bearing felsic paragneiss

an independent set of reactions has been calculated

average PT (for a(H2O) = 0.5)

a phl ann crd fcrd py gr alm
0.146 0.0119 0.666 0.0338 0.0239 2.60e-5 0.277
sd(a)/a 0.22890 0.53599 0.05000 0.39968 0.45628 1.01266 0.15000
a an ksp q H2O sill
0.263 0.843 1.00 0.500 1.00 0
sd(a)/a 0.16293 0.05000 0 0 0

reactions

1) 5phl + 6crd = 9py + 5ksp + 5H2O + 3sill
2) 2phl + 9q + 6sill = 3crd + 2ksp + 4H2O
3) 4ann + 3fcrd + 3q = 6alm + 4ksp + 2H2O
4) 5ann + 6fcrd = 9alm + 5ksp + 5H2O + 3sill
5) 2ann + 3fcrd + gr = 4alm + 3an + 2ksp + 2H2O

calculations

	P(T)	sd(P)	a	sd(a)	b	c	ln_K	sd(ln_K)
1	4.6	1.26	566.35	9.01	-0.22815	-29.874	-22.411	4.281
2	6.3	0.36	159.63	3.69	-0.28618	15.309	2.284	0.492
3	7.6	1.36	159.53	7.12	-0.19523	-17.261	19.507	2.624
4	7.4	1.01	159.82	9.06	-0.17311	-33.664	30.081	3.849
5	7.9	1.57	81.01	3.77	-0.16547	-11.391	20.103	2.055

average PT (for a(H2O) = 0.5)

T = 728°C, sd = 43,
P = 6.0 kbars, sd = 0.6, cor = 0.876, f = 0.85

Sample P13, Spinel-bearing metapelitic paragneiss

Sample P13, Spinel-bearing metapelitic paragneiss

an independent set of reactions has been calculated

average pressures (suggested T = 800°C) (for a(H2O) = 0.5)

a phl ann crd fcrd py gr alm
0.124 0.0239 0.641 0.0390 0.0281 1.50e-5 0.283
sd(a)/a 0.25143 0.45607 0.05000 0.38644 0.43617 1.21644 0.15000
a ksp an sp q
0.801 0.333 0.284 1.00
sd(a)/a 0.05000 0.13356 0.15388 0
H2O sill

a 0.500 1.00 0
sd(a)/a

reactions

1) 10phl + 9crd = 15py + 10ksp + 3sp + 10H2O
2) 5phl + 6crd = 9py + 5ksp + 5H2O + 3sill
3) 2sp + 5q = crd
4) 4ann + 3fcrd + 3q = 6alm + 4ksp + 4H2O
5) 5ann + 6fcrd = 9alm + 5ksp + 5H2O + 3sill
6) 2ann + 3fcrd + gr = 4alm + 2ksp + 3an + 2H2O

calculations

	P(T)	sd(P)	a	sd(a)	b	c	ln_K	sd(ln_K)
1	3.9	1.57	1087.83	17.52	-0.55806	-41.595	-34.766	7.056
2	4.4	1.29	563.41	9.01	-0.23176	-29.379	-20.188	4.140
3	8.2	0.80	-10.76	0.88	-0.03239	3.654	2.075	0.312
4	5.3	1.31	159.62	7.12	-0.19906	-16.910	16.210	2.350
5	6.0	0.99	160.16	9.06	-0.17973	-33.193	25.670	3.530
6	7.0	1.66	81.67	3.77	-0.16902	-11.199	19.517	2.046

average pressures (for a(H2O) = 0.5)

T°C 800

av P 6.7

sd 0.38

f 1.5

Sample P13, Spinel-bearing metapelitic paragneiss

an independent set of reactions has been calculated

average temperatures (suggested P = 7.0 kbar) (for a(H2O) = 0.5)

a phl ann crd fcrd py gr alm
0.124 0.0239 0.641 0.0390 0.0281 1.50e-5 0.283
sd(a)/a 0.25143 0.45607 0.05000 0.38644 0.43617 1.21644 0.15000

a ksp an sp q

0.801 0.333 0.284 1.00

sd(a)/a 0.05000 0.13356 0.15388 0

a H2O sill

0.500 1.00 0

sd(a)/a

reactions

1) 10phl + 9crd = 15py + 10ksp + 3sp + 10H2O
2) 4phl + 3crd + 3q = 6py + 4ksp + 4H2O
3) 5phl + 6crd = 9py + 5ksp + 5H2O + 3sill
4) 4ann + 3fcrd + 3q = 6alm + 4ksp + 4H2O
5) 2ann + 9q + 6sill = 3fcrd + 2ksp + 2H2O
6) 2phl + 3crd + gr = 4py + 2ksp + 3an + 2H2O

calculations

	T(P)	sd(T)	a	sd(a)	b	c	ln_K	sd(ln_K)
1	696	113	1090.65	17.52	-0.48067	-46.224	-34.766	7.056
2	692	100	430.56	6.87	-0.21629	-16.233	-12.661	2.815
3	671	170	565.29	9.01	-0.18482	-31.727	-20.188	4.140
4	569	112	153.97	7.12	-0.16478	-18.724	16.210	2.350
5	724	45	160.42	3.68	-0.25256	14.994	-2.712	1.479
6	730	101	260.18	3.82	-0.16541	-10.445	-1.432	2.229

average temperatures (for a(H2O) = 0.5)

p 7.0
av T 750
sd 19
f 1.0

Sample P13, Spinel-bearing metapelite paragneiss

an independent set of reactions has been calculated

average PT (for a(H2O) = 0.5)

a phl ann crd fcrd py alm
sd(a)/a 0.25143 0.45607 0.05000 0.38644 0.43617 1.21644 0.15000

a ksp an sp q
sd(a)/a 0.801 0.333 0.284 1.00 0

a H2O sill
sd(a)/a 0.05000 0.13356 0.15388 0

a H2O sill
sd(a)/a 0.500 1.00 0

reactions

- 1) 10phl + 9crd = 15py + 10ksp + 3sp + 10H2O
- 2) 5phl + 6crd = 9py + 5ksp + 5H2O + 3sill
- 3) 2sp + 5q = crd + 3ksp + 4H2O
- 4) 4ann + 3fcrd + 3q = 6alm + 4ksp + 4H2O
- 5) 5ann + 6fcrd = 9alm + 5ksp + 5H2O + 3sill
- 6) 2ann + 3fcrd + gr = 4alm + 2ksp + 3an + 2H2O

calculations

P(T) sd(P) a sd(a) b c ln_K sd(ln_K)

1	5.5	1.47	1092.96	17.52	-0.55220	-42.552	-34.766	7.056
2	5.4	1.22	566.35	9.01	-0.22815	-29.874	-20.188	4.140
3	8.2	0.76	-10.67	0.88	-0.03324	3.679	2.075	0.312
4	5.7	1.23	159.53	7.12	-0.19523	-17.261	16.210	2.350
5	6.1	0.93	159.82	9.06	-0.17311	-33.664	25.670	3.530
6	7.3	1.56	81.01	3.77	-0.16547	-11.391	19.517	2.046

average PT (for a(H2O) = 0.5)

T = 755°C, sd = 49,
P = 6.3 kbars, sd = 0.6, cor = 0.840, f = 1.19

Sample P11, Corundum-bearing metapelite paragneiss

Sample P11, Corundum-bearing metapelite paragneiss

an independent set of reactions has been calculated

average PT (for a(H2O) = 0.9)

a phl ann crd fcrd py alm
sd(a)/a 0.28109 0.45190 0.05081 0.35487 0.48484 0.15000

4 5.2 0.80 -63.28 1.99 -0.11017 16.986 10.132 1.585
5 4.9 1.19 -54.28 3.54 -0.08047 22.877 2.746 3.160

average PT (for a(H2O) = 0.5)

T = 672°C, sd = 101,
P = 5.0 kbars, sd = 0.8, cor = 0.891, f = 1.10

Sample P11, Corundum-bearing metapelitic paragneiss

an independent set of reactions has been calculated

average temperatures (suggested P = 7.0 kbar) (for a(H2O) = 0.5)

a phl ann crd fcrd py alm
sd(a)/a 0.0996 0.0247 0.588 0.0533 0.0188 0.299
0.28109 0.45190 0.05081 0.35487 0.48484 0.15000

a sp cor H2O sill ksp
sd(a)/a 0.291 1.00 0.500 1.00 0.840
0 0.05000

reactions

- 1) 10phl + 9crd = 15py + 3sp + 10H2O + 10ksp
- 2) 3phl + 3crd = 5py + cor + 3H2O + 3ksp
- 3) 5phl + 6crd = 9py + 5H2O + 3sill + 5ksp
- 4) 3ann + 3fcrd = 5alm + cor + 3H2O + 3ksp
- 5) 2ann + 15sill = 3fcrd + 9cor + 2H2O + 2ksp

calculations

	T(P)	sd(T)	a	sd(a)	b	c	ln_K	sd(ln_K)
1	677	121	1093.64	17.52	-0.46469	-46.888	-37.191	7.839
2	660	145	333.63	5.28	-0.12683	-16.340	-11.874	2.576
3	644	188	566.91	9.01	-0.17450	-32.076	-21.908	4.601
4	354	162	101.60	5.39	-0.08651	-18.442	13.331	1.886
5	827	43	167.42	3.84	-0.24229	13.563	-1.744	1.400

average temperatures (for a(H2O) = 0.5)

P 7.0
av T 785
sd 21
f 1.0

Sample P11, Corundum-bearing metapelitic paragneiss

an independent set of reactions has been calculated

average PT (for a(H2O) = 0.5)

a phl ann crd fcrd py alm
sd(a)/a 0.0996 0.0247 0.588 0.0533 0.0188 0.299
0.28109 0.45190 0.05081 0.35487 0.48484 0.15000

a sp herc cor H2O sill
sd(a)/a 0.291 0.589 1.00 0.500 1.00
0 0.15077 0.10000 0 0

reactions

- 1) 5py + 10cor = 3crd + 9sp
- 2) phl + alm = ann + py
- 3) 3ann + 5py + 10cor = 3phl + 3crd + 9herc
- 4) ann + 3py + 6sill = phl + 3crd + 3herc
- 5) 2ann + 5py + 10cor = 2phl + 3fcrd + 9sp

calculations

	P(T)	sd(P)	a	sd(a)	b	c	ln_K	sd(ln_K)
1	4.4	1.06	-53.61	3.64	-0.10718	22.641	7.164	2.782
2	6.4	17.04	45.32	0.60	-0.01205	0.369	-4.157	0.735
3	5.8	1.19	-124.20	3.62	-0.14996	22.010	17.687	3.043