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**TECTONOMETAMORPHIC INVESTIGATION OF THE
UVAUK COMPLEX, NUNAVUT, CANADA**

by

ANDREA J. MILLS, B.Sc. (Hons.)

A thesis submitted to
the Faculty of Graduate Studies and Research
in partial fulfillment of
the requirements for the degree of
Master of Science

Department of Earth Sciences

Carleton University

Ottawa, Ontario

May, 2001

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Chairman, Department of Earth Sciences

Thesis Supervisor

Thesis Supervisor

Carleton University

May 2001

ABSTRACT

The Uvauk complex (UC) comprises granulite-facies rocks with the geometry of an east-west trending, doubly-plunging synform cored by anorthosite and bound by ultramylonites. U-Pb geochronology indicates that some anorthosite is ca. 2711 Ma in age, and that tectonometamorphic events occurred within the UC at ca. 2.62-2.60 Ga, 2.53-2.50 Ga (~8.6 kbar, 730 °C), 1.94-1.91 Ga (11.4-14.3 kbar, 720-865 °C), and 1.83-1.75 Ga (~6 kbar, 625-695 °C). Similar conditions (8.1-9.2 kbar, 705-770 °C at ca. 2.5 Ga; ~11 kbar, 690-755 °C at ca. 1.9 Ga) are recorded by surrounding rocks. Clockwise calculated P-T paths suggest at least ~10-15 km of crustal thickening associated with ca. 1.9 Ga metamorphism (the extent of ca. 2.5 Ga exhumation is unconstrained). Gabbro dykes were synkinematically emplaced during granulite-facies mylonitization at ca. 1.9 Ga. Metamorphic data suggest that some anorthosite may be similar in age. Post-tectonic garnet breakdown occurred during ca. 1.83 Ga granite plutonism.

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ORIGINAL CONTRIBUTIONS

During the summer of 1998, I devoted six weeks to the mapping of the westernmost part of the Uvauk complex. In addition, areas northwest and east of the 1998 map area were mapped in 1999, and the data and observations from these three areas were compiled in the production of a 1:100 000 scale regional map by Tella et al. (in press) of the MacQuoid Lake-Gibson Lake-Akunak Bay Area, a product of work conducted for the 1997-2001 Western Churchill NATMAP Project. Lithological and structural mapping by the author were conducted at a 1:25 000 scale (see figures in Appendix I), and samples were collected for geochronology and geothermobarometry. I cut and polished thin sections and these were ground and mounted by Mike Jackson at Carleton University. I completed thin section studies, and generated SEM images with the help of Pat Hunt at the Geological Survey of Canada. Points for Electron Microprobe analyses were selected by the author and the location of these points was entered into computer software that allowed for automated analyses to be gathered by Katherine Venance. I used the TWEEQU software (Berman, 1988; 1991) for geothermobarometric analyses. Samples for geochronology were crushed and ground by Jerry Gagnon and Ron Christie at the Geological Survey of Canada, and magnetic separation, picking of mineral fractions, and abrading were completed by the author. U-Pb and Th-Pb chemistry was completed by U-Pb geochemists (Jack McRae, Dianne Bellerive, and Carole Lafontaine) at the Geological Survey of Canada, and mass spectrometry was performed by Carole Lafontaine, also at the Geological Survey of Canada.

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Chapter 1: Introduction

1.1 The Western Churchill Province

The Western Churchill Province (WCP) lies between the Slave Province to the northwest, and the Superior Province to the southeast (Figure 1.1). It is underlain by Paleo- to Neoarchean (ca. 3.5 to 2.6 Ga) deformed orthogneisses, northeast trending belts of Neoarchean (ca. 2.9 to 2.6 Ga) metavolcanic and metasedimentary rocks, and Paleoproterozoic plutonic and sedimentary rocks (Hoffman, 1988; Tella et al., 2000). Paleoproterozoic rocks include 1850 to 1810 m.y. old Hudson Suite monzonites (Peterson et al., 2000), 1765 to 1750 m.y. old Nuelin Suite granites (Loveridge et al., 1988; Peterson et al., 2000) and the ca. 2.45 to <1.91 Ga platformal Hurwitz and Amer groups (Aspler et al., 2000; Davis et al., 2000b). Unlike its neighbours, the Superior and the Slave provinces, the WCP is characterised by extensive Paleoproterozoic reworking and reactivation of older discrete structures related to Paleoproterozoic collisions between the WCP and the Slave Province (ca. 1.9-2.0 Ga Thelon orogen) and Superior Province (ca. 1.8 Ga Trans-Hudson orogen) (Hoffman, 1988).

The WCP is transected by an anastomosing, northeast-trending, geophysical lineament known as the Snowbird Tectonic zone (STZ) (Figure 1.1; Hoffman, 1988). This lineament is evident on both horizontal gravity gradient and magnetic anomaly maps of Canada (Walcott and Boyd, 1971; Gibb and Halliday, 1974; Sharpton et al., 1987) and extends 2000 km from Hudson Bay to the foothills of the Rocky Mountains. Discontinuous occurrences of granulite-facies rocks and fault zones coincide with the geophysical expression of the STZ (e.g.: Lewry and Sibbald, 1977; Tella and Eade, 1986; Hanmer et al., 1994; Hanmer, 1997). The STZ divides the Western Churchill Province

into two domains: the Rae and Hearne subprovinces (Hoffman, 1988). Evidence for older continental crust in the Rae domain includes platform-type sequences. Their absence in the Hearne domain, in conjunction with commonly-occurring rocks of oceanic affinity, support this subdivision of the WCP. Preliminary petrologic and isotopic studies support this division, and demonstrate that the Hearne province is characterized by predominantly late Archean, juvenile oceanic crust, whereas the Rae domain is underlain by more evolved continental crust (Sandeman et al., 2000).

The significance of the STZ remains controversial due, primarily, to conflicting interpretations of geochronological constraints on the timing of deformation and metamorphism along the length of the STZ. Two proposed interpretations are that: 1) the STZ represents a suture zone (Gibb and Haliday, 1974; Hoffman, 1988; Ross et al., 1995), and 2) the STZ marks a lithologic-thermotectonic transition (Lewry and Sibbald, 1980; Crocker et al., 1993; Bickford et al., 1994). Suggestions that this feature represents a tectonic suture were first made in the early 1970's, based solely on geophysical evidence (Walcott and Boyd, 1971; Gibb and Haliday, 1974). Hoffman (1988) suggested that the STZ represents a Paleoproterozoic suture. He further bracketed the age of collision between 1.92 Ga, the age of the Taltson magmatic zone, which appears truncated in the subsurface by the STZ, and 1.85 Ga, the age of the Baker Lake basin, which onlaps the central part of the STZ (Hoffman, 1988). However, detailed structural mapping and timing constraints from U/Pb geochronology studies led Hanmer et al. (1995a and b) to conclude that a significant portion of the STZ (i.e. the Striding-Athabasca segment) was located in the interior of the western Churchill continent by Late Archean, a conclusion that does not preclude the possibility of mid-Archean suturing. In contrast, Lewry and

Sibbald (1980) have interpreted the STZ to represent a remobilization front that marks the limit of deformation related to the Trans-Hudson orogen to the southeast. Similarly, Loveridge et al. (1988) viewed the STZ as a crustal reactivation front, related to collision of the Slave Province to the northwest and the Superior Province to the southeast. In this scenario, the STZ represents an overlap region of thermotectonically reworked crust overriding a northwestern, southeast-subducting and a southeastern, northwest-subducting plates.

Within the WCP, regional tectonometamorphic events are interpreted to have occurred at >2.92 Ga, ~2.68 Ga, ~2.60 Ga, ~2.55 Ga, ~1.90 Ga, ~1.83 Ga and ~1.75 Ga (Hanmer et al., 2000, and references within). Many of these correlate to significant magmatic events (e.g. ca. 2.68 Ga volcanism and plutonism, ~1.83 Ga Hudson Suite and 1.75 Ga Nuelin Suite granites; Hanmer et al., 2000; Peterson et al., 2000). A magmatic correlation is not evident for all events (e.g. 2.5 Ga). Previous division of the WCP into the Rae and Hearne domains, separated by the STZ (Hoffman, 1988), has recently been refined (Davis et al., 2000a; Hanmer et al., 2000). Evidence for a strong ca. 2.5-2.55 Ga tectonometamorphic overprint along a “central corridor” through the WCP (Figure 1.1), has contributed to the proposal of a new, tri-partite division of the WCP (Hanmer et al., 2000). Constituents of this division include: a southern block, characterized by juvenile crust (Sandeman et al., 2000) and an entirely Archean (pre-2.68 Ga) deformation history (Hanmer et al., 2000); a northern block comprising more evolved continental crust where ca. 2.60 Ga deformation appears to be well preserved; and a “central corridor”, or northern Hearne, extending at least from the Yathkyed Lake area to the MacQuoid Lake area (Figure 6.1), which is characterized by ca. 2.5 Ga deformation and metamorphism,

with common, though not ubiquitous, overprinting by ca. 1.9 Ga metamorphism (Berman et al., 2000).

New magnetotelluric and seismic studies show evidence for a major crustal boundary beneath the Chesterfield Inlet segment of the STZ (Jones et al., 2000). One interpretation involving underthrusting of Rae crust beneath the northern Hearne is plausible, although sound evidence is lacking (Jones et al., 2000). Significantly, Jones et al. (2000) highlight major crustal differences farther to the north, between crust northwest and southeast of a northern extension of the northeast-trending segment of the STZ (Figure 1.1), suggesting that the northern (Chesterfield Inlet) segment may not be linked to the STZ.

Evidence for ca. 2.5 Ga metamorphism and deformation, characteristic of the northern Hearne domain, has been documented for the Southern shear zone south of the KC (Sanborn-Barrie, 1999), for the Big Lake shear zone south of the Cross Bay complex (Ryan et al., 2000), for Yathkyed Lake (MacLachlan et al., 2000), and for the MacQuoid-Parker lakes area (Stern and Berman, 2001; Berman et al., 2000). Berman et al. (2000) suggest that this ca. 2.5 Ga event may correspond to a major collisional event during amalgamation of the Late Archean Kenorland supercontinent, and that the effects of this event may extend as far south as the Wollaston domain of NE Saskatchewan and northeast to northern Baffin Island.

Evidence for a tectonothermal event at ca. 1.9 Ga occurs predominantly within the central corridor, where Rae crust is interpreted to underlie Hearne crust (Jones et al., 2000). In the Parker and MacQuoid lakes areas, two thermotectonic events have been identified and are characterized by distinct P-T conditions (Berman et al., 2000). The

first is characterized by the presence of large (1-3 cm), low CaO ($X_{Ca} \sim 0.03$) garnet porphyroblasts that are wrapped by the main regional D2 foliation, and yield P-T conditions of ~ 4.6 - 5.6 kbar and ~ 630 - 690°C . In contrast, the second is marked by micro-porphyroblasts (<0.1 mm diameter) and thin (<0.1 mm) inclusion-rich overgrowths with markedly higher CaO ($X_{Ca} \sim 0.08$ - 0.11), and yields P-T conditions of 10.3 - 10.6 kbar and 660 - 690°C . The age for the earlier event is interpreted to be ca. 2.49 - 2.56 Ga, as indicated by U/Pb dating of monazite inclusions in garnet porphyroblasts (Stern and Berman, 2001). Ca. 1.9 Ga matrix monazite was interpreted to represent the age of the second, high-pressure metamorphic event. In addition, preservation of only the high-P metamorphism in ca. 2.19 Ga dykes is consistent with high-P metamorphism at ca. 1.9 Ga (Berman et al., 2000). Ca. 1.9 Ga magmatism, metamorphism and deformation are interpreted to have occurred at ~ 12 kbar and 800°C at the Kramanituvar complex (Figure 1.2) and within the Cross Bay structure (Sanborn-Barrie, 1999; Berman et al., 2000).

Several high-grade complexes, many of which contain high-pressure granulite-facies mylonitic rocks, outcrop along the STZ. Three such complexes are exposed along the Striding-Athabasca segment, and comprise mafic to intermediate middle Archean rocks that are bounded by high-grade ribbon mylonites. Geochronological studies and field relations conducted on the southernmost of these complexes indicate that deformation within the segment occurred during mid-Archean (3.13 Ga) and late Archean (2.62 - 2.60 Ga) times (Hanmer et al., 1995b). In contrast, Paleoproterozoic ages have been documented for movement along the northern segment of the STZ (Figure 1.2; Sanborn-Barrie, 1999; Gordon, 1988).

The northern, or Chesterfield Inlet, segment of the STZ is characterized by a number of complexes that share similar lithologic, structural and metamorphic features. These lie between Baker Lake and Hudson Bay (Figure 1.2), and include the Daly Bay complex (Gordon, 1988), the Kramanituar complex (Sanborn-Barrie et al., 1997; Sanborn-Barrie, 1999), the Uvauk complex (Tella et al., 1993; Mills et al., 1999), the Hanbury Island shear zone (Tella and Annesley, 1988) and disparate anorthosite exposures on Walrus and Coats Islands. It has been suggested that these complexes may collectively represent an east-trending, stacked succession of gneisses, termed the Aqxaameq gneisses (includes those occurring along the northern segment of the STZ only; Figure 1.1 and 1.2; Schau and Tella, 1993). Sanborn-Barrie (1999) and Berman et al. (2000) suggest that magmatism and metamorphism along the northern segment of the STZ may have been related to formation of the Thelon orogen. Alternatively, Sanborn-Barrie (1999) further suggests that the complexes may have formed in response to migration of a mantle plume. This is based on the inferred e-MORB mantle-source for the Kramanituar complex (Sanborn-Barrie, 1999), in conjunction with the apparent younging of mafic magmatism and granulite metamorphism from east to west (ca. 2067-1950 Ma at Daly Bay, Gordon, 1988; ca. 1910-1902 Ma at Kramanituar complex, Sanborn-Barrie, 1999), with the centrally located Uvauk complex giving an intermediate age for granulite metamorphism (ca. 1920-1940 Ma, Tella et al., 1994b). This study is an investigation of a highly strained granulite-facies metamorphic suite known as the Uvauk complex (UC) (Tella et al., 1994a), named after Uvauk Inlet on the north shore of Chesterfield Inlet (Figure 1.2). This complex is lithologically and structurally similar to other high-strain zones along the northern segment of the STZ.

1.2 Previous work

The Uvauk complex comprises anorthosite, gabbroic anorthosite, leucogabbro, diorite, gabbro and quartzofeldspathic gneiss that have been strongly deformed and metamorphosed at granulite facies (Tella et al., 1994a and b). Regional mapping and preliminary geochronological studies by Tella et al. (1994b) indicated that the UC comprises rocks of two structurally juxtaposed granulite-facies mylonitic events: a late Archean (ca. 2.59 Ga) event and an Early Proterozoic (ca 1.94 Ga) event. Thermobarometric studies were interpreted to indicate that Archean metamorphism occurred at pressures of >10 kb (30 km) whereas the latter event correlates to exhumation of 6-9 km (2-3 kb) (Tella et al., 1994b). Tella, (1994b) suggested that the Uvauk complex (including both anorthosite and related rocks and the granulite-facies mylonites to the south) represents an allochthonous Archean block that was emplaced onto lower grade quartzofeldspathic gneisses in the late Archean. They further suggest that the Rae-Hearne boundary in this part of the WCP represents a significant Archean tectonic feature that has been overprinted by Early Proterozoic ductile deformation.

1.3 Scope and objectives of this study

This thesis was carried out within the framework of the 1997 - 2002 Western Churchill NATMAP project, a collaborative venture involving the Geological Survey of Canada, the Government of the Northwest Territories and Indian and Northern Affairs Canada. Bedrock mapping of the WCP, conducted at 1:50 000 scale, in conjunction with 1:125 000 compilation, provides a framework for extensive geological research, including important contributions from a variety of specialised fields including geochemistry, geochronology, petrology and metallogeny.

During the summer of 1998, six weeks were devoted to mapping the westernmost region of the UC, on the north side of Chesterfield Inlet in Nunavut (area A of Figure 1.3). Ten days and two weeks in 1999 were spent mapping areas B and C, respectively. Fieldwork included mapping each area at a scale of 1:25 000, and collection of samples for thermobarometry, geochronology and geochemistry studies. Although area A is the main focus of this study, reference to area C is necessary as it comprises similar rock types, and the geometry and orientation of structural fabrics in the area have important regional implications.

The purpose of this study is to reconstruct the tectonometamorphic history of this segment of the Snowbird tectonic zone by providing a detailed map of critical portions of the UC, documenting variations in structures and metamorphic assemblages, and characterizing pressures and temperatures. U/Pb geochronological studies of selected rocks from the UC were conducted to further constrain the timing of magmatism, deformation and metamorphism in the area. Major questions that will be addressed are:

- 1) the nature of multiple metamorphic and deformation events;
- 2) differences in P-T conditions recorded by different rocks within the complex;
- 3) timing and magnitude of displacement across the bounding shear zones to the complex.

1.4 General geological setting of the Uvauk complex

The UC comprises predominately anorthosite, anorthositic gabbro and mafic granulite rocks, which will be referred to herein as the anorthosite suite. The UC also includes highly strained rocks (ultramylonites) north and south of the anorthosite suite (as defined by Tella et al., 1994a and b), herein referred to as the Northern and Southern

bounding shear zones (refer to Figure 3.1). These rocks are predominantly quartzofeldspathic and are interpreted to have been deformed and metamorphosed at granulite-facies during one or more events. The complex is highly strained; the strain gradient decreases to the north and south, and is highly variable within the anorthosite suite. Granulite-facies assemblages are commonly preserved within the complex and in rocks north of the complex. However, much of the UC has been affected by pervasive amphibolite-facies retrograde metamorphism interpreted to have occurred after deformation (see Chapter 2). Granulite-facies assemblages are best preserved in the shear zone south of the anorthosite suite.

The geometry of the UC resembles an east-west oriented belt that structurally overlies predominantly tonalitic country rocks. In the easternmost part of the UC (i.e. area C, Figure 1.3), the anorthositic suite is bounded at its structural base by a nearly flat ($\sim 10\text{-}15^\circ$ west-dipping) shear zone. When considered in conjunction with the geometry of the western part of the UC (i.e. steeply south-dipping shear zones and rock units), several questions are raised: a) are the Northern and Southern bounding shear zones separate shear zones, or are they part of the same shear zone that has been folded? b) are the fabrics within the UC parallel to those in the surrounding rocks or do they differ? Examination of the conditions, timing and magnitude of displacement along the shear zones in the western part of the UC will help to address these questions.

Contacts between the UC and surrounding rocks have been interpreted as ductile faults based on the presence of well-developed mylonites north and south of the anorthosite suite (Tella et al., 1993). The northern contact is largely unexposed, but anorthositic and mafic rocks there contain well-developed mylonite fabrics. The well-

exposed southern contact is marked by the ~1 km thick Southern bounding shear zone, which comprises predominantly quartzofeldspathic and dioritic rocks, but also includes anorthositic rocks. The magnitude of displacement accommodated by the shear zones is unknown; this has implications for whether the Uvauk complex should be viewed as an allochthon.

A relatively young (Proterozoic), magnetite-rich, granite pluton cuts the Uvauk complex to the east of the study area (Figure 1.3); the contact is a well-defined, sharp, intrusive contact (Tella et al., 1993). A contact aureole is characterized by fine-grained, sugary-textured plagioclase in the anorthositic gneisses adjacent to the pluton (Tella et al., 1993). It is not known if dykes and/or fluids related to the pluton have affected the anorthosite suite outside of the aureole. Similarity between this pluton and others in the WCP suggests that it is part of the ca. 1.82 Ga granite suite (Tella et al., 1993; Hudson Suite monzonites of Peterson et al., 2000), and thereby provides an important lower age limit for shearing.

Figure 1.1 Simplified geology of the western Canadian Shield (Snowbird TZ = Snowbird tectonic zone; modified after Wheeler et al., 1996).

Figure 1.2 Simplified geology of the east segment of the Snowbird tectonic zone
(Modified after Stern and Berman, 2001).

Legend

unmetamorphosed rocks

- sedimentary rocks
- ~1.82 granite

greenschist-facies rocks

- rocks
- Archean

amphibolite-facies rocks

- felsic gneiss
- supracrustal rocks

granulite-facies rocks

- Archean
- Proterozoic
- contact
- fault

Figure 1.3 General geology of the Uvauk complex and surrounding rocks, showing location of the three areas mapped in 1998 and 1999 (A, B, and C; modified after Tella et al., 1993).

Chapter 2: Lithology

Figures:

Figure 2.1: Lithological map of the western part of the Uvauk complex

Figure 2.2: Rocks of the Uvauk anorthosite suite

Figure 2.3: Primary, igneous and secondary, metamorphic mafic phases in the Uvauk anorthosites

Figure 2.4: Dykes of the Uvauk complex: two varieties

Figure 2.5: Diorite gneiss south of the anorthosite suite

Figure 2.6: Garnet-quartzofeldspathic gneiss

Figure 2.7: Tonalite straight gneiss and tonalite orthogneiss

Figure 2.8: Par-autochthonous anorthosite suite: locally-derived pelitic xenoliths and anorthosite veins within deformed country rock

Figure 2.9: Rocks north of the Uvauk complex: interlayered biotite-hornblende tonalite and amphibolite, enderbite, mafic granulite, and charnockite

Chapter 2: Lithology

This chapter contains descriptions of rock units from within the UC and the surrounding country rocks. The UC comprises a metamorphosed and sheared composite igneous complex as well as the bounding shear zones. Rocks of the anorthosite suite, although penetratively deformed and metamorphosed, are named in accord with their igneous origin. Use of metamorphic nomenclature is avoided, both for simplicity and to convey an inferred petrogenetic link between anorthosites, anorthositic gabbros, gabbroic anorthosites and leucogabbros. Metamorphic nomenclature is retained where protoliths are unknown. Description of the rocks within and surrounding the UC is critical to an understanding of the origin of the UC as it may provide clues to whether the UC was locally derived or allochthonous.

The anorthosite suite of the UC is a composite metaigneous belt approximately 30 km long, and varies from ~20 km wide at the easternmost part of the complex, to ~1 km at the westernmost part of the belt, near Uvauk Inlet, north of Chesterfield Inlet (Figure 1.3; Figure 2.1). Its original geometry has been extensively modified by post-intrusion deformation. Rocks of the suite dip steeply to the south in the west (area A of Figure 1.3), but dip gently to the west at the eastern extremity (area C of Figure 1.3). Contacts between the anorthosite and the surrounding geology are everywhere concordant; no crosscutting relations are preserved. Amphibolite-facies mineral assemblages (both prograde and retrarograde) are most common, but granulite-facies assemblages are locally preserved.

2.1 Anorthosite suite

Plagioclase-rich rocks are the dominant lithology of the metaplutonic suite, although voluminous mafic dykes constitute a significant proportion (35-40%). Nomenclature of Windley et al. (1973), based on proportion of mafic minerals, is used for the anorthosite and related rocks, as this nomenclature is both more descriptive than standard IUGS classification, and it implies a direct petrogenetic relationship between anorthositic rocks with varying mafic proportions. Anorthosite (0-10% mafic minerals; Figure 2.2a) is the most common composition within the suite, but gabbroic anorthosite (10-20% mafic minerals), anorthositic gabbro (20-35% mafic minerals), and gabbro (>35% mafic minerals) are present, in decreasing abundance respectively. A rock containing 35-50% mafic minerals is herein referred to as a “leucogabbro” in order to differentiate it from more melanocratic rocks.

The distribution of igneous phases within the suite is outlined in Figure 2.1. Gabbroic anorthosite (Figure 2.2b) is most common in the core of the suite (Figure 2.1). The southernmost part of the suite (e.g. M193 on Figure 2.1) is typically very coarse-grained (Figure 2.2c) and ranges in composition from anorthositic gabbro to leucogabbro. At the northern margin of the anorthosite suite, garnet-bearing anorthosite predominates (Figure 2.1; Figure 2.2d), but is intercalated with garnet-free anorthosite toward the southern extent of the map unit. Well banded, fine-grained gabbroic anorthosite, with cm-scale layering of mafic, felsic and intermediate bands, outcrops along the outermost margins of the suite and is garnet-bearing along the northern and western margins. This marginal phase varies in thickness, but is generally <5 m thick.

Concordant, garnet-bearing, mafic layers are also common throughout the anorthosite suite. Rarely preserved chilled margins and crosscutting of the foliation in the

anorthosite indicate that these mafic layers originated as dykes (see Chapter 3). In the northern part of the anorthosite suite, thin (10-20 cm), conformable mafic dykes are abundant. In contrast, the dykes are thicker (5-50 m) in the central and southern parts of the suite.

2.1.1 Anorthosite

The anorthosite unit is relatively well exposed and thickens in map view from ~1km in the west, to about 5 km in the easternmost part of the local map area (A of Figure 1.3; Figure 2.1). Although compositional layering is locally discernible, no unequivocal top indicators (e.g.: graded layering, trough scouring) were observed. Anorthosite is dominantly white weathering, with <5-15% mafic material (Figure 2.2a). Fresh surfaces are white where grain size is relatively coarse, but vary to dark grey with decreasing grain size. Deformation within the map unit is highly variable (see section 3.2 for description of structures).

Mafic minerals are variable in mode, mineralogy and grain size, throughout the suite. The range in grain size is extreme, with mafic crystals (or polycrystalline aggregates that replaced crystals) ranging from <1 cm to greater than 1 m, as observed at one locality (M193 on Figure 2.1). Mafic minerals include hornblende, actinolite, biotite, and cummingtonite; all of these hydrous minerals are interpreted to have replaced either a granulite-facies or the original igneous assemblage (Figure 2.3a). Characteristic of this map unit are polycrystalline aggregates of amphiboles, including hornblende and cummingtonite, with minor biotite, which are interpreted to pseudomorph primary pyroxene. Similar features have been noted by McGregor and Friend (1997), who suggested that the amphiboles are likely retrograde products after orthopyroxene.

Therefore, orthopyroxene may have been the main primary igneous phase in the anorthosites (see also MacDonald, 1999). However, occurrence of pyroxene is limited to the southernmost outcrops of anorthositic gabbro and leucogabbro (Figure 2.1).

2.1.2 Anorthositic gabbro to gabbroic anorthosite (GbAn)

The core of the anorthosite suite comprises grey, medium to coarse-grained, variably deformed gabbroic anorthosite to anorthositic gabbro. The map unit ranges up to 1 km in thickness (map view) in the east, thinning to its termination towards the west. The deformation is extremely heterogeneous; 1-10 m wide bands of highly strained, ultramylonitic, gabbroic anorthosite are intercalated with less deformed, protomylonitic, flaser-anorthosite (see section 3.3.4). The term 'flaser' refers to a protomylonite in which elliptical porphyroclasts lie within a finer mylonite matrix (Passchier and Trouw, 1998).

Grain size of the most deformed gabbroic anorthosite is difficult to assess but probably averages ~2 cm in the less deformed, flaser-anorthosite. The average mode for anorthositic gabbro is ~70% plagioclase, ~25% hornblende, ~3% cummingtonite, and 2% biotite, with trace titanite, ilmenite, rutile and zircon. Rare clinopyroxene, in the leucogabbro along the southern margin of the anorthosite suite, with exsolution lamellae of orthopyroxene (Figure 2.3b) indicate that clinopyroxene may have been the dominant igneous phase (MacDonald, 1999).

2.1.3 Mafic dykes of the Uvauk complex

Mafic, black, fine-grained, east-striking, granulite-facies dykes are ubiquitous throughout the anorthosite suite, and they are spatially associated with, if not restricted to the anorthosite suite. The thickness of the dykes is variable, ranging from 10's of cm in the north but up to 50 m in the southern and central parts of the anorthosite. Most dykes

are concordant with the foliation in the anorthosite, and exhibit a similar foliation and lineation to that found in the anorthosite. However, rarely preserved chilled margins and crosscutting of the foliation in surrounding anorthosite support the interpretation that the concordant mafic layers originated as dykes.

The dykes, herein referred to as the Uvauk dykes, comprise >90% of all dykes within the anorthosite suite. They contain ~35% clinopyroxene, ~35% garnet, ~20% plagioclase, and ~5% quartz, with minor hornblende and trace ilmenite. Relict orthopyroxene is evident from scanning electron microscopy (SEM) studies and is commonly mantled by clinopyroxene (see Chapter 4 for details and illustrations). The average grain size of clinopyroxene, and of orthopyroxene mantled by clinopyroxene, is ~0.5 mm. Hornblende is generally restricted to narrow zones near microfractures. Grain size of garnet porphyroblasts ranges from ~0.4 mm to ~1 mm. Plagioclase moats and plagioclase-hornblende symplectite coronae commonly surround garnet, especially in the westernmost part of the area, and in the north (i.e. the garnet-bearing anorthosite).

The majority of Uvauk dykes comprise a granulite-facies mineral assemblage (Figure 2.4), although hornblende-bearing assemblages are not uncommon, especially to the north and west. The hornblende-bearing dykes are interpreted to have undergone retrograde metamorphism (hydration?) after deformation (see Chapter 3). Thermobarometric studies reveal that the retrograde assemblage formed at high temperatures (~800°C; Chapter 4). Millimetre-scale tonalitic stringers are locally present within the dykes, suggesting possible incipient partial melting. The possibility that the stringers may be related to a separate magmatic phase seems less likely, as the

stringers are only present locally, where hydration is indicated by a retrograde assemblage.

Other dykes found in the map area include concordant orthopyroxene-garnet-bearing dykes, clinopyroxenites, and one crosscutting diabase dyke. Orthopyroxene-garnet-bearing dykes (Figure 2.4a and b) are rare, isotropic, and are concordant with the foliation in the anorthosite. These dykes are coarse-grained relative to the Uvauk dykes, with both garnet and orthopyroxene averaging ~2 mm in size. Rare pyroxenite dykes are also present, and are variably altered to anthophyllite, most notably at their margins. These dykes appear within and south of the anorthosite suite, but were not observed north of the anorthosite suite. One fine-grained, unfoliated, grey, diabase dyke, crosscuts the anorthosites and other mafic dykes at a high angle. This dyke comprises randomly oriented amphibole needles in a fine-grained plagioclase-rich matrix, and is interpreted to have been emplaced after deformation and metamorphism.

Preliminary geochemical studies reveal that the gabbro dykes of the UC exhibit *e*-MORB-like, or flat to very slightly light rare earth element (LREE)-enriched, chondrite normalized REE patterns (Sun and McDonough, 1989). Three of the five dykes fall within the field of the Paleoproterozoic gabbros of the KC. This petrochemical similarity is consistent with a common origin and age for gabbroic rocks within both of these Aqxaarneq gneiss-bearing complexes.

2.2 Country rocks south of the anorthosite suite

South of the anorthosite suite, country rocks are predominantly metaplutonic, and range in composition from quartzofeldspathic (generally tonalitic) to dioritic, with minor rocks of more mafic composition (amphibolites). Subordinate rocks of supracrustal

origin are intercalated with quartzofeldspathic to tonalitic rocks. The following rock descriptions are presented in order from north (at the UC contact) to south.

2.2.1 Diorite-monzodiorite straight gneiss (Di)

An east-striking belt of dioritic to monzodioritic gneiss outcrops south of the anorthosite suite and ranges in thickness from 10's of metres to 100's of metres, possibly reflecting map-scale boudins (green unit on map insert A). A ~50 m thick panel of diorite also outcrops north of the anorthosite suite (Figure 2.1). The diorite is homogeneous, green-grey to buff weathering, and intensely deformed. It typically displays a well developed, mylonitic to ultramylonitic foliation and lineation, and is, for that reason, herein referred to as a straight gneiss. South of the anorthosite suite, it is intercalated with sheets of quartzofeldspathic garnet-bearing gneiss, tonalite mylonite, amphibolite and subordinate well-foliated leuconorite. Most of the rock unit is well banded, diorite straight gneiss or ultramylonite (Figure 2.5a). However, a small proportion is somewhat less deformed; rare low-strain pods preserve an isotropic texture, facilitating identification of the protolith as a plutonic rock (Figure 2.5b).

The modal mineralogy of the diorite is ~60% plagioclase, ~20% orthopyroxene, 15% clinopyroxene, and <5% quartz. Pyroxene grains are generally < 3 mm, and plagioclase matrix grains average ~50 µm in diameter. Monzodiorite comprises ~20% orthopyroxene, 10-15% clinopyroxene, 50% plagioclase, up to 10% K-feldspar, ~5% quartz and minor garnet, biotite, and ilmenite. K-feldspar porphyroclasts occur sporadically and may have been introduced as quartzofeldspathic veins, which have subsequently been disaggregated by deformation.

2.2.2 Garnet-quartzofeldspathic gneiss (QFgn)

The dominant rock type in this map unit is a cream to buff weathering, garnet-bearing quartzofeldspathic gneiss (qf gneiss; Figure 2.6), comprising a 500-800 m thick lithologic unit south of, and intercalated with, the diorite straight gneiss (refer to map insert A). The rock typically has a pockmarked surface due to garnets that have weathered out of the rock (Figure 2.6a and b). The gneiss is well banded, with mm-scale mafic-rich bands alternating with cm-scale felsic bands. Other rock types in the map unit include garnet poor quartzofeldspathic (qp) gneiss, garnet amphibolite, diorite gneiss, and leuconorite (described below).

The garnet-quartzofeldspathic gneiss consists of ~30% garnet, ~40% plagioclase, ~20% quartz (half of which is interpreted to be vein quartz), ~5% biotite, and <5% orthopyroxene or cummingtonite (the two are mutually exclusive), with trace ilmenite, pyrite, rutile, monazite, and zircon. Garnets from the gneiss are generally ~2-3 mm in diameter, anhedral, highly fractured, and rimmed by biotite. Locally, the rock contains pink to lavender, unaltered garnets ranging in size from 2-5 mm.

Garnet amphibolite to garnet granulite occurs as conformable layers 10's of cm to several metres in thickness. Although most rocks are very fine-grained, some coarser grained, more plagioclase-rich parts within laterally continuous horizons contain clinopyroxene with garnet coronae. These rocks locally display a strong mylonite foliation, concordant with layering, but are more commonly granoblastic. The latter texture is interpreted to be a function of recrystallization.

Another lithology found within the qf gneiss, although not volumetrically important, is a leuconorite that is variable in grain size and modal mineralogy. The more leucocratic variety contains larger (up to 1 cm), slightly flattened, orthopyroxene crystals,

and ~65-75% plagioclase. The finer-grained variety typically contains more abundant and smaller orthopyroxene crystals (up to 3 mm), and about 50-65% plagioclase. The relationship between the leuconorite and the quartzofeldspathic gneiss is not clear where deformation is intense, but in one outcrop the leuconorite crosscuts the quartzofeldspathic gneiss. It is not known whether the norite is petrogenetically related to the anorthosite.

A well-developed foliation is ubiquitous in the quartzofeldspathic gneiss and is defined by quartz and plagioclase ribbons, compositional banding, and pressure shadows about garnet. Elongate quartz and plagioclase ribbons are parallel to compositional banding, but biotite-rich pressure shadows locally define a shape fabric oblique to the main foliation (see section 3.6.2). A strong extension lineation is also exhibited and is defined by elongate aggregates of quartz and plagioclase.

The protolith of the quartzofeldspathic gneiss is difficult to determine. Rare rafts of biotite-rich, tourmaline-bearing pelitic gneiss (Figure 2.6c) were observed within this rock unit, especially towards its southern margin where the rafts occur within a biotite-hornblende tonalite orthogneiss. The presence of tourmaline led to the tentative field interpretation of the garnet-quartzofeldspathic gneiss as a paragneiss derived from intense deformation and partial melting of supracrustal rocks. However, petrography and SEM imaging have revealed that the quartzofeldspathic gneiss is deficient in potassic phases. Rare microscopic grains of orthopyroxene were found in two thin sections. The presence of cummingtonite in sections that lack orthopyroxene suggests that this amphibole *may* be a retrograde product of orthopyroxene, but the two phases are mutually exclusive so no supporting textural evidence exists. The protolith is thought to be either a supracrustal

rock of intermediate composition or, perhaps more likely, a mixed rock: mylonitized migmatite or diatexite.

2.2.3 Tonalite straight gneiss (*Tstr*)

The main rock type in this map unit is biotite-hornblende $\pm$ clinopyroxene-tonalite gneiss with abundant inclusions of metasedimentary rocks, amphibolite, and conformable dykes of pyroxenite and mafic granulite (Figure 2.7a). It is generally poorly exposed, but good shoreline outcrop was observed at location M220 (Figure 2.1). The tonalite straight gneiss is white to light grey, medium-grained, and possesses extremely well-developed straight compositional banding at the mm- to 10's of cm-scale. Pyroxenite layers are generally 10 to 30 cm in thickness, extend for >50 m, and are locally boudinaged. Alteration of pyroxenite boudins to hornblendite or to clay minerals is common. Lenses of leuconorite and amphibolite (up to metre-scale thickness) are also present locally.

Within the tonalite orthogneiss, a range of sedimentary lithologies comprise 10-30 m thick sequences that were observed at three different locations across strike, but could not be traced along strike. These include a) brown to beige weathering, layered, tremolite and magnetite-bearing carbonate iron formation; b) white plagioclase-rich, tremolite-bearing (with minor magnetite) granofels; c) white, 5-10 cm thick quartzite layers; and c) dark brown to grey, tourmaline-garnet-biotite-semi-pelitic schist.

2.2.4 Tonalite orthogneiss (*Togn*)

Foliated, biotite $\pm$ hornblende-bearing, tonalite orthogneiss lies immediately south of the tonalite straight gneiss (Figure 2.1; Figure 2.7b). It is grey, medium-grained, and contains abundant concordant inclusions of amphibolite and semipelitic schist to

paragneiss. It also contains veins of hornblende-bearing monzogranite with potassium feldspar porphyroblasts that are locally dismembered from the vein from which they were derived, lending the tonalite a gneissic appearance. Minor medium to coarse-grained, white, foliated, hornblende-tonalite is generally concordant to, and exhibits a weak foliation parallel to the foliation of the biotite-tonalite. The hornblende-tonalite is coarse-grained, and locally contains hornblende xenocrysts that may have been mechanically assimilated from amphibolite inclusions. Rare crosscutting relations indicate that the hornblende-tonalite is intrusive into the biotite-tonalite. Semipelitic rafts are common in the northern part of the map unit, and become less voluminous farther south, while amphibolite rafts become more voluminous, although thinner, to the south.

Gabbro dykes, sills, and pods within the biotite-tonalite orthogneiss in the southern part of the map unit are different from dykes seen in the tonalitic orthogneiss or any other rock unit. These ENE-striking dykes range from 10 to 30 metres in thickness. The dykes are coarse-grained and consist of clinopyroxene and plagioclase. One leucogabbro dyke contains up to 50% plagioclase and another dyke contains coronitic garnet. Although the dykes are unconformable, the foliation in the surrounding tonalite appears to deflect around them, suggesting that they were emplaced before the last phase of deformation.

Although the tonalite straight gneiss (section 2.2.3) may represent a highly strained equivalent of the biotite-tonalite gneiss, they are interpreted to have different origins. In contrast to the tonalite straight gneiss, this map unit (Togn vs. Tstr, respectively, of map insert A) generally contains a higher proportion of mafic minerals, lacks the diverse assemblage of supracrustal inclusions, does not contain concordant,

granulite-facies, mafic dykes, and does not exhibit deformation fabrics indicative of intense ductile strain. The tonalite gneiss and straight gneiss may be part of the same igneous suite, but they exhibit different mineralogy, structures, and have different rock associations.

2.2.5 Plagioclase porphyroblastic, garnet-biotite semipelitic schist

Buff to rust weathering, moderately foliated, plagioclase porphyritic, garnet-biotite schist (Figure 2.6c) outcrops as a panel north of the tonalite orthogneiss, the former of which appears to thicken toward the east (see map insert 1). The schist is up to ~200 m thick in map view and is not well exposed, but sporadic outcrops along strike constrain its trend and thickness. The southwestern peninsula (in yellow on map insert 1) is underlain by foliated, kyanite-garnet-biotite-pelitic to semipelitic schist, intercalated with medium grey to black, granular amphibolite and garnet amphibolite.

Semipelitic schist typical of this map unit has a modal mineralogy comprising ~40% feldspar, ~20% quartz, ~20% garnet, and ~15-20% biotite, which is locally chloritized. Plagioclase porphyroblasts are typically ~ 5 mm, but are locally up to several cm in diameter. Garnet generally does not exceed 5 mm. Kyanite was observed in only one thin section (M339 on Figure 2.1) and was not discerned in the field (see also Chapter 4).

2.2.6 Inferred relationships between anorthosite and mylonitized country rocks

Although primary contact relationships between the anorthosite and mylonitized country rocks to the south are obscured by deformation, several observations suggest that the original contact may have been an intrusive one, an interpretation that pertains to the question of whether the UC is allochthonous. At several localities, pelitic schist (Figure

2.8a) and quartzofeldspathic gneiss were found within the anorthosite suite. These are common rock types within the country rock south of the UC (Figure 2.8b), and are therefore interpreted to be xenoliths of local derivation.

Layers of anorthosite, were also found, albeit rarely, within the shear zone south of the anorthosite suite (Figure 2.8c). Anorthosite layers are generally ~15 cm in thickness, can be traced for >50 m along strike, and are parallel to the foliation in the surrounding rock. The ~3:1 aspect ratio of mafic aggregates in the anorthosite bands is substantially less than the ~10:1 aspect ratio of mafic aggregates in the surrounding ultramylonitized country rocks. The anorthosite thus appears somewhat less strained than the surrounding ultramylonites, suggesting that intrusion of the former may have been late synkinematic with respect to deformation of the latter. However, rheological differences between the two rock types could also explain the apparent contrast in strain.

Because no crosscutting features are preserved, the mode of emplacement of the anorthosite suite is difficult to assess. Tectonic emplacement seems an unlikely mechanism for the intercalation of 10 cm wide veins that can be traced for >50m along strike. Although the original nature of the contacts is obscured by deformation, the relationships described above are consistent with intrusion of the anorthosite into the surrounding rocks, and subsequent modification of lithological contacts by deformation. Field observations, although equivocal, are consistent with an intrusive contact between the anorthosite and the surrounding geology, implying that displacement along the shear zone was not regionally significant. However, the amount of ductile thinning may have been significant. In addition, the possibility for two ages of anorthosite (see Chapter 5)

complicates interpretations, as tectonic emplacement of one of the two cannot be precluded.

2.3 Country Rocks North of the Anorthosite suite

2.3.1 Interlayered tonalite gneiss and amphibolite (Tamp)

North of the anorthosite, tonalite gneiss and amphibolite appear to be the dominant rock types, although assessment is limited by poor exposure. Grey, medium-grained, well-foliated, hornblende-biotite tonalite contains abundant 2-10 cm thick, continuous and conformable amphibolite layers, and locally garnet-amphibolite inclusions (Figure 2.9a). Amphibolite volume ranges from 30-60% of the exposure and is intercalated with the tonalite gneiss, lending the rock a highly layered appearance. Rocks of this map unit comprise amphibolite-facies assemblages. Mineralogy of the tonalite gneiss comprises plagioclase, hornblende, quartz and biotite. Variable amounts of biotite observed in amphibolite and in the garnet amphibolite *may* be secondary, possibly related to potassic fluids introduced during intrusion of the regional 1.83 Ga Hudson or 1.76 Ga Nuelin granites.

Other rock types present in the map unit include intermediate hornfels, monzogranite and quartzofeldspathic gneiss. Medium to fine-grained, intermediate hornfels outcrops locally, and comprises plagioclase, hornblende, epidote, and minor biotite. The hornfels is texturally distinctive, with subhedral plagioclase crystals up to 5 mm in diameter, and a fine-grained matrix of hornblende-plagioclase with common euhedral, randomly oriented plagioclase laths. Although the texture is likely due to recrystallization, a volcanogenic origin is interpreted for the hornfels, based on the intermediate composition, fine grain size, cm-scale layering and the presence of rare

plagioclase-porphyritic horizons. Monzogranite is present in rare veins that display a protomylonitic fabric. A 10-15 m thick panel of quartzofeldspathic, garnet-bearing gneiss, comparable to the quartzofeldspathic gneiss south of the anorthosite, is also present north of the anorthosite suite.

2.3.2 Mafic granulite, enderbite and charnockite

The northernmost map unit comprises white to grey, medium-grained, foliated biotite±orthopyroxene-tonalite (tonalite to enderbite; Figure 2.9b), mafic granulite, less common pyroxenite and amphibolite, and minor, orthopyroxene charnockite to enderbite. The tonalite to enderbite weathers recessively, forming topographic lows between ~50-100 m diameter plugs of dark green, fine- to medium-grained, mafic granulite and pyroxenite, which form rocky hills. White to pink, orthopyroxene-bearing, charnockite to enderbite (the proportion of K-feldspar is variable) occurs as veins within the mafic granulite and pyroxenite (Figure 2.9c). The veins are typically unfoliated, although a weak foliation may be preserved along vein margins (see Figure 3.10a).

Charnockitic to enderbitic veins contain large (~1 cm) orthopyroxene crystals with a complex coronitic structure and very fine-grained matrix plagioclase and quartz (see Figure 3.11a). Amphibolite is confined to elongate rafts and lenses ~10 cm to 1 m in thickness in the low-lying regions of tonalite to enderbite.

The granulite-facies mafic rocks are green-weathering and occur as two main varieties: 1) fine-grained clinopyroxene-garnet±plagioclase granofels; and 2) less abundant medium-grained clinopyroxene granofels. The first is generally very fine grained, with clinopyroxene averaging <0.2 mm and smaller garnets. Its modal mineralogy is ~70% clinopyroxene, 15% garnet, 10% plagioclase (higher in rare, more leucocratic layers), and

<5% opaques. The second granofels is a nearly monomineralic clinopyroxenite with a uniform grain size of ~1 mm. Compositional banding was observed only locally, and is characterised by 2-3 cm layers with variable plagioclase to clinopyroxene ratios.

The granofels that outcrop in the northern part of the map area are interpreted to be xenoliths within the surrounding tonalite to enderbite. The orientation of compositional banding locally displayed by the mafic granofels is inconsistent both between adjacent blocks and with the surrounding tonalite. The foliation in the tonalite appears to wrap around the large (50-100 metre thick) mafic xenoliths, while the smaller amphibolite inclusions form concordant rafts and lenses within the tonalite. White to pink, orthopyroxene-bearing, charnockite to enderbite crosscuts and brecciates the mafic xenoliths (intrusion breccia; Figure 2.9c) and is thought to be related to the surrounding tonalite. The tonalite is only locally pyroxene-bearing, and may represent retrogressed enderbite.

The contact between the enderbite-mafic granulite unit and the tonalite-amphibolite to the south is unexposed, and therefore unclear. The presence of tonalitic and amphibolitic rocks within both map units suggests a gradational contact, with the northern unit containing thicker mafic inclusions that preserve granulite-facies assemblages. Possibly, the rocks of the tonalite-amphibolite unit have been thoroughly affected by retrograde metamorphism. On the other hand, the tonalite-amphibolite unit (Tamp) is well foliated, rarely displays a lineation, and exhibits amphibolite-facies mineral assemblages whereas the enderbite-mafic granulite unit displays a well-developed, down-dip lineation, is moderately foliated, and contains granulite-facies

assemblages. The difference in metamorphic grade, concordance of foliation, and down-dip lineation across the contact suggest the contact may be a cryptic fault.

2.4 Pegmatite

Dykes of pink to white, undeformed, biotite-muscovite granite pegmatite crosscut all rocks in the field area. Pegmatite dykes are present throughout the anorthosite suite, and are generally ~10 cm wide. They typically occupy shear zones and fractures oriented at a high angle to the main foliation. Lack of a penetrative foliation and lineation, and isotropic texture exhibited by the pegmatite indicates that the dykes were emplaced after deformation. Though not volumetrically significant, the pegmatite represents one of the youngest rocks in the area as it is among the few rock types that have not been deformed or metamorphosed. The pegmatite dykes may be related to the Hudson (ca. 1.83 Ga) or Nuelin (ca. 1.75 Ga) granites described in Chapter 1.

Figure 2.1 Lithological map of the western Uvauk complex. Sample numbers indicate locations described in the text, and shear zones are overlain by grey transparency.

Legend

Anorthosite suite

- Anorthosite
- Anorthositic gabbro, leucogabbro
- Garnet-anorthosite
- Mafic granulite, enderbite, charnockite
- Tonalite gneiss, amphibolite

Surrounding geology

- Diorite to monzodiorite gneiss
- Garnet-quartzofeldspathic gneiss
- Tonalite straight gneiss, mixed metasedimentary rocks
- Tonalite orthogneiss
- Garnet-biotite semipelitic schist

Figure 2.2 Rocks of the Uvauk anorthosite suite: a) anorthosite, b) gabbroic anorthosite, c) coarse-grained anorthositic gabbro (with loony for scale), and d) garnet-bearing gabbroic-anorthosite.

Figure 2.3a) Polycrystalline amphibole aggregate, interpreted to have completely replaced pyroxene, in gabbroic anorthosite (field of view ~5.2 mm; A = actinolite; B = biotite; H = hornblende; P = plagioclase); b) clinopyroxene crystal (C) with orthopyroxene exsolution lamellae, surrounded by plagioclase (P) clinopyroxene and garnet (G) matrix, from leucogabbro at southernmost part of the anorthosite suite (field of view ~5.2 mm).

Figure 2.4 Dykes of the Uvauk complex: a) hand sample of medium-grained, orthopyroxene-garnet-bearing, mafic dyke; b) photomicrograph of 1-2 mm garnet and orthopyroxene crystals, probably of igneous origin. The rock lacks a well-developed structurally imposed fabric, although undulose extinction and minor subgrain development indicate that the rock has suffered some component of strain (field of view ~5.2 mm); c) Hand sample of fine-grained, clinopyroxene-garnet-bearing, mafic dyke; d) Photomicrograph of 1-2 mm garnets, <500 μ m clinopyroxene porphyroclasts, with polycrystalline aggregates of clinopyroxene and plagioclase defining a well-developed foliation (Field of view ~5.2 mm).

Figure 2.5 Diorite south of anorthosite suite: a) very fine-grained, diorite straight gneiss or ultramylonite, with well-developed compositional banding; b) foliated diorite from low strain lens.

Figure 2.6a) Garnet-quartzofeldspathic gneiss, b) well-layered garnet-qf gneiss, c)
tourmaline-bearing semipelitic schist associated with garnet-qf gneiss and tonalite
orthogneiss.

Figure 2.7a) Shoreline exposure of tonalite straight gneiss with concordant pyroxenite dykes; b) tonalite orthogneiss with pink, monzogranite veins concordant to the gneissosity.

Figure 2.8a) Layer of gabbroic anorthosite within country rocks south of main anorthosite suite; b) plagioclase porphyroblastic semipelitic schist, inferred to be a xenolith within meta-anorthosite; c) typical semipelitic schist south of the Uvauk complex (Loony for scale in each photograph).

Figure 2.9 Rocks north of UC: a) interlayered biotite-hornblende tonalite and amphibolite with dismembered porphyroclasts from coarse-grained to pegmatitic monzogranite; b) foliated orthopyroxene-bearing tonalite (enderbite); c) mafic granulites crosscut and brecciated by charnockite.

Chapter 3: Structural geology of the Uvauk complex

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Chapter 3: Structural geology of the Uvauk complex

3.1 Introduction

The Uvauk complex (UC) is an ENE-trending triangular belt, approximately 30 km in length and up to 20 km wide (map view) in the east (Figure 1.3). The complex comprises a composite meta-igneous suite and bounding shear zone rocks, both of which developed under granulite-facies conditions. The western part of the UC (Area A of Figure 1.3) is the focus of this study, however, the results of mapping in the eastern extremity (Area C of Figure 1.3) provide important constraints on the overall geometry of the complex.

The western part of the UC (Area A of Figure 1.3) comprises very steeply south-dipping ($\sim 75^\circ$) lithologic units and, to facilitate discussion, has been subdivided into three domains: southern, central and northern (Figure 3.1). These domains correspond to the anorthosite suite and predominantly metaplutonic rocks that flank the UC to the north and south. The UC comprises the central domain of anorthosite and the bounding mylonites that separate the central domain from the northern and southern domains. It is unclear whether the mylonite belts accommodated significant displacement, or represent strain-accommodating structures with only minor displacement.

The structural geology of the study area is first described in order to provide an overview of the deformation history; interdomainal differences are subsequently highlighted. The two domain-bounding high strain zones, designated the Northern and Southern bounding shear zones (Figure 3.1), in addition to two satellite shear zones that outcrop further to the south of the UC, are described below. A comparative discussion follows to highlight differences between the shear zones.

The chapter finishes with a discussion of mafic magmatism in the UC and attempts to constrain the timing of emplacement with respect to latest deformation. Field relationships suggest that Uvauk dykes (described in section 2.1.4) were predominantly deformed synchronously with their host rocks, however primary intrusive features and crosscutting relationships are locally preserved. Details of this relationship, and its potential significance, are outlined in the latter part of this chapter.

3.1.1 Structural overview of the Uvauk complex

Deformation of the Uvauk complex occurred at granulite-facies conditions, the evidence for which (opx-grt-cpx-pl assemblage) is well-preserved in most Uvauk dykes and in rocks north of the UC. However, widespread retrograde metamorphism has produced upper-amphibolite assemblages that prevail throughout much of the complex. In addition to its polymetamorphic history, the structural evolution of the UC has been further complicated by intense ductile deformation (mylonitization) and thorough overprinting of early structures (see below). Two shear zones, each averaging ~1 km in width transect the map area from east to west. In contrast to the penetrative mylonitic fabrics that characterize the shear zones, the rocks between the shear zones are *variably* foliated and lineated. Due to substantial overprinting, older structures are difficult to discern, and it is possible that there may be more than one generation of structures that predate the main deformation fabrics. In light of complicated overprinting relationships, designation for “generational” structures are as follows. Primary igneous layering (locally preserved) is denoted S_0 . The dominant penetrative foliation is denoted S_x whereas earlier foliations are denoted S_{x-1} , and later ones are designated S_{x+1} . Transposition of older structures, through tight to isoclinal folding and progressive

elimination of fold closures, and, most commonly, by mylonitization, is evident in the shear zones; the mylonitic foliation is therefore designated S_m , and S_m is generally parallel to S_x . It is not readily apparent whether S_m developed synchronously with, or subsequent to, the development of the regional foliation, S_x , the predominant foliation within the complex and in the domains to the north and south. Pre- to syn-transposition folds are collectively denoted F_x , whereas younger folds that fold the main tectonic foliation are denoted F_{x+1} (or F_{m+1} , since most observed folds are within the shear zones). Lineations in the study area are denoted L_x .

3.1.2 Introduction to mylonites

Although a rigorous microstructural analysis was not completed, microtextural observations and the dominant processes responsible for their production are briefly summarized. Since the shear zone rocks are the products of mylonitization, a brief definition of a mylonite is appropriate: a mylonite is a foliated, commonly lineated, rock, which occurs in planar zones of intense deformation and is finer grained than the surrounding rocks, into which it grades (Bell and Etheridge, 1973). The production of porphyroclasts (parent grains), surrounded by subgrains and new grains (daughter grains to the original parent grain), commonly referred to as a core-and-mantle structure, is characteristic of mylonites, and occurs in a ductile regime (White et al., 1980). Core-and-mantle structures form through progressive diminution of grain size, which occurs as the result of dynamic recovery and recrystallization processes, or subgrain rotation and grain boundary migration in particular (Tullis and Yund, 1987; Hirth and Tullis, 1992; Trimby et al., 1998; Urai et al., 1986). In general, the proportion of recrystallized matrix grains increases with progressive shear strain (White et al., 1980), although, if strain is

partitioned into the matrix, then porphyroclasts may survive large increments of finite strain; this can result from rheological differences due to contrasting mineral composition (e.g. Hanmer, 1987).

The term ‘mylonitic foliation’ (*sensu* Passchier and Trouw, 1998) is used throughout the chapter. For clarity, a mylonitic foliation is defined as a spaced foliation composed of alternating layers and lenses with different mineral composition or grain size (i.e. the S planes or ‘schistosité’).

3.2 Southern domain

3.2.1 Foliations

S_x is the dominant foliation present within the southern domain, and consists of compositional layering parallel to grain-scale shape-preferred orientation. S_x dips steeply south-southwest, with an average orientation of 109/66°SSW (Figure 3.2a). Within the biotite-tonalite orthogneiss unit, the shape fabric is defined by oblate quartzofeldspathic aggregates and the parallel alignment of sparse biotite grains. Compositional banding in the unit ranges from mm- to dm-scale, and is characterized by alternating layers of dark amphibolite and coarse-grained, pink hornblende-monzogranite veins (Figure 3.3a). In contrast to the continuous foliation within the biotite-tonalites, the semi-pelitic schist exhibits an anastomosing to rough disjunctive foliation. Here, S_x is defined by microlithons with randomly oriented to weakly oriented quartzofeldspathic minerals, separated by short (mm-scale), discontinuous septae defined by parallel alignment of biotite (Figure 3.3b). Amphibolites of the southern domain range from well-foliated and compositionally banded, defined by variable plagioclase content (Figure 3.3c), to unfoliated.

3.2.2 Lineations

A rare polycrystalline extension lineation was observed in rocks of the southern domain. Where present, it is defined by elongate, generally polymineralic aggregates of hornblende and biotite or of quartz and plagioclase. The average orientation of the lineation is $40^{\circ} \rightarrow 121$ (Figure 3.2b).

3.2.3 Shear-sense indicators

Shear-sense indicators in the southern structural domain include rotated δ -porphyroclasts and boudins, strain-insensitive fabrics, and rare asymmetric extensional shear bands (Figure 3.4). All shear-sense indicators were observed in, or at a low angle to, the flow plane. The δ -porphyroclasts rotate more quickly than the soft polycrystalline mantle, and viscous drag causes proximal parts of the mantle to be entrained and to deflect, forming a wing from which the sense of drag can be directly read to give shear-sense, in this case dextral (Figure 3.4a; Hanmer and Passchier, 1991). The strain-insensitive fabrics (S_b) are interpreted to represent the kinematic strain axes and the gneissosity (S_x) is taken to roughly parallel the flow plane (Hanmer, 1984; Hanmer and Passchier, 1991). The orientation of S_b $\sim 30^{\circ}$ counterclockwise to S_x , is consistent with a dextral sense of shear (Figure 3.4b). At one locality, weakly developed asymmetric extensional shear bands were indicative of sinistral sense of shear. However, shear bands were not penetratively developed at this locality, and may not be regionally significant. All shear-sense indicators described above are developed in amphibolite facies rocks, with no evidence for overprinting of or by granulite-facies assemblages.

3.3 Central domain

3.3.1 Foliations

The central domain is characterized by an east-striking, steeply south-dipping continuous foliation (S_x), with an average orientation of 079/77°S (Figure 3.2c). Much of the anorthosite exhibits a spaced foliation, defined by flattened porphyroclasts (commonly replaced by polycrystalline aggregates) of mafic minerals in a fine-grained, plagioclase matrix. A mylonitic foliation (S_m), defined by flattened lenses of polycrystalline mafic minerals with alternating bands of polycrystalline plagioclase, is heterogeneously developed throughout the central domain (Figure 2.2b, c, and d). This intense foliation is thought to be contemporaneous with S_x , based on consistent metamorphic grade through strain gradients (see below).

Granulite-facies assemblages, including nonhydrous phases such as pyroxenes, are only locally preserved within the central domain, and are best preserved along the southernmost margin of the anorthosite suite (see Figure 2.4a). The foliation in anorthosites is a banding that is defined, predominantly, by randomly- to radially-oriented, post-tectonic amphibolite-facies assemblages (Figure 3.5a). Intergrowths of hornblende, cummingtonite, biotite, and rare actinolite with plagioclase commonly pseudomorph pyroxene (see section 2.1.1), suggesting that the original fabric-forming mineral was pyroxene (Figure 3.5b).

Mafic granulites also have a coarse, continuous foliation, which is generally poorly developed with respect to that in the anorthosites, but, based on concordance of foliations and continuity of metamorphic grade, is interpreted to have been produced by the same S_x -forming deformation event. S_x in the mafic granulites is defined by elongate aggregates of clinopyroxene and plagioclase (Figure 2.4d). Some mafic granulitic rocks are not foliated (Figure 2.4a and b) and these may represent dykes that postdate S_x . Post-

tectonic growth of hydrous phases is locally evident in the granulite-facies dykes (see Chapter 4).

The presence of two foliations was only noted at one locality in the central domain. An early cleavage (S_{x-1}), observed at locality M191 (Figure 3.1), is defined by the alignment of biotite and hornblende, and is oriented at $131/64^\circ$ SW. It is oblique to, and crosscut by S_x , which is also defined by alignment of biotite and hornblende, and is oriented at $104/84^\circ$ S. S_x is parallel to, and probably the same generation as, the main foliation exhibited by adjacent rocks (Figure 3.5c). The local preservation of S_{x-1} suggests the possibility that there may be a significant, though poorly preserved, early foliation in the study area.

3.3.2 Lineations

A well-developed, penetrative, east-plunging, extension lineation (L_x) characterizes the central structural domain, and has a mean orientation of $43^\circ \rightarrow 080$ (Figure 3.2d). The lineation is defined by elongate aggregates and rods of plagioclase and amphiboles and occurs on surfaces of the main foliation, S_x (Figure 3.6a).

Locally (e.g. at locality 98TXM193 on Figure 3.1), elongate dm- to m-scale orthopyroxene crystals define a mineral lineation in a homogeneous, recrystallized, sugary (*sensu* Hanmer, 1987) plagioclase matrix that lacks any tectonic fabric. The crystal form generally lacks the tails associated with tectonic lineations, and is interpreted to represent an early, ultra-high temperature magmatic lineation (Figure 3.6b).

3.3.3 Folds

F_x folds were observed predominantly in the western part of the central domain where strain gradients are high, and the dominant fabrics are mylonitic (Figure 3.7a).

The mylonitic foliation parallels compositional banding and is tightly to isoclinally folded with axial planes parallel to S_m , and steeply plunging fold axes parallel to L_x . These folds should not necessarily be regarded as a separate phase of deformation that folded S_m , but may represent the effect of local perturbations in the flow field during mylonitization (Passchier and Trouw, 1998). Rare sheath folds were observed, suggesting that some folding or rotation of fold axes occurred during shearing in the central domain (Figure 3.7b).

3.3.4 Strain variability

Meter-scale variations in strain, most evident within the core of the central domain, are manifested as differences in the degree of foliation development. The proportion of dynamically recrystallized matrix ranges from ~5 to ~100%, providing further evidence of substantial strain variation across this domain (Figure 3.8a, b, and c). Rocks in the core of the central domain show evidence for two main styles of deformation: 1) heterogeneously sheared, internally boudinaged and rotated blocks of mylonite to ultramylonite (Figure 3.9a); interleaved with 2) more homogeneously deformed augen- or flaser-textured (protomylonite) panels (Figure 3.9b). Mafic minerals in the former are extremely elongate, forming thin bands or ribbons, whereas those in the latter can still be recognized as individual grains despite elongation. It is unclear whether these two deformation styles result from strain partitioning, or from an age difference in the rocks. If the latter holds, then the question remains: do the more intensely deformed rocks represent an older igneous suite with a longer deformation history or a younger, warmer, and therefore softer synkinematic intrusion? This question could not be answered in this study.

3.4 Northern domain

3.4.1 *Foliations*

The northern structural domain is characterized by a northeast-striking, steeply dipping foliation (S_x) with a mean orientation of 060/75 SE (Figure 3.2e). S_x is a spaced foliation defined by flattened mafic and quartzofeldspathic aggregates. Compositional banding is evident where thin (~10 cm wide), amphibolite lenses lie parallel to S_x (Figure 2.10). Large mafic lenses, up to 100 m in diameter, are interpreted to be xenoliths. These are wrapped by the foliation but themselves do not contain a planar or linear fabric, with the exception of rare compositional banding, defined by variable plagioclase to clinopyroxene ratios. Veins of charnockitic to enderbite composition are ubiquitous throughout the mafic inclusions and locally contain a weak foliation along their margins (Figure 3.10a). The rock is interpreted as an “intrusion breccia”, produced by injection of charnockitic magma into fine-grained mafic rocks.

Contact relationships between the enderbite and mafic inclusions are interpreted to be a function of size and aspect ratio of the inclusions. Thin, elongate inclusions accommodate strain and rotate toward parallelism with the flow plane, whereas the large, thick inclusions act as rigid blocks deflecting the foliation around them (Hanmer and Passchier, 1991, and references therein). Preservation of granulite-facies assemblages is common in inclusions in the northern domain, but amphibolite retrograde assemblages prevail in the long, narrow mafic inclusions.

3.4.2 *Lineations*

A heterogeneously distributed, steeply southeast-plunging lineation (Figure 3.2f) characterizes the northern structural domain. The mean orientation of this lineation is

75°→149, and it is defined by elongate aggregates of quartz and plagioclase and of biotite ± hornblende ± orthopyroxene. The lineation is developed in the enderbite to tonalitic rocks that wrap around the mafic xenoliths and by the charnockitic phases within the xenoliths, but is not developed within mafic xenoliths. The charnockites are generally well-developed L-tectonites (Figure 3.10b and c), except at vein margins where a foliation is present (Figure 3.10a). The surrounding enderbite to tonalitic rocks are generally L>S tectonites.

3.4.3 Microstructural observations

Petrographic observations reveal that the unfoliated charnockite (Figure 3.11a) contains a nearly 100% recrystallized, foam textured, plagioclase-rich matrix (Figure 3.11b). Insertion of the quartz plate reveals aggregates of either slow- or fast-oriented crystals. These are irregularly shaped and probably represent whole crystals that have been recrystallized. Because the orthopyroxene crystals appear undeformed and no planar or linear fabrics are evident in the rock, recrystallization is interpreted to have occurred under low strain or static conditions. Recrystallization, involving the creation of new grain boundaries and subsequent reduction in internal energy, was likely the result of highly mobile dislocations. Such features are indicative of extremely high temperatures (Hirth and Tullis, 1992).

3.5 Southern bounding shear zone

The Southern bounding shear zone is ~1 km thick, east-striking belt of quartzofeldspathic to dioritic mylonites with an average foliation orientation of 091/79°S (Figure 3.2g). Most rocks within the belt display a well-developed mylonitic foliation and lineation, in stark contrast to the undeformed rocks that are rarely preserved within

the shear zone. The southern margin of the shear zone (Figure 3.1) is marked by the transition from heterogeneously mylonitized, quartzofeldspathic gneiss (south of the shear zone) to homogeneous quartzofeldspathic mylonite with a finely laminated, flaggy appearance (ultramylonite or shear zone rocks). The northern margin of the shear zone is delineated by the transition from the heterogeneously deformed meta-anorthosite to homogeneously mylonitized meta-anorthositic and -dioritic rocks.

3.5.1 Foliations

Primary igneous layering (S_0) was observed in one low strain zone within the diorite gneiss, and is defined by variability in the modal abundance of plagioclase and clinopyroxene. The main foliation developed in the shear zone is a mylonitic foliation (S_m), defined by compositional banding and a parallel shape fabric comprising ribbons and oblate, polycrystalline, monomineralic aggregates (Figure 3.12a). Quartz veins are abundant in some of the mylonites, and, in the field, are virtually indistinguishable from flattened quartz grains. S_m is a strongly developed, straight and planar foliation, indicative of very high strain. Toward the southern margin of the shear zone, the foliation becomes progressively more wavy and irregular, as ribbons deflect around plagioclase porphyroclasts or rigid garnets (Figure 3.12b).

S_m developed under granulite-facies conditions, as indicated by the presence of orthopyroxene in garnet pressure shadows (quartzofeldspathic gneiss) and of garnet porphyroblasts in pressure shadows of clinopyroxene porphyroclasts (diorite gneiss). Post-tectonic amphibolite-facies conditions are indicated by cummingtonite crystals that cut across S_m , and by the presence of randomly oriented biotite adjacent to, and within fractures of, garnet porphyroblasts (quartzofeldspathic gneiss).

An oblique, or strain-insensitive, foliation (S_b) is heterogeneously developed and typically oriented at $\sim 25\text{-}35^\circ$ counterclockwise to S_m . The strain-insensitive foliation is defined by the obliquity of plagioclase and quartz aggregates, or biotite grains, relative to compositional banding (Figure 3.13). S_b is interpreted to have been formed when a balance was achieved between foliation-destroying and foliation-forming processes, i.e., foliation-forming flattened grains rotated, tracking the finite strain ellipsoid and subsequently recrystallized, destroying the foliation. The cycle continues so that subsequent strain resulted in mildly flattened grains with long axes roughly parallel to the orientation of the maximum instantaneous extension direction (Hanmer, 1984; Hanmer and Passchier, 1991). The orientation of biotite aggregates rimming, and possibly replacing, garnet shows similar obliquity to S_m , which is itself defined by compositional banding and quartz ribbons (Figure 3.13b). In this case, however, biotite probably formed late with respect to deformation, since biotite has not rotated into the flow plane.

Evidence for progressive transposition is best illustrated in the thicker of the satellite shear zones south of the Southern bounding shear zone but the process of transposition is herein inferred to be synchronous with mylonitization. S_m is defined by mm- to dm-scale banding between alternating plagioclase-rich and mafic-rich bands (Figure 2.7a and 2.8a and b). The mean orientation of S_m , $091/79^\circ$ S, is roughly parallel to the margins of the shear zone (Figure 3.2g). S_T (a transposition foliation) can be seen in centimetre- to decimetre-scale compositional banding in Figure 3.14a where the axial plane of an east-plunging antiform is parallel to S_m . Tight to isoclinal folding of the S_m mm-scale compositional bands can be discerned in areas of lower strain (Figure 3.14b), and the axial planes and limbs are parallel to an apparently younger transposition

foliation (S_T). S_T is probably a stage in the development of S_m , and these folds are likely the result of local heterogeneities (Passchier and Trouw, 1998), and not of a separate folding event. Given that folds were found to decrease in amplitude laterally and that no fold trains were obvious, it seems likely that the observed folding is simply the result of progressive transposition of fabric elements during mylonitization.

3.5.2 Lineations

Extension lineations are ubiquitous throughout the Southern bounding shear zone and are defined by elongate quartz and plagioclase aggregates. The average orientation is $24^\circ \rightarrow 101$, but moderate to shallow, west-plunging lineations are also present in the shear zone (Figure 3.2h). Both the east- and west-plunging lineations were observed together at one locality, however, it could not be determined if the lineation porpoises about the horizontal, or if they represent discrete generations of lineation development.

3.5.3 Folds

A low strain zone within the monzodiorite preserves F_m folding of primary igneous layering (S_0). Fold axes plunge steeply to the west ($\sim 60^\circ \rightarrow 282$). These folds are interpreted to be shear-related, incipient folds that had not rotated into the flow plane. The foliation is more intense in the limbs of the fold and progressively intensifies to become the main mylonitic fabric (S_m) away from the hinge. The mylonitic foliation is itself folded by F_{m+1} with Z-fold geometry (Figure 3.15c). This fold sequence is consistent with the aforementioned hypothesis that folding in the Southern bounding shear zone occurred syn-mylonitization, and that S_m is a transposition fabric, formed by progressive strain associated with strain gradients.

3.5.4 Satellite shear zones

Two high-strain zones south of the Southern bounding shear zone occur within the Southern domain (Figure 3.1). The southernmost high-strain zone is apparently <200 m wide, and the northern one is ~500 m wide; the eastern lateral extent of these is unknown due to lack of exposure. The two high strain zones south of the main shear zone may be satellites of the Southern bounding shear zone, compatible with their orientation and the orientation of their constituent structural fabrics. However, the satellite shear zones preserve amphibolite-facies assemblages, and lack evidence for an earlier, granulite-facies metamorphism. If the satellite shear zones are temporally linked to the Southern bounding shear zone, then they either formed at slightly higher structural levels or have been thoroughly reactivated and recrystallized under amphibolite-facies conditions. Alternatively, the shear zones may not be coeval with the Southern bounding shear zone, but may post-date it and may record different P-T conditions. The fact that the orientation of the shear zones is concordant to the regional foliation provides no constraints on the timing of shear zone development, which may post-date development of the regional foliation.

3.5.5 Shear-sense indicators

Shear-sense indicators are rarely preserved within the Southern bounding shear zone, but are more common within the northernmost of the two “satellite” shear zones. Shear-sense indicators were observed in sections nearly perpendicular to the foliation, and at a low angle to the extension lineation; all indicated dextral shear-sense. Progressive dextral shear was inferred from the Z-fold geometry described above; this determination is considered to be reliable because of the high angle between the fold axis and lineation orientation, indicating that the fold axis had not rotated into the flow plane.

The orientation of the strain-insensitive fabric (S_b), at $\sim 25\text{--}35^\circ$ counterclockwise to the main foliation, S_m , is indicative of dextral shear, as S_b represents the instantaneous stretching direction (Figure 3.13). The geometry of a minor fold train in a plagioclase-rich band within the tonalitic straight gneiss is consistent with dextral shear sense as well. Shear-sense determination is based on the clockwise rotation of axial planes toward the inferred flow plane with progressive strain (Figure 3.16a). Dextral shear-sense is also indicated by a rotated δ -type garnet porphyroblast, and by δ - and σ -type rotated boudins (Figure 3.16b and c). The mechanism behind δ -type porphyroclasts is described in section 3.2.3. Shear-sense is further indicated by σ -type rotated boudins based on stair-stepping wings, which step up in the direction of movement of the upper block (Passchier and Trouw, 1996).

3.5.6 Microstructural observations

Rocks within the Southern bounding shear zone exhibit features indicative of both grain boundary migration and subgrain rotation, consistent with dynamic recrystallization under high temperatures (or low strain rates) (Hirth and Tullis, 1992). Irregular grain boundaries, patchy extinction, and bulging of one grain into an adjacent grain, are indicative of grain boundary migration (Figure 3.17a). Core-and-mantle structures are the main evidence for subgrain rotation, which occurs by progressive misorientation of subgrains as a result of climb-accommodated dislocation creep (Urai et al., 1986; Figure 3.17b). Core-and-mantle structures in pyroxene (Figure 3.18a) are thought to require high temperatures in order to initiate and sustain dislocation climb (Brodie and Rutter, 1985).

Several microstructural features are indicative of mylonitization under granulite-facies conditions. The presence of garnet porphyroblasts in pressure shadows of clinopyroxene porphyroclasts (in dioritic gneiss; Figure 3.18a) and of orthopyroxene in garnet pressure shadows (in quartzofeldspathic gneiss; Figure 3.18b) provide compelling evidence that deformation occurred at granulite-facies. Lenticular clinopyroxene layers are ~3-5 centimetres in length and comprise polycrystalline aggregates interpreted to have been derived from grain size refinement of single crystals (Figure 3.19a and b). Phase dispersion (*sensu* Hanmer, 1984) is evident in some thin sections, and may have formed by grain boundary sliding to reduce crystal surface energy or by flow-induced grain boundary dilation during metamorphism and deformation (Hanmer, 2000). Elongate orthopyroxene porphyroclasts are common in the diorite gneiss map unit (map insert A) and these may represent low strain relict grains (Hanmer, 2000) or attenuated ribbon grains formed by dislocation glide (Brodie and Rutter, 1985). These textural features are similar to those described by Hanmer (2000) from granulite-facies mylonites of the Striding-Athabasca segment of the Snowbird tectonic zone.

The strain gradients, microstructures, and textures described above are all consistent with mylonitization under granulite-facies conditions. A detailed study of the deformation processes and microstructural interpretation (MacDonald, 1999) provides rigorous documentation of microstructural features preserved in rocks of the UC (Southern bounding shear zone, in particular). MacDonald (1999) also concluded from her studies that deformation occurred under granulite-facies conditions. SEM imaging reveals randomly oriented minute biotite laths in the quartzofeldspathic gneiss and minor

actinolite occupying fractures within clinopyroxene crystals of the diorite gneiss, indicative of a late, post-tectonic amphibolite facies metamorphic event.

3.6 Northern bounding shear zone

The Northern bounding shear zone trends ENE and ranges between ~500 m to 1 km in width (Figure 3.1). A gradational, northward increase in strain toward the shear zone is suggested by the progressive increase in aspect ratio of flattened polycrystalline mafic aggregates. Metamorphic grade within the Northern bounding shear zone is granulite facies (see chapter 4 for full description and quantification of P-T conditions).

3.6.1 Foliations

Rocks within the Northern bounding shear zone generally exhibit cm- to dm-scale banding, parallel to S_m , between anorthosite and mafic granulite or amphibolite. Within anorthositic bands, S_m is defined by alternating bands of plagioclase and bands, or elongate discontinuous lenses, of amphibole (Figure 3.20a). The average foliation orientation within the shear zone is $067/71^\circ$ SE, roughly parallel to the strike of the shear zone (Figure 3.2i). Primary igneous layering (S_0) in diorite, defined by the variation of plagioclase to clinopyroxene, is preserved in a low strain zone within the Northern bounding shear zone and is oriented perpendicular to S_m (see section 3.7.4 for context and further description).

3.6.2 Lineations

A well-developed extension lineation, with a mean orientation of $57^\circ \rightarrow 090$, is ubiquitous throughout the Northern bounding shear zone (Figure 3.2j). The lineation is defined by elongate polycrystalline aggregates of plagioclase and mafic minerals,

generally hydrous phases. In the western part of the Northern bounding shear zone, the lineation is locally quite steep and is associated with extremely heterogeneous strain.

3.6.3 Folds

S_m is folded by F_{m+1} folds at the westernmost part of the north-bounding shear zone (Figure 3.21a). Fold axes are generally parallel to L_x although, locally, a high angle between the two features allowed for shear-sense determination. Within the anorthositic bands, the mylonite foliation is folded, but no general axial planar foliation is apparent. Locally, however, retrogressed mafic granulite exhibits an axial planar foliation defined by hornblende and elongate hornblende-plagioclase intergrowths, interpreted to be garnet pseudomorphs (Figure 3.21b and c). Rare clinopyroxene, rimmed and surrounded by hornblende, indicates that the mafic bands were mafic granulite that has been retrogressed to amphibolite. Axial planar development of hornblende and garnet pseudomorphs in mafic layers indicates that retrograde metamorphism was syntectonic with respect to mylonitization.

3.6.4 Shear-sense indicators

Only four shear sense indicators were observed within the north-bounding shear zone, and all but one suggest a sinistral shear sense component. All shear-sense indicators were observed in sections nearly perpendicular to the foliation, and at a low angle to the extension lineation. The sinistral shear sense indicators include a sinistrally rotated boudin, back-rotated asymmetric pull-aparts, and an S-fold within the anorthosite (Figure 3.22). Sinistral shear-sense inferred from the fold is valid because of the high angle between the mineral lineation and the fold axis, indicating that the fold axis had not rotated into the flow plane. Figure 3.22b illustrates a back-rotated anorthositic boudin

(with long axis oriented $\sim 10^\circ$ clockwise to the main mylonite foliation). Within the boudin, shear bands separate back-rotated mafic segments, and these are indicative of sinistral rotation. One dextrally rotated δ -porphyroclast was noted at the westernmost part of the shear zone where deformation is complicated by local heterogeneities and tight to isoclinal folding. Its significance is thus somewhat ambiguous.

Sinistral shear sense was also determined in the 20–40 metre wide sheet of diorite within the North-bounding shear zone. Primary igneous layering displayed in a 3 metre wide low strain inclusion is perpendicular to the ENE-striking mylonite fabric displayed by the surrounding material (Figure 3.22c). Rotation of this fabric into the flow plane is interpreted to be a consequence of transposition of primary layering into the mylonite foliation by progressive rotation and attenuation of primary layering. The sigmoidal form of the deflected banding in the low strain pod is interpreted to reflect counterclockwise rotation of the fabric into the flow plane, consistent with sinistral sense of shear (Figure 3.22d). The inclusion possesses an aspect ratio of >6 , and the rotational behaviour of such objects is essentially the same as that of a passive marker (i.e. such objects do not rotate once aligned in the flow plane; Hanmer and Passchier, 1991). Sinistral shear sense in the reference frame described above suggests the shear zone is an oblique normal fault, with north-side-up, and sinistral strike-slip components.

3.6.5 Microstructure

Petrography revealed a high matrix-to-porphyroclast volume ratio in rocks from the Northern bounding shear zone relative to rocks of the Central domain, indicating a strain gradient which increases to the north (Figure 3.1). Core-and-mantle structures indicate that recovery and dynamic recrystallization processes accommodated

intracrystalline deformation. Along the northern margin of the Northern bounding shear zone, rocks exhibit fine grain size, phase dispersion (*sensu* Hanmer, 1984), and polygonal texture, suggesting the possibility that the rocks have annealed, or have been modified by post-deformation recrystallization. Alternatively, Hanmer (2000) suggests that, in the Striding-Athabasca segment of the STZ, extreme phase dispersion may have developed synkinematically under high temperatures.

Overprinting of older structures is evident in thin section (locality M353; Figure 3.1), where S_m is defined by alternating biotite- and plagioclase-rich bands within which rootless intrafolial folds are preserved (Figure 3.20b). The limbs and axial planes of the folded S_m foliation are parallel to the transposed mylonitic foliation. The “transposition foliation” is likely akin to the mylonite foliation, and is interpreted to have developed during the same deformation event.

Clinopyroxene is rarely preserved in the shear zone, and hornblende development along clinopyroxene rims and cleavage is indicative of fluid influx. Hornblende and plagioclase symplectitic intergrowths around garnets are suggestive of decompression (e.g.: $12 \text{ Qtz} + 3 \text{ Ts} + 2 \text{ Prp} + 4 \text{ Grs} = 12 \text{ An} + 3 \text{ Tr}$; equation written with High P assemblage on the left).

3.7 Emplacement of mafic magmas

The Uvauk dyke swarm constitutes ~40%, by volume, of the composite meta-igneous belt, which is dominated by the Uvauk anorthosite suite (Figure 3.23a). The texture and mineralogy of the dykes are described in section 2.1.4. Many dykes have granulite-facies assemblages, although amphibolite-facies assemblages are pervasively developed in the Northern bounding shear zone. The trend of the dykes is roughly east-

west, concordant with the foliation in the anorthosite. The dykes typically contain a foliation parallel to dyke margins and to the foliation of the anorthosite host. Rare chilled margins are preserved and provide unequivocal evidence that the ubiquitous black bands are in fact dykes (Figure 3.23b).

Deformation of the dykes is highly variable; differences in deformation style are interpreted to reflect variations in the extent of deformation due to timing of emplacement with respect to deformation, rather than rheological differences. Most dykes display a penetrative foliation, and the variability in intensity is consistent with their intrusion throughout a deformation event. The style of deformation also varies among the dykes; some dykes are boudinaged (Figure 3.23c), whereas others are isoclinally folded along with the surrounding anorthosite (Figure 3.23d). However, rarely preserved crosscutting features indicate that some dykes have not been significantly deformed (Figure 3.24a). Preservation of bayonette-like apophoses (Figure 3.24b), and obliquity between the foliation in anorthosite and the dyke margins, indicate that some dykes were emplaced late with respect to deformation. Rare crosscutting of anorthosite foliation, and of conformable dykes, by texturally and mineralogically similar dykes is also indicative of post-tectonic emplacement of some dykes (Figure 3.24c), suggesting that dyke emplacement initiated *prior to* and continued *after* deformation.

The spatial association of mafic dykes with anorthositic rocks begs the question of a petrogenetic link between the two. Although a petrogenetic link has not been chemically established between dykes and anorthositic rocks of the UC, Sanborn-Barrie (1999) has demonstrated a cogenetic relationship between gabbroic anorthosite and gabbroic rocks (possibly equivalent to UC dykes) within the KC through geochemical

modeling. If consanguinity between the anorthosites and gabbroic dykes of the UC can be established, then the anorthosite must also have been emplaced *during* deformation.

3.8 Summary

The central anorthositic domain of the UC resembles a flat sheet in the east (area C of Figure 1.3), that may have been affected by two phases of Proterozoic folding (F3 and F4 folds, as described by Tella et al., 1993). The geometry of the westernmost part of the Uvauk complex has been interpreted as a tight, moderately east-plunging synform (Tella et al., 2000b; area A of Figure 1.3). Between these two areas, exposure is very poor and the transition of foliation orientations between the two is difficult to interpret. However, air photo interpretation and topology provide no evidence for faulting; the transition is apparently gradational, and may be controlled by map-scale folding. It is unclear whether the shear zones represent one contiguous shear zone because they each comprise different rock types and contain different metamorphic textures and assemblages.

The central domain (Figure 3.1) is bounded by two principal shear zones: the Northern and Southern bounding shear zones. The former dips moderately toward SSE and the average lineation orientation is moderate to the east; the latter dips steeply to the south and the average lineation orientation plunges shallowly to the east. Foliations in the westernmost part of the UC (i.e. the area shown in Figure 3.1) are east-striking and dip steeply to the south (079/77°S). North and south of the anorthosite suite (Figure 3.2a and e), foliations change progressively to more northeast-striking, and southeast-striking, respectively (Appendix D). The distribution of foliation orientations does not require that the shear zones represent discrete faults separating structurally distinct domains. The

overall geometry could result from strain partitioning, with the bulk of strain accommodated by the shear zones, with or without significant displacement.

Lineations are least common in the southern domain, where they plunge moderately to the southeast. Lineations in the central domain plunge moderately to the east, whereas lineations in the northern domain plunge steeply to the southeast. The difference in lineation orientation may reflect a different strain path associated with the different domains.

Among the three domains (excluding the map-scale shear zones delineated on Figure 3.1), strain is greatest within the central domain where it is highly heterogeneous due to strain partitioning. In contrast, rocks north and south of the central domain exhibit a homogeneously developed penetrative foliation. Tight to isoclinal folding is evident in the central domain only, and is compatible with localization of strain within this domain.

Granulite-facies assemblages are best preserved in the northern domain, especially within mafic to ultramafic protoliths. Rocks of the anorthosite suite locally preserve granulite-facies assemblages, but more commonly contain a post-tectonic retrograde amphibolite-facies assemblage. The southern domain hosts amphibolite-facies rocks, and preserves no granulite-facies assemblages.

The Southern bounding shear zone is a ~1 km wide, E-trending belt of ultramylonites. The average orientation of foliations (091/79°S), lineations (24°→101), and the predominance of dextral shear-sense indicators suggest that the shear zone may represent an oblique, dextral strike-slip fault with a south-side-up component, indicative of transpression. The magnitude of displacement across shear zones could not be estimated due to the lack of piercing points. Based on evidence cited above, deformation

within the shear zone is interpreted to have occurred under granulite-facies conditions (Figures 3.19a and b). Petrographic evidence also exists for a late, post-tectonic, amphibolite-facies retrograde metamorphic event.

The Northern bounding shear zone is an ENE-trending shear zone that thickens in map view from 200-300 m in the west to ~1 km wide toward the east. The average orientation of foliations within the shear zone is 067/71°SSE, and lineations have an average orientation of 57°→090. Sinistral shear sense in this reference frame suggests the shear zone represents an oblique normal fault, with north-side-up, and sinistral strike-slip components. Deformation occurred at granulite-facies conditions (see Chapter 4), with a late synkinematic to post-kinematic hydration resulting in the formation of hornblende at the expense of clinopyroxene.

Differences between the Southern and Northern bounding shear zones include overall orientation of the belts, orientation of lineations and foliations, metamorphic grade, and the interpreted sense of shear. The former trends E whereas the latter trends ENE, a minor difference that may be insignificant. Lineations plunge more shallowly and foliations dip more steeply in the Southern relative to those in the Northern bounding shear zone. Anhydrous, granulite-facies assemblages are well preserved in the Southern bounding shear zone. In contrast, pervasive hornblende-development suggests that rocks within the Northern bounding shear zone have been affected by either retrograde metamorphism or hydration at granulite facies. The axial planar alignment of hornblende indicates that hydration or retrograde metamorphism occurred *during* deformation. The sense of shear indicated is dextral/reverse and sinistral/normal for the Southern and Northern bounding shear zones, respectively.

The significance of the Southern and Northern bounding shear zones is considered in terms of three possibilities. The two shear zones may have formed independently during two temporally distinct events, consistent with the difference in kinematics and metamorphic grade displayed by the two shear zones. Alternatively, the two shear zones may have formed during the same deformation event, but bifurcated around the central anorthosite-dominated domain. Finally, the two shear zones may be part of one shear zone that has been folded by an E-W fold system, so that the shear zone outcrops along the north and south limbs of an east-plunging synform but is contiguous beneath the anorthosite suite. The presence of shallowly west-dipping anorthositic rocks in the eastern part of the complex (Appendix I) suggests that the orientation of foliations and rock units may have been modified by folding. The presence of a marginal phase on the southern and northern margins of the anorthosite suite, and of diorite gneiss north and south of the suite, are also consistent with a fold geometry, but do not *require* folding. The significance of the shear zones is revisited in the final chapter.

The anorthosite suite of the UC is cut by abundant E-W trending gabbroic dykes. The spatial association of the dykes and anorthosites at UC is similar to that described at KC, where geochemical modelling has established a petrogenetic link between the two rock types (Sanborn-Barrie, 1999). Similarity of the UC dykes to the KC dykes suggests that the two dyke sets may be the same age (Appendix III). If this is the case, then there may be a component of 1.9 Ga magmatism at the UC, and the formerly described syn-tectonic emplacement of the dykes would further imply that deformation within the central anorthosite domain occurred at ca. 1.9 Ga.

Figure 3.1 Generalized map of the western part of the Uvauk complex showing structural domains with shear zones shaded in black. The average foliation orientation for each domain is indicated.

Figure 3.2 Equal-area, lower hemisphere stereoplots of structural domains of the Uvauk complex: a) Poles to foliation planes of the southern domain; b) Lineations of the southern domain; c) Poles to foliation planes of the central domain; d) Lineations of the central domain; e) Poles to foliation planes of the northern domain; f) Lineations of the northern domain; g) Poles to foliation planes of the Southern bounding shear zone; h) Lineations of the Southern bounding shear zone; i) Poles to foliation planes of the Northern bounding shear zone; j) Lineations of the Northern bounding shear zone.

Southern domain

Central domain

Northern domain

Southern bounding
shear zoneNorthern bounding
shear zone

N

N

N

N

N

Figure 3.3 Foliations in the southern domain: a) gneissosity in tonalite orthogneiss defined, in part, by concordant pink monzogranitic bands (lens cap for scale); b) disjunctive foliation in semipelitic schist (field of view is 5.2 mm); c) compositional banding in amphibolite (penny for scale).

Figure 3.4 Shear-sense indicators in the southern domain : a) dextrally-rotated K-feldspar δ -porphyroclast in tonalite orthogneiss (top = north; foliation = $093/76^\circ$ S; lineation from adjacent outcrop = $32^\circ \rightarrow 110^\circ$); b) strain-insensitive foliation in tonalite orthogneiss, indicating dextral sense of shear (section is parallel to the extension lineation and perpendicular to the foliation; field of view is 5.2 mm).

Figure 3.5 Foliation in anorthosite of the central domain: a) compositional banding in gabbroic anorthosite. Radial amphibole arrangement is post-tectonic and is interpreted to have replaced orthopyroxene; b) pre-tectonic clinopyroxene with orthopyroxene exsolution lamellae; c) early, penetrative foliation (S_{x-1}) cut by later foliation ($S_{x/m}$; S_m foliation = 104/84°SSW; S_{x-1} foliation = 131/64°SW). Both foliations are defined by fine-grained biotite.

Figure 3.6 Lineations in the central domain: a) the lineation is typically an extension lineation as shown in the tonalitic veinlet within granulite-facies gabbro dyke and is defined by the alignment of mafic minerals and of polycrystalline plagioclase aggregates (view to the north; lineation plunges east); b) the alignment of clinopyroxene crystals in a sugary plagioclase matrix may have developed under ultra-high temperatures (view to the north; lineation plunges east).

Figure 3.7 Folding in the central domain: a) isoclinal folds in anorthosite and mafic granulite, thought to have occurred during mylonitization; b) sheath fold in gabbroic anorthosite is thought to have formed during mylonitization.

a **b**

Figure 3.8 Variability in strain within the central domain: a) low strain: primary plagioclase crystals with deformation twins and incipient development of subgrains at grain boundaries; b) moderate strain: protomylonite, approaching mortar texture, with oblate plagioclase porphyroclasts parallel to the foliation; c) high strain: reduced of plagioclase grains as a result of ductile processes (field of view for each photomicrograph is 5.2 mm).

Figure 3.9 Deformation styles exhibited in the core of the central anorthosite suite: a) heterogeneously sheared, internally boudinaged, and rotated blocks of mylonite to ultramylonite, interleaved with b) more homogeneously deformed augen- or flaser-textured (protomylonite) panels.

Figure 3.10 Fabric elements of charnockite from the northern domain: a) charnockite vein in mafic granulite lacks foliation except at vein margins; b) hand sample of charnockite with no notable foliation; c) well-developed lineation in charnockite hand sample.

Figure 3.11 Microstructural observations of charnockite from the northern domain: a)
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b) well-developed granoblastic or foam texture in plagioclase-quartz matrix (field
of view for both is 5.2 mm).

Figure 3.12 Trace of foliation in the Southern bounding shear zone: a) mylonites to ultramylonites may exhibit plagioclase ribbons parallel to compositional banding; b) protomylonites typically contain plagioclase porphyroclasts, around which the foliation must deflect (field of view for both is 5.2 mm).

Figure 3.13 Oblique strain-insensitive foliations (S_b) in the Southern bounding shear zone: a) plagioclase and quartz aggregates define S_b in tonalite orthogneiss, S_b is aligned $\sim 30^\circ$ counterclockwise to the main foliation (S_m); b) biotite aggregates rimming garnet in garnet-quartzofeldspathic gneiss define a shape fabric (S_b) that is $\sim 45^\circ$ counterclockwise to the main foliation (S_m).

Figure 3.14 Transposition of compositional banding in tonalite straight gneiss: a) tight, upright, isoclinal folding of banding in tonalite straight gneiss; b) rootless fold closures in straight gneiss, indicative of an older composition-defined foliation.

Figure 3.15 Folds in the Southern bounding shear zone: a) compositional banding (S_1) is folded, with axial planes of cm-scale parasitic folds parallel to axial plane of outcrop-scale fold (partially covered by lichen). Note axial planar metamorphic mineral growth, indicating that metamorphism here outlasted deformation; b) compositional bands in quartzofeldspathic gneiss folded to produce an upright, open antiform; c) folding of mylonite foliation in dioritic ultramylonite.

Figure 3.16 Shear-sense indicators in the Southern bounding shear zone: a) progressive rotation of axial planes of minor folds toward the inferred flow plane is consistent with dextral sense of shear; b) dextral shear-sense is inferred from rotated δ -type garnet porphyroblast and by c) δ - (right) and σ - (left) type rotated boudins. Arrows point to embayments adjacent to δ -type rotated boudin (all photographs are from tonalitic straight gneiss; for a) and b): top = south; $S_m = 090/80^\circ$; $L_m = 56^\circ \rightarrow 118^\circ$; for c): top = north; $S_m = 094/72^\circ$; $L_m = 37^\circ \rightarrow 106^\circ$).

Figure 3.17 Intracrystalline deformation-accommodating processes operative in the Southern bounding shear zone: a) grain boundary migration, or bulging, indicated by the apparent protrusion of an unstrained grain into an adjacent grain with higher dislocation density (field of view ~2 mm); b) core-and-mantle structure, or deformed single crystal surrounded by a mantle of fine-grained material of the same mineral. The structure is thought to form by dynamic recrystallization of the outer grain in response to migration of dislocations into subgrain walls during progressive deformation (field of view ~5.2 mm).

Figure 3.18 Features of granulite-facies leucogabbro: a) clinopyroxene ribbons are lenticular and range from ~1-8 cm; b) ribbons comprise a fine-grained mosaic and are bounded by coronitic garnets. Note the dispersion of opaque minerals throughout the plagioclase-rich matrix (width of field of view is 5.2 mm).

Figure 3.19 Metamorphic grade under which deformation occurred: a) orthopyroxene grains in garnet pressure shadows suggest that deformation occurred at granulite facies. Randomly oriented biotite and cummingtonite (not shown) crystals indicative of post tectonic hydration or retrograde metamorphism; b) porphyroclasts, subgrains and new grains of clinopyroxene and orthopyroxene, suggestive of high P-T conditions required for dislocation climb to accommodate intracrystalline deformation (see section 3.6.1). ~50 μm garnet neoblasts in pressure shadow of clinopyroxene porphyroclasts also indicates that deformation occurred at high grade.

Figure 3.20 Foliations in garnet anorthosite of the Northern bounding shear zone: a) main foliation (S_m), defined by compositional banding; b) S_m is axial planar to small-scale intrafolial rootless folds that are locally preserved (field of view is 5.2 mm).

Figure 3.21 **Folds in the Northern bounding shear zone: a) folded anorthosite (white) and amphibolite (retrogressed mafic granulite-facies dyke, black). The mylonite foliation within the former is folded, whereas the latter possesses an axial planar metamorphic mineral alignment. The mafic granulite is interpreted to have recrystallized during retrograde (hornblende-producing) metamorphism, thereby losing the foliation formed under granulite-facies conditions, whereas the anorthosite retained the earlier foliation (axial plane is parallel to width of photograph). b) relict garnet is surrounded by a plagioclase-rich moat that is oblate parallel to hornblende alignment and to the axial plane of the fold (axial plane is parallel to width of photomicrograph; field of view is 5.2 mm), c) the two plagioclase aggregates within the mafic layer are garnet pseudomorphs that are elongate parallel to the axial plane (trace of fold is illustrated by thin, plagioclase-rich band; axial plane is parallel to width of photomicrograph; field of view is 5.2 mm).**

Figure 3.22 Shear-sense indicators in the Northern bounding shear zone: a) sinistrally rotated mafic boudin; b) back-rotated mafic blocks in a back-rotated anorthosite boudin. Long axis of boudin is oriented $\sim 10^\circ$ clockwise to the flow plane (assumed to be roughly parallel to the nearly vertical foliation). Back-rotation of the mafic blocks produced synthetic shear bands from which shear-sense can be determined; c) primary modal layering (S_1) of low-strain pod (foreground) perpendicular to surrounding mylonite foliation (background, parallel to hammer); d) modal layering, (S_1), is transposed into the flow plane (lower part of photograph) by progressive rotation and attenuation of its extremities (view to the south; foliation = $084/70^\circ S$; lineation = $24^\circ \rightarrow 193^\circ$).

Figure 3.23 Emplacement of mafic magma in the Uvauk complex: a) east-trending, concordant, fine-grained, granulite-facies mafic dykes comprise ~40% of the composite anorthosite-gabbro suite and may be petrogenetically related to the anorthosite; b) rarely preserved chilled margins provide good evidence for hypabyssal emplacement of the dykes; c) some dykes have been highly boundinaged; d) other dykes have been folded along with the anorthosite. The difference in deformation style may refect different rheological behaviour due to timing of emplacement with respect to deformation.

Figure 3.24: Cross-cutting features exhibited by some Uvauk dykes: a) in the easternmost part of the Uvauk complex, a near-vertical dyke crosscuts the shallowly, west-plunging foliation in the anorthosite; b) obliquity between the dyke margin and the foliation in surrounding anorthosite and the presence of bayonette-like apophoses are consistent with the interpretation that the mafic layers are in fact dykes; c) mafic dyke cross-cutting foliation-parallel dykes, lines running from lower left toward upper right are late brittle fractures.

Chapter 4: Metamorphic Petrology and Thermobarometry

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CHAPTER 4. Metamorphism and Geothermobarometry

This chapter deals with the metamorphism of the Uvauk complex, focusing on the interpretation of petrographic features and chemical variations that are directly related to geothermobarometric analysis. The chapter begins with an introduction to some of the theory involved in geothermobarometry and, in particular, how peak P-T estimates can be retrieved from granulite-facies rocks. After a brief description of the methods used, the distribution of selected samples is outlined with respect to lithology and metamorphic grade. The following section describes, for each sample in turn, the results of petrographic and thermobarometric analysis. First order interpretations of textural relationships and inferred metamorphic reactions are discussed first, followed by descriptions and interpretations of mineral chemistry, and thermobarometric results. Second order interpretations and implications are discussed in the final section.

4.1 Introduction to geothermobarometry

Geothermobarometry is the assessment of pressure (P) and temperature (T) conditions under which a rock equilibrated based on the measured compositions of coexisting minerals. While particular metamorphic mineral assemblages can be used to infer, qualitatively, the metamorphic facies, or grade, mineral compositions can be used to quantify metamorphic conditions. The fundamental equation that is solved is:

$$\Delta G = 0 = \Delta H_{(298,1)} + \int_{(298-T)} \Delta C_P dT + \int_{(1-P)} \Delta V dP - T \{ \Delta S_{(298,1)} + \int_{(298-T)} \Delta C_P / T dT \} + RT \ln K_{eq}$$

where ΔG , ΔH , ΔS , ΔV , and ΔC_P represent change in (Δ) free energy, enthalpy, entropy, volume, and heat capacity, and K_{eq} is the equilibrium constant of reaction, respectively (see Spear, 1993). The required thermodynamic data can be taken from appropriate, internally consistent compilations. The equilibrium constant is determined from

measured mineral compositions, combined with activity coefficients calculated with available activity models.

This study utilizes the Thermobarometry With Estimation of EQUilibration state (TWEEQU) software of Berman (1991) and mineral compositions measured using the Cameca Camabax electron microprobe at the Geological Survey of Canada. The software computes lines of constant K_{eq} in P-T space (see Figure 4.3, for example). The equilibrium constant is determined by:

$$K = [(a_C)^2(a_D)^3 / (a_B)^4(a_A)] \quad \text{for a given reaction: } A + 4B = 2C + 3D$$

where “a” is the activity of a mineral end member. The activity is equal to the mole fraction of the component raised to the power of its site multiplicity. The mole fraction of a given species is computed from the microprobe analyses.

4.1.1 Problems with geothermobarometry of granulite-facies rocks

Most geothermometers are based on exchange reactions because they show considerable temperature sensitivity (large ΔS) and small pressure sensitivity (small ΔV), resulting in steep equilibrium constant curves on a P-T diagram (Spear, 1993). A critical assumption in exchange thermometry is that the distribution of cations, once established, does not change (or reset) upon cooling (Frost and Chacko, 1989). The temperature at which cation exchange for a particular reaction effectively ceases is referred to as the “closure temperature” for that reaction. The main problem with thermobarometry in granulite grade rocks results from their formation at temperatures in the 700° - 1000°C range, which is above the closure temperature for Fe-Mg exchange (600°-700° C; Lasaga, 1983; Spear and Florence, 1992; Bégin and Pattison, 1994). Thermometers based on Fe-Mg exchange are thus susceptible to “resetting”, or post-metamorphic peak (late) cation

exchange. The compositions of minerals that have undergone late exchange will reflect the temperature at which exchange ceased rather than the temperature at which peak metamorphism occurred.

Geobarometers are based on net transfer reactions which involve relatively large volume changes and have a pressure sensitive equilibrium constant (Spear, 1993). Because net-transfer reactions involve mass transfer of Al and Si, and breaking of Si-Al bonds, they have a closure temperature significantly higher than that of exchange reactions (Frost and Chacko, 1989). However, many barometers contain ferromagnesium phases that undergo Fe-Mg exchange after the barometer reaction has closed. Therefore, the apparent equilibrium constants for the net-transfer reactions are changed as a result of continued Fe-Mg exchange after Al-Si diffusion closure (Fitzsimmons and Harley, 1994; Pattison and Bégin, 1994). The extent to which the equilibrium constant is altered by Fe-Mg “feedback” depends on the sensitivity of the calibrated reaction to the Fe-Mg distribution coefficient (K_D), but the most significant error in pressure determination stems from the barometer being intersected at lower P by the “reset” exchange thermometer.

From this discussion, it is clear that the interpretation of chemical zoning is extremely important in geothermobarometry. Not only is this interpretation critical to assess which mineral compositions may have been in equilibrium at a specified time, but chemical zoning can also provide insight into changing metamorphic conditions. Growth zoning develops due to a change in the composition of the material supplied to the surface of the growing crystal (Tracy, 1982). Diffusion zoning, or the transport of matter driven by a chemical potential gradient, involves cation transfer between a crystal's

surface with either an adjacent crystal or with the matrix via an intergranular exchange medium (Pattison and Bégin, 1994). This process can significantly alter mineral compositions from their peak metamorphic conditions, obscuring attempts to reconstruct peak metamorphic conditions (Spear, 1992; Spear and Florence, 1992).

Pattison and Bégin (1994) investigated zoning patterns in garnet, and their observations led to significant implications for geothermometry of granulites. Garnet zoning patterns are described in terms of the four principle divalent cations of garnet end members (Ca, Fe, Mg, Mn). Garnet from the granulite facies Minto terrain (northern Quebec) show uniform Mn and weak concentric zoning of Ca. In contrast, Fe and Mg zoning (represented by $\text{Fe}/(\text{Fe}+\text{Mg})$ due to the antithetic variation of Fe relative to Mg) shows zoning patterns highly dependent on the mineral species of neighbouring minerals. Garnet $\text{Fe}/(\text{Fe}+\text{Mg})$ shows minimal variation adjacent to plagioclase and quartz, but strong variations adjacent to biotite and orthopyroxene. Their observations led to the following conclusions: 1) Fe-Mg exchange continues after (i.e. at lower temperatures) Ca in garnet and Al in orthopyroxene have ceased to change; 2) garnet grains with no adjacent ferromagnesium phases remain apparently unaffected by late Fe-Mg exchange but their chemistries have been homogenized by intracrystalline exchange, as indicated by homogeneity across individual garnets, and by intercrystalline exchange, as inferred from the uniformity of $\text{Fe}/(\text{Fe}+\text{Mg})$ between different grains. While the former may be induced by high temperature cooling or retrograde metamorphism, the latter suggests the presence of an intergranular exchange medium, such as a melt; 3) the abrupt transition from effective intergranular exchange to localized Fe-Mg exchange is probably a consequence of the loss of an intergranular exchange medium.

Several authors have proposed methods to retrieve accurate peak metamorphic conditions from high grade rocks in light of retrograde cation exchange. Frost and Chacko (1994) suggest that accurate estimates can be acquired using “fossil thermobarometry”. Samples should be collected from low strain zones where high temperature exsolution lamellae and relict mineral assemblages are likely to be preserved. The GASP (Anorthite = Grossular + Kyanite + Quartz) barometer is recommended as this barometer does not involve exchange between ferromagnesium phases and its equilibrium constant will, therefore, not be greatly affected by retrograde Fe-Mg exchange. Finally, they recommend using intersecting barometers with different slopes to confirm the results of geothermometry.

Bégin and Pattison (1994) present a method of peak P-T retrieval which attempts to reconstruct the Fe/(Fe+Mg) ratios of garnet and orthopyroxene at the point where net transfer reactions close. The positions of an Fe-Mg exchange thermometer (Almandine + Enstatite = Pyrope + Ferrosilite) and an Al-thermometer (Pyrope = Mg-Tschermakite + Enstatite) are recalculated with incremental changes in Fe/(Fe+Mg) ratios of garnet and orthopyroxene. The garnet and orthopyroxene compositions that generate coincident P-T intersections for both thermometers are taken to represent the compositions that existed at peak metamorphic conditions. Similar recalculations are applied to barometers to acquire an intersection with the averaged thermometer that should represent near-peak metamorphic conditions.

Results of Bégin and Pattison (1994; also Fitzsimmons and Harley, 1994) were based on calculations in the MgO-Al₂O₃-SiO₂ (MAS) system. However, recent studies have demonstrated a large effect of Fe on Al₂O₃ solubility in orthopyroxene (Harley,

1984; Lee and Ganguly, 1988; Aranovich and Berman, 1997). Aranovich and Berman (1997) suggested that recent attempts to calibrate Al solubility in orthopyroxene in the FAS (FeO-Al₂O₃-SiO₂) more closely approximates orthopyroxene occurrences in crustal rocks. They offer a new garnet-orthopyroxene thermometer (Almandine = Al₂O₃ in Opx + 3 Ferrosilite) based on Al₂O₃ solubility in orthopyroxene in the FAS system, and argue that the larger enthalpy and entropy of FAS equilibrium relative to the MAS equilibrium results in a considerably more robust thermometer, although its slope is less steep than that of the Fe-Mg exchange equilibrium. In addition, the authors suggest that composition corrections are unnecessary where computed differences given by the two thermometers are similar to the combined uncertainties of the thermometric calculations.

Both convergence techniques (Bégin and Pattison, 1994; Fitzsimmons and Harley, 1994) recognize the importance of ferromagnesium mineral modes in relation to P-T estimates. The ferromagnesium mineral of lesser volumetric abundance will be more strongly affected by Fe-Mg exchange. This fact should be considered when modifying the compositions of garnet and orthopyroxene in a convergence technique to reconstruct near-peak mineral compositions. In practice, however, results are not overly sensitive to whether garnet, orthopyroxene, or both is compositionally modified.

4.2 Methodology

Retrieval of peak conditions for rocks of the Uvauk complex incorporated as many of the above recommendations as possible. Samples of largest possible grain size were selected, although samples from within zones of high ductile strain were ubiquitously fine grained. At thin section scale, larger grains were preferred over small ones to reduce the potential effects of intragranular diffusional re-equilibration.

Ferromagnesian phases separated by quartz or plagioclase were analyzed to minimize the results of resetting caused by intergranular diffusion. The TWEEQU multi-equilibrium approach (Berman, 1991) was used for P and T calculations, providing the appropriate mineral assemblage was present.

The following set of independent equilibria were chosen in this study as the most robust. Reactions are written with the higher-*T* assemblage on the right, using mineral abbreviations of Kretz (1983), in addition to: Al-Opx = Al-bearing orthopyroxene, Ri = richterite, and Ts = Mg-tschermakite.

Thermometers

- 1) Alm = Al-Opx + 3 Fsl
- 2) 3 En + 2 Alm = 3 Fsl + 2 Prp
- 3) Alm + 3 Di = Prp + 3 Hd
- 4) 3 Tr + 5 Alm = 5 Prp + FeTr
- 5) Alm + Phl = Prp + Ann
- 6) Ed + Ab = Ri + An
- 7) Ed + 4Qtz = Tr + Ab

Barometers

- 8) 3 Qtz + 2 Grs + Alm = 3 An + 3 Hd
- 9) 3 Qtz + Grs + 2 Alm = 3 An + 6 Fsl
- 10) 12 Qtz + 3 Ts + 2 Prp + 4 Grs = 12 An + 3 Tr
- 11) Grs + 2 Ky + Qtz = 3 An
- 12) Grs + Qtz + 2 Rt = 2 Ttn + An

Fe-pyroxene end-members were used, as these equilibria are considered to be more robust than those involving Mg-endmembers. Mg-endmember amphibole equilibria were preferred because of the large uncertainties involved in estimating Fe³⁺ in amphiboles (Spear and Kimball, 1984; Holland and Blundy, 1994). Composition X-ray maps were

prepared for samples 98TXM178d, 98TXM201, 98TXM304, and 99TXM561b). Careful microprobe traverses were conducted across most phases so that any zoning might be discerned.

Correction of orthopyroxene - garnet temperatures was applied only when a large disparity was found between the temperature recorded by the Fe-Mg exchange thermometer (#2) and the Al-Opx thermometer (#1; Aranovich and Berman, 1997), which is based on the solubility of Al_2O_3 in orthopyroxene. $\text{Fe}/(\text{Fe} + \text{Mg})$ was increased in orthopyroxene in order to shift equilibrium #2 until it intersected equilibrium #1 at the pressure determined from equilibrium #9. The results of convergence of equilibria #1 and #2 are discussed where applicable in section 4.4.

All thermobarometric calculations with the exception of equilibria 1, 6 and 7 were computed using TWEEQU version 1.02, since only this version includes amphibole end-members that are present in equilibria #4 and #10. Version 1.02 incorporates the following activity models: Berman (1990), Fuhrman and Lindsley (1988), McMullin et al. (1991), Newton (1983), and Mader and Berman (1992) for garnet, plagioclase, biotite, pyroxene and amphibole, respectively.

The garnet-orthopyroxene thermometer of Aranovich and Berman (1997; equilibrium #1) was calculated using TWQ 2.02. This procedure does not introduce significant error because differences in thermodynamic parameters of orthopyroxene and garnet used by the two versions are very small (see discussion in Sanborn-Barrie, 1999). End-member and solution properties used in TWQ 2.02 are from Berman and Aranovich (1996).

Equilibria #6 and #7 were calculated using the program of Holland and Blundy (1994) and applied to all rocks containing amphibole. These equilibria were generally preferred to #4, because unrealistically high T estimates were calculated with the latter for many samples. Equilibrium #4 is based on Fe-Mg exchange between garnet and hornblende, and may be affected by problems in estimating Fe^{3+} in amphibole. Equilibrium #6 was used for anorthositic rocks since this equilibrium does not require the presence of quartz, and is recommended for high temperature samples with anorthite-rich feldspars and low silica activity (Holland and Blundy, 1994). Equilibria #7 was used for silica-saturated samples, such as amphibolites north and south of the meta-anorthosite suite.

Preparation for microprobe analysis involved the selection of points to be analysed, and programming the coordinates on these points in each of three micro-domains to be analysed by GSC micro-analyst, Katherine Venance. In general, the points were selected to traverse across garnet in order to detect any evidence of chemical zoning. Selected probe points were more closely spaced toward garnet rims in order to detect small compositional changes close to the rims. Where possible, garnet rims were analysed adjacent to plagioclase, in order to reduce the effects of late Fe-Mg exchange, and to determine the composition of garnet most likely to be in equilibrium with plagioclase. Additional points were analysed in plagioclase adjacent to and progressively farther from garnet to detect any zoning or compositional heterogeneity. Analysis of mafic minerals was less rigorous, with typically three points analysed from core to rim. Where possible, mafic minerals not immediately adjacent to another Fe-Mg mineral were analyzed. Fe-Mg zoning is less pronounced in other mafic minerals than in garnet, due to

the higher diffusivity of the former (Pattison and Bégin, 1994). Selected probe analyses are included in Appendix II.

P-T results are reported as either “core” or “rim” conditions. Thermobarometric results using rim compositions are generally considered most robust, as chemical equilibrium is most likely to be attained between the rims of proximal grains. In order to estimate latest P-T conditions, rims of adjacent garnet and plagioclase were selected, along with rims of proximal mafic phases separated from each other by quartz and/or plagioclase. In order to estimate early metamorphic conditions, garnet cores were used in conjunction with appropriate phases that occur as inclusions within garnet. However, where inclusions are absent, garnet core compositions were used in conjunction with foliation-forming matrix minerals that are assumed to have been in equilibrium with garnet cores. The similarity between compositions of hornblende and plagioclase inclusions, where present, compared to matrix hornblende and plagioclase compositions lends good support to the application of this assumption to other microdomains or samples where garnet inclusions are not present.

4.3 Sample selection and location

All samples for geothermobarometry were collected from the western portion of the Uvauk complex (Area A of Figure 1.3). Sample selection was based on: 1) presence of metamorphic assemblages for which P and T can be calculated, 2) lack of discernable alteration or retrograde metamorphism, and 3) relatively coarse grain size.

Within the anorthosite suite, two mafic dikes (98TXM333 and 98TXM342) as well as three garnet-anorthosites (98TXM304, 99TXM561b, 99TXM564a) were sampled for geothermobarometry. Sample 99TXM564a is a garnet-bearing anorthositic gabbro

from the southernmost margin of the anorthosite, collected ~20 m from the location of the leucogabbro (sample 98TXM334b) dated at ca. 2711 Ma (see chapter 5). The other two samples collected from within the anorthosite complex are from the northern margin.

Two samples from wall rocks north of the anorthosite suite were analysed. Sample 98TXM320 is a garnet-biotite amphibolite from ~200m north of the anorthosite (Figure 4.1). The protolith is not apparent, and there is no evidence for any earlier higher grade assemblage. This amphibolite-facies sample lies between granulite-facies rocks to the north and south, and was analysed with the hope of shedding light on this apparent pattern. Sample 98TXM248 is a charnockite that forms veins in mafic granulites ~2.5 km north of the anorthosite suite (Figure 2.10; Figure 4.1). A genetic link between this rock and the anorthosite of the UC is possible but was not tested for this study.

Two mylonites from the Southern bounding shear zone were sampled for thermobarometry. Sample 98TXM350d is a two-pyroxene monzodiorite gneiss and sample 98TXM178d is a garnet-quartzofeldspathic gneiss. These two rock types are abundant in the South-bounding shear zone (Figure 3.1), are intercalated, and are interpreted to have been deformed and metamorphosed together.

Three samples from the Southern domain were analysed for thermobarometry. Because the southern domain comprises dominantly tonalite and amphibolite (commonly garnet-free), which lack appropriate mineral assemblages to calculate both P and T, less abundant rock types were sampled. Two garnet amphibolites (98TXM201 and 98TXM339a) and one semipelite (98TXM339b) were used to characterise the P-T conditions of this domain, and to compare with results from south of Chesterfield Inlet.

4.4 Results

The main petrographic features of each sample are first described, followed by a description of chemical variations, and the results of P-T calculations are then discussed (Table 4.1). Results for rocks from within the complex are presented first, followed by results for rocks south and north of the complex.

4.4.1 Rocks within the Anorthosite suite

a) Garnet-bearing leucogabbro (99TXM561b)

A sample was collected from the Northern bounding shear zone, at the northernmost margin of the meta-anorthosite suite. The rock is well foliated, with compositional banding between plagioclase- and hornblende-rich layers of mm-scale thickness. The foliation envelops garnet porphyroblasts that range in diameter from 1-2 cm. The modal mineralogy is 50% plagioclase, 30% garnet, 10% hornblende, 6% titanite, 2% quartz, 1-2% clinopyroxene, ilmenite, and rutile. Clinopyroxene ranges up to ~200 μm in size and is invariably rimmed by amphibole. Garnet contains abundant inclusions of plagioclase, hornblende, titanite, ilmenite, quartz, and rare rutile. Porphyroblasts are subhedral, highly fractured, and are surrounded by ~1 millimetre wide hornblende coronae or hornblende-plagioclase symplectite. In leucocratic bands, anhedral garnet porphyroblasts average ~500 μm in diameter. These bands contain 5-10% quartz which forms elongate aggregates intergrown with zoned plagioclase with low-An cores and high-An rims (Figure 4.2a).

All mineral phases are pre-tectonic, with the exception of symplectitic plagioclase and hornblende. Symplectitic intergrowths of plagioclase and hornblende radiate around scalloped and embayed garnet rims, and are best developed in pressure shadows. They are interpreted to have formed from garnet-breakdown via an idealized reaction such as:

or the reaction may involve garnet and sodic plagioclase $\pm$ clinopyroxene reacting to produce hornblende and anorthitic plagioclase. Given the large volume change of any garnet-plagioclase reaction, the symplectite most likely formed during decompression. Although only rare, relict clinopyroxene was noted in the sample, its presence indicates that the rock probably once contained the high-P granulite-facies assemblage: garnet-clinopyroxene-plagioclase $\pm$ quartz. Hydration of clinopyroxene produced common hornblende rims.

One large (~ 2 cm diameter) garnet porphyroblast (garnet #1), partially rimmed by well-developed symplectite, was analysed in detail. The garnet is strongly zoned, and contains a low CaO core and high CaO rims (Figure 4.2b). Core analyses range in grossular content between $X_{\text{Grs}} = 0.26$ -0.30. High CaO rims are best preserved where hornblende-plagioclase symplectitic intergrowths are not well developed around the margin of the garnet. High CaO rims are ~ 500 -750 μm wide, show an increase in X_{Grs} from 0.35 to as high as 0.49, and decrease toward the outer rim to as low as 0.33. $\text{Fe}/(\text{Fe}+\text{Mg})$ is typically ~ 0.78 -0.81 within the core, and increases from ~ 0.77 at the core-rim interface (~ 700 μm from outermost rim) to ~ 0.84 at the garnet rim. In contrast, where hornblende-plagioclase symplectite is well developed at the garnet rim, X_{Grs} does not exceed ~ 0.37 , and $\text{Fe}/(\text{Fe}+\text{Mg})$ does not increase, but decreases slightly to about 0.77 (Figure 4.2b). X_{Sps} is also slightly elevated in the core (~ 0.04), relative to the rim (~ 0.008).

One microporphyroblast of garnet (~ 400 μm in diameter; garnet #2) from a leucocratic matrix band was analysed to compare with the large porphyroblast described

above. This anhedral garnet is not rimmed by a hornblende-plagioclase symplectite, but is associated with the previously described aggregates of quartz and zoned plagioclase (Figure 4.2a). The microporphyroblast possesses very high grossular content, with X_{Grs} that decreases from the core (0.59) to the rim (0.48). $\text{Fe}/(\text{Fe}+\text{Mg})$ is high relative to the 2 cm porphyroblast, ranging from 0.89 in the core to 0.91 at the rim.

Matrix plagioclase has a composition of $\sim\text{An}_{48-53}$. Anorthite content increases within the symplectite structure. From the present data set, it is not clear whether anorthite content peaks in the middle of the symplectite at $\sim\text{An}_{82}$, and then decreases to $\sim\text{An}_{67}$ adjacent to the garnet rim, or whether there is a heterogeneous distribution of plagioclase compositions within the symplectite. Where the symplectite is not well developed, higher anorthite content (An_{64}) appears to characterize the matrix plagioclase within ~ 1 mm of the garnet rim while lower anorthite ($\sim\text{An}_{75}$) plagioclase occurs adjacent to hornblende ~ 500 μm from the garnet rim. Plagioclase within a 250 μm radius of the garnet has a uniform composition of An_{62} , and this is significantly lower than the anorthite content within well developed symplectite ($\sim\text{An}_{82}$). These observations where the symplectite is not well developed suggests that the distribution of plagioclase compositions in the well-developed symplectite is due to heterogeneity, not to a systematic decrease in An content towards the garnet.

Plagioclase inclusions are abundant within the large garnet porphyroblast, and are generally unzoned, with the exception of a very minor increase in An content immediately adjacent to the garnet. Inclusions within the low CaO core of the garnet range in composition between An_{42} and An_{48} , somewhat lower than that observed in matrix plagioclase. Two plagioclase inclusions, ~ 200 μm in diameter, from the outer 500

µm high CaO rim of the garnet contrast markedly in composition: An₅₈ and An₇₀. The inclusions are similar in size and in distance to the outer rim of garnet. The only notable difference is that the inclusion with higher An content is farther from the low-high CaO interface. The fact that rare rutile was found along with calcic plagioclase as inclusions within the high-CaO garnet rim suggests that the increase in CaO in both garnet and plagioclase may be related to titanite breakdown via a reaction such as:

The significance of rutile occurring within the high CaO garnet rim is unknown, but increases in grossular content have been similarly attributed to the breakdown of other Ca-phases such as epidote or clinozoisite in lower grade rocks (Menard and Spear, 1993; Crawford, 1977).

Minor clinopyroxene is present in the matrix, and is rimmed by hornblende to actinolitic hornblende. All amphiboles are calcic amphiboles, with generally high Fe content. Foliation-forming elongate polycrystalline aggregates of matrix amphibole range in composition between Fe-hornblende and actinolitic hornblende. Amphiboles associated with garnet rims where symplectitic intergrowths are poorly developed are generally ~500 µm in length, and have compositions between Fe-hornblende and Fe-actinolitic hornblende (Leake, 1978). In contrast, symplectitic amphiboles are 50-100 µm in length, and range in composition between Fe-tschermakite and ferro-edinite. Amphibole inclusions within the two-centimetre garnet porphyroblast are tschermakitic in composition, contain the lowest Fe/(Fe+Mg) and highest Na and Al of all grains analysed from the sample. Na and Al are higher in symplectitic hornblende than matrix hornblende.

Three sets of *P-T* conditions were calculated for the rock. Conditions under which the garnet core formed were estimated based on the composition of the low-CaO garnet core and an inclusion of hornblende and plagioclase within this core. The inclusion comprises a very slightly zoned (An₄₇ core – An₄₉ rim) plagioclase grain, ~200 µm in length, and a tschermikitic amphibole grain that is ~50 µm in diameter. The *P-T* estimate uses plagioclase from the core of the inclusion to avoid the effects of later reaction with the surrounding garnet. Hornblende farthest from the surrounding garnet, adjacent to plagioclase was used to reduce the potential effects of late Fe-Mg exchange between garnet and amphibole.

Conditions under which the high CaO garnet rim formed were estimated using an unzoned (An₅₈) plagioclase inclusion, ~200 µm in diameter, from within the ~750 µm wide, high-CaO garnet rim. Because no amphibole inclusions were found within this part of the garnet, the amphibole inclusion from the inner inclusion, described above, was used. Since chemical equilibrium between these compositions is uncertain, the result can only be considered an approximation. Based on the similarity of garnet compositions, high-CaO rim conditions are likely also given by the core of the microporphyroblast described above.

Conditions of garnet breakdown were estimated based on the composition of symplectitic plagioclase and hornblende within 200 µm of the rim of the large garnet porphyroblast (Figure 4.2b). In addition, the extreme rim of the garnet microporphyroblast (400 µm in diameter), together with rims of nearby hornblende and plagioclase, were used to extract late metamorphic conditions.

Thermobarometric determinations are based on equilibria #4, #6, and #10. An average of the results indicated by thermometers #4 and #6 was used where the two agree within $\sim 100^\circ\text{C}$. The garnet core, and plagioclase and hornblende inclusion from within the garnet core, yield conditions of 11.0 kbar, and 740°C based on thermometer #4. The same compositions yield 11.7 kbar and 785°C based on thermometer #6. The average P-T conditions for the core of the large garnet porphyroblast are ~ 11.3 kbar and 760°C (Figure 4.3a).

Conditions for the outer, high-CaO garnet rim are estimated at 14.3 kbar and $\sim 780^\circ\text{C}$, based on the average of results using barometer #10 together with thermometers #4 and #6 (which yield 12.6 kbar, 745°C and 14.4 kbar and 840°C , respectively; Figure 4.3b), and using thermometer #4 with barometer #12 (which yields 15.8 kbar and 780°C). Results using the higher anorthite inclusion, adjacent garnet, and the same hornblende inclusion from the garnet core yield average conditions of 15.1 kbar and 800°C . The core of the garnet microporphyroblast (garnet #2) yields 12.5 kbar and 840°C with equilibria #4 and #10, and 10.8 kbar and 755°C with equilibria #3 and #8. However, the high Fe/(Fe+Mg) ratio (0.89 at the core, 0.91 at the rim), relative to the Fe/Mg of the garnet #1 rim (<0.84) suggest the possibility that garnet #2 has re-equilibrated due to late stage Fe-Mg exchange.

Calculation of garnet breakdown conditions, determined using symplectite and garnet #1 rim compositions, yields ~ 11.5 kbar and 790°C with equilibria #4 and #10 (Figure 4.3c). This likely represents a maximum for the P-T estimate of garnet breakdown, since the garnet #1 rim elsewhere has a lower grossular content ($X_{\text{Grs}} = 0.30$ - 0.33 vs. 0.37 for g34). Temperatures calculated using the lower garnet X_{Grs} are

unrealistically high ($>900^{\circ}\text{C}$), due to the composition of nearby amphibole used in the calculation. The rim of the garnet microporphyroblast (with $X_{\text{Grs}} = 0.48$), and nearby matrix plagioclase and hornblende yield conditions of 10.0 kbar and 800°C , with thermometers #4 and #6 in good agreement (Figure 4.3d).

The following general P-T-t picture emerges from the P-T conditions and textural relationships described above. Earliest conditions recorded by the rock are ~ 11.3 kbar and 760°C . The high CaO garnet rims formed at higher pressure, possibly as high as 14.3 kbar, and a slightly higher temperature ($\sim 780^{\circ}\text{C}$), although this is not well constrained due to the absence of hornblende inclusions in the garnet rim. Latest metamorphic conditions are marked by post-tectonic breakdown of garnet to produce the hornblende-plagioclase symplectite at slightly lower pressures and higher temperatures: 10.0 kbar and 800°C . The above conditions and textures could have been produced during one, protracted event, but similarity with textures in other rocks described below suggest formation during at least two events. Pre-tectonic garnet with low-CaO cores may have formed under prograde conditions during the first event. The high-CaO garnet rims formed at higher pressures (~ 3 kbar) and slightly higher temperatures ($\sim 20^{\circ}\text{C}$), possibly during a second event with subsequent post-tectonic, ~ 10 kbar decompression at high temperatures producing the symplectitic plagioclase and hornblende intergrowths.

b) Garnet-bearing gabbro anorthosite (98TXM304)

This sample was collected from the Northern bounding shear zone, and displays a well-developed foliation, defined by a grain-scale shape fabric parallel to compositional banding. The average modal mineralogy comprises $\sim 60\%$ plagioclase, $\sim 17\%$ hornblende, $\sim 15\%$ garnet, $\sim 5\%$ titanite, 2% quartz, 1% ilmenite, trace apatite and biotite. Garnet

porphyroblasts range up to ~8 mm in diameter, and are rimmed by hornblende-plagioclase intergrowths, which, in turn, are rimmed by fine-grained hornblende coronae. Locally, complete breakdown of garnet to plagioclase and hornblende is evident as garnet pseudomorphs up to 1 mm in diameter that are wrapped by the foliation, but consist of radial intergrowths of plagioclase and hornblende.

Titanite occurs in the matrix aligned within the foliation, and as inclusions in some garnet porphyroblasts. Titanite is interpreted to be pre-tectonic, and in this sample shows no evidence for the breakdown of titanite to produce rutile.

The random orientation of amphiboles in garnet pseudomorphs is indicative of post-tectonic growth. However, much of the hornblende in mafic bands is within the foliation, suggesting that hornblende formation occurred during deformation, and continued after deformation ceased. Although no clinopyroxene was discerned in this sample, it may have been present and have reacted to form amphibole. Evidence for this is present in sample 99TXM561b where clinopyroxene is ubiquitously rimmed by hornblende, and this texture is interpreted to reflect the hydration of clinopyroxene to form hornblende.

As with 99TXM561b, quartz inclusions, although rare, occur in some garnets, indicating the presence of quartz prior to garnet formation. The texture documented for 99TXM561b, comprising polygonal aggregates of quartz with reversely zoned plagioclase also occurs in this sample although at a much finer scale (Figure 4.4a).

Plagioclase and garnet are zoned, whereas hornblende compositions are homogeneous at thin section scale. The composition of matrix plagioclase is ~An₄₀₋₄₅. Within the hornblende-plagioclase coronae, plagioclase compositions are high (An₈₀₋₈₅)

in distal parts of the corona, and decrease to $\sim\text{An}_{65-70}$ adjacent to the garnet (Figure 4.4). Plagioclase inclusions within garnet are not zoned and have an average composition of An_{42} , identical to the matrix plagioclase composition. This indicates that matrix plagioclase was likely in equilibrium with garnet cores, and lends greater confidence to core P-T calculations based on matrix plagioclase compositions for other microdomains where plagioclase inclusions are absent. Plagioclase compositions used for each core P-T estimate range from An_{42-51} , whereas symplectitic plagioclase compositions, used for rim determinations range from An_{64-85} .

Hornblende composition is predominantly Fe-tschermakitic hornblende to Ferro-tschermakite, with compositions rarely plotting in the Fe-pargasitic hornblende field (Leake, 1978). Hornblende rims are slightly higher in Al content than cores, but are generally homogeneous at the scale of the thin section. There is no notable chemical difference between matrix hornblende and symplectitic hornblende.

Compositional profiles for different garnet porphyroblasts exhibit a subtle difference in CaO, but differ markedly in $\text{Fe}/(\text{Fe}+\text{Mg})$. Garnet B has a flat $\text{Fe}/(\text{Fe}+\text{Mg})$ profile (0.81-0.83), and exhibits a minor gradual decrease in CaO from $X_{\text{Grs}} = 0.29$ at the core to 0.26 at the rim. Garnet C has $X_{\text{Grs}} = 0.25-0.26$ and possesses a $<20\ \mu\text{m}$ rim which exhibits an increase in $\text{Fe}/(\text{Fe}+\text{Mg})$ from 0.82 to 0.91 at the outer rim adjacent to well-developed symplectite (hornblende within $25\ \mu\text{m}$ of garnet rim). Garnet A exhibits a decrease in $\text{Fe}/(\text{Fe}+\text{Mg})$ from 0.85 at the core, to 0.73-0.76 within the outer $\sim 500\ \mu\text{m}$ of the 4 mm garnet grain. X_{Grs} is consistently between 0.30-0.31, but decreases to as low as 0.25 within $\sim 50\ \mu\text{m}$ of the rim where symplectite is not well developed (Figure 4.4b). There is no evidence of high-CaO garnet rims in this rock.

The most significant difference between the three garnets is size: garnet A is 4 mm in diameter, whereas garnets B and C have diameters of 1 mm and 450 μm , respectively. It is not clear whether the 500 μm low Fe/(Fe+Mg) rim of garnet A was produced during garnet growth or during diffusive re-equilibration at high temperature. The fact that garnet A also shows a pronounced drop in CaO within 50 μm of the rim that remains *after* partial garnet breakdown suggests that both compositional features were produced by high temperature diffusive re-equilibration. In this case, the small garnets would have had their Fe/(Fe+Mg) completely reset, producing homogeneous low Fe/(Fe+Mg) profiles. The increase in Fe/(Fe+Mg) exhibited by the outer 20 μm of garnet C is probably an effect of late cation exchange with adjacent symplectitic hornblende.

The larger garnet from microdomain A has the highest potential for preserving prograde metamorphic conditions, and may also provide information on the P-T-t path incurred by the rock. Early prograde P-T conditions are based on the high Fe/(Fe+Mg) garnet core together with the cores of matrix plagioclase and hornblende. Later prograde conditions are based on the garnet inner rim, with the symplectitic plagioclase and hornblende. Conditions of garnet breakdown were determined using compositions of the outermost garnet rim where it shows a decrease to lower X_{Grs} content, nearby symplectitic plagioclase with high anorthite content, and the inboard rim of symplectitic hornblende. Symplectitic hornblende was used with garnet core and matrix plagioclase for late prograde estimates from microdomains B and C because matrix hornblende was not analysed in these domains. As discussed above, the cores of garnets B and C are considered equivalent to the rims of garnet A due to re-equilibration.

Thermobarometric results were calculated using equilibria #6 and #10. Results for microdomain A are 11.3 kbar and 730°C for the core, 12.2 kbar and 720°C for the inner rim, and 10.8 kbar and 780°C for garnet breakdown, as indicated by symplectitic compositions. Core conditions are 12.1 kbar and 730°C, and 11.3 kbar and 750°C, for microdomains B, and C, respectively. As expected, these conditions are similar to those yielded by the outer rim of garnet A. Conditions based on garnet rim compositions are 10.9 kbar and 790°C, and 9.5 kbar and 745°C, for microdomains B, and C, respectively. The lowest P-T is recorded by the rim of garnet C (9.5 kbar and 745°C), which is the only calculation for which both thermometers agree. This may represent a lower limit on the P-T conditions under which the symplectite formed, or, perhaps more likely, the garnet rim may have been modified by late Fe-Mg exchange. The higher temperatures recorded by the larger garnets A and B (~10.8 kbar and 780°C) are therefore preferred. The results are consistent with garnet growth during one prograde event initiated at ~11.3 kbar and 730°C, continuing to ~12.2 kbar and 720°C, and culminating with garnet breakdown at 10.8 kbar and 780°C (Figure 4.4b).

c) Leucogabbro - garnet-bearing anorthositic gabbro (99TXM564a)

This sample was collected from the Southern bounding shear zone, near the site of the metaleucogabbro dated in this study (Figure 4.1; Figure 5.1). Its modal mineralogy comprises ~55% plagioclase, ~28% garnet, ~10% clinopyroxene, ~5% orthopyroxene, ~2% hornblende, trace quartz, biotite, pyrite, and alumino-silicate. Compositional layering and elongate aggregates of mafic minerals in mafic bands define a well developed foliation. Subhedral garnets average ~400 µm, contain inclusions of quartz and alumino-silicate and are rimmed by a ~10 µm rind of symplectitic orthopyroxene and

plagioclase, which is best developed in pressure shadows. Rare clinopyroxene with exsolution lamellae are rimmed by 50 μm orthopyroxene subgrains. These textural relationships suggest that clinopyroxene is a primary igneous mineral, whereas orthopyroxene probably formed during metamorphism.

Chemical zoning is evident in garnet, plagioclase, and pyroxenes. From core to rim, garnet decreases in X_{Grs} from ~ 0.25 to ~ 0.20 , and increases in $\text{Fe}/(\text{Fe}+\text{Mg})$ from 0.48 to 0.62 (Figure 4.6). Matrix plagioclase has a composition of $\sim \text{An}_{40-45}$ but An content increases up to An_{55} toward garnet. Both pyroxenes have slightly higher $\text{Fe}/(\text{Fe}+\text{Mg})$ ratios at rims not touching another ferromagnesian phase. Rims next to ferromagnesian phases were not analysed.

Reported P-T values for cores are based on the compositions of garnet cores, and the cores of matrix plagioclase, orthopyroxene and clinopyroxene. Rim P-T values use rim compositions of all phases, with mafic phases separated by plagioclase where possible, in order to avoid the effects of late cation exchange between ferromagnesian minerals. Because orthopyroxene rims clinopyroxene, there is likely to have been Fe-Mg exchange between the two. The small amount of orthopyroxene present in the mafic aggregates is likely to have been more significantly modified by exchange than the more abundant clinopyroxene. Fe-Mg exchange between adjacent pyroxenes during cooling would have resulted in diffusion of Fe into orthopyroxene and Mg into clinopyroxene, effectively shifting equilibria #2 and #3 to higher and lower temperatures, respectively. Although P-T results are based on equilibrium #1, this thermometer is also likely to have been shifted to higher temperature, although to a lesser extent than the shift of

equilibrium #2. For this reason, all P-T estimates from sample 99TXM564a should be regarded as maxima.

Equilibria #1 and #9, involving orthopyroxene, were used to estimate P-T conditions based on both core and rim compositions. Equilibria involving clinopyroxene are also plotted in Figures 4.7a and b, but are not considered as reliable because this phase is interpreted to be igneous in origin and, therefore, likely not in equilibrium with garnet cores. Similarly, core calculations based on orthopyroxene compositions may not be reliable because orthopyroxene, which rims clinopyroxene, may have formed *after* garnet growth and may, therefore, not have been in equilibrium with garnet cores. Based on equilibria #1 and #9, cores yield the following P-T conditions: 13.6 kbar and 1020°C, 13.2 kbar and 990°C, and 13.3 kbar and 1040°C, for microdomains A, B, and D, respectively (Figure 4.7a). These high temperature estimates are interpreted to be an erroneous artifact resulting from likely disequilibrium between phases and from Fe-Mg exchange between adjacent orthopyroxene and clinopyroxene during cooling. Rim compositions yield lower P-T conditions: 11.7 kbar and 840°C, 10.8 kbar and 880°C, and 12.7 kbar and 835°C, for microdomains A, B, and D, respectively (Figure 4.7b). Although these calculations may have been modified by Fe-Mg exchange, they are considered to reflect maximum metamorphic conditions since orthopyroxene and garnet rims are most likely to have been in equilibrium. Since microdomain B is from a more melanocratic part of the sample, and more likely to have had pyroxene compositions perturbed by late Fe-Mg exchange, an average of results based on the rim compositions of microdomains A and D is taken as the best P-T estimate for the sample (12.2 kbar, 840°C).

d) Mafic dykes (98TXM333 and 98TXM342)

Thermobarometric data were obtained for two mafic dykes, 20–40 m wide, that intrude the anorthosite suite, but have been deformed and metamorphosed along with their host rocks. Most dykes exhibit a strong foliation and lineation and may be folded along with the anorthosite. However, the range in deformation textures and rare crosscutting relationships indicate that the dykes were emplaced during deformation (see Chapter 3). Both dykes (98TXM333 and 98TXM342) contain ~35% garnet, ~35% clinopyroxene, ~25% plagioclase (An_{40-50}), <5% quartz, minor orthopyroxene and hornblende, and trace ilmenite. Relict 100 μm orthopyroxene is present in both thin sections, and is generally mantled by clinopyroxene, up to ~500 μm in diameter. Hornblende is generally restricted to narrow zones (~1 mm) near microfractures. Garnet porphyroblasts are larger in 98TXM342, with average diameters of ~1 mm in 98TXM342, whereas they average ~0.4 mm in 98TXM333. Both rocks exhibit a moderately developed foliation, which is defined by elongated aggregates of clinopyroxene and plagioclase. Sample 98TXM342 also contains asymmetric extensional shear bands oriented at about 30° to the main foliation.

Garnet profiles for the two samples show similar trends in $\text{Fe}/(\text{Fe}+\text{Mg})$ ratios, but a notable difference in CaO trends (Figure 4.8 and 4.9). $\text{Fe}/(\text{Fe}+\text{Mg})$ gradually increases from 0.60–0.62 at the core to 0.67–0.68 at the rim in 98TXM342, and from ~0.67 at the core to 0.73–0.76 at the rim in 98TXM333. CaO profiles are flat ($X_{\text{Gr}} = 0.18$) across the smaller garnets of sample 98TXM333. Sample 98TXM342, however, shows higher CaO content in the cores ($X_{\text{Gr}} = 0.23$), decreasing to $X_{\text{Gr}} = 0.18$ at the rims.

The difference in CaO trends exhibited by the two samples may be a product of the timing of metamorphic mineral growth along a P-T path. The higher CaO cores of 98TXM342 are consistent with initial growth at higher pressures, with the lower CaO rims recording growth at lower P. The CaO composition of garnet in 98TXM333 is comparable to that of the rims of the 98TXM342 garnets, suggesting that garnets in 98TXM333 may have started growing after those of sample 98TXM342. Because diffusion of Ca (by way of NaSi – CaAl exchange) in plagioclase has a characteristic length scale of $<1\mu\text{m}$ at 600°C in 100 m.y. (Spear, 1993), it is unlikely that the smaller garnets of 98TXM333 were homogenized by diffusive re-equilibration. The occurrence of Fe/(Fe+Mg) zoning in both samples does, however, indicate late stage re-equilibration during cooling because of the greater diffusivity of Fe and Mg in garnet.

Plagioclase compositions vary slightly more for 98TXM342 (An_{40-50}) than for 98TXM333 (An_{40-45}). The anorthite content of plagioclase generally increases toward garnet. Orthopyroxene has a similar composition in both rocks, with 25-27% FeO and 18-19% MgO, and in both rocks is rimmed by clinopyroxene. Hornblende is generally developed at the interface between orthopyroxene and clinopyroxene, where it forms a rind that is $<10\mu\text{m}$ thick (Figure 4.9b).

Mantling of orthopyroxene by clinopyroxene indicates that orthopyroxene was stable before clinopyroxene growth initiated. However, it is not known whether orthopyroxene was stable during initial garnet growth, or *prior to* garnet growth as an igneous phenocryst, with clinopyroxene becoming the stable mafic phase at the onset of high-grade metamorphism. Because of this ambiguity, P-T results are based on clinopyroxene.

P-T estimates for both dykes were obtained using equilibria #'s 3 and 8. Peak conditions were based on core compositions of garnet and matrix phases between 200-500 μm from the garnet. Sample 98TXM333 yields P-T conditions of 10.0 kbar and 795°C, 9.5 kbar and 810°C, and 11.5 kbar and 860°C, for the cores of microdomains A, B, and C, respectively (Figure 4.10a). Sample 98TXM342 yields P-T conditions of 13.1 kbar and 870°C, 13.2 kbar and 910°C, and 11.4 kbar and 820°C, for the cores of microdomains A, B, and C, respectively (Figure 4.11a). The higher P-T conditions recorded by 98TXM342 relative to 98TXM333 are interpreted to reflect earlier growth of the former under higher pressures. The larger garnet size and better developed foliation in 98TXM342 are also compatible with this interpretation. The rim analyses from both samples yield slightly lower conditions. The rims of 98TXM333 yield 7.2 kbar and 660°C, 8.1 kbar and 765°C, and 8.6 kbar and 785°C, for microdomains A, B, and C, respectively (Figure 4.10b). The rims of 98TXM342 yield: 7.5 kbar and 720°C; 8.7 kbar and 790°C; and 6.3 kbar and 620°C; for microdomains A, B, and C, respectively (Figure 4.11b). Conditions calculated from rim compositions of 98TXM342 are low relative to those determined based on core compositions, compatible with modification of rim compositions by Fe-Mg exchange upon cooling while the cores of these garnets retain near-peak compositions. The average conditions determined from the cores of 98TXM342 (12.6 kbar and 865°C) is interpreted to represent maximum P-T conditions under which the dykes were metamorphosed.

4.4.2 Rocks of the Southern bounding shear zone

a) Quartzofeldspathic gneiss (98TXM178d)

The garnet-quartzofeldspathic gneiss, south of the anorthosite suite (Figure 4.1), exhibits a strong foliation and lineation predominantly defined by elongate polycrystalline plagioclase aggregates. The rock consists of ~30% garnet, ~40% plagioclase, ~20% quartz (half of which is interpreted to be vein quartz), ~5% biotite, <5% orthopyroxene or cummingtonite (the two are mutually exclusive), and trace ilmenite, pyrite, rutile, monazite, and zircon. Anhedral garnet porphyroblasts range from 500 μm to 1 mm, are wrapped by quartz veins and ribbons, and commonly contain quartz inclusions. Orthopyroxene is typically <100 μm , generally occurs near garnet, especially within pressure shadows (Figure 4.12), and is interpreted to reflect the metamorphic grade under which deformation occurred. Cummingtonite is randomly oriented and may have replaced orthopyroxene, as it occurs only in samples that lack orthopyroxene. Fine-grained (20-50 μm) biotite is randomly oriented, indicative of post-tectonic growth, within bands ~100 μm in thickness. Plagioclase porphyroclasts range up to 1 mm and are mantled by subgrains and new grains (neoblasts) indicative of grain size reduction by deformation in a ductile regime (see Chapter 3).

Garnet compositional profiles for the gneiss show low CaO cores ($X_{\text{Grs}} = 0.05\text{--}0.07$), and ~50-100 μm wide high CaO rims ($X_{\text{Grs}} = 0.08\text{--}0.11$) that decrease toward the extreme rims ($X_{\text{Grs}} = 0.06\text{--}0.08$) (Figure 4.12a). The increase in CaO appears generally quite abrupt on X-ray maps, but is locally more diffuse and ranges between ~50-100 μm in thickness (Figure 4.12b). Deflection of a quartz ribbon around the garnet is evident in Figure 4.12b, indicating pre-tectonic growth of the garnet core with respect to most recent deformation. Apparent truncation of the quartz ribbon (lower right corner) is consistent with growth of the high-CaO garnet rim *after* deformation, but this relationship

was not commonly observed. It is unclear whether the garnet was resorbed prior to growth of the high-CaO rims, or if the older low-CaO core was overgrown by high-CaO rims and subsequently resorbed. $\text{Fe}/(\text{Fe}+\text{Mg})$ mimics the CaO trend, with $\text{Fe}/(\text{Fe}+\text{Mg})$ increasing gradually from 0.60-0.62 at the cores to 0.66-0.69 at the rims.

Plagioclase compositions vary between An_{21-29} , with no apparent relation to garnet proximity or from core to rim within individual grains. Orthopyroxene is generally unzoned, but locally exhibits a marked increase in Al and slight decrease in $\text{Fe}/(\text{Fe}+\text{Mg})$ toward the rim.

Application of equilibria #'s 1, 2 and 9 yields consistent results, with higher temperatures generally recorded by the Al-opx net transfer reaction (equilibrium #1), most notably for analyses using rim compositions. This is compatible with some degree of late cation exchange, to which equilibrium #2 is more susceptible. For this reason, equilibrium #1 is preferred for estimation of near-peak conditions.

P-T estimates based on core compositions are consistently lower than those based on rim (especially inner rim) compositions. Equilibria calculated from core compositions use the core of garnet, and orthopyroxene and plagioclase from the matrix. P-T conditions based on core compositions are 8.6 kbar and 725°C, 9.2 kbar and 755°C, and 7.5 kbar and 695°C, for microdomains A, B, and C, respectively (Figure 4.13a). Rims give higher P-T values of 10.7 kbar and 775°C, 10.7 kbar and 795°C, and 12.9 kbar and 850°C, for microdomains A, B, and C, respectively (Figure 4.13a). Although the upper end of core results approaches the lower end of rim results, the abrupt transition from low to high CaO in garnet, and from lower to higher Al_2O_3 in opx, are more compatible with two metamorphic events.

b) Monzodioritic mylonite gneiss (98TXM350d)

Green to grey monzodioritic mylonite gneiss is intercalated with garnet-quartzofeldspathic gneiss and shares the same tectonic fabric orientations and strain intensity. It contains ~25% clinopyroxene, 20% orthopyroxene, 28% plagioclase (An_{45-48}), 20% k-feldspar, 3% biotite and opaques, 2% quartz, with rare subhedral garnet microporphyroblasts, and trace actinolite, apatite and zircon. Grain size is generally <50 μm , but some porphyroclasts of clinopyroxene and K-feldspar may be as large as 2 mm. Garnet typically occurs as ~50 μm , subhedral neoblasts within pressure shadows of pyroxene porphyroclasts (Figure 4.14). Ribbons of plagioclase, K-feldspar, and elongated pyroxene grains with winged subgrain extensions define the foliation. A well-developed extension lineation is defined by plagioclase and K-feldspar aggregates as well as elongate aggregates of pyroxene. All phases appear to have grown prior to deformation except randomly oriented microlaths of biotite.

Garnet exhibits flat CaO profiles ($X_{Gr\text{ts}} = 0.15$) (Figure 4.14) and either a flat Fe/(Fe+Mg) profile, or a slight increase toward the rim. Plagioclase is slightly more calcic at rims adjacent to clinopyroxene or garnet (An_{48-50} adjacent to garnet; $\sim An_{45-47}$ in the matrix). Myrmekitic texture, comprising fine vermiform K-feldspar and plagioclase, occurs in areas with little to no mafic minerals. These intergrowths must have formed after deformation in order for the randomly oriented texture to be preserved, and thus points to a metamorphic origin for the myrmekitic texture. Porphyroclasts of clinopyroxene and orthopyroxene are rimmed by a ~10-50 μm thick veneer of actinolite, suggesting late fluid infiltration at low-grade conditions.

P-T estimates from sample 98TXM350d are similar to those determined from the cores of 98TXM178d, but significantly lower than those determined based on rim compositions. The garnet-orthopyroxene exchange thermometer (equilibrium #2) plots 100-200°C lower than the Al-opx thermometer (equilibrium #1) for two of the three microdomains. This suggests the rock re-equilibrated under post-peak P-T conditions, consistent with the short time required to diffusively homogenize Fe/(Fe+Mg) within a 50 μm diameter garnet at 700-800°C (see below; Spear, 1993). The third of the microdomains yields thermometers in close agreement.

In order to retrieve peak conditions from the monzodiorite, recalculation of P and T was carried out as discussed in the introduction to this chapter (section 4.1), based on the assumption that Fe-Mg exchange was the only process to have affected the rocks after peak metamorphic conditions. Fe/(Fe +Mg) was increased in orthopyroxene in order to shift equilibrium #2 significantly to higher T, while equilibrium #1 shifts only slightly to higher T. Orthopyroxene was selected instead of garnet, but use of either phase gives similar results. The P-T estimate obtained from microdomain D (Figure 4.15) without modification of Fe/(Fe+Mg) is 8.4 kbar and 750°C. Attempts to extract peak P-T conditions required a decrease in Mg, and an equal increase in Fe. This modification was first made with TWQ v. 2.02, to change the thermometers used. Similar calculations were then made with TWQ v. 1.02 so that the same Fe/(Fe+Mg) ratios were used in the barometer calculation. Equilibria #1 and #2 converge at 8.5 kbar and 800°C (Figure 4.15), a temperature increase of 50°C and a pressure increase of 0.1 kbar. This recalculated P-T is interpreted to best represent conditions under which the rock equilibrated. Application of the above recalculation yields 8.1 kbar and 785°C for

microdomain A, compared to 8.1 kbar and 630°C for uncorrected compositions. Microdomain B yields 8.8 kbar and 730°C, and does not require recalculation because equilibrium #2 plots very close to equilibrium #1.

4.4.3 Rocks south of the complex

a) Banded garnet-amphibolite (98TXM201)

A lens of banded garnet-amphibolite was sampled from the tonalite orthogneiss map unit, south of the highly strained rocks of the Uvauk complex. Concordance between the gneissosity of the tonalite gneiss and the amphibolite indicate that the amphibolite experienced the main tectonothermal event affecting the orthogneiss. The gneissosity in the amphibolite is defined by alternating melanocratic and leucocratic bands of ~2 cm width. The former comprise ~40% hornblende, ~30% garnet, ~20% plagioclase, 10% quartz, and minor biotite. Leucocratic bands comprise ~60% plagioclase, 25% quartz, 10% garnet, ~5% hornblende and trace ilmenite, rutile and pyrite. Garnet porphyroblasts are pre-tectonic, range from ~2 mm to ~1 cm in diameter, are highly fractured, and commonly contain large (0.1-0.5 mm in diameter) quartz inclusions. The coarse grain size of foliation-forming minerals relative to the thin high-CaO garnet rims (described below) prevented determination of the relative timing of growth of the garnet rims relative to foliation development. Porphyroblasts are surrounded by thin (~150 µm) discontinuous moats comprising an inner, ~50 µm wide plagioclase zone, surrounded by a ~100 µm wide hornblende zone (Figure 4.16). This texture is variably developed, and generally does not completely enclose garnet porphyroblasts.

Trends in mineral chemistry are somewhat uncertain because of the variability in the limited data set. Plagioclase composition varies between An_{28-38} with maximum An in plagioclase 20-50 μm from garnet rims, and slightly lower An content immediately adjacent to garnet. An content appears to decrease away from garnet, with matrix plagioclase compositions, An_{28} , occurring $\sim 150 \mu m$ from garnet rims. Proximity to hornblende appears to have the same effect, so that plagioclase adjacent to hornblende generally has low An content. The CaO content of garnet increases from core ($X_{Grs} = 0.19-0.20$) to rim ($X_{Grs} = 0.26-0.32$), but the nature of this increase is inconsistent between grains. Two of the three analyzed garnet porphyroblasts (C and G), both ~ 3 mm in diameter, show an abrupt transition to $\sim 150-200 \mu m$ wide, high CaO rims, whereas the other, ~ 4 mm garnet (A), exhibits a continuous increase over a distance of $\sim 600 \mu m$. Fe/(Fe+Mg) ratios are relatively flat at 0.83-0.85, with a minor increase to 0.87-0.93 at the extreme rim of all three garnet grains. One garnet porphyroblast (from microdomain G; Figure 4.16) also shows an increase in Fe/(Fe+Mg) at the edge of its core (defined as the last probe point with $X_{Grs} < 0.2$) which may be due to proximity to a fracture. All amphiboles are calcic amphiboles, and range from Fe-tschermakite to Fe-tschermakititic hornblende, with one amphibole plotting in the ferro-actinolite field (Leake, 1978).

P-T results use equilibrium #10 and Holland and Blundy's (1994) thermometer, equilibrium #7. Equilibria #7 is preferred to #6 because the rock contains abundant quartz and may not have been metamorphosed to extremely high P-T conditions (see discussion in section 4.2). P-T results are 8.7 kbar and 730°C for microdomain A, and 9.2 kbar and 705°C for microdomain C (Figure 4.17b), based on garnet core compositions, matrix plagioclase situated $\sim 100-500 \mu m$ from the garnet rim, and

hornblende $\pm$ biotite at least 500 μm from the garnet rim. P-T results for rims are based on the high-Ca rims of garnet grains ($X_{\text{GrS}} = 0.26\text{-}0.32$), the slightly more calcic matrix plagioclase within 100 μm from garnet rims, and the rims of hornblende $\pm$ biotite ~ 200 μm from garnet rims. These yield 10.1 kbar and 705°C, and 11.4 kbar and 690°C, for microdomains A, and C, respectively. Three sets of conditions were calculated for microdomain G: low-CaO core conditions, high-CaO rim conditions, and conditions of garnet breakdown to hornblende and plagioclase (Figure 4.17b). Core compositions yield conditions of 9.7 kbar and 685°C. The high CaO garnet rim, in conjunction with matrix plagioclase and the rim of matrix hornblende less than 1 mm from the garnet, yield 12.1 kbar and 690°C (Figure 4.16). Mineral compositions adjacent to the extreme rim of garnet yield 9.7 kbar and 705°C, which likely represents a maximum P-T, given that the garnet was not analyzed in detail to locate portions that may have changed to lower X_{GrS} .

b) Melanocratic amphibolite (98TXM339a)

Melanocratic, garnet-bearing amphibolite is intercalated with kyanite-garnet-biotite semipelite to pelite on the 1 km long peninsula in the southern part of the map area (Figure 4.1). The amphibolite comprises $\sim 70\%$ hornblende, $\sim 14\%$ quartz, $\sim 5\%$ plagioclase, $\sim 5\%$ garnet, $\sim 3\%$ biotite, $\sim 2\%$ clinozoisite, and $\sim 1\%$ ilmenite. The rock is massive and roughly equigranular with an average grain size of $\sim 0.5\text{-}1$ mm. Garnet instability is indicated by garnet pseudomorphs, which constitute $\sim 20\text{-}25\%$ of the rock, consisting of predominantly plagioclase, hornblende, biotite and clinozoisite (Figure 4.18). Relict garnet comprises only $\sim 5\%$ of the rock, and consists of resorbed, anhedral grains up to 1 mm in diameter, surrounded by ~ 1 mm wide moats of plagioclase, hornblende, biotite and clinozoisite, interpreted to have formed from garnet breakdown.

The pseudomorphs suggest that ~60-70% of the original garnet has been destroyed. The remaining garnet grains typically have flat CaO profiles, with $X_{\text{Grs}} \sim 0.26$ -0.30, and a marked X_{Grs} decrease (to ~0.22-0.25) near hornblende or plagioclase inclusions. In some microdomains, X_{Grs} decreases from core to rim, but the analysed “rims” are not the rims of the original garnet grains. The Fe/(Fe+Mg) ratio is generally ~0.90, and is flat across most grains. Low Fe/(Fe+Mg) ratios, ~0.85, are exhibited in two areas: in the relatively high CaO core of microdomain A, and at a relatively high CaO rim and a portion of the core of microdomain B (Figure 4.18). The significance of this is difficult to interpret due to limited preservation of garnet. However, the low Fe/(Fe+Mg) garnet rim in microdomain B is close to the edge of the pseudomorph structure (Figure 4.18), indicating that it may have been close to the original garnet rim.

Plagioclase is generally very calcic, with ~60-70% of the plagioclase compositions being ~An₈₀ or higher. An content appears to increase toward garnet or hornblende. Plagioclase with significantly lower An content (An₅₂₋₅₈) is present in only one of the three microdomains (microdomain A), and is situated at the outermost part of the garnet pseudomorph. A plagioclase inclusion from microdomain B has similar anorthite content to that in the surrounding zone of garnet breakdown (Figure 4.18).

Amphibole in 98TXM339a is relatively homogeneous calcic hornblende throughout the thin section, whereas the inclusion in garnet shows slightly lower Al, Na, and Fe/(Fe+Mg).

All P-T results use plagioclase from within the pseudomorph structure, with the exception of one plagioclase inclusion in garnet. The little plagioclase that occurs in the matrix is highly altered. Cores and rims of relic garnet grains are used. Matrix

hornblende is generally used for core analyses, whereas hornblende within the pseudomorph is used for rim analyses, although their compositions are similar. Therefore, P-T results using inclusions are considered the best representation of early prograde conditions, whereas conditions under which the pseudomorphs formed may be best represented as an average of the three domains studied.

Thermobarometric analysis use equilibria #'s 4, 7 and 10 (Figure 4.19). P-T estimates based on equilibrium #7 consistently yield higher P-T conditions than those determined using equilibrium #4. Core analyses yield 9.1 kbar and 690°C, 8.1 kbar and 770°C, and 6.7 kbar and 705°C, for microdomains A, B, and C, respectively. Rim analyses yield 5.2 kbar and 645°C, 5.8 kbar and 730°C, and 6.5 kbar and 710°C, for microdomains A, B, and C, respectively. The core conditions determined from microdomain B are considered most robust, as this garnet contained hornblende and plagioclase inclusions, although both do not occur in the same part of the garnet (Figure 4.18). Core compositions yield considerably higher pressures compared to rim compositions which are interpreted to record the conditions under which garnet breakdown occurred. Because the original rims of garnet have been destroyed, it is unknown whether they once possessed the high CaO rims seen in the amphibolite sample 98TXM201.

c) Kyanite-garnet-biotite semipelitic schist (98TXM339b)

The kyanite-garnet-biotite semipelitic schist is intercalated with the garnet-amphibolite, described above. The semipelite is buff to tan in colour and is composed of ~30% of each of plagioclase and quartz, ~20% garnet, ~15% biotite, <5% kyanite and trace ilmenite, pyrite, chalcopyrite, rutile, titanite and monazite. Garnet porphyroblasts

are pink to lavender, range up to ~8 mm in diameter, and are wrapped by the moderately developed foliation. Kyanite may be up to ~3 mm, plagioclase averages ~ 1 mm, biotite is typically <1 mm, and polycrystalline quartz ribbons may be several cm in length. The foliation is defined by both biotite, and elongate grains and aggregates of plagioclase and quartz. Rare kyanite blades appear to crosscut the foliation.

Mineral profiles reveal very little zoning in most minerals analysed (Figure 4.20). Cores of large garnet porphyroblasts (~6 mm diameter) have a higher grossular component ($X_{\text{Grs}} \sim 0.09$) than rims ($X_{\text{Grs}} \sim 0.05$). Garnet is almandine-rich with $\text{Fe}/(\text{Fe}+\text{Mg}) \sim 0.85$ across garnet cores and a slight increase to 0.87 at the rims, although the smallest garnet analysed (microdomain B; ~400 μm in diameter) has nearly uniform $\text{Fe}/(\text{Fe}+\text{Mg}) \sim 0.87$ -0.8. Plagioclase composition ranges from An_{41-45} with no notable change with proximity to garnet. $\text{Fe}/(\text{Fe}+\text{Mg})$ in biotite is generally ~0.5.

Thermobarometric results were calculated using equilibria #5 and #11. The highest P-T conditions, 6.4 kbar and 630°C, were retrieved from the core of a large (~6 mm) garnet (microdomain A) (Figure 4.21). However, this garnet yields low P-T conditions for its rim (3.3 kbar, 550°C). Smaller garnet, such as in microdomain B, yields lower P-T from the cores (5.0 kbar, 590°C) but yields similar rim conditions (4.0 kbar, 570°C). The ~6 mm garnet of microdomain C gives results similar to those of microdomain A, with 5.6 kbar and 620°C recorded by cores, and 1.9 kbar and 500°C retrieved from rim compositions.

Only the P-T result from the core of microdomain A falls within the kyanite stability field. Results from microdomain C fall on the kyanite-sillimanite univariant line, and results from microdomain B fall slightly below the univariant line. One of the

rim results falls within the sillimanite field and two plot in the andalusite field. The lower P-T conditions retrieved from the semipelite relative to the amphibolite with which it is intercalated (98TXM339a) suggest that the semipelite has not retained mineral compositions that correspond to peak metamorphic conditions, possibly related to fluid-assisted diffusion due to preferential fluid infiltration through the more permeable metasedimentary rocks. If the temperature determined from the amphibolite (98TXM339a) is assumed for this rock as well, a pressure of 8.5 kbar, which is comparable to that yielded from the amphibolite, can be interpolated for the semipelitic schist. It therefore appears that only those elements that directly affect the position of the thermometer (Fe, Mg) have been affected significantly. There is no evidence of any resetting of Ca in the semipelitic schist.

4.4.4 Rocks north of the complex

a) Charnockite (98TXM248)

Charnockite occurs as veins within mafic granulite of the tonalite-enderbite gneiss map unit north of the anorthosite suite (Chapter 2). It generally has a well-developed lineation, and only locally exhibits a foliation along its margins. This particular sample (98TXM248) appears undeformed, although microtextural evidence suggests that it has been deformed under low strain or static conditions and extremely high temperatures (Chapter 3). Its modal mineralogy comprises ~45% plagioclase, 20% orthopyroxene, 20% quartz, minor hornblende, garnet, K-feldspar, clinopyroxene and trace biotite, magnetite, and apatite. Orthopyroxene is rimmed by a complex coronitic structure composed mainly of hornblende, garnet, K-feldspar, clinopyroxene and quartz (Figure 4.22). Occurrence of clinopyroxene in coronae suggests formation of garnet and

clinopyroxene via the pressure-sensitive reaction: $An + Opx = Cpx + Grt + Qtz$. Orthopyroxene grains average ~8 mm in length, and their coronae average ~300 μm in thickness. The fine-grained plagioclase and quartz matrix is described in section 2.3.2.

Most phases display relatively homogenous compositional profiles. Coronitic garnets generally have flat Ca profiles ($X_{Grs} = \sim 0.17-0.18$), with a minor increase in Fe/(Fe+Mg) from 0.69 on the outboard side (away from the orthopyroxene) to 0.76 inboard (towards the orthopyroxene) where it contacts hornblende (Figure 4.22). Orthopyroxene compositional profiles are flat from core to rim, with Fe/(Fe+Mg) ~ 0.4 . Matrix plagioclase has a composition of An_{25-28} , while plagioclase within the coronitic structure has higher anorthite content, $\sim An_{35-45}$. P-T results based on core compositions use matrix plagioclase; those based on rim compositions use plagioclase within the coronitic structure. All P-T estimates are based on equilibrium #1 and the average of the barometers #8 and #9.

Higher P-T conditions are indicated by the core compositions and matrix plagioclase relative to compositions within the coronitic structures. The cores of microdomain A and B yield pressures of 11.7 kbar and 9.9 kbar, and temperatures of 790°C and 715°C, respectively. Rim results are 10.5 kbar and 785°C for microdomain A, and 7.2 kbar and 685°C for microdomain B. The difference between conditions determined based on core and rim compositions is interpreted to be related to decompression after garnet formation, compatible with the presence of calcic plagioclase in the coronae.

b) Garnet-biotite-amphibolite (98TXM320)

Garnet-biotite-amphibolite occurs immediately north of the Northern bounding shear zone, in the southern part of the tonalite and amphibolite map unit (Figure 4.1). The map unit is poorly exposed, but is interpreted to be ~2 km thick, and comprises predominantly garnet-free rock types that are not suitable for geothermobarometry. Rare foliated, garnet-bearing biotite-rich amphibolite occurs along the southern margin of the map unit. It consists of ~30% garnet, ~25% biotite, ~23% quartz, ~17% plagioclase, ~5% hornblende, and trace apatite. Garnet grains are subhedral, average ~2 mm in diameter, and typically contain abundant, randomly oriented quartz inclusions. Rare hornblende grains approach 1 mm in length and are invariably rimmed by biotite. Biotite averages ~500 μm in length, is the major foliation-defining mineral, and envelops garnet. Plagioclase and quartz aggregates are elongate parallel to the foliation.

Minor zoning is evident in most mineral phases. Garnet Ca profiles are either flat, with $X_{\text{Grs}} = 0.13$, or decrease slightly from core ($X_{\text{Grs}} = 0.15$) to rim ($X_{\text{Grs}} = 0.13$; Figure 4.24). From core to rim, $\text{Fe}/(\text{Fe} + \text{Mg})$ increases in garnet from 0.86 to 0.90. Matrix plagioclase generally has the composition An_{42-47} but typically decreases to as low as An_{32} adjacent to garnet. A small Or component is present in some plagioclase rims touching garnet (Or_{06}). Hornblende increases in Al, Na, and $\text{Fe}/(\text{Fe} + \text{Mg})$ from core to rim and has an average $\text{Fe}/(\text{Fe} + \text{Mg})$ of ~0.68. Biotite composition is constant throughout the thin section, with $\text{Fe}/(\text{Fe} + \text{Mg}) \sim 0.62$.

Core P-T determinations are based on the compositions of garnet cores, and the cores of matrix plagioclase, hornblende, and biotite. Rim analyses use garnet rims, adjacent plagioclase rims, and the rims of matrix biotite and hornblende within 1 mm of

the garnet. P-T determinations use equilibria #4, #5, #7 and #10. The first three are thermometers that yield similar results, although #4 typically yields slightly lower temperatures (by ~25 °C). The quoted P-T estimate is an average of the intersection of #5 and #7 with the barometer (equilibrium #10).

Cores yield P-T conditions of 7.4 kbar and 740°C, 7.8 kbar and 770°C, and 6.5 kbar and 740°C, for microdomains B, D, and E, respectively (Figure 4.25). Rims yield lower conditions for all microdomains: 4.6 kbar and 615°C, 6.2 kbar and 670°C, and 6.0 kbar and 700°C, for microdomains B, D, and E, respectively. The rock records lower P-T conditions compared to anorthositic rocks of the Northern bounding shear zone and the charnockite to the north. It is possible that rocks north and south of sample 98TXM320 (Figure 4.1) may have been exhumed from depth to higher crustal levels. However, it is also possible that 98TXM320 has been thoroughly reset, or that its mineral assemblage grew, during a later, lower grade event. The ubiquitous rimming of hornblende by biotite in this sample suggests that biotite formation occurred late with respect to hornblende formation and supports the latter interpretation.

4.5 Discussion

A polymetamorphic history for the UC has been suggested by previous work (Tella et al., 1994b), and is consistent with both metamorphic observations described above and with geochronological data presented in Chapter 5. Discussion of thermobarometric results begins with rocks from south of the anorthosite suite, as these are interpreted to have experienced most, if not all, of the metamorphic events affecting the area, and they also provide some of the most important textural relationships and robust P-T estimates. Rocks north of the anorthosite suite are discussed next, followed by rocks from within the

anorthosite suite. For clarity of discussion, M1, M2, and M3 are used to refer to metamorphic events related to garnet core growth (M1), high-CaO garnet rim growth and symplectite development (subdivided into M2a and M2b, respectively, where possible), and later, retrograde metamorphism that, at least locally, occurred after the cessation of most recent deformation (M3; e.g. sample 98TXM339a). P-T estimates are summarized in Table 4.1.

Rocks south of the anorthosite suite

The granulite-facies quartzofeldspathic gneiss, sample 98TXM178d, provides strong evidence for two discrete metamorphic events. A high 50-100 mCaO rim that follows the irregular shape of garnet porphyroblasts suggests that the garnet may have been embayed and resorbed *prior to* development of the high CaO rim. Alternatively, the low-CaO garnet core may have been overgrown by high CaO garnet and underwent subsequent resorption (Figure 4.12a). Formation of the high CaO rim by diffusional re-equilibration seems unlikely, because the penetration distance of Ca in garnet is negligible due to the low diffusivity of Ca in garnet (Spear, 1993). Although inclusions are not present to constrain conditions of core formation, results for other rocks with inclusions indicate that the assumption of equilibrium between garnet core and matrix phases is likely reliable. The results calculated for three microdomains are in good agreement (Figure 4.13a), and yield an average value of 8.4 kbar and 725°C for M1. The high CaO rims are interpreted to have formed during M2 at conditions of ~11.4 kbar and 805°C, based on the average of the three microdomains analysed (Figure 4.13b). Since the foliation in the quartzofeldspathic gneiss wraps garnet porphyroblasts, M1 is interpreted to be pre-tectonic with respect to mylonitization. Truncation of the foliation

by the high CaO M2 garnet rim in one sample (98TXM178d; Figure 4.12b) is interpreted to indicate that M2 was post-tectonic with respect to mylonitization, although more textural evidence is needed to corroborate this interpretation. Conditions under which resorption occurred were not calculated, but are interpreted to reflect M3 conditions.

Interpretation of the other granulite-facies sample, 98TXM350d, is complicated by the effects of late Fe-Mg exchange on the small (~50 μ m) garnet porphyroblasts. Only one of the initial P-T estimates from this sample showed good agreement between equilibria #1 and #2. Results from this microdomain (B) are 8.8 kbar and 730°C, and are interpreted to best represent metamorphic conditions for this sample. Equilibrium #1 plots somewhat higher (50-100°C) than equilibrium #2 for the other two microdomains analysed, suggesting that re-equilibration may have caused the Fe-Mg exchange thermometer (#2) to reset at lower temperatures (Bégin and Pattison, 1994; Fitzsimons and Harley, 1994). Application of the convergence technique of Bégin and Pattison (1994) yields 8.5 kbar and ~800°C and ~8.1 kbar and 785°C for microdomains D and A, respectively. The temperature yielded by this method is comparable to the M2 temperatures determined from the quartzofeldspathic gneiss, although the pressures are lower, whereas microdomain B gives P-T results comparable to M1. Interpretation of these results is ambiguous: the diorite mylonite may preserve M1 conditions only, in which case the dry assemblage may have prevented re-equilibration during M2. Alternatively, garnet growth may have occurred during M2, but was subsequently affected by diffusive re-equilibration and homogenization at lower pressures. The pressures recorded by the rock are more consistent with the former possibility (M1 metamorphism), and this interpretation is preferred because pressure comparisons are

more reliable indicators than temperature comparisons due to the problems introduced by late Fe-Mg exchange.

Further south, the amphibolite-facies sample 98TXM201 contains garnet porphyroblasts with high CaO rims, similar to those of 98TXM178d. It is interpreted to have a similar metamorphic history, although conditions derived from core compositions indicate that M1 occurred at slightly higher pressure and comparable temperature (average values = 9.2 kbar and 705°C for 98TXM201 vs. ~8.4 kbar and 725°C for 98TXM178d). High CaO garnet rim compositions yield average M2 conditions of 11.2 kbar and 695°C, similar to pressures yielded for 98TXM178d. M1, or core conditions for 98TXM201 are an average of three microdomains, based on garnet core and matrix plagioclase and hornblende compositions, and yield results consistent within $\pm 25^\circ\text{C}$ and 0.5 kbar. M2 estimates were based on the high-CaO garnet rim and proximal matrix plagioclase and hornblende compositions. Results from three microdomains agree within $\pm 10^\circ\text{C}$ and 1 kbar. Garnet breakdown conditions, designated M2b, were determined from hornblende-plagioclase moats of one microdomain analysed in great detail, and indicate decompression at 705°C and 9.7 (maximum value) kbar subsequent to formation of the high CaO garnet rims. The lower temperature yielded by 98TXM201 relative to 98TXM178d may reflect the closer proximity of the latter to a local heat source (i.e. the anorthosite suite and associated mafic magmas).

Samples 98TXM339a and b were collected from the southernmost locality, farthest from the anorthosite suite. The amphibolite, 98TXM339a, contains garnet pseudomorphs with only minor relict garnet, which may or may not have possessed high CaO. Conditions determined from core compositions using inclusions from one

microdomain indicate that metamorphism occurred at 8.1 kbar and 770°C. Comparable pressures between sample 98TXM339a and the core conditions yielded by the quartzofeldspathic gneiss suggest that the cores of 98TXM339a reflect M1 conditions. M2 metamorphic conditions can not be determined, nor can M2 be precluded for this sample due to the consumption of most of the garnet (30-40%). Garnet breakdown during M3 is interpreted to have occurred at 5.8 kbar and 695°C, based on the average of three microdomains which agree within $\pm 50^\circ\text{C}$ and 0.7 kbar.

The semipelitic schist from the same locality yields P-T conditions of 6.0 kbar and 625°C, based on core compositions of garnet together with matrix plagioclase and biotite from the two microdomains with garnet microporphyroblasts >5 mm. These results agree within $\pm 5^\circ\text{C}$ and 0.4 kbar. If the semipelite and intercalated amphibolite (98TXM339a) have a shared metamorphic history, then it is reasonable to question whether the semipelitic rock has been significantly affected by Fe-Mg diffusional re-equilibration. If the temperature is assumed to be equal to the temperature reflected by the amphibolite, then a pressure of 8.5 kbar would result. This is comparable to M1 pressure determinations from the amphibolite, and is consistent with garnet growth in the semipelite during M1, with subsequent preferential lower-grade thermal overprinting of the more permeable semipelite, possibly due to fluid-assisted diffusional re-equilibration during M3. Garnet from 98TXM339b lacks high CaO rims. The lack of evidence for M2 metamorphism is unlikely to be the result of CaO diffusion because the diffusivity of Ca in the most probable CaO source, plagioclase, is extremely low. A cryptic structural break north of locality 98TXM339a (Figure 4.1) may account for the discrepancy

between the thermotectonic histories indicated by samples 98TXM339a and 98TXM339b relative to the samples from farther north, discussed above.

Rocks north of the anorthosite suite

In contrast with the pre-tectonic garnet porphyroblasts in rocks south of the anorthosite suite, garnet occurs as a coronitic phase surrounding orthopyroxene. Plagioclase chemistry was not analysed rigorously enough to provide a clear understanding of the chemical variations in this rock, and for this reason, P-T results are not considered reliable. The occurrence of garnet in post-tectonic coronitic structures (surrounding orthopyroxene) suggests that garnet growth probably occurred during M2. M1 in other samples has been determined from core compositions, but garnet in 98TXM248 is clearly not in equilibrium with orthopyroxene cores. M2 in other samples has been determined from rim compositions that are considered to have been in equilibrium; the garnet in 98TXM248 appears to be in textural equilibrium with the rim of orthopyroxene. It is possible that the coronitic garnet grew during the tail end of M1, but the preferred interpretation is that this charnockite may have intruded during M1, but the garnet-forming reaction (orthopyroxene + plagioclase = clinopyroxene + garnet + quartz) occurred during M2. The average of P-T conditions determined from two microdomains from the charnockite are 10.8 kbar and 755°C, and the pressure calculations are consistent with M2 conditions determined for other rocks in the area.

The amphibolite-facies sample 98TXM320 contains pre-tectonic garnet porphyroblasts that lack high-CaO rims. The lack of zoning within ~2 mm diameter porphyroblasts cannot be a function of diffusional re-equilibration due to the immobility of Ca within garnet, even at high temperature (Spear, 1993). The average calculated P-T

conditions (~7.2 kbar and ~750°C) are comparable to M1 conditions determined from the cores of 98TXM178d. The lack of evidence for a second event is problematic, however, since the rock does not contain an anhydrous assemblage like 98TXM350d that would be most resistant to later re-equilibration. One explanation for the lack of evidence for M2 in the rock is that it may represent part of a down-dropped block that was juxtaposed next to higher-grade rocks after M2. The lack of exposure in this area north of the anorthosite suite limits field-based constraints. Alternatively, the post-M1 mineralogy may not have been capable of forming new garnet during M2, although it is difficult to determine from what reactants the garnet formed.

Discussion of rocks within the anorthosite suite

The relatively simple zoning patterns of garnet in sample 98TXM304 are most consistent with prograde conditions evolving through one metamorphic event. Initial conditions of 11.3 kbar and 730°C are indicated by the compositions of the core of the largest garnet porphyroblast analysed, along with a plagioclase inclusion and matrix hornblende. Matrix plagioclase was used for two other microdomains and yields similar P-T conditions. Its compositional similarity to inclusion plagioclase lends credibility to the use of matrix plagioclase for M1 P-T estimates in other samples. A marginal increase in CaO toward the rim of one of the garnet porphyroblasts indicates a pressure increase to 12.2 kbar at ~720°C. The high pressures recorded by the rock are very similar to M2 conditions calculated for 98TXM178d and 98TXM201, and are thus thought to reflect M2 conditions. Envelopment of M2 garnets by the mylonite foliation indicates that M2 was coeval with, or predates, latest deformation. Based on the average conditions determined for microdomains A and B, which contain the larger garnet porphyroblasts,

garnet breakdown to symplectitic plagioclase and hornblende is interpreted to have occurred at ~10.8 kbar and 785°C ($\pm 5^\circ\text{C}$ and 0.1 kbar), indicating decompression at elevated temperatures. The smaller (450 μm diameter) garnet microporphyroblast is not included in this calculation as its Fe/(Fe+Mg) is more likely to have been reset. The low Fe/(Fe+Mg) in the outer 500 μm of the large garnet porphyroblast (microdomain A) is interpreted to have resulted from diffusion at the higher temperature, lower pressure conditions under which garnet breakdown occurred. High temperature Fe-Mg exchange and symplectite development could represent a later, separate retrograde event, or may represent the retrograde part of one progressive P-T path. The latter interpretation is preferred because all conditions determined from this sample are compatible with M2 conditions, and there is no evidence for M1 or M3 effects. The interpretation that the crosscutting symplectite developed during M2 stipulates that M2 must be syntectonic with respect to latest deformation, i.e. latest deformation postdated M2 garnet growth and ceased before M2 symplectite developed.

Sample 99TXM561b records similar P-T conditions to those yielded by sample 98TXM304, but the large (2 cm diameter) garnet porphyroblast (garnet #1) analysed exhibits a chemical profile that is markedly different from garnet profiles of the latter sample. The outer 500-750 μm rim is characterized by high CaO, with slightly lower Fe/(Fe+Mg) evident only where symplectite is well developed. Conditions of 11.3 kbar and 760°C were calculated using plagioclase and hornblende inclusions within the low CaO garnet core. The high-CaO garnet rim, in conjunction with an inclusion from this high-CaO rim and matrix hornblende, yields conditions of 14.3 kbar and 780°C, based on an average of conditions determined using different equilibria. This estimate is not

considered reliable due to the lack of hornblende inclusions in the outer garnet, and the uncertainty that equilibrium had been established between matrix hornblende and the garnet rim. Latest metamorphic conditions of ~10.0 kbar and 800°C are based on calculations using the core of a garnet microporphyroblast (garnet #2) in conjunction with nearby matrix hornblende and plagioclase. The garnet microporphyroblast is considered texturally equivalent to the rim of the large garnet porphyroblast. The low CaO core of garnet #1 may correspond to M1 growth at higher pressures than recorded by rocks south of the anorthosite suite. Alternatively, the low-CaO core may have grown during early prograde M2 metamorphism. High-pressure conditions determined from the high CaO rim are interpreted as near-peak conditions of M2. Garnet breakdown conditions are comparable to those calculated for 98TXM304, and are interpreted to indicate late-stage M2b decompression at elevated temperatures.

Sample 99TXM564a, from the same locality as the leucogabbro dated in this study at ca. 2711 Ma (see Chapter 5), yields maximum P-T conditions of ~13.2 kbar and 990°C. This estimate, an average of calculations from three microdomains based on core compositions, is considered unrealistically high, possibly because of disequilibrium between phases and Fe-Mg exchange between adjacent orthopyroxene and clinopyroxene which would shift equilibria #1 and 2 to higher temperature. The average of conditions determined based on rim compositions for the two microdomains from leucocratic layers is 12.2 kbar and 840°C. The domain located within a more melanocratic layer is omitted from this calculation since its pyroxenes are more likely to have been affected by late Fe-Mg exchange. This estimate is considered more reliable since rim compositions are more

likely to have been in equilibrium. These conditions are comparable to M2 conditions determined for other rocks.

It is unclear whether the garnet porphyroblasts in 99TXM564a grew during M1 or M2. If the rock records M1 conditions, then a cryptic fault must exist, along which the rock would have been uplifted and juxtaposed against rocks that record lower pressure. If the rock records M2 conditions, then why does the rock record no evidence for earlier metamorphism? One possibility is that M1 pressures were not sufficiently high enough to induce garnet growth in this rock, which has a more Mg-rich bulk composition compared to other analyzed samples (compare Mg content of phases analysed, Appendix II). Garnet stability is greatly expanded with increasing Fe/(Fe+Mg), and garnet in 99TXM564a contains significantly lower Fe/(Fe+Mg) than in 99TXM561b (0.5 vs. 0.8, respectively). Garnet growth could have been initiated at the lower M1 pressures in rocks with higher Fe/(Fe+Mg) (e.g. 99TXM561b), while the more Mg-rich rocks (e.g. 99TXM564a) did not crystallize garnet until higher M2 pressures were attained.

Based on field observations, the mafic dykes are interpreted to have been emplaced during latest deformation (see Chapter 3), which was likely coeval with M2 (see discussion of 98TXM304, above). The dyke interpreted to have been emplaced earlier in the tectonothermal history (98TXM342) records P-T conditions of ~12.6 kbar and 865°C, and the calculated temperature should be considered a *maximum* estimate in light of evidence for late Fe-Mg exchange between adjacent clino- and orthopyroxene. The values quoted are averages based on the core compositions of three microdomains in the sample with the larger garnet porphyroblasts. Average P-T conditions of 8.1 kbar and 755°C based on rim compositions for three microdomains within 98TXM342 may

provide a minimum estimate for conditions under which the dykes have been metamorphosed. Alternatively, rim compositions represent latest metamorphic conditions sustained by the rock, and this may correspond to conditions under which garnet breakdown occurred in other rocks (e.g. 98TXM304).

Summary of metamorphic events: M1

Three metamorphic events are interpreted to have occurred, based on textural observations and on P-T results retrieved from rocks of the UC. M1 textures include garnet porphyroblasts of variable size that are enveloped by the main foliation, S_x . M1 conditions are inferred from rocks of the Southern bounding shear zone (98TXM178d and 98TXM350d), as well as by rocks north and south of the UC. The most convincing evidence for polymetamorphism comes from the abrupt transition from low- to high-CaO zoning in garnet porphyroblasts of the quartzofeldspathic gneiss (98TXM178d) from the Southern bounding shear zone. Based on the pre-tectonic assemblage comprising the low-CaO garnet core, foliation-forming matrix plagioclase, and orthopyroxene from garnet pressure shadows, the sample records M1 conditions of 8.4 kbar and 725°C. The main source of uncertainty is the orthopyroxene composition, which may have been modified during M2 metamorphism, and thus may no longer preserve its M1 composition. Appropriate inclusions (for thermobarometric analysis) within the M1 cores would enhance confidence in P-T calculations. The diorite gneiss (98TXM350d) of the Southern bounding shear zone records conditions of 8.8 kbar and 730°C, and, despite the lack of supporting textural evidence, these conditions are interpreted to reflect M1 conditions due to the close agreement with M1 results from 98TXM178d. Whereas the high-CaO garnet rims of 98TXM178d provide evidence of a second, higher pressure

metamorphic event (M2), 98TXM350d shows no evidence of M2, possibly because the dry assemblage prevented re-equilibration.

Amphibolite south of the UC (98TXM201) comprises large (2 mm to 1 cm) pre-tectonic garnet porphyroblasts enveloped by a foliation that is defined, primarily, by compositional banding. Low-CaO cores of the garnet porphyroblasts, in conjunction with matrix plagioclase and hornblende, yield conditions of 9.2 kbar and 705°C. These conditions are considered to reflect M1 conditions, based on the interpretation of the low-CaO garnet cores as products of M1 metamorphism. High-CaO garnet rims are interpreted to have grown during M2. As in the case for 98TXM178d, confidence in these results could be enhanced if appropriate inclusions were present, as chemical equilibrium between the M1 garnet core and the post-M2 matrix phases can not be established. Amphibolite farther south (98TXM339a) yields M1 conditions of 8.1 kbar and 770°C, although significant alteration of garnet and plagioclase is evident in the rock. Amphibolite north of the anorthosite suite (98TXM320) contains ~2 mm garnet porphyroblasts that are wrapped by the foliation. These garnet porphyroblasts lack high-CaO rims that are elsewhere interpreted as products of M2 garnet growth (e.g. 98TXM178d, 98TXM201). Thermobarometry yielded P-T conditions of 7.2 kbar and 750°C, interpreted to reflect M1 conditions. It is unclear why this sample shows no evidence of M2 metamorphism.

M1 is pre-tectonic (i.e. predates both S_m and S_x), is interpreted to have occurred at ~7-9 kbar and 705-770°C and exhibits no systematic variation in P or T with respect to location (Figure 4.26). The timing of M1 paragenesis, and subsequent S_x development are not easily established. Based on the U-Pb (monazite) age from quartzofeldspathic

gneiss of the Southern bounding shear zone (sample 98TXM352; see Chapter 5), M1 may correspond to the earliest thermal event evident in the area (ca. 2.62-2.60 Ga). Alternatively, M1 paragenesis in the UC and surrounding rocks may have occurred at ca. 2.53-2.50 Ga (see Chapter 5; see also Berman et al., 2000), or this may represent the age of intense thermal overprinting. The enveloping S_x foliation is likely correlative to the regional S_2 foliation (Tella et al., 2000b), a transposition foliation thought to have formed at ca. 2.55-2.50 Ga (Berman et al., 2000; Ryan et al., 2000b).

Summary of metamorphic events: M2

M2 is characterized by syn- to post-tectonic garnet growth, commonly resulting in high CaO garnet rims produced at high pressures (10.8-14.3 kbar) and high, though variable temperatures (700-865°C). High temperatures were sustained throughout post-tectonic decompression to ~10 kbar (refer to results for samples 98TXM304, 99TXM561b, 98TXM210 and Discussion section). Decompression is interpreted to mark the culmination of the high-P M2 event, and not a distinct event following a temporal gap, because a significant time gap would allow time for cooling. However, it is possible that decompression-related symplectite development occurred post-M2 (i.e. during an M3 event), but the simpler explanation is preferred.

The effects of M2 metamorphism are somewhat variable and are therefore best understood through inspection of samples from a range of localities. South of the anorthosite suite, garnet porphyroblasts from two samples (98TXM178d and 98TXM201) show evidence for two metamorphic events. Garnet porphyroblasts with low-CaO cores and an abrupt transition to high-CaO rims are consistent with two distinct phases of garnet growth, as has been suggested for rocks elsewhere in the Hearne domain (e.g.

Berman et al., 2000). P-T calculations yield M2 conditions of 11.4 kbar and 805°C for 98TXM178d and 11.2 kbar and 695°C for 98TXM201. The latter sample exhibits discontinuous hornblende-plagioclase coronae indicative of garnet breakdown. Garnet breakdown phases indicate decompression to 9.7 kbar at sustained high temperatures (705°C), also interpreted as the terminal phase of M2.

P-T conditions calculated for a garnet anorthosite (98TXM304) from the Northern bounding shear zone provide evidence for part of the M2 P-T path. Core compositions yield early prograde conditions of 11.3 kbar and 730°C. A marginal increase in CaO in garnet is inferred to imply a minor pressure increase to 12.2 kbar at 720°C. Garnet breakdown to symplectitic plagioclase and hornblende occurred under conditions of 10.8 kbar and 785°C, also consistent with high temperature decompression. The sample contains high-CaO garnet ($X_{\text{Gr}} \approx 0.3$) with no evidence for low-CaO garnet cores, and is consistent with an entirely M2-related tectonothermal history. The absence of M1 effects suggests that this sample may represent an anorthosite of Proterozoic age.

Another garnet-anorthosite from the Northern bounding shear zone (99TXM561b) contains garnet porphyroblasts that indicate a more complex history. A large (>1 cm) garnet porphyroblast that is wrapped by the S_m foliation contains a low-CaO core, which, with plagioclase and hornblende inclusions, gives a P-T estimate of 11.3 kbar and 760°C (based on the average of two thermometers with one barometer). It is unclear whether this determination represents M1 or early M2 conditions. If, as for other samples with pronounced CaO zoning, the low-CaO garnet core is assumed to have formed during M1, then this sample may represent an Archean component of anorthosite. However, the garnet porphyroblast may represent a xenocryst from surrounding country rocks.

Calculations using the high-CaO garnet rim yield a near-peak M2 estimate of 14.3 kbar and 780°C. Garnet breakdown conditions, based on the outer garnet rim and surrounding symplectitic plagioclase and hornblende, are estimated at 11.3 kbar and 760°C, and provide further evidence for high-temperature decompression interpreted to represent the culmination of M2. Garnet microporphyroblasts (400 µm in diameter) are unzoned with respect to CaO and contain high X_{Grs} (~0.4-0.6). P-T calculations based on a microporphyroblast are inferred to reflect late M2 (i.e. M2b) conditions of 10.0 kbar and 800°C.

Leucogabbro from the Southern bounding shear zone (99TXM564a) displays textures consistent with a single episode of garnet growth at 12.2 kbar and 840°C, interpreted to reflect M2 conditions. Absence of textural evidence for M1 metamorphism is attributed to the Mg-rich nature of the sample, which would require higher pressures to induce garnet growth (Berman, pers. comm., 2000).

Two gabbro dykes (98TXM333 and 98TXM342) also exhibit textural features consistent with one high-grade metamorphic event. A U-Pb (zircon) age of ca. 1936 Ma was determined for gabbroic rocks (similar to these dykes) and anorthosite, both from the westernmost part of the UC, and this is interpreted as the age of granulite-facies mylonitization (Roddick and Tella, unpublished data, 1993; Tella et al., 1994b). Emplacement of the gabbro dykes is herein interpreted to be coeval with mylonitization (see Chapter 3). Given the evidence for a Proterozoic age of the gabbro dykes, these rocks are expected to show a simpler history than Archean country rocks and anorthosite. The dykes record conditions of 10.3 kbar and 820°C for sample 98TXM333 and 12.6 kbar and 865°C for sample 98TXM342. The latter is interpreted to represent peak P-T

estimate, as the former is interpreted to have been emplaced later in the deformation event. The smaller garnets (with lower CaO cores) and the more weakly developed foliation exhibited by 98TXM333, relative to 98TXM342, is consistent with emplacement of 98TXM342 earlier in the Paleoproterozoic tectonothermal event. Pressures determined from these samples are comparable to M2 pressures calculated for 98TXM178d and 98TXM201. The higher temperatures may be related to closer proximity to a local heat source (i.e. themselves).

A charnockite (sample 98TXM248) from north of the anorthosite suite yields P-T conditions of 10.8 kbar and 755°C. Textural evidence and comparability to M2 pressures calculated from other samples support the argument that these conditions reflect M2 metamorphism. The rock contains no evidence for earlier metamorphism.

Based on late-M2 symplectites, which locally exhibit axial planar elongation where isoclinal folding is preserved (see Chapter 3), M2 is interpreted to be synchronous with latest deformation (ca. 1936 Ma; Roddick and Tella, unpublished data, 1993; see Chapter 5). The regional extent of Paleoproterozoic ductile deformation is unclear; it appears to be focused primarily within the Northern bounding shear zone, but is likely heterogeneously developed throughout the UC. Folding of the UC anorthosite suite into a tight, easterly-plunging syncline (see also Tella et al., 2000b) may have occurred in the Proterozoic.

The spatial variation in M2 conditions is such that the highest temperatures come from the core of the UC. This is consistent with the interpretation that the gabbro dykes were emplaced during M2, and may have provided the heat for M2 metamorphism in the UC and surrounding areas. It is unclear whether the rocks of the UC represent a discrete

tectonothermal domain or whether they provided the heat for contact metamorphism. Pressure differences for M2 metamorphism appear consistent with a gradual decrease in depth away from the UC, although further work is required to test this supposition. Thermobarometry on samples across 10 km transects to the north and south of the UC anorthosite suite is recommended for clarification.

Based on the garnet-pseudomorph texture exhibited by sample 98TXM339a, M3 is interpreted to have occurred at P-T conditions of 5.8-6 kbar and 625-695°C, and to be largely post-tectonic. Fluid-assisted diffusion may have reset Fe-Mg ratios in the ferromagnesian minerals of 98TXM339b. M3 is likely related to the regional emplacement of Paleoproterozoic granites (e.g. the 1.83 Ga Hudson suite granites). The production of Hudson granitoids from near-solidus, water-rich magmas (Peterson et al., 2000) supports the above interpretation.

Figure 4.1: Location map for thermobarometric samples.

Figure 4.2a: SEM image of garnet-bearing gabbroic anorthosite (sample 98TXM561b). Note the intergranular network, or high-An plagioclase-quartz intergrowths developed at grain boundaries between plagioclase of moderate anorthite content.

Figure 4.2b: Leucogabbro (99TXM561b) – BSE image showing edge of large (2 cm diameter) garnet porphyroblast (garnet #1; Grt; microdomain E) with hornblende (Hbl)-plagioclase (Pl) coronitic structure, quartz (Qtz) and rutile (Ru) inclusions, and titanite (Ti) in the matrix. Note compositional profiles across part of the garnet, the coronitic structure, and matrix plagioclase. Points analysed are indicated on the image, and those used for P-T calculations are indicated.

Geothermometers

Geobarometers

PT results use equilibria #10 and #4, or an average of #4 and #6, where the two are within 100°C.

Figure 4.3: Leucogabbro (99TXM561b) - P-T results, summarized from different microdomains. Points from which calculations are based, and equilibria used, are indicated.

Figure 4.4a: Detail of microtexture in gabbroic anorthosite (sample 98TXM304), showing polygonal matrix plagioclase with high calcic rims (paler grey) and minor interstitial quartz.

Figure 4.4b: Gabbroic anorthosite (98TXM304) – BSE images showing microdomain A of garnet porphyroblast (~4 mm in diameter) and ~500 μ m thick hornblende-plagioclase corona (top). Detail of poorly developed part of the corona (bottom) showing garnet (Grt), hornblende (Hbl), plagioclase (Pl) and very fine, interstitial quartz (Qtz). Compositional profiles show variation across mineral grains; analysed points used in P-T calculations are indicated.

98TXM304 - Metagabbro-anorthosite: Microdomain A

Core conditions:
g10 p13 h15
PT: 11.3 kbar,
730°C

Rim conditions:
g26 p25 h16
PT: 12.2 kbar,
720°C

Conditions of garnet
breakdown:
g29 p22 h7
PT: 10.8 kbar, 780°C

Geothermometers

4) $3 \text{ Tr} + 5 \text{ Alm} = 5 \text{ Prp} + 3 \text{ FeTr}$

6) $\text{Ed} + \text{Ab} = \text{Ri} + \text{An}$

Geobarometers

10) $12 \text{ Qtz} + 3 \text{ Ts} + 2 \text{ Prp} + 4 \text{ Grs} = 12 \text{ An} + 3 \text{ Tr}$

PT results use equilibria 6 and 10.

Figure 4.5a: Gabbroic anorthosite (98TXM304) - P-T results from microdomain A. Points on which calculations are based and equilibria used are indicated.

Geothermometers

Geobarometers

PT results use equilibria 6 and 10.

Figure 4.5b: Gabbroic anorthosite (98TXM304) - P-T results from microdomains B and C. Points on which calculations are based and equilibria used are indicated.

Figure 4.6: Leucogabbro (99TXM564a) – BSE image showing microdomain B - garnet porphyroblast (~0.5 mm in diameter), clinopyroxene and orthopyroxene aggregate, and matrix plagioclase. Note the thin corona of symplectitic orthopyroxene and plagioclase rimming the garnet. Compositional profiles show variation across mineral grains; analysed points used in P-T calculations are indicated.

98TXM564a - Microdomain B

Geothermometers

- 1) Alm = alOpx + 3Fsl
- 2) 3En + 2Alm = 3Fsl + 2Prp
- 3) Alm + 3Di = Prp + 3Hd

Geobarometers

- 8) 3Qtz + 2Grs + Alm = 3An + 3Hd
 - 9) 3Qtz + Grs + 2Alm = 3An + 6Fsl
- PT results using equilibria #1 and #9.*

Figure 4. 7a: Leucogabbro (99TXM564a) - P-T results from the cores of three microdomains.

Geothermometers

- 1) Alm = alOpx + 3Fsl
- 2) 3En + 2Alm = 3Fsl + 2Prp
- 3) Alm + 3Di = Prp + 3Hd
- 4) 3Tr + 5Alm = 5Prp + FeTr
- 5) Alm + Phl = Prp + Ann

Geobarometers

- 8) 3 Qtz + 2 Grs + Alm = 3 An + 3 Hd
 - 9) 3 Qtz + Grs + 2 Alm = 3 An + 6 Fs
 - 10) 3 Tsc + 12 Qtz + 2 Prp + 4 Grs = 12 An + 3 Tr
- PT results using equilibria #1 and #9.*

Figure 4.7b: Leucogabbro (99TXM564a) - P-T results from the cores of three microdomains.

Figure 4.8: Mafic granulite-facies dyke (98TXM333) – BSE image showing microdomain B - garnet porphyroblast (~0.4 mm in diameter), orthopyroxene rimmed by clinopyroxene, matrix plagioclase, and minor quartz. Compositional profiles show variation across mineral grains; analysed points used in P-T calculations are indicated.

98TXM333 - Microdomain B

Figure 4.9a: Mafic granulite-facies dyke (98TXM342) – BSE image showing microdomain A - garnet porphyroblast (~0.8 mm in diameter), orthopyroxene rimmed by clinopyroxene, matrix plagioclase, and a 1 mm long quartz ribbon. Compositional profiles show variation across mineral grains; analysed points used in P-T calculations are indicated.

98TXM342 - Microdomain A

Figure 4.9b: Incipient hornblende development: at the interface between orthopyroxene and clinopyroxene, $< 10\ \mu\text{m}$ hornblende veneer rims the two phases.

Geothermometers

- 1) $\text{Alm} = \text{aOp} + 3\text{Fsl}$
- 2) $3\text{En} + 2\text{Alm} = 3\text{Fsl} + 2\text{Prp}$
- 3) $\text{Alm} + 3\text{Di} = \text{Prp} + 3\text{Hd}$

Geobarometers

- 8) $3\text{Qtz} + 2\text{Grs} + \text{Alm} = 3\text{An} + 3\text{Hd}$
- 9) $3\text{Qtz} + \text{Grs} + 2\text{Alm} = 3\text{An} + 6\text{Fsl}$

PT results use equilibria 1 and the average of 8 and 9.

Figure 4.10a: Mafic granulite-facies dyke (98TXM333) – P-T results based on core compositions from three microdomains.

Geothermometers

- 1) $\text{Alm} = \text{aOp} + 3\text{Fsl}$
- 2) $3\text{En} + 2\text{Alm} = 3\text{Fsl} + 2\text{Prp}$
- 3) $\text{Alm} + 3\text{Di} = \text{Prp} + 3\text{Hd}$

Geobarometers

- 8) $3\text{Qtz} + 2\text{Grs} + \text{Alm} = 3\text{An} + 3\text{Hd}$
 - 9) $3\text{Qtz} + \text{Grs} + 2\text{Alm} = 3\text{An} + 6\text{Fsl}$
- PT results use equilibria 1 and 8.*

Figure 4.10b: Mafic granulite-facies dyke (98TXM333) – P-T results based on rim compositions from three microdomains.

Geothermometers

- 1) $\text{Alm} = \text{aOp} + 3\text{Fsl}$
- 2) $3\text{En} + 2\text{Alm} = 3\text{Fsl} + 2\text{Prp}$
- 3) $\text{Alm} + 3\text{Di} = \text{Prp} + 3\text{Hd}$

Geobarometers

- 8) $3\text{Qtz} + 2\text{Grs} + \text{Alm} = 3\text{An} + 3\text{Hd}$
 - 9) $3\text{Qtz} + \text{Grs} + 2\text{Alm} = 3\text{An} + 6\text{Fsl}$
- P-T results use equilibria #1 and the average of #8 and #9.*

Figure 4.11a: Mafic granulite-facies dyke (98TXM342) – P-T results based on core compositions from three microdomains.

Geothermometers

- 1) Alm = alOpx + 3Fsl
- 2) 3En + 2Alm = 3Fsl + 2Prp
- 3) Alm + 3Di = Prp + 3Hd

Geobarometers

- 8) 3Qtz + 2Grs + Alm = 3An + 3Hd
 - 9) 3Qtz + Grs + 2Alm = 3An + 6Fsl
- PT results use equilibria #1 and #8.*

Figure 4.11b: Mafic granulite-facies dyke (98TXM342) – *P-T* results based on rim compositions from three microdomains.

Figure 4.12a: Quartzofeldspathic gneiss (98TXM178d) – BSE image showing microdomain C - garnet porphyroblast (~0.8 mm in diameter), orthopyroxene in pressure shadow, matrix plagioclase, quartz ribbons, ilmenite (ilm) and biotite-rich layers, and randomly oriented biotite patches. Compositional profiles show variation across mineral grains; analysed points used in P-T calculations are indicated.

98TXM178d - Microdomain C

Figure 4.12b: X-ray maps of garnet based on variation in: a) Mg; b) Ca; c) Mn, and d) Fe. Garnet is $\sim 0.9 \times 0.7$ mm.

Geothermometers

- 1) $\text{Alm} = \text{alOpx} + 3\text{Fsl}$
- 2) $3\text{En} + 2\text{Alm} = 3\text{Fsl} + 2\text{Prp}$

Geobarometers

- 9) $3\text{Qtz} + \text{Grs} + 2\text{Alm} = 3\text{An} + 6\text{Fsl}$

PT results use equilibria #1 and #9.

Figure 4.13a: Quartzofeldspathic gneiss (98TXM178d) – *P-T* results based on core compositions from three microdomains.

Geothermometers

1) Alm = alOpx + 3Fsl

2) 3En + 2Alm = 3Fsl + 2Prp

Geobarometers

9) 3Qtz + Grs + 2Alm = 3An + 6Fsl

PT results use equilibria #1 and #9.

Figure 4.14: Monzodioritic protomylonite (98TXM350d) – BSE image showing microdomain A - Clinopyroxene porphyroblast (~2 mm in diameter), orthopyroxene in the pressure shadow, subgrains and new grains of orthopyroxene at rims of clino- and orthopyroxene, actinolitic hornblende developed at interface between the two pyroxenes. Large (length > field of view) polycrystalline K-feldspar (Ksp) megacryst may not be inherent to the rock, but was probably introduced during deformation by attenuation and boudinage of granitic veins. Garnet forms tiny (~50 μm) subhedral microporphyroblasts within pressure shadows. Compositional profiles show variation across mineral grains, and analysed points used in P-T calculations are indicated.

98TXM350d - Microdomain A

Figure 4.15: Monzodioritic protomylonite (98TXM350d) – P-T results from the core of microdomain D. Lack of agreement between equilibria #1 and #2 is taken to indicate late Fe-Mg exchange. The method used to extract near-peak conditions involves iterative modification of the Fe/Mg ratio in either garnet or orthopyroxene to get the two thermometers to converge. This technique resulted in a temperature increase of 50°C, and a nominal pressure increase of 0.1 kbar, yielding a P-T estimate of 8.5 kbar and 800°C for the rock.

Sample 350d: Core of microdomain D (unmodified).

Geothermometers

- 1) Alm = aOp + 3Fsl
 - 2) 3En + 2Alm = 3Fsl + 2Prp
 - 5) Alm + Phl = Prp + Ann
- Geobarometers
- 9) 3An + 6Fsl = 3Qtz + Grs + 2Alm

Sample 350d: Core of microdomain D (modified to account for late cation exchange).

Opx composition (as calculated by twq v. 2.02):			
	unmodified	modified	
Mg(M1)	0.5858	0.5458	
xFe	0.3897	0.4297	
xAl ₂	0.0245	0.0245	

Opx composition (as calculated by twq v. 1.02):			
	Mg(M1)	Fe(M1)	Mg(M2)
unmodified	0.5814	0.3868	0.5862
modified	0.5419	0.4256	0.5463

Point A: 5.6 kbar, 580°C - estimate based on thermometer #5 and barometer #9.

Point B: 7.5 kbar, 695°C - estimate based on thermometer #2 and barometer #9; lack of agreement between thermometers #1 and #2 may be a result of late Fe-Mg exchange, causing #1 to yield low temperatures.

Point C: 8.4 kbar, 750°C - estimate based on thermometer #1 and barometer #9.

Point D: **8.5 kbar, 800°C** - Convergence of equilibria #1 and #2 by modification of Fe/Mg ratio in opx to retrieve peak conditions.

Equilibria #'s 5, and 9 calculated using TWEEQU v. 1.02. Equilibria #1 and #2 calculated using TWEEQU v. 2.02 (for modified and unmodified opx).

Figure 4.16: Banded garnet-amphibolite (98TXM201) – BSE image showing microdomain G - garnet porphyroblast (>2 mm in diameter), matrix plagioclase, hornblende and quartz. Note the thin hornblende grains surrounding part of the garnet, in places separated by fine-grained plagioclase. Compositional profiles show variation across mineral grains; analysed points used in P-T calculations are indicated.

Geothermometers

Geobarometers

PT results use equilibria 7 and 10.

Figure 4.17a: 98TXM201 – P-T results, based on calculations from microdomain G.

Geothermometers

Geobarometers

PT results use equilibria 7 and 10.

Figure 4.17b: 98TXM201 – P-T results, based on calculations from microdomains A and C.

Figure 4.18: Melanocratic garnet-amphibolite (98TXM339a) – BSE image showing microdomain B – relict garnet porphyroblast (>1 mm in diameter; top), surrounded by thick (~1 mm) pseudomorph structure interpreted to have formed by partial garnet breakdown reaction. Some structures contain no relict garnet and are interpreted to be post-tectonic garnet pseudomorphs. Relict garnet (below) contains abundant inclusions of quartz, plagioclase, biotite and hornblende. Compositional profiles show variation across mineral grains; analysed points used in P-T calculations are indicated.

Geothermometers

4) $3\text{Tr} + 5\text{Alm} = 5\text{Prp} + \text{FeTr}$

5) $\text{Alm} + \text{Phl} = \text{Prp} + \text{Ann}$

7) $\text{Ed} + 4\text{Qtz} = \text{Tr} + \text{Ab}$

Geobarometers

10) $3\text{Tsc} + 12\text{Qtz} + 2\text{Prp} + 4\text{Grs} = 12\text{An} + 3\text{Tr}$

PT results using equilibria 7 and 10.

Figure 4.19a: Melanocratic garnet-amphibolite 98TXM339a – P-T results are based on core compositions from three microdomains.

Geothermometers

Geobarometers

PT results using equilibria 7 and 10.

Figure 4.19b: Melanocratic garnet-amphibolite 98TXM339a – P-T results are based on rim compositions from three microdomains.

Figure 4.20: Kyanite-garnet-biotite semipelitic schist (98TXM339b) – BSE image
showing microdomain B - garnet porphyroblast (~0.4 mm in diameter), matrix
plagioclase, kyanite, biotite, and quartz. Compositional profiles show variation
across mineral grains, and analysed points used in P-T calculations are indicated.

98TXM339b - Microdomain B

Core analysis:
g7 p7 b5
PT: 5.0 kbar, 590°C

Rim analysis:
g9 p4 b6
PT: 4.0 kbar, 570°C

Geothermometers

5) $\text{Ph} + \text{Alm} = \text{Ann} + \text{Prp}$

Geobarometers

11) $2\text{Ky} + \text{Grs} + \text{Qtz} = 3\text{An}$

PT results use equilibria #5 and #11.

Figure 4.21a: Kyanite-garnet-biotite semipelitic schist (98TXM339b) – P-T results are based on core compositions from three microdomains.

Geothermometers

5) $\text{Ph} + \text{Alm} = \text{Ann} + \text{Prp}$

Geobarometers

11) $2\text{Ky} + \text{Grs} + \text{Qtz} = 3\text{An}$

PT results use equilibria #5 and #11.

Figure 4.21b: Kyanite-garnet-biotite semipelitic schist (98TXM339b) – P-T results are based on rim compositions from three microdomains.

Figure 4.22: Charnockite (98TXM248) – BSE image showing microdomain B –

orthopyroxene grain (~1 cm in diameter), with a complex coronitic structure comprising garnet, with hornblende, clinopyroxene, and quartz on the inboard side of garnet, and a thin veneer of K-feldspar outboard of the garnet.

Compositional profiles show variation across mineral grains; analysed points used in P-T calculations are indicated.

98TXM248 - Microdomain B

Figure 4.23: Charnockite (98TXM248) – P-T results determined for two microdomains.

Figure 4.24: Garnet-biotite amphibolite (98TXM320) – BSE image showing microdomain D – garnet porphyroblast (~2 mm in diameter), with matrix plagioclase, quartz, biotite, and hornblende.

98TXM320 - Microdomain D

Geothermometers

- 4) $3\text{Tr} + 5\text{Alm} = 5\text{Prp} + \text{FeTr}$
- 5) $\text{Alm} + \text{Phl} = \text{Prp} + \text{Ann}$
- 7) $\text{Ed} + 4\text{Qtz} = \text{Tr} + \text{Al}$

Geobarometers

- 10) $3\text{Tsc} + 12\text{Qtz} + 2\text{Prp} + 4\text{Grs} = 12\text{An} + 3\text{Tr}$

PT results based on the average of thermometers #5 and #7, and barometer # 10.

Figure 4.25a: Garnet-biotite amphibolite (98TXM320) – *P-T* results, based on core compositions from three microdomains.

Geothermometers

Geobarometers

PT results using thermometers #5 and #7, and barometer #10.

Figure 4.25b: Garnet-biotite amphibolite (98TXM320) – *P-T* results, based on rim compositions from three microdomains.

Note: True north is 0°13' counter-clockwise from grid north.

P-T estimates are colour-coded according to metamorphic event they are interpreted to reflect:

Figure 4.26: Summary of P-T results from the westernmost Uvauk complex.

P-T conditions from the Uvauk complex and surrounding rocks				
Sample	M event	Pressure (kbar)	Temperature (°C)	
<u>Rocks within the anorthosite suite</u>				
98TXM304	core	M2	11.3	730 (best estimate; microdomain A)
	rim	M2	12.2	720 (best estimate; microdomain A)
	grt breakdown	M2	10.8	785 (average of 2 microdomains)
99TXM561b	core	M1	11.3	760 (best estimate; garnet #1)
	high Ca rim	M2	14.3	780 (best estimate; garnet #1)
	grt breakdown	M2	10.0	800 (best estimate; garnet #2)
99TXM564a	rim	M2	12.2	840 (average of 2 microdomains)
98TXM333	core	M2	10.3	820 (average of 3 microdomains)
98TXM342	core	M2	12.6	865 (average of 3 microdomains)
<u>Southern bounding shear zone</u>				
98TXM178d	core	M1	8.4	725 (average of 3 microdomains)
	rim	M2	11.4	805 (average of 3 microdomains)
98TXM350d	core	M1	8.8	730 (best estimate; microdomain B)
<u>Rocks south of the anorthosite suite</u>				
98TXM201	core	M1	9.2	705 (average of 3 microdomains)
	rim	M2	11.2	695 (average of 3 microdomains)
	gnt breakdown	M2	9.7	705 (best estimate; microdomain G)
98TXM339a	core	M1	8.1	770 (best estimate; microdomain B)
	gnt breakdown	M3	5.8	695 (average of 3 microdomains)
98TXM339b	core	M3	6.0	625 (average of 3 microdomains)
<u>Rocks north of the anorthosite suite</u>				
98TXM320	core	M1	7.2	750 (average of 3 microdomains)
98TXM248	core	M2	10.8	755 (average of 2 microdomains)

Table 4.1: Summary of P-T estimates.

Chapter 5: Geochronology of the Uvauk complex

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Chapter 5: Geochronology

5.1 Introduction

A U/Pb geochronological study was conducted at the Geological Survey of Canada laboratory in Ottawa on three samples to determine the age of the Uvauk meta-anorthosite suite and the timing of tectonothermal events. The samples selected include: (1) leucogabbro mylonite from the southernmost margin of the anorthosite complex; (2) tonalitic mylonite from the diorite gneiss map unit; and (3) garnet-quartzofeldspathic (qf) gneiss from the Southern bounding shear zone (Figures 3.1 and 5.1). The leucogabbro was collected in order to determine the magmatic age of the anorthosite suite of the Uvauk complex (UC). The two rocks from the Southern bounding shear zone were studied in order to constrain the timing of metamorphism and deformation south of the anorthosite suite. As well as constraining timing of tectonothermal events within the UC, the geochronological data are compared with those for nearby, and potentially related, anorthosite complexes, including the Kramanituvar (KC) and the Daly Bay (DBC) complexes, to further the current understanding of the northern segment of the Snowbird tectonic zone.

5.2 Previous geochronology

Prior to this study, a limited number of age determinations were available from the study area, indicating an Archean age for anorthositic rocks and providing evidence for both Archean and Proterozoic metamorphic events (Tella et al., 1994b; Roddick and Tella, unpublished data, 1993; Figure 5.1). A three-point Sm-Nd isochron based on three whole rock samples (1. highly deformed anorthosite from the western part of the complex; 2. mafic granulite, also from the western part; and 3. undeformed gabbroic

anorthosite from the eastern part of the complex), yielded a Sm-Nd isochron age of 2770 $\pm$ 60 Ma, interpreted to be the igneous crystallization age of the anorthosite of the Uvauk complex (Roddick and Tella, unpublished data, 1993). A mylonitic, garnet-bearing anorthosite (sample "1.", above), from the westernmost part of the anorthosite suite yielded ages of ~1936 Ma (U-Pb, zircon) and ~1903 Ma (U-Pb, titanite) (Tella et al., 1994b; Figure 5.1). The zircon age was inferred to reflect the timing of granulite-facies metamorphism, an interpretation based in part on zircon morphology. The titanite age was interpreted to represent the timing of cooling through the blocking temperature of titanite (660-700°C; Scott and St-Onge, 1995). A garnet-quartzofeldspathic gneiss sample located 2 km east and less than 500 metres across strike to the south of the 1936 Ma anorthosite (Figure 5.1) yielded ages of ~2599 Ma (U-Pb, zircon) and ~2595 Ma (U-Pb, monazite). These ages were interpreted to reflect the timing of granulite-facies metamorphism and synchronous mylonitization of these rocks.

Reconnaissance geothermobarometry (Tella et al., 1994b) conducted on the dated samples is consistent with the interpretation of two distinct granulite-facies tectonometamorphic events. A high pressure, 10 kb (30 km depth), tectonometamorphic event was interpreted to have occurred in the Neoproterozoic (ca. 2.59 Ga). A second mylonitization event at ca. 1.9 Ga was interpreted to have occurred at 2-3 kbar (6-9 km) lower pressure. This study aims to further investigate the polymetamorphic history of the Uvauk complex and surrounding rocks, and to consider the potential effects of the younger thermal event on the older one. Specifically, the tonalite and quartzofeldspathic gneiss collected from the mylonite belt south of the anorthosite were selected to evaluate the previously derived metamorphic age of ~2595 Ma, and to look for effects of the

Proterozoic metamorphism which had apparently affected the anorthosite and mafic granulites in the anorthosite complex (Tella et al., 1994b; Roddick and Tella, unpublished data, 1993).

The Kramanituur complex (KC), located ~100 km west of the UC, is lithologically and structurally similar to the UC; both coincide with a regional geophysical anomaly, the Snowbird tectonic zone (STZ). As inferred for the UC, there is evidence for ca. 2.55 and 1.9 Ga events. A detailed study of the KC by Sanborn-Barrie (1999), involving structural, geochronological, thermobarometric, and geochemical analyses, provides a useful database with which to compare data from the UC. Gabbroic anorthosite from the central part of the KC yielded an age of 1902 ± 6 Ma (U-Pb, zircon), interpreted to be the age of magmatic crystallization on the basis of euhedral to subhedral prismatic zircon morphology, euhedral internal zonation of many zircons, low U contents (28 to 86 ppm), and high Th/U ratios. Metamorphism in the KC is interpreted to have initiated at ~ 1910 Ma, based on monazite data, *prior to* the emplacement of voluminous gabbroic anorthosite.

Evidence for Archean tectonometamorphism within rocks located directly south of the KC comes from a ca. 2550 Ma syenogranite pegmatite dike that crosscuts a belt of metasedimentary rocks. Sanborn-Barrie (1999) suggested that the Archean metasedimentary and plutonic rocks to the south of the complex were not affected by Proterozoic events within the KC, but were subsequently juxtaposed along a structural break active during the ca. 1901 Ma exhumation of the complex.

5.3 Analytical procedure

U/Pb geochronological samples were prepared and analysed at the Geological Survey of Canada (GSC). All samples collected were ~30 kg in weight, subsequently crushed in a steel jaw crusher and pulverized in a disk mill grinder. A heavy mineral separate was concentrated using a Wilfley™ table and heavy liquid (methyl iodine; specific gravity=3.3) separation procedures. Magnetic grains were removed from the mineral separate using a rare earth magnet. The remaining, less magnetic portion was passed through a Frantz™ isodynamic separator at increasing magnetic strengths; the chute through which the grains passed beneath the magnet was oriented at a side slope of 10°. After each run, the magnetic portion was extracted and bottled. The nonmagnetic reservoir was then run at maximum amperage (1.8Å) with the side slope of the chute altered for each run through 5°, 3°, 1°, 0° and -½°, and the magnetic portion from each run was separated, bottled and labelled.

Zircons, monazites, and rutiles were selected for analyses using a binocular microscope. In general, top lighting was used for initial grain selection based on morphology (discussed in detail below), and bottom lighting was used to further discern any flaws in the grains. Turbid and fractured grains, or those with inclusions or visible overgrowths were avoided where possible. Grain selection was based primarily on morphology, although colour and size were taken into account in some cases. Representative populations, or fraction types, were isolated and photographed.

Samples were then air abraded using a pyrite medium in order to remove the outer part of the zircon (Krogh, 1982). Some zircons were abraded to remove very thin (~10-25 µm) visible overgrowths. Multiple zircon grains of a given population were abraded together with the exception of zircon fragments thought to represent metamorphic

overgrowths. These were abraded individually so that, if broken upon abrasion, the pieces were known to have come from one fragment, thus reducing the possibility of analysing grains of different age and obtaining an arbitrary age that represents a mixture of zircon age populations. Zircons were abraded for between 1 and 10 hours depending primarily on their size and the degree to which they fractured. Monazites were not abraded, and only one of three rutile fractions was abraded for 20 minutes to remove surficial impurities.

Abraded zircons were photographed, weighed, and prepared for dissolution chemistry. Measurement utilized image analysis software that is designed to estimate the weight of a zircon based on its size and density (Matthews and Davis, 1999). Final selection of the best quality zircons from a variety of populations was then carried out, and selected fractions were transferred into clean Petri dishes and pipetted into a clean beaker.

Zircon and monazite fractions were spiked with a mixed $^{205}\text{Pb}/^{235}\text{U}/^{233}\text{U}$ tracer and dissolved in Teflon™ micro-capsules (Parrish, 1987). The chemical procedure followed for U and Pb extraction was similar to that described by Roddick et al. (1987). Pb and U were loaded separately onto Re filaments using Si-gel. Analyses were conducted using a Finnigan-Mat 261 multi-collector thermal ionization mass spectrometer in static mode (Roddick et al., 1987). Error propagation methods were those of Roddick et al. (1987). Linear regressions were calculated using a variation of the method of York (1969) and age errors are quoted at the 95% confidence level. Results for U/Pb isotopic analyses for the three samples dated are presented in Table 5.1.

Zircons representing the fraction types analysed were selected for cathodoluminescence (CL) and backscattered electron (BSE) imaging. Grains were mounted in an epoxy puck which was polished for optimal optical imaging. The puck was then carbon-coated and imaged using scanning electron microscope (SEM) at the GSC to observe the internal structure of representative zircon morphological types.

5.4 Sample and population description

5.4.1 Sample 98TXM334b – *Leucogabbro mylonite*

Leucogabbro mylonite was collected from the anorthositic metagabbro map unit (Figure 2.1) in order to determine the magmatic age of the anorthosite suite (Figure 5.1 and 5.2a; Table 5.1). The rock comprises ~55-60% plagioclase, ~30-35% clinopyroxene, and ~10% garnet, with minor orthopyroxene, quartz, ilmenite, magnesio-hastingsite, actinolite, and biotite. It is a fine-grained ultramylonite with a very well developed L-S mylonite foliation, and large ($>5 \times >5 \times <0.5$ cm), flattened aggregates of clinopyroxene indicative of coarse grain size prior to deformation (Figure 5.2a). Clinopyroxene aggregates have aspect ratios of up to ~20:1; the average aggregate size is ~5 cm in length and ~3 mm in width. A ~250 μm wide garnet corona typically surrounds clinopyroxene, and a <100 μm wide plagioclase moat commonly lies inboard of the garnets. The matrix is composed of fine-grained polycrystalline plagioclase, which, together with the clinopyroxene aggregates, is interpreted to have undergone dynamic recrystallization (Figure 5.2b).

Zircon was the only phase separated from the leucogabbro for U/Pb analysis. Five different populations were characterized (Table 5.2), and two of these were subsequently analysed (Figure 5.3): (1) Colourless, euhedral, prismatic grains that lack visible cores,

have slightly resorbed tips, and aspect ratios of $>3:1$ are the most common morphology in the leucogabbro (~80-90% of grains); and (2) a small population of equant, multifaceted, subrounded zircons. Back scattered electron (BSE) images of some zircons show subdued, low luminescent internal zoning parallel to crystal faces, which suggests that the zircons are magmatic (e.g. Pidgeon, 1992; Figure 5.4). More commonly, zircons are internally homogeneous or display patchy internal zoning, as seen using cathodoluminescence (CL) imaging. The lack of well-developed, oscillatory zoning may be due to post-emplacement diffusion of lead from the crystal lattice (Mezger and Krogstad, 1997). Complete lack of internal zoning, on the other hand, may be due to zircon recrystallization during deformation (e.g.: Wayne and Sinha, 1988), and this recrystallization may or may not have been accompanied by Pb-loss.

5.4.2 Sample 98TXM352 - Garnet-quartzofeldspathic gneiss

Garnet-quartzofeldspathic gneiss was collected from the qf gneiss map unit (Figure 5.1) in order to constrain the timing of granulite-facies metamorphism. The location, fabrics and mineralogy of the sample are representative of the ~2599 Ma sample dated by Tella et al. (1994b; Roddick and Tella, unpublished data, 1993). The rock comprises ~30% garnet, ~40% plagioclase, ~20% quartz (half of which is interpreted to be vein quartz), ~5% biotite, and <5% cummingtonite (interpreted to have formed via orthopyroxene breakdown), with trace ilmenite, pyrite, rutile, monazite, and zircon. The protolith is interpreted as either a supracrustal rock of intermediate composition (such as a metavolcaniclastic rock) but, perhaps more likely, the rock was a migmatite or diatexite prior to mylonitization. Compositional banding in the gneiss (Figure 5.8) is defined by layers of biotite, cummingtonite, and opaque minerals. The banding is oriented parallel

to the well-developed mylonitic foliation, defined by flattened aggregates of quartz and plagioclase, and by the preferred alignment of biotite. A ubiquitous, well-developed, shallowly east-plunging extension lineation is defined by ribbons of quartz and plagioclase and elongated mafic aggregates.

Descriptions of zircon and monazite extracted for geochronology are summarized in Table 5.3. Zircons are predominantly multi-faceted, nearly spherical, clear and colourless, with coarsely pitted surfaces (Z2-Z4) (Figure 5.6); rare prismatic crystals were noted but not analysed (Z1). Rare fragments with concave surfaces resembling metamorphic overgrowths that had been spalled off crystals, are also present (Z5). Some of the fragments have a cup-shaped morphology that more closely resembles spalled overgrowths (ie: #16 and #17 of Figure 5.6). These fragments and the spherical grains were selected for analysis in order to try to determine the timing of metamorphism. Multi-faceted, spherical to ovoid zircons, with pitted surfaces, are commonly interpreted to have grown in a metamorphic environment (van Breeman et al., 1987). The three monazite morphological types analysed were: 1) tabular and somewhat angular crystals (M1); 2) nearly spherical, smooth crystals (M2); and 3) equidimensional, angular grains or grain fragments (M5) (Figure 5.6c and d; Table 5.2).

CL and BSE images of zircons from the quartzofeldspathic gneiss generally reveal simple internal compositional zoning (Figure 5.7). Some Z2-Z4 grains show signs of possessing inherited cores (Figure 5.7a and c). Faint, concentric zoning is rarely present in the zircon spheres; however, when present, it is suggestive that these grains may have grown in a melt (eg: Pidgeon, 1992; Figure 5.7c). The majority of Z2-Z4 zircons lack obvious cores and are either homogeneous or possess weak, patchy internal zoning

(Figure 5.7b, d, and e). Unzoned crystals may be the result of metamorphic crystal growth, or recrystallization (Pidgeon, 1992). In contrast with the nearly euhedral prismatic zircons separated from the metaleucogabbro, these multi-faceted spherical to ovoid zircons have often been interpreted in terms of metamorphic growth (van Breemen et al., 1987, and references therein). It cannot, however, be established whether unzoned zircons represent metamorphic crystal growth or if the zircons had formed earlier and recrystallized during a thermal overprint (ie. 'metamorphic' vs. 'metamorphosed' zircons; Mezger and Krogstad, 1997). Most Z2-Z4 grains also possess thin (10-20 μm), featureless overgrowths, similar to zircons described from other granulite terrains (Hanchar and Rudnick, 1995), and these appear bright in BSE images relative to the crystal interiors (Figure 5.7b, d, and e). Fragments, thought to be metamorphic overgrowths, exhibit a homogeneous internal structure.

5.4.3 Sample 98TXM298b – Tonalite mylonite

Tonalite mylonite (Figure 5.5) occurs in 10-50 cm thick sheets that are concordant with the foliation in the surrounding diorite gneiss. The mylonite foliation and granulite-facies assemblage indicate that the rock has a shared metamorphic and deformation history with the quartzofeldspathic gneiss and may help to constrain the timing of granulite-facies tectonism within the Southern bounding shear zone (Figure 5.1). The rock is leucocratic, composed of up to 90% quartz and feldspar, with only ~5% biotite, ~3% orthopyroxene, and trace hornblende, ilmenite, garnet, zircon and rutile. Plagioclase porphyroclasts average <2 mm in diameter, and comprise <10% of the rock. The matrix is composed of fine-grained (<200 μm) plagioclase interpreted to have formed by progressive rotation of subgrains derived from porphyroclasts. Orthopyroxene occurs as

sparse elongate polycrystalline aggregates or locally interspersed between fine-grained matrix plagioclase. The low proportion of porphyroclasts and the fine grain size are interpreted to be a product of grain size reduction as a result of ductile deformation. Polycrystalline quartz aggregates range up to ~3 mm in thickness and have extremely large (~20:1) aspect ratios. Concordant quartz veins may also be present.

Zircon and rutile were separated for geochronological analysis. Descriptions of both minerals are presented in Table 5.4. Zircons are predominantly nearly equant, ellipsoidal to roughly cubic and well-faceted (Population 2, Fraction I.D. Z2, Z3, and Z4 of Table 5.4) (Figure 5.9). Subhedral stubby prisms (aspect ratio <3:1), though not abundant, are also present (Population 1, Fraction I.D. Z1 of Table 5.4). BSE images reveal that the majority of all zircons display thin (~10-20 μm) overgrowths with no discernable internal structure (Figure 5.7a and b). These internally homogeneous zircons are interpreted to have grown in a metamorphic environment.

Rutile was separated based primarily on colour, although characteristics such as grain size were also used to group the grains. Large, black, vitreous grains and fragments were most common. Dark red to rust-coloured rutile fragments were less common and some of these had dark inclusions, thought to be intergrowths of ilmenite. Deep orange to amber rutile fragments were least abundant and commonly had intergrowths of ilmenite which were physically removed by cracking the grains. The black rutile comprise the largest grain size (up to 400 μm), red were intermediate and the orange fragments were smallest (Table 5.4).

5.5 Results

5.5.1 Sample 98TXM334b – *Leucogabbro mylonite*

Nine zircon fractions from the leucogabbro were analysed, and the data vary in discordance (0.3 – 3.0%), with most $^{207}\text{Pb}/^{206}\text{Pb}$ ages falling in the 2720-2636 Ma range. Two fractions (#1, #2) of the Z1 fraction (Table 5.2; Figure 5.3) yielded significantly older $^{207}\text{Pb}/^{206}\text{Pb}$ ages than the other seven (Table 5.1). Fraction #1, consisting of 12 grains, is 3% discordant and gave a $^{207}\text{Pb}/^{206}\text{Pb}$ age of ca. 2788 Ma. Fraction #2 (three crystals) was 1.2% discordant and yielded a $^{207}\text{Pb}/^{206}\text{Pb}$ age of ca. 2770 Ma. Fraction #3 was a multi-grain fraction, comprising 12 grains, and yielded a slightly discordant (0.4% discordance) age of ca. 2706 Ma. Fractions #4 and #5 were both single grain fractions from the Z5 fraction (Table 5.2; Figure 5.3), and gave slightly discordant (0.3-0.4% discordance) ages of ca. 2705 Ma and ca. 2702 Ma, respectively. Slightly discordant results (0.4%-0.5% discordance) were also determined for fraction #6 (6 grains), fraction #7 (4 grains) and fraction #8 (2 grains), yielding $^{207}\text{Pb}/^{206}\text{Pb}$ ages of ca. 2699, 2699 and 2696 Ma, respectively. Fraction #9 is 1.5% discordant, with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of ca. 2669 Ma.

Four fractions, #'s 1, 2, 3, and 7, contain high quantities of common Pb (>100 pg), resulting in larger errors associated with the $^{207}\text{Pb}/^{206}\text{Pb}$ age of each of these fractions. Contamination in the lab may have been the source of high common Pb, and these analyses were calculated accordingly. No discernible link between age and abundance of common Pb could be drawn.

A Yorkfit linear regression (York, 1969; Roddick, 1987) through six fractions (#4 - #9, inclusive; see Table 5.5 for details of regression output) yields an upper intercept of 2711 $\pm$ 5/-4 Ma and a lower intercept of 1933 $\pm$ 86 Ma (Figure 5.11), with a mean

standard weighted deviation (MSWD) of 1.44. The upper intercept is interpreted as the age of magmatic crystallization; the lower intercept age suggests some component of Paleoproterozoic Pb-loss, renewed zircon growth, or recrystallization. Either Pb-loss or recrystallization is more likely since no overgrowths were detected on the zircons. The lower intercept, although poorly constrained due to the clustering of data points toward the upper intercept, is similar to the ca. 1936 zircon Ma U/Pb age determined by Tella et al. (1994b) from anorthosite and mafic granulite ~4 km to the west (Figure 5.1). The distribution of data suggests that multi-grain fractions #1 and #2 contain a component of inherited zircon, although no cores were optically detected in the grains from either fraction. Possibly a discrete grain within each fraction was older, and is responsible for the inherited component. The older $^{207}\text{Pb}/^{206}\text{Pb}$ ages are not an artifact of a high blank, as no reasonable modification of the amount of blank Pb produces significantly younger $^{207}\text{Pb}/^{206}\text{Pb}$ ages.

5.5.2 Sample 98TXM352 - Garnet-Quartzofeldspathic Gneiss

Eight zircon fractions and four monazite fractions were analysed from the qf gneiss (Table 5.4). Zircon and monazite contained less than 5 and 8 pg common Pb, respectively, except for two zircon fractions, #13 and #14, which contain 16 and 17 pg, respectively. All zircon fractions contain low U concentrations (<90 ppm). All but one fraction (#15) analysed were single-grain fractions.

Monazite data

All monazite fractions (#18 - #21) were single-grain fractions. Results are variably discordant (0.8 - 5.0%) with $^{207}\text{Pb}/^{206}\text{Pb}$ ages ranging between 2606 Ma and 2298 Ma. The oldest monazite contains notably high U (3905 ppm), exceeding that of all

other fractions by an order of magnitude. Fractions #18, #19, #20, and #21 yielded $^{207}\text{Pb}/^{206}\text{Pb}$ ages of ca. 2558, 2558, 2542, and 2298 Ma, respectively.

Regression of three of the four monazite fractions (excluding fraction #19) yields the best-fit chord with an upper intercept of ca. 2629 Ma and a lower intercept of ca. 1921 Ma (MSWD = 0.78; Figure 5.11a; Table 5.5). This regression is preferred over the regression of all data, which gave similar upper and lower intercepts, but with significantly higher MSWD (23.5). The monazite fraction discarded from the preferred regression plots slightly above discordia and may have been affected by early Pb-loss or recrystallization. The other three monazite fractions are interpreted to represent mixtures of ca. 2606-2629 Ma monazite and ca. 1921 Ma monazite. The upper intercept is interpreted as the maximum age of Archean metamorphism based on monazite morphology and thermobarometric evidence that temperatures of the earliest metamorphic event to have affected the Uvauk gneiss (Chapter 4) were in excess of the closure temperature of monazite (700°C, Heaman and Parrish, 1991, and references therein). The lower intercept is compatible with either renewed metamorphic growth of monazite, recrystallization or Pb-loss at ca. 1920 Ma. This interpretation of the monazite data is consistent with the interpretation by Tella et al. (1994b) of two events of granulite-facies metamorphism: one at ca. 2.6 Ga and a second at ca. 1.9 Ga.

Zircon data

The zircon data are discordant (0.4 to 2.6%) with $^{207}\text{Pb}/^{206}\text{Pb}$ ages that range from ca. 2584 to ca. 2492 Ma. Fractions #10 and #11 (Table 5.1) are 0.4% discordant and yielded $^{207}\text{Pb}/^{206}\text{Pb}$ ages of ca. 2584 Ma and ca. 2578 Ma, respectively. Fraction #12 is 1.0% discordant and yielded a $^{207}\text{Pb}/^{206}\text{Pb}$ age of ca. 2574 Ma. The remaining five

fractions yielded $^{207}\text{Pb}/^{206}\text{Pb}$ ages between ca. 2528 and ca. 2492 Ma. Fractions #13 and #14, a round grain and a fragment, yielded similar $^{207}\text{Pb}/^{206}\text{Pb}$ ages of ca. 2528 and ca. 2527 Ma, and are 0.6 and 1.2% discordant, respectively. The one multi-grain fraction (#15) gave a $^{207}\text{Pb}/^{206}\text{Pb}$ age of ca. 2513 Ma and is 0.8% discordant. Fractions #16 and #17, both fragments interpreted to represent overgrowths, are 0.8 and 2.6% discordant, and yielded the youngest $^{207}\text{Pb}/^{206}\text{Pb}$ ages of ca. 2501 and ca. 2492 Ma, respectively.

Although all zircon fractions are discordant, most are <1% discordant and plot below concordia between ca. 2500 and ca. 2600 Ma. No correlation exists between apparent age and grain size, grain morphology (with the exception of some grains thought to represent overgrowths), or grain chemistry (Figure 5.12). Two fragments thought to represent overgrowths, #16 and #17, yield the youngest $^{207}\text{Pb}/^{206}\text{Pb}$ ages, and these possess the most convincing overgrowth morphology of all fragments picked as overgrowths. All zircons contain relatively low U concentrations (36-88 ppm) and Th/U ratios (0.4-0.9).

Interpretation of the zircon data is complicated because the data do not define a single age and can only be explained by superposition of more than one process. Data distributions similar to that obtained for zircons from the Uvauk quartzofeldspathic gneiss are common in many granulite-facies terranes (e.g. Ashwal et al., 1999; Vavra et al., 1996; Corfu et al., 1994; van Breemen et al., 1987). They have been interpreted to reflect: 1) growth of metamorphic zircon over older crystals (Corfu et al., 1994; van Breemen et al., 1987); 2) continuous Pb-loss or recrystallization due to protracted high-grade metamorphism (Ashwal et al., 1999); and 3) episodic zircon growth or recrystallization due to polymetamorphism (Corfu et al., 1994; Vavra et al., 1996; Pan et

al., 1998). Differentiating between these possibilities is difficult because all can produce a similar data distribution, and it is also possible to have superimposed effects of two or more of these processes. Furthermore, the effects of Pb loss, either before or after high-grade metamorphism, compound the aforementioned complexities (Mezger and Krogstad, 1997). The zircon data will be considered in light of these three possibilities.

Possibility 1: Growth of metamorphic zircon over older crystals

Similar data distributions have been interpreted as the result of overgrowth of metamorphic zircon on older crystals (e.g. van Breemen et al., 1987). Zircons that are the products of these processes would be expected to contain visible cores detectable using CL or BSE imagery (e.g. van Breemen et al., 1987; Vavra et al., 1996; Schaltegger et al., 1999). Three possible interpretations follow if cores are observed, and the preferred interpretation will consider the data distribution as well as the protolith to the sample: 1) In the case of an orthogneiss, the upper and lower intercepts of a linear regression would be interpreted to represent the age of the protolith and metamorphism, respectively. 2) If, however, metamorphic zircon overgrew cores of variable age (e.g. a sedimentary protolith), then a lower intercept representing the age of metamorphism would have multiple upper intercepts reflecting the age of inheritance for each fraction. 3) Alternatively, the SEM images may reflect polymetamorphism with metamorphic zircon overgrowth on cores that formed during an earlier metamorphic event. Although zircons from the Uvauk gneiss do not exhibit detectable cores, the three possibilities outlined above must be considered in any viable interpretation. Further, if the zircons have recrystallized, the internal morphological details may not be preserved. The best interpretation of this equivocal data set will consider the zircon data in conjunction with

interpretation of the monazite data (above). In accord with the interpretation of monazite data, the upper and lower intercepts of a linear regression more likely represent the age of two metamorphic events. Possibly, the upper intercept corresponds to the age of recrystallized *metamorphic* cores, while the lower intercept corresponds to the minimum age of subsequent metamorphism resulting in either renewed zircon growth or, perhaps more likely, recrystallization resulting in the lack of preservation of internal features.

Possibility 2: Continuous Pb-loss or recrystallization due to protracted high-grade metamorphism

Geochronological studies of crustal granulites from southwest Madagascar resulted in an age spectrum of over 80 m.y. (Ashwal et al., 1999). Ashwal et al. (1999) interpreted this type of data distribution for granulite-facies anorthositic rocks to result from high-temperature Pb-loss during one or more protracted periods of metamorphism. Features leading to this conclusion include concordant $^{207}\text{Pb}/^{206}\text{Pb}$ ages and a correlation between grain size, internal features, and apparent age. In particular, Ashwal et al. (1999) found that the oldest zircons are the largest grains that show relict magmatic zoning, whereas the youngest grains are small fragments that contained high-U crack networks resembling a bright webbing in SEM images. The Madagascar data are interpreted to result from Pb-loss by volume diffusion in originally magmatic zircons during one or more protracted periods of granulite-facies metamorphism.

Zircons of the Uvauk gneiss show a distribution of $^{207}\text{Pb}/^{206}\text{Pb}$ ages ranging over 80 m.y., similar to the range described by Ashwal et al. (1999). However, zircons from the Uvauk gneiss do *not* exhibit those internal features described by Ashwal et al. (1999). Interpretation of protracted granulite-facies metamorphism would be supported by a

chemistry-age correlation, such as higher U concentration in either younger or more discordant zircons. However, zircons of the Uvauk gneiss do not exhibit this kind of correlation either (Figure 5.12d). In addition, recent studies on Pb diffusion in zircon suggest that diffusion rates are too slow to significantly alter Pb isotope ratios by volume diffusion in crystalline zircon (Cherniak and Watson, 2001), suggesting that continuous Pb-loss must occur by a mechanism other than volume diffusion (e.g. fracture-assisted diffusion, annealing, or recrystallization).

Possibility 3: Episodic zircon growth or recrystallization due to polymetamorphism

A mechanism of episodic or diachronous zircon growth or recrystallization has been invoked to explain similar data distributions from granulite-facies terranes elsewhere (eg: Corfu et al., 1994; Vavra et al., 1996; Pan et al., 1998) and is the preferred interpretation for this data set. For example, Vavra et al. (1996) suggested that three episodes of granulite-facies metamorphism occurred in the Ivrea Zone of the southern Alps between ca. 296 Ma and ca. 226 Ma. The first is interpreted as having resulted from dehydration melting, the second due to decompression melting, and the third to have been induced by fluid-infiltration. CL images of zircons from Ivrean rocks show complex internal zoning, in which Vavra et al. (1996) were able to identify various internal domains based on the type of zoning displayed, and to subsequently interpret the origin of most of the domains (eg: detrital cores, unaltered magmatic protoliths, recrystallized inherited cores, and overgrowths exhibiting a range of zoning styles). Although zircons of the quartzofeldspathic gneiss do not exhibit such complexity in internal structure, episodic zircon growth under two or more granulite-facies conditions remains possible, as such features may have been destroyed by recrystallization.

Interpretation of Uvauk qf gneiss

The zircon data from the quartzofeldspathic gneiss are interpreted in terms of episodic growth, consistent with the evidence of metamorphic overgrowths (discussed below). Regression of the least discordant data (Fractions 10, 11, 13, 15, 16; Table 5.5) yields an upper intercept of ca. 2620 Ma and a lower intercept of ca. 2319 Ma (chord 1 on Figure 5.11) with an MSWD of only 0.4. Interpretation of the upper intercept (ca. 2620 Ma) as a metamorphic age is compatible with interpretation of the upper intercept from the monazite data (ca. 2630 Ma) as representing the age of initial metamorphism. However, this interpretation is not considered viable because of the effects of multiple processes such as Pb-loss as indicated by those fractions that lie below chord 1, and all zircons may have been variably affected by such processes. Processes such as Pb-loss would effectively rotate the chord from its original position (e.g. Mezger and Krogstad, 1997). Interpretation of lower intercepts is often uncertain as it cannot be assumed that the zircons did not lose any Pb after formation or recrystallization, nor can it be assumed that zircons from metamorphic rocks have lost Pb only during a discrete period of metamorphism, particularly in old, polymetamorphic terrains (Mezger and Krogstad, 1997). If the zircons have been affected by Pb-loss, then the data points may have been offset to a position below the original mixing line. This would result in a downward shift of the discordia line, yielding a younger lower intercept and an older (or possibly younger) upper intercept (Mezger and Krogstad, 1997). The net effect, then, of late to recent Pb-loss would be to generate an inaccurate upper intercept (likely the earliest metamorphic event) and a lower intercept (the second metamorphic event) that may not

be geologically significant. For these reasons, the ages indicated by chord 1 are not considered geologically significant.

Thin (10-20 μm) overgrowths are evident on most zircons from the qf gneiss (Figure 5.7b, d, and e), and although these were likely to have been removed during the 5-10 hours of abrasion with pyrite, they may represent a component of later (*after* ca. 2319 Ma, possibly ca. 1900 Ma) zircon growth. The 1921 ± 13 Ma lower intercept of the monazite data indicates that either a thermal event ($>700^\circ\text{C}$, based on monazite closure temperature; Heaman and Parrish, 1991, and references therein) or Pb-loss, affected the rock at this time. Although no ca. 1920 Ma thermal effect is apparent from the zircon data, there is compelling evidence that Paleoproterozoic high-grade metamorphism did occur within various places in the region. Granulite-facies metamorphism of the UC anorthosite suite (1) was interpreted to have occurred at ca. 1936 Ma (Tella et al., 1994b), (2) is consistent with the lower intercepts for zircon data from the leucogabbro and for monazite data from the quartzofeldspathic gneiss, (3) has been documented in the KC at ca. 1900-1910 Ma, along with coeval anorthosite magmatism (Sanborn-Barrie, 1999), and (4) has been documented regionally within the Hearne domain (Stern and Berman, 2000). The effect of a ca. 1900 Ma event on zircons from the quartzofeldspathic gneiss may be cryptic, especially if the rock was *already* a dry, granulite-facies rock. Interestingly, zircons from ultramylonites from the Brevard fault zone were interpreted to have suffered total Pb-loss during deformation under amphibolite-facies conditions and were not affected by subsequent deformation events (Wayne and Sinha, 1988). It is possible that the effects of a ca. 1900 Ma episode on zircons from the gneiss produced the fine metamorphic overgrowths (Figure 5.7), and/or caused minor Pb-loss from the existing

zircons. Either of these processes would have the aforementioned effect of shifting the discordia chord *downward*, resulting in a young lower intercept with uninterpretable significance.

Regression of the oldest fractions (# 10, 11 and 12) through a lower intercept of ca. 1900 Ma results in a discordia curve (chord 2 on Figure 5.11; MSWD = 1.7), that is compatible with a component of ca. 1900 Ma zircon growth, recrystallization or Pb-loss. The upper intercept (ca. 2595 Ma) would then be interpreted to represent a minimum age of early metamorphism. This interpretation is consistent with the interpretation of the upper intercept derived from the monazite data as indicative of the age of early metamorphism. Although the lower intercept was forced to ~1900 Ma, the MSWD suggests the discordia line is a good fit. The data are compatible with, but do not necessarily argue for, minor zircon growth or Pb-loss at ca. 1900 Ma.

Fractions 13-16 plot to the left (or above) the discordia described above, and, therefore, require a pre- 1.9 Ga event to explain their distribution. If zircons from the Uvauk gneiss sample have been affected by a ca. 2600 Ma early metamorphism and a late 1900 Ma thermal event, what, then, would be the age of the second event? Widespread metamorphism at ~2550-2500 has been documented throughout the northern Hearne domain (Stern and Berman, 2000; Berman et al., 2000) and south of the KC (Sanborn-Barrie, 1999). Of the zircon fragments analysed, fractions #16 and #17 most closely resemble metamorphic overgrowths (Figure 5.6). These define a chord (chord #3 on Figure 5.11) with an upper intercept of 2505 Ma, suggesting that these metamorphic overgrowths may have grown at that time. The lower intercept, ~800 Ma, is interpreted to represent a component of late to recent Pb-loss and may be a geologically irrelevant

age. In this scenario, the ca. 2527 Ma $^{207}\text{Pb}/^{206}\text{Pb}$ ages of fractions #13 and #14, and upper intercepts of regressions 3 and 4 (Figure 5.11; Table 5.5), are compatible with zircon growth between 2505-2530 Ma, in further support of another high-grade metamorphic event at that time.

In summary, the zircon age data are consistent with metamorphic growth or recrystallization of zircon between ca. 2595 and 2620 Ma, between 2500-2530 Ma, and possibly again at ca. 1900 Ma. Lack of correlation between apparent age and morphology, grain size, Th/U, or U content, is consistent with the hypothesis that individual grains represent recrystallized crystals containing mixed ages of radiogenic Pb (Figure 5.12).

5.5.3 Sample 98TXM298b – Tonalitic mylonite

Tonalitic mylonite is found intercalated with dioritic to monzodioritic mylonite and quartzofeldspathic gneiss, and contains a granulite-facies assemblage dominantly consisting of plagioclase and quartz, with minor garnet and orthopyroxene. Due to similarities in metamorphic grade and deformation fabrics of the tonalitic mylonite with the dioritic to monzodioritic mylonite and quartzofeldspathic gneiss, these rocks are thought to have a shared tectonothermal history. The interpretation of the quartzofeldspathic gneiss is highly relevant to the tonalitic mylonite because the latter exhibits a similar, but more closely constrained set of age data, ranging from ~2560 Ma to ~2520 Ma.

Six single-grain zircon fractions and three multi-grain rutile fractions were analysed from the tonalitic mylonite (Table 5.1). Two zircon fractions (#22 and #24) yielded results that are <0.3% discordant, three are between 0.5-0.9% discordant (#23,

#26, and #27), and another (#25) is reversely discordant (-1.8%). Fractions #22 and #24 yield $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 2561 Ma and 2548 Ma, respectively. Fraction #23 yields a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2556 Ma. Fraction #25 yields a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2544 Ma. Fractions #26 and #27 yield $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 2532 Ma, and 2523 Ma, respectively. The error associated with each analysis varies considerably, from as high as ± 19.5 m.y. (for fraction #24) to within ± 1.4 m.y. (for fraction #22).

Rutile fractions yielded discordant $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 1647 Ma, 1614 Ma and 1392 Ma for fractions #28, #29 and #30, respectively. The extremely large error ellipses associated with these analyses result from the very low concentrations of U and radiogenic Pb within the samples (Table 5.4). The young $^{207}\text{Pb}/^{206}\text{Pb}$ ages of rutile relative to the zircon ages most likely reflect the significantly lower closure temperature of rutile (400°C for rutile vs. >800-900°C for zircon; Heaman and Parrish, 1991, and references therein; Cherniak and Watson, 2001). The rutile data are interpreted to indicate that the rock cooled below 400°C, the temperature at which U and Pb diffusion is thought to cease in rutile at, or before, ca. 1600 Ma.

Analysis of U-Pb data from the quartz ribbon-bearing tonalitic mylonite demonstrates that significant zircon growth or resetting occurred during the Neoarchean. Like the age data for the quartzofeldspathic gneiss, the age data for the tonalitic mylonite do not define a simple two-component history. No clear-cut correlation exists between apparent age and grain size, morphology or chemistry (Figure 5.12). U concentrations and Th/U values do not indicate any change in zircon chemistry with apparent age. If continuous or episodic new growth of metamorphic zircon occurred, it occurred under similar conditions thereby producing zircon with roughly the same chemistry.

Regression through all zircon fractions (#'s 22-17) yielded a discordia chord with an upper intercept of ca. 2592 Ma and a lower intercept of ca. 2324 Ma (MSWD = 0.78; Figure 5.13; Table 5.5). The simplest interpretation of this data is that the upper intercept represents either the age of protolith crystallization, or inheritance, and that the lower intercept represents the age of metamorphism. However, the sample must be interpreted in light of the interpretation of the quartzofeldspathic gneiss because the two rocks are interpreted to have a shared metamorphic history. The upper intercept here is consistent with the age of Neoproterozoic metamorphism inferred from the quartzofeldspathic gneiss (ca. 2600 Ma). The lower intercept is consistent with a U-Pb (titanite) age from a gabbro that cuts supracrustal rocks south of the KC, interpreted as a metamorphic and/or cooling age (Sanborn-Barrie, 1999). However, as was the case for the qf gneiss, this age may not be geologically relevant but may reflect the combined effects of zircon growth or recrystallization at ca. 2500-2530 Ma and zircon growth, recrystallization or Pb-loss at ca. 1900 Ma. The lack of any obvious Paleoproterozoic thermal effect on the zircons may be related to the difficulty in resetting the U-Pb systematics in hot, dry rocks.

5.6 Summary of conclusions

Geochronology of rocks from the UC places constraints on the magmatic age of the anorthosite suite and provides evidence for both Neoproterozoic and Paleoproterozoic metamorphism. An igneous crystallization age of 2711 Ma was determined for the leucogabbro (sample 98TXM334b), based on a cluster of seven nearly concordant fractions. The sample is thought to represent a more mafic constituent of the anorthosite suite and precludes at least part of the UC anorthosite from being related to the ca. 1902 Ma anorthosite at Kramanituvar (Sanborn-Barrie, 1999). However, the possibility for a

separate Paleoproterozoic phase of anorthosite or mafic dyking within the UC cannot be ruled out.

Interpretation of metamorphic ages from U-Pb data sets for both the qf gneiss (98TXM352) and the tonalite mylonite (98TXM298b) is complicated because the data sets do not define single colinear arrays. The visually homogenous zircon interiors may have contained multiple age domains that are no longer discernable due to recrystallization (Figure 5.7). This severely limits the use of SEM and CL imagery as interpretive aids. Nonetheless, geochronological data from the quartzofeldspathic gneiss and tonalitic mylonite are interpreted to reflect a polymetamorphic history. Zircon growth and/or recrystallization at ca. 2595-2630 Ma is indicated by the upper intercepts of zircon data (98TXM352 and 98TXM298b) and monazite data (98TXM352). Renewed zircon growth is interpreted to have occurred at ca. 2500-2530 Ma, based on the age of zircon overgrowths in the qf gneiss (98TXM352). Further zircon growth, recrystallization, or Pb-loss is interpreted to have occurred at ca. 1920 Ma, based primarily on the lower intercept determined from monazite data for the qf gneiss (98TXM352). An event is also indicated at this time by the lower intercept calculated for zircon data from the leucogabbro (98TXM334b). A component of late to recent Pb-loss is indicated by the more discordant zircon fractions (e.g. fraction #17).

The level of confidence in the interpretations varies, and in some cases is based primarily on consistency arguments. A high level of confidence is attributed to the interpretation of 2711 ± 5 Ma crystallization of anorthosite, although this may reflect the age of only a part of the suite and the presence of a Paleoproterozoic anorthosite component cannot be precluded. The interpretation of ca. 2.6 Ga metamorphism is based,

primarily, on the interpretation that the monazite in the Uvauk qf gneiss is metamorphic in origin. The protolith is thought to be either a supracrustal rock, or possibly a tonalite, neither of which commonly contains primary monazite. Metamorphism of Southern bounding shear zone rocks at ca. 2.6 Ga is also consistent with the field interpretation that the ca. 2711 Ma leucogabbro-anorthosite is autochthonous (or parautochthonous) and was magmatically emplaced into the surrounding, predominantly tonalitic country rocks. Ca. 2.50-2.53 Ga metamorphism is interpreted based on both the presence of ca. 2.5 Ga zircon overgrowths, and on synchronicity with a regionally significant metamorphic event documented at Parker and MacQuoid lakes (Stern and Berman, 2000; Berman et al., 2000), at Yathkyed Lake (MacLachlan et al., 2000), and within the Southern shear zone at the KC (Sanborn-Barrie, 1999). Ca. 1900 Ma metamorphism is based on lower intercept data for the leucogabbro (98TXM334b) and the qf gneiss (98TXM352), and is consistent with the age of metamorphism documented in other parts of the Hearne domain (e.g. Sanborn-Barrie, 1999; Stern and Berman, 2000; Berman et al., 2000).

Rutile data from the tonalitic mylonite indicates that the rock cooled through ~400°C, the closure temperature of rutile (Heaman and Parrish, 1991), by ca. 1600 Ma. On the basis of the current data set, it is not possible to distinguish whether the rocks were kept at temperatures above 400°C until this time or experienced a distinct Proterozoic thermal pulse. Disturbance of the rutile Pb systematics could possibly be related to intrusion of the regional Nuelin Suite granites (1765-1750 Ma; Peterson et al., 2000) with subsequent cooling through 400°C by ca. 1600 Ma.

The significance of a 1900 Ma event recorded by rocks of the Uvauk complex

Roddick and Tella (unpublished data, 1993; Tella et al., 1994b) yielded a U-Pb (zircon) age of 1936 Ma, interpreted as the age of metamorphism, for garnet anorthosite and garnet amphibolite (likely a retrogressed Uvauk dyke) from the western UC (Figure 5.1). Both rock types are mylonitic and contain garnets with well-developed coronae of symplectitic plagioclase and hornblende, interpreted to have formed by garnet breakdown during decompression (see Chapter 4). In the garnet amphibolite, garnet has typically been completely replaced by intergrowths of hornblende and plagioclase. These rheologically weak garnet pseudomorphs behaved ductilely and have locally been deformed to define an axial planar foliation. Hornblende in the amphibolite is also aligned, and where the mylonite foliation is folded, hornblende and the garnet pseudomorphs define an overprinting, axial planar foliation (refer to section 3.7.3; Figure 3.21). Based on textures such as relict foliation-forming clinopyroxene rimmed by axial planar hornblende and replacement of garnet by axial planar hornblende and plagioclase, the axial planar foliation is interpreted to have developed during retrograde metamorphism (characterized by garnet breakdown and replacement of clinopyroxene by hornblende). Abundant titanite (2-5%) is present in anorthosites and garnet amphibolites that are interpreted to have undergone retrograde metamorphism at upper amphibolite to lower granulite facies. Roddick and Tella (unpublished data, 1993) determined an age of ca. 1903 Ma (U-Pb, titanite) from anorthosite and mafic granulite of the Northern bounding shear zone (Figure 5.1), and this may represent a high temperature part of the cooling path.

The 1936 Ma interpreted age of metamorphism may not correspond to the timing of garnet growth. Possibly, the zircon dated by Roddick and Tella (unpublished data,

1993) was produced as Zr-bearing phases such as garnet (Fraser et al., 1997) and clinopyroxene (Heaman and Parrish, 1991) were destroyed. The ca. 1936 Ma age (U-Pb, zircon; Roddick and Tella, unpublished data, 1993) may thus mark the time of garnet and/or clinopyroxene breakdown (Roberts and Finger, 1997), with subsequent cooling below the closure temperature of titanite (660-700°C, Scott and St-Onge, 1995) at ca. 1903 Ma (U-Pb, zircon; Roddick and Tella, 1993, unpublished data).

Recommendations

In-situ microprobe dating of zircon cores and metamorphic overgrowths from both the quartzofeldspathic gneiss and the tonalitic mylonite is recommended to better constrain the timing of multiple metamorphic events. In particular, the age of the bright zircon overgrowths (e.g. zircons from 98TXM352 and 98TXM298b; Figure 5.7) and the age of any relict domains or visually undetectable cores could be determined this way. In addition, chemical dating of monazite, in conjunction with the production of zoning maps, may provide further constraints on the timing of deformation and metamorphism.

Legend

Anorthosite suite	
	Anorthosite
	Anorthositic gabbro, leucogabbro
	Garnet-anorthosite
Surrounding geology	
	Mafic granulite, enderbite, charnockite
	Tonalite gneiss, amphibolite
	Diorite to monzodiorite gneiss
	Garnet-quartzofeldspathic gneiss
	Tonalite straight gneiss, mixed metasedimentary rocks
	Tonalite orthogneiss
	Garnet-biotite semipelitic schist
	Sample location
	Limit of mapping

Figure 5.1: Location map for geochronological samples from the western Uvauk complex. Note data summarized from Roddick and Tella (1993, unpublished data), and localities of samples dated in this study.

Figure 5.2: Leucogabbro mylonite from the Uvauk complex. a) coarse-grained leucogabbro mylonite with elongate clinopyroxene aggregates (sample 98TXM334b, dated in this study), b) photomicrograph of leucogabbro mylonite. Polycrystalline aggregates of clinopyroxene and plagioclase are interpreted to have been dynamically recrystallized. Note coronitic garnet surrounding elongate clinopyroxene aggregates and thin plagioclase between the two (field of view is 5.2 mm).

Figure 5.3: Zircon representative of the dated leucogabbro mylonite (sample 98TXM334b; see Table 5.2 for descriptions).

Figure 5.4: BSE image of zircons from the dated leucogabbro mylonite (sample 98TXM334b): a) subrounded crystal with faint, irregular, and patchy zoning; b) prismatic crystal showing faint zoning parallel to crystal faces; c) subrounded crystal showing faint, patchy zoning and slightly darker rim; d) subrounded, elongate crystal showing faint zoning parallel to crystal faces, truncated at zircon edge at bottom of plate.

Figure 5.5: Garnet quartzofeldspathic gneiss (sample 98TXM352): a) outcrop photo, showing compositional banding and pock-markings from weathered-out garnets; b) photomicrograph showing biotite foliation wrapping subhedral garnet, and abundant biotite in pressure shadows (field of view is 5.2 mm).

Figure 5.6: Zircon and monazite from garnet quartzofeldspathic gneiss (sample 98TXM352): a) spherical to ovoid zircon crystals with pitted surfaces (Z3); b) zircon fragments, thought to represent metamorphic overgrowths spalled free from cores (Z5). The cup-shaped morphologies of #16 and #17 most closely resemble overgrowths; c) oblate, well-rounded monazite with smooth surface (M2); d) thin, platy to tabular monazite crystals and fragments (see Table 5.3 for descriptions).

Figure 5.7: BSE images of zircons from garnet quartzofeldspathic gneiss (98TXM352) and tonalite mylonite (98TXM298b): a) subhedral, multi-faceted prisms of type Z1 from tonalite mylonite showing faint evidence of a slightly brighter zircon core; b) clear, colourless spheres from type Z2 of tonalite mylonite, exhibiting thin metamorphic overgrowths; c) clear, colourless spherical grain from type Z3 of quartzofeldspathic gneiss displaying oscillatory zoning, suggesting the zircon originally grew in a melt; d) and e) clear, colourless spheres from type Z3 of quartzofeldspathic gneiss showing homogenous cores and thin (<20 μm) metamorphic overgrowths, similar to b).

Figure 5.8: Outcrop photograph of tonalite mylonite (sample 98TXM298b) with well-developed mylonite foliation defined by elongate aggregates and ribbons of quartz and plagioclase.

Figure 5.9: Zircon and rutile populations from tonalite mylonite (sample 98TXM298b; see Table 5.3 for descriptions).

Figure 5.10: Concordia diagram of zircon data from leucogabbro mylonite (sample 98TXM334b; see text for interpretation).

Figure 5.11: Concordia diagrams of (a) monazite and (b) zircon data from garnet quartzofeldspathic gneiss (sample 98TXM298b; see text for interpretation).

Figure 5.12: Comparison of zircon ages to chemical properties: a) plot of Th/U vs. age (Ma); b) plot of U (ppm) vs. age (Ma); c) plot of zircon mass (pg) vs. age (Ma); and d) plot of U concentration (ppm) vs. % discordance. Each plot shows a lack of correlation between zircon grain size and chemistry with age, or % discordance.

Figure 5.13: a) concordia diagram of zircon and rutile data from quartz-ribbon tonalite mylonite (98TXM298b); b) detail showing zircon data (see text for interpretation).

Table 5.4 U/Pb analytical data for samples collected from the Uvauk complex

Fraction ^a	Wt. ¹ U (μg)	Pb ^b (ppm)	Pb ^b (pg)	Atomic ratios ^d						Apparent Age (Ma)	
				$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$	$\frac{^{206}\text{Pb}}{^{206}\text{Pb}}$	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$	$\frac{^{206}\text{Pb}}{^{207}\text{Pb}}$		
				$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$	$\frac{^{206}\text{Pb}}{^{206}\text{Pb}}$	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$	$\frac{^{206}\text{Pb}}{^{207}\text{Pb}}$		
98TXM334b (Z5822): Uvauk complex leuco gabbro (sample location: 63°51.644'N, 92°40.630'W)											
1. Z1a(12),nm1°,eu,pr,incl,co	24	104	73	333	259	0.362	0.5249 ±.55%	14.141 ±.50%	0.19537 ±.24%	2720	2787.9 ±7.8
2. Z1c(3),nm1°,eu,pr,incl,co	18	136	97	197	366	0.362	0.5304 ±.13%	14.130 ±.42%	0.19324 ±.33%	2743	2769.9 ±11.0
3. Z2a(12),nm0°,eu,pr,cl,co	28	95	64	311	293	0.319	0.5196 ±.47%	13.320 ±.43%	0.18592 ±.22%	2697	2706.4 ±7.3
4. Z5b(1),nm0°,fg,eq,cl,co	5	80	51	4	3768	0.233	0.5195 ±.10%	13.308 ±.13%	0.18579 ±.05%	2697	2705.3 ±1.7
5. Z5c(1),nm0°,fg,eq,cl,co	6	120	77	5	3991	0.237	0.5191 ±.09%	13.275 ±.12%	0.18547 ±.05%	2695	2702.4 ±1.5
6. Z2c(6),nm0°,eu,pr,cl,co	3	78	52	5	1518	0.301	0.5178 ±.14%	13.215 ±.16%	0.18509 ±.06%	2690	2699.1 ±2.1
7. Z1b(4),nm1°,eu,pr,incl,co	32	69	47	105	694	0.338	0.5172 ±.22%	13.198 ±.22%	0.18508 ±.11%	2687	2698.9 ±3.5
8. Z4(2),nm0°,fg,spr,cl,co,	8	99	66	3	7257	0.313	0.5172 ±.10%	13.174 ±.12%	0.18473 ±.04%	2688	2695.8 ±1.4
9. Z2b(4),nm0°,eu,pr,cl,co	18	139	92	9	7688	0.331	0.5053 ±.09%	12.659 ±.11%	0.18172 ±.04%	2636	2668.6 ±1.5
98TXM352 (Z5949): Garnet-quartz-feldspathic gneiss (sample location: 63°51.316'N, 92°41.311'W)											
10. Z3b(1),dia,eq,sph,cl,co	10	69	38	5	4264	0.12	0.4913 ±.09%	11.699 ±.10%	0.17270 ±.04%	2576	2584.0 ±1.8
11. Z5d(1),dia,fg,og,fr,cl,co	8	78	44	5	3861	0.172	0.4896 ±.10%	11.619 ±.12%	0.17212 ±.05%	2569	2578.4 ±1.5
12. Z5e(1),dia,fg,og,cl,co	5	41	24	3	2179	0.203	0.4858 ±.16%	11.499 ±.10%	0.17168 ±.10%	2552	2574 ±3.3
13. Z3a(1),dia,eq,sph,cl,co	6	88	51	16	966	0.249	0.4773 ±.13%	10.990 ±.15%	0.16698 ±.08%	2516	2527.6 ±3.0
14. Z5a(1),dia,fg,og,cl,co	9	69	39	27	608	0.232	0.4742 ±.12%	10.915 ±.19%	0.16694 ±.14%	2502	2527.2 ±4.8
15. Z4a(5),dia,eq,sph,cl,co	17	74	42	3	13129	0.233	0.4730 ±.09%	10.799 ±.12%	0.16557 ±.04%	2497	2513.4 ±1.4
16. Z5b(1),dia,fg,og,cl,co	9	69	38	5	3396	0.181	0.4701 ±.09%	10.653 ±.10%	0.16436 ±.04%	2484	2501.1 ±1.6
17. Z5c(1),dia,fg,og,fr,cl,co	18	36	20	4	4162	0.211	0.4599 ±.10%	10.367 ±.12%	0.16351 ±.05%	2439	2492.3 ±1.7
18. M2b(1),1.5A,eq,py,cl	4	3905	5611	4	104941	2.185	0.4945 ±.09%	11.934 ±.11%	0.17503 ±.04%	2590	2606.4 ±1.4
19. M5(1),1.5A,eq,dy,incl,fr	13	332	8179	7	17644	57.97	0.4808 ±.10%	11.272 ±.12%	0.17002 ±.05%	2531	2557.9 ±1.6
20. M2a(1),1.5A,eq,py,cl	2	227	6278	3	4216	66.34	0.4722 ±.16%	10.965 ±.17%	0.16842 ±.05%	2493	2542.0 ±1.8
21. M1(1),1.5A,irr,py,cl	15	296	5829	8	13736	54.64	0.4068 ±.10%	8.184 ±.12%	0.14590 ±.05%	2200	2298.4 ±1.6
98TXM298b (Z5824): Quartz-ribboned tonalitic mylonite (sample location: 63°51.624'N, 92°41.166'W)											
22. Z2a(1),dia,eq,sph,cl,co	11	83	46	4	6234	0.155	0.4863 ±.09%	11.421 ±.12%	0.17032 ±.04%	2555	2560.8 ±1.4
23. Z3b(1),dia,eq,sph,cl,co	6	26	15	12	382	0.27	0.4824 ±.32%	11.300 ±.30%	0.16987 ±.18%	2538	2556.4 ±6.2
24. Z1b(1),dia,spr,cl,co	3	25	15	3	770	0.253	0.4850 ±.28%	11.301 ±.73%	0.16898 ±.58%	2549	2547.6 ±19.5
25. Z4b(1),dia,eq,sph,cl,co	3	49	28	34	150	0.205	0.4784 ±.89%	11.122 ±.83%	0.16862 ±.50%	2520	2544.0 ±16.8
26. Z3a(1),dia,eq,sph,cl,co	8	54	29	5	2504	0.145	0.4789 ±.10%	11.053 ±.13%	0.16740 ±.05%	2522	2531.8 ±1.8
27. Z1a(1),dia,spr,cl,co	7	58	33	19	677	0.214	0.4763 ±.19%	10.939 ±.23%	0.16656 ±.15%	2511	2523.4 ±5.1

28. R1a(40)0.5A,fg,or,cl	44	0.61	0.16	38	28	-0.058	0.2939 ±2.7%	4.102 ±9.6%	0.10123 ±8.2%	1661	1646.8 +277.7/-341.2
29. R2a(35)0.5A,fg,bk,inc	198	0.15	0.04	61	23	-0.103	0.2786 ±3.7%	3.820 ±17.3%	0.09945 ±15.1%	1584	1613.7 +478.1/-704.7
30. R1b(40)0.5A,fg,rd,cl	79	0.65	0.16	89	25	-0.065	0.2847 ±3.1%	3.471 ±14.6%	0.08842 ±12.7%	1615	1391.7 +422.6/-587.2

^aMineral analyzed: Z, zircon, followed by morphology type. The number in parenthesis indicates the number of grains included in the fraction.

Magnetic properties on FrantzTM: dia, diamagnetic at 1.8A and -0.5 side slope; nm0*, nonmagnetic at indicated side slope in degrees.

Grain characteristics: eu, euhedral; lr, irregular; fg, fragment; fr, fractured; og, overgrowth; cl, clear; tur, turbid; incl, inclusions; co, colourless; py, pale yellow; dy, dark yellow; bk, black; rd, red; or, orange; eq, equant; pr, prismatic (>3:1); spr, prismatic (<3:1); sph, spherical.

^bRadioactive Pb.

^cTotal common Pb in analysis corrected for spike and fractionation.

^dZircon fractions not weighed; weight estimated from grain size, therefore U and Pb concentration data are qualitative only. Concentration uncertainty varies with sample weight: >10% for sample weights <10 µg, and <10% for sample weights above 10 µg.

Population	Fraction I.D.	Description
1	Z1 (Fractions 1, 2, and 7)	Large (~200-400 μm), colourless, euhedral, elongate prisms, with aspect ratios >3:1, extracted from the magnetic fraction at maximum magnetic current (1.7Å) and a side slope of 1°. Most grains lack visible cores, exhibit fractures and may contain colourless inclusions. Rare turbid grains and grains with inclusions were avoided for analysis.
1	Z2 (Fractions 2, 6, and 9)	Large (~300-400 μm), colourless, euhedral, elongate prisms, with aspect ratios >3:1, and no visible cores. These were extracted from the magnetic fraction at 1.7Å and side slopes of 0° and -½°.
2	Z3	Clear, colourless, small prisms (<150 μm), void of cores, approaching acicular habit, with generally sub-rounded tips. Extracted from the magnetic fraction at 1.7Å and side slopes of 0° and -½°.
1	Z4 (Fraction 8)	Clear, colourless, ~200-300 μm fragments of euhedral prisms, without visible cores, extracted from the magnetic separate at 1.7Å and 0° side slope.
3	Z5 (Fractions 4 and 5)	Clear, colourless, rounded, multi-faceted, equant grains with irregular surfaces, variable in size (~100-300 μm in diameter), from the magnetic separate at 1.7Å and 0°. Fractures and inclusions are rare and no visible cores were detected.
4	Z6	“Cigar”-shaped, or prolate grains with rounded edges and tips, and no cores, extracted from the 1.7Å magnetic fraction at 0° or -½° side slope.
5	Z7	Clear, colourless, subrounded, almost equant (aspect ratio~2:1), subhedral prisms (~100-300 μm long). Grains generally contain no fractures, cores, overgrowths or inclusions, and were extracted from the magnetic fraction at 1.7Å and 0° or -½° side slope.

Note: Bold labels indicate analysed fractions.

Table 5.2: Description of zircon populations from leucogabbro mylonite (sample 98TXM334b).

Popu- lation	Fraction I.D.	Description
1	Z1	Elongate, sub- to euhedral prisms. Crystals are colourless, but generally turbid with abundant fractures and inclusions common. No cores or overgrowths noted.
2	Z2	Large (~200 µm), nearly spherical grains with irregular or pitted surfaces. Most grains are clear, colourless, lack fractures and inclusions.
2	Z3	Large (~120 µm), nearly spherical grains with either smooth or pitted surfaces. Most grains are clear, colourless, and lack inclusions and fractures.
2	Z4	Small (<100 µm), round to slightly flattened grains with smooth surfaces. Grains are clear, colourless and lack inclusions or fractures.
3	Z5	Large (~200 µm) fragments thought to represent metamorphic overgrowths which have been separated from their cores. Clear, colourless, irregular in shape but with a concave surface that may form the inside of the overgrowth.
4	Z6	Spherical, clear and colourless zircons with a discernable, but incomplete, metamorphic overgrowth. No fractures or inclusions.
5	Z7	Clear and colourless, ovoid or egg-shaped, with no fractures or inclusions.
1	M1	Large (~200-300 µm), flat or platy, angular, pale yellow grains with irregular surfaces. Minor fractures and no inclusions or overgrowths.
2	M2	Small (~100 µm), slightly flattened, round, pale yellow grains with smooth surfaces. No inclusions or fractures.
3	M3	Deep yellow, nearly spherical grains with smooth surfaces and minor tiny opaque inclusions in some grains.
4	M4	Large (~300 µm), angular, pale yellow fragments. Minor fractures and rare surficial inclusions.
5	M5	Large (~300 µm), angular, pale yellow fragments. Minor fractures and rare surficial inclusions.

Note: Bold labels indicate analysed fractions. All zircons were extracted from the diamagnetic fraction, monazite from the magnetic fraction at 1.5 Å.

Table 5.3 Description of zircon populations from garnet-quartzofeldspathic gneiss (sample 98TXM352).

Popu- lation	Fraction I.D.	Description
1	Z1	Subhedral, clear, stubby prisms (aspect ratio ~2:1), averaging 235 μm . No fractures or inclusions, but rare apophyses, possibly the remains of overgrowths.
2	Z2	Large (~260 μm), clear, spheres with probable thin metamorphic overgrowths. No fractures or inclusions.
2	Z3	Clear, colourless, spheres, averaging ~150 μm . No notable fractures, inclusions or overgrowths.
2	Z4	Small (typically <100 μm), clear and colourless, flattened spheres with smooth surfaces. No fractures, inclusions or overgrowths.
3	Z5	Clear, colourless "sheets", "plates" or "wafers". Generally >300 μm in two dimensions and flat (~50 μm thickness) with no fractures or inclusions.
4	Z6	Possible metamorphic overgrowths with some "core" still attached. Not analysed for geochronology.
1	R1	Very tiny (<100 μm on average), orange, angular, vitreous fragments, extracted from 0.5Å magnetic separate. Opaque inclusions (ilmenite intergrowths?) common but avoided as much as possible in fraction for analysis. No fractures or overgrowths.
2	R2	Red to rust, vitreous, ~150 μm angular fragments, extracted from 0.5Å magnetic separate. Ilmenite intergrowths either less common or less discernable. No fractures or overgrowths.
3	R3	Large (150-400 μm), black, vitreous, angular fragments, extracted from 0.5Å magnetic separate. Many resemble polycrystalline clusters but these are avoided in the selection process.

Note: Bold labels indicate analysed fractions. All zircons were extracted from the diamagnetic Frantz™ separate.

Table 5.4: Description of zircon populations from tonalitic mylonite (sample 98TXM298b).

Table 5.5: Summary of discordia lines discussed in text.

Chord #	Upper Intercept (Ma)	Lower Intercept (Ma)	Fractions used	MSWD	Forced Y-intercept
98TXM334b: Leuco-metagabbro mylonite					
1	2711 ± 4	1933 ± 86	4, 5, 6, 7, 8, 9	1.44	No
98TXM352: Garnet quartzofeldspathic gneiss					
1	2620 + 22/-17	2319 + 323/-46	10, 11, 13, 15, 16	0.42	No
2	2595 + 9/-6	1909 + 191/-188	10, 11, 12,	1.71	To ~1900 Ma
3	2505 ± 3	800 ± 168	16, 17	N/A	No
4	2530 ± 7	1899 ± 39	16	N/A	To ~1900 Ma
5	2629 ± 5	1921 ± 13	18, 20, 21	0.78	No
98TXM298b: Tonalitic mylonite					
1	2593 + 71/-24	2324 + 131/-171	22, 23, 24, 25, 26, 27	0.78	No

Chapter 6: Tectonometamorphic history of the Uvauk complex

Table:

Table 6.1: Summary of tectonometamorphic events in the UC and surrounding area

Chapter 6: Tectonometamorphic history of the Uvauk complex

6.1 Summary of conclusions

This multidisciplinary study involved field mapping of three detailed areas within the UC (Map areas A, B, and C of Figure 1.3; Appendix I), north of Chesterfield Inlet, Nunavut. Petrographical, thermobarometric, and geochronological studies were carried out on samples from area A (Figure 1.3), the best exposed and main area of interest, to provide insight into the tectonometamorphic history of the UC. First order conclusions are outlined below, followed by a discussion that integrates all of the data.

6.1.1 Lithology

The UC is cored by anorthosite and related rocks herein referred to as the Uvauk anorthosite suite. Rocks of the anorthosite suite define a ~30 km long, east-striking belt that thickens in map view from ~1 km in the west to nearly 20 km in the east. The suite comprises ~60% anorthosite, gabbroic anorthosite, anorthositic gabbro and leucogabbro, and ~40% fine- to medium grained, mafic (gabbro) dykes. The anorthosite characteristically contains post-tectonic pseudomorphs comprising radially and randomly oriented hornblende $\pm$ actinolite $\pm$ biotite. These amphibolite-facies aggregates, interpreted to replace pyroxene (section 2.1.1), predominate in the anorthosite wherein pre-tectonic, granulite-facies assemblages are only rarely preserved. The granulite-facies assemblage, clinopyroxene-orthopyroxene-garnet-plagioclase, is common within the gabbro dykes, although amphibolite-facies overprinting is locally evident.

Dominantly metaplutonic rocks outcrop north and south of the anorthosite suite. Diorite gneiss, garnet-quartzofeldspathic gneiss, and amphibolite are the main rock types immediately south of the anorthosite suite, within the Southern bounding shear zone (see

Map A, Appendix I and Figure 3.1). Granulite-facies tonalite straight gneiss and amphibolite-facies orthogneiss, with subordinate semipelitic schist occur south of the Southern bounding shear zone. North of the anorthosite suite, amphibolite-facies tonalite gneiss and minor amphibolite also show no evidence for granulite-facies metamorphism. Granulite-facies assemblages are evident in enderbite (orthopyroxene-bearing tonalite orthogneiss) and mafic granulite (clinopyroxene $\pm$ garnet $\pm$ plagioclase) in the northernmost map unit (Figures 2.9, 3.11).

6.1.2 Structure

The main foliation developed throughout the map area (Map A of Appendix I) strikes roughly ENE to ESE and dips steeply to the south (see Figures 3.1 and 3.2 for details). Foliation development is variable, and the main foliation (S_x) is interpreted to overprint at least one older foliation (S_{x-1} ; refer to section 3.3.1; Figure 3.5c).

Two steeply south-dipping shear zones were delineated in the main field area (Figure 3.1), and are herein designated the Northern and Southern bounding shear zones owing to their occurrence along the margins of the anorthosite suite (Map A). The Northern bounding shear zone is characterized by moderately southeast-dipping foliations, moderately east-plunging lineations and rare sinistral shear-sense indicators, consistent with oblique, north side-up, normal faulting with a sinistral strike-slip component. The secondary mineral assemblages displayed are hydrous (hornblende-bearing), but were formed within the granulite facies (see Chapter 4). The Southern bounding shear zone is characterized by steeply south-dipping foliations, shallowly east-plunging lineations, and common dextral shear-sense indicators, consistent with oblique, dextral strike-slip faulting with a south side-up component. Early granulite-facies

assemblages are commonly preserved and incipient amphibolite-facies overprinting is evident by randomly oriented biotite laths and microlaths in the quartzofeldspathic gneiss and by a thin veneer of actinolite rimming and filling fractures in pyroxene of the diorite gneiss (Figure 3.19). The timing of shear zone development is discussed below (see section 6.2.2).

Overprinting relationships are unclear, and there are several possible relationships between the main foliation, S_x , and the mylonite foliation (S_m), which is best developed within the shear zones (discussed below). Coeval development of S_x and S_m is possible and may have resulted from strain partitioning. Alternatively, S_m may significantly postdate S_x , and the nearly parallel average orientations of the two foliations may be fortuitous. The two shear zones may also have formed independently during two temporally distinct events, consistent with the difference in kinematics and metamorphic grade displayed by the two shear zones

The central domain and the bounding shear zones exhibit tight to isoclinal folding of compositional banding and the layer-parallel foliation. Axial planes are parallel to the mylonite foliation, S_m (Figures 3.7, 3.20, and 3.21), which is consistent with F_m fold development before or during mylonitization. Sheath folds, or folds with a tubular shape parallel to the mylonite lineation, also occur in a highly strained part of the central domain and these are commonly produced as a result of local distortions in the flow field during mylonitization (Passchier and Trouw, 1998). However, the minor folds may also reflect a major fold structure in the area. Hornblende alignment and flattening of hornblende + plagioclase garnet pseudomorphs parallel to the axial planes of folds (Figure 3.21) indicate that retrograde metamorphism accompanied mylonitization in the

Northern bounding shear zone. Foliation orientations, the distribution and orientation of minor folds and repetition of lithologic units (section 3.8) in the westernmost part of the UC (Map A, Appendix I) are consistent with folding of the anorthositic suite into a tight, moderately east-plunging synform (see also Tella et al. 1994a and b; 2000b). However, it is not clear whether the bounding shear zones represent one continuous, folded shear zone, or if folding is restricted to the central domain and Northern bounding shear zone, with the older, Southern bounding shear zone behaving as a stiff buttress. Map unit geometry and foliation orientations within the easternmost part of the UC (Map C, Appendix I) are consistent with an open, gently NW-plunging synform. Outcrop-scale E-W trending, open and upright minor folds as well as north-trending, open and upright folds are common in Map Area C, and interference patterns are locally present (see also Tella et al., 1993).

6.1.3 Metamorphism and thermobarometry

Based on textural observations, chemical zoning, and P-T calculations, three metamorphic events (M1, M2, M3) are interpreted to have affected rocks of the UC and surrounding area (Figure 4.26). M1 occurred at upper amphibolite- to granulite-facies conditions and is preserved in rocks of the Southern bounding shear zone, in amphibolites north and south of the anorthosite suite, and possibly in one of the two garnet-anorthosite samples (98TXM561b) from the Northern bounding shear zone. M1 is characterized by a pre-tectonic mineral assemblage comprising low-CaO (ranging from $X_{\text{Gr}} = 0.05$ -0.07 in sample 98TXM178d to $X_{\text{Gr}} = 0.26$ -0.30 in sample 99TXM561b), subhedral garnet porphyroblasts that are wrapped by a foliation defined by the preferred alignment of biotite $\pm$ hornblende (e.g. Figures 4.12 and 4.24). P-T conditions for M1 range between

7-9 kbar and 705-770°C without systematic spatial variation in pressure or temperature. The current data set does not help to constrain the amount of exhumation associated with this M1 event.

M2 occurred under granulite-facies conditions. It is characterized by syn- to post-tectonic garnet growth that, in rocks also containing M1 garnet, is manifested as high-CaO garnet rims (ranging from $X_{\text{Gr}} = 0.08-0.11$ in sample 98TXM178d from the Southern bounding shear zone to $X_{\text{Gr}} = 0.35-0.49$ in sample 99TXM561b from the Northern bounding shear zone) that are variable in thickness (~50-100 μm in sample 98TXM178d vs. ~500-750 μm in sample 99TXM561b; refer to Figures 4.2 and 4.12).

The gabbro dykes are interpreted to have undergone a single-stage, M2 history. They have 0.4-1.0 mm garnet porphyroblasts that exhibit either flat ($X_{\text{Gr}} = 0.18$; 98TXM333) CaO profiles, or a continual decrease from core ($X_{\text{Gr}} = 0.23$) to rim ($X_{\text{Gr}} = 0.18$; 98TXM342), and contain relict orthopyroxene mantled by clinopyroxene. P-T conditions determined from the dykes are similar to those determined from rocks containing garnet porphyroblasts with high-CaO M2 rims (e.g. 12.6 kbar and 865°C for 98TXM342 vs. 12.2 kbar and 840°C for 98TXM178d).

Peak P-T conditions determined for M2 for rocks within the UC range between 11.4-14.3 kbar and 720-865°C, followed by decompression (M2b) to 10.0-10.8 kbar under sustained high temperatures (785-800°C). Similar conditions (~11 kbar, 690-755 °C) are recorded by surrounding rocks. The highest M2 temperatures (865°C) occur within the Uvauk anorthosite suite, with lower temperatures occurring to the north (755°C), and the lowest M2 temperatures occurring in the southernmost sample

containing diagnostic M2 textures (~3 km south of the anorthosite suite; 695°C). No systematic spatial change in pressure is evident (Figure 4.26).

Garnet breakdown textures (symplectites, coronitic structures) and thermobarometry indicative of decompression at high temperatures may have either marked the culmination of the high-P M2 event or been produced during a later event. The first interpretation is favoured because geochronological data summarized below do not indicate an event between the high-P M2 and the low-P M3 (described below). In accordance with this interpretation, clockwise M2 P-T paths are described by all samples for which garnet breakdown calculations were made (e.g. 98TXM201, 98TXM304, 98TXM561b).

M3 effects are more pervasive in rocks south of the anorthosite suite. Based on the garnet-pseudomorph texture exhibited by an amphibolite sample (98TXM339a: Figure 4.18), M3 is interpreted to be post-tectonic. M3 represents a retrograde event at middle to lower amphibolite facies (5.8-6.0 kbar and 625-695°C).

Cryptic faults may be required to account for samples, both north and south of the anorthosite suite, that show evidence for M1, but lack evidence for M2 metamorphism (e.g. 98TXM320, 98TXM339a and b). Alternatively, retrograde M3 features, such as the replacement of garnet in 98TXM339a, suggest that high CaO garnet characteristic of M2, may have been irradiated by processes such as garnet-consuming reactions (in 98TXM339a) or fluid-enhanced re-equilibration of garnet zoning (in 98TXM339b). Finally, it is possible that the mineralogies of one or more of these samples were not fertile for garnet growth at the time of M2 metamorphism.

Geochronological evidence for Archean anorthosite (discussed below) and thermobarometric studies presented herein suggest that the emplacement of anorthositic magma was diachronous within the UC. While one of the three thermobarometric samples from the anorthosite suite appears to have sustained Archean M1 and Proterozoic M2 events (99TXM561b), the other two show evidence for only M2 metamorphism (98TXM304 and 99TXM564a). Zoning profiles and thermobarometric analysis of leucogabbro to anorthositic gabbro (99TXM564a) from the same locality as the ca. 2711 Ma leucogabbro (93TXM334b; Figure 4.1 and 5.1) provides evidence for M2 metamorphism only. Possibly, the Mg-rich phases were not conducive to garnet growth during M1 (section 4.5). Zoning profiles and thermobarometric results of garnet-bearing gabbro anorthosite (98TXM304) from within the Northern bounding shear zone are also consistent with a solely Proterozoic (M2) metamorphic history, although the more Fe-rich minerals (relative to 99TXM564a) suggest that its mineral chemistry did not inhibit garnet growth in this rock. It is thus likely that sample 98TXM304 did not sustain Archean (M1) metamorphism, suggesting that the rock may be Proterozoic in age. In contrast, a garnet-bearing leucogabbro (99TXM561b) from farther east in the Northern bounding shear zone contains a large garnet porphyroblast (~2 cm) with a low-CaO core ($X_{\text{Gr}} < 0.30$) relative to its high-CaO rims ($X_{\text{Gr}} < 0.35\text{-}0.49$). Thermobarometric calculations based on the garnet core composition and inclusions within the garnet core yield 11.3 kbar and 760°C and are interpreted to reflect M1 conditions. This is the only anorthosite sample analysed that appears to have undergone Archean M1 metamorphism. It is likely, then, that two temporally distinct components of anorthosite are present within the UC, although they may not be discerned in the field. In addition, the gabbro

dykes are also interpreted to have undergone M2 metamorphism only. They may also represent a younger (Proterozoic) magmatic component of the UC, consistent with preliminary geochemical data discussed below.

6.1.4 U-Pb geochronology

Geochronological analyses of three samples collected from the UC provides evidence for at least a component of Archean anorthosite within the UC, and indicates that the UC has endured a polymetamorphic history. A magmatic age of $2711 \pm 5/-4$ Ma (U-Pb, zircon) was determined for leucogabbro thought to represent a mafic phase of the UC anorthosite suite (see Figure 5.1 for sample locations), but does not preclude a Paleoproterozoic magmatic component within the UC.

Based on results of this study, the age of the Southern bounding shear zone cannot be clearly defined. However, results are consistent with multiple episodes of deformation and metamorphism as has been documented for other parts of the WCP. Two samples were collected from the Southern bounding shear zone for geochronological analyses, the quartzofeldspathic gneiss (98TXM352) and the tonalite mylonite (98TXM298b). Thermal events are interpreted to have occurred at ca. 2.62-2.60 Ga, ca. 2.53-2.50 Ga, and ca. 1.94-1.90 Ga (Chapter 5). The earliest event is reflected by the upper intercept of monazite data from quartzofeldspathic gneiss (98TXM352) and by the upper intercepts of zircon data from quartzofeldspathic gneiss (98TXM352) and tonalite mylonite (98TXM298b) from the Southern bounding shear zone. A thermal event at ca. 2.53-2.50 Ga is indicated by the upper intercept of a discordia chord through two discordant metamorphic zircon overgrowths (Fraction #'s 16 and 17, sample 98TXM352; Table 5.1; Figure 5.11). Evidence for a thermal disturbance at ca. 1.94-1.90 Ga comes from lower

intercepts of discordia chords through monazite data from the quartzofeldspathic gneiss (98TXM352; 1921 ± 13 Ma), through zircon data from the leucogabbro (98TXM334b; 1933 ± 86 Ma) and from regressions of zircon data from the quartzofeldspathic gneiss whose lower intercepts were forced to ca. 1900 Ma, and yielded low MSWD's (see Table 5.1). Rutile data indicate that rocks of the Southern bounding shear zone cooled through 400°C at ~ 1500 Ma (see section 5.5.3). Based on these results, the Southern bounding shear zone is herein interpreted to be an Archean structure that has undergone Paleoproterozoic reworking, consistent with the interpretation of Tella et al. (1994b). The implications of this interpretation are discussed below.

6.1.5 Geochemistry of the UC anorthosite suite

Although a thorough geochemical investigation was not completed for this study, whole rock geochemistry was acquired for 20 samples from the westernmost part of the UC, and 19 of those are rocks of the anorthositic suite (Appendix III). The gabbro dykes of the UC typically exhibit *e*-MORB-like, or flat to very slightly light rare earth element (LREE)-enriched, chondrite normalized REE patterns (Sun and McDonough, 1989). Three of the five dykes fall within the field of the Paleoproterozoic gabbros of the KC. This petrochemical similarity is consistent with a common origin and age for gabbroic rocks within both Aqxaarq complexes. In contrast, the anorthosites and gabbroic anorthosites of the UC exhibit lower abundances of the REE and exhibit greater positive Eu anomalies relative to similar rock types from the KC (Appendix III). It is unclear whether these differences reflect differences in source characteristics, possibly resulting from different magmatic ages, or whether they reflect greater fractionation from a more depleted source at the UC relative to the KC.

6.2 Discussion

6.2.1 *Origin of the UC anorthosite suite*

An allochthonous origin has been proposed for the UC (Tella et al., 1994b; 2000b), but no fault imbrication or cross-cutting relationships were discerned in the field to support this suggestion. Intense deformation at the margins of the anorthosite suite renders field observations equivocal. However, the presence of anorthositic veins within the surrounding country rock and xenoliths of local origin within the anorthosite are consistent with a parautochthonous origin for at least the youngest component of the UC anorthosite suite. Magmatic emplacement of Proterozoic anorthosite was probably accompanied by deformation that was heterogeneously partitioned in the interior and most intense along the margins of the intrusion and in the adjacent Archean gneisses. An allochthonous relationship between Archean anorthosite and Archean tonalitic country rocks, however, can not be precluded.

6.2.2 *Tectonothermal history of the UC*

The oldest thermal event indicated by U-Pb systematics of zircon and monazite from quartzofeldspathic gneiss of the Southern bounding shear zone (98TXM352) occurred at about ca. 2.62-2.60 Ga. P-T conditions for this event are unknown. It likely corresponds to the oldest regional fabric, which Ryan et al. (1999) describe in the MacQuoid-Gibson lakes area as a layer-parallel S_1 foliation that is poorly bracketed in age between ca. 2.68, the age of a rhyodacite containing an S_1 foliation, and ca. 2.55-2.50 Ga, the age of regional metamorphism and S_2 development.

The orientation of the main foliation in tonalitic gneisses south of the UC (S_x) is consistent with the S_2 foliation developed in rocks south of Akunak Bay, Chesterfield

Inlet (see Tella et al., 2000b), and is thought to be Neoarchean in age (ca. 2.55-2.50 Ga; Berman et al., 2000). S_2 is a composite foliation (S_1/S_2) developed by the transposition of layering and a layer-parallel foliation (S_1 ; Ryan et al., 1999) and wraps the type I, low-CaO garnet porphyroblasts (comparable to M1 garnet at the UC) described by Berman et al. (2000) at Parker and MacQuoid lakes. Monazite inclusions within type I garnet give ages between 2490-2560 Ma, and the subhedral to euhedral morphology is interpreted to indicate that type 1 garnet growth (M1 metamorphism) occurred synchronous with S_2 development (Berman et al., 2000). Correlation of S_x (within the UC) with S_2 (regional) is consistent with the interpretation of a ca. 2.53-2.50 Ga tectonothermal event (based on U-Pb geochronological data; this study; see section 5.5.2) and is compatible with coeval development of S_m (in the Southern bounding shear zone) and S_x (in the Southern domain). Granulite-facies mylonitization and synchronous mafic-anorthosite magmatism in the Big Lake shear zone have been dated at ca. 2.50-2.49 Ga (Ryan et al., 2000), and the Southern bounding shear zone at UC may have similarly developed during late-stage D2 deformation. This regional event took place at 7-9 kbar and 705-770°C in the UC and at 5-6 kbar and 630-690°C at MacQuoid and Parker lakes (Berman et al., 2000; Ryan et al., 2000), indicating greater structural depth of the former during Neoarchean times. Speculation that the S_2 regional foliation and Big Lake shear zone originated as a shallowly dipping structure subsequently reoriented more steeply by two regional folding events (F3 and F4; Ryan et al., 2000b; 2000c) may also apply to the Southern bounding shear zone at the UC.

S_m in the Northern bounding shear zone is interpreted to have developed at ca. 1.9 Ga, for the following reasons. Previous geochronological studies of rocks from the

westernmost part of the UC, considered to be part of the Northern bounding shear zone, yielded a U-Pb (zircon) age of ca. 1936 Ma, interpreted to reflect the timing of deformation and granulite-facies metamorphism (Roddick and Tella, unpublished data, 1993; Tella et al., 1994b), herein interpreted to have occurred at 10.8-14.3 kbar and 695-865°C in the UC. U-Pb (titanite) data from the same sample indicate cooling through 660-700° (T_c of titanite, Scott and St. Onge, 1995) at ca. 1903 Ma. Based on data collected in the present study, ca. 1936 is correlated to M2 metamorphism and coeval mylonitization in the Northern bounding shear zone and central domain. Emplacement of the syntectonic E-trending Uvauk dykes (e.g. samples 98TXM333 and 98TXM342; see section 3.7) that are spatially associated with the anorthosite suite is interpreted to have been coeval with ca. 1936 Ma granulite-facies mylonitization. The dykes and their magmatic source may have been the heat source for M2 metamorphism, consistent with the documented M2 temperature decrease to the north and south of the Uvauk anorthosite suite. Ca. 1936 Ma anorthosite magmatism probably accompanied the granulite-facies mylonitization and mafic magmatism (see section 6.2.1, above).

The ca. 1936 Ma M2 event appears to coincide with regional F3 folding. Within the northern Hearne domain, east-northeast-trending F3 folds are poorly bracketed between 2.19 Ga, the age of folded diabase dykes, and 1.83 Ga, the age of granite veins unaffected by F3 folding (Ryan et al., 2000b). A link between these structures and the ca. 1.9 Ga high-pressure metamorphism described by Berman et al (2000) seems logical but has not yet been clearly demonstrated. The link between S_m in the Northern bounding shear zone and M2 (in the UC) supports a ca. 1.9 Ga age for the regional D3. D3 involved widespread fold development, with typically little to no new penetrative fabric

(Ryan et al., 2000b) and D3 mylonitization within the UC is probably localized within the Northern bounding shear zone.

Folding within the Northern bounding shear zone, and of rocks in the UC, is interpreted to be coeval with ca. 1936 Ma mylonitization, based on parallelism between the axial planes of folds and the mylonite foliation, S_m (Figures 3.7, 3.20, and 3.21), which is consistent with F_m fold development during mylonitization (see section 3.6.3 and 6.1.2). This is consistent with the interpretation that the UC represents a tight, shallowly ENE-plunging map-scale F_3 synform (Tella et al., 2000b). It is therefore possible that the Northern and Southern bounding shear zones may be part of the same structure that has been repeated through folding so that the shear zone outcrops along the north and south limbs of an east-plunging synform and is continuous beneath the anorthosite suite. The presence of a marginal phase on the southern and northern margins of the anorthosite suite, and of diorite gneiss north and south of the suite, are also consistent with a fold geometry. Shallowly west-dipping anorthositic rocks in the eastern part of the complex (Appendix I) further indicate the considerable modification of foliation and rock unit orientations by Proterozoic folding.

Although it is possible that the Southern bounding shear zone is part of the same structure as the Northern bounding shear zone, there are no direct observations to prove that the Southern shear zone has undergone deformation at ca. 1.9 Ga. Thermal effects are evident in the modest disturbance of U-Pb systematics in the Southern bounding shear zone, and this may be due to the focusing of ca. 1.9 Ga deformation within the anorthosite suite. The intense deformation fabrics exhibited by some anorthosite rocks are consistent with synkinematic emplacement, since deformation would likely have

partitioned into the surrounding quartz-bearing rocks if the anorthosite had been cold and rheologically stiff.

The clockwise M2 P-T path determined for some samples (e.g. 98TXM304, 98TXM201, 99TXM561b) suggests that M2 involved at least ~10-15 km of crustal thickening as implied by the 3-5 kbar pressure increase from burial depths indicated by M1 conditions (the extent of M1 exhumation is unconstrained) to those indicated by M2 conditions. Subsequent decompression to ~10 kbar (30 km depths) marking the final stages of M2/D3 at ca. 1903 Ma (U-Pb, titanite), may have resulted from orogenic collapse as a result of gravitational instability.

M3 metamorphism has mainly affected rocks south of the UC, occurred at P-T conditions of ~6 kbar and 625-695°C, and is characterized by garnet replacement textures with no associated tectonic fabric (Figure 4.18). This retrograde metamorphism may be related to widespread granite plutonism at ca. 1.83 Ga (Peterson et al., 2000), as the nearby Promise Point granite (south of the UC) is thought to be a part of this suite (see Figure 1.3 and Tella et al., 2000 and 2000b). Rocks of the UC cooled through 400°C by ca. 1.78-1.83 Ga (U-Pb, rutile; Roddick and Tella, 1993, unpublished data; Table 6.1). The large error (± 300 Ma and greater; Table 5.1) associated with U-Pb (rutile) ages of ca. 1650 Ma, determined for tonalite mylonite from the Southern bounding shear zone (sample 98TXM298b), allow the younger ages determined in this study to possibly correlate to the older rutile ages determined by Roddick and Tella (1993, unpublished data). Post-tectonic, greenschist-facies retrogression is also indicated by ca. 1.75 Ga anhedral monazite in chlorite-filled fractures of rocks from Parker Lake (Stern and Berman, 2001) and by titanite from a 1.83 Ga granite that crosscuts the Big lake shear

zone (Davis et al., 2000, unpublished data; Tella et al., 2001). The results are compatible with widespread greenschist-facies hydrothermal activity related to the ca. 1.75 Ga Nuelin suite granites (Peterson et al., 2000).

A significant, but unresolved issue deals with the timing of crustal thickening in the northern Hearne domain. It has not yet been established if thickening occurred at ca. 1.9 Ga, or whether the thickening occurred at ca. 2.5 Ga with subsequent thermal activity at ca. 1.9 Ga causing regional metamorphism (see also Berman et al., 2000; Ryan et al., 2000). High-pressure metamorphism of ca. 2.19 Ga MacQuoid dykes (Ryan et al., 2000 and 2000c) is consistent with a high-pressure event at ca. 1.9 Ga. The Uvauk gabbro dykes are interpreted to have been synkinematically emplaced at ca. 1.94 Ga and these give the same high pressures as the high-CaO garnet rims in other rocks. Thickening at ca. 1.9 Ga is the preferred interpretation, based primarily on the clockwise P-T paths calculated for M2 metamorphism in several samples, which are consistent with crustal thickening and exhumation at that time (England and Thompson, 1984). The lack of observations to support higher pressures associated with the M1 event is also compatible with thickening at ca. 1.9 Ga.

6.2.3 Implications of UC history for interpretation of the Aqxaarq gneisses

The UC is thought to be related to other granulite-facies, anorthositic complexes that outcrop along Chesterfield Inlet ("Aqxaarq gneisses"; Tella and Annesley, 1988). These were thought to be Archean in age and tectonically emplaced to their present positions during the Proterozoic (Shau and Tella, 1993). A detailed investigation of one of these complexes, the KC (Sanborn-Barrie et al., 1997; Sanborn-Barrie, 1999), documents Archean deformation at ca. 2.55 Ga and Paleoproterozoic (ca. 1910 Ma)

granulite-facies metamorphism followed by emplacement of voluminous anorthositic magma at ca. 1902 Ma. Similarly, in the UC there is evidence for both an Archean (leucogabbro magmatism at ca. 2711 Ma, metamorphism at ca. 2.62-2.60 Ga and at ca. 2.53-2.50 Ga; this study) and a Paleoproterozoic history (mafic $\pm$ anorthosite magmatism at ca. 1.94 Ga). Polymetamorphism at these times may be a characteristic of the Aqxaarneq gneisses.

6.3 Regional Implications

The data documented herein provide evidence for polyphase metamorphism and deformation within the UC. The earliest thermal event is interpreted to have occurred at ca. 2.63-2.60 Ga, with subsequent events occurring at ca. 2.53-2.50 Ga and ca. 1.94-1.91 Ga. The earliest event broadly corresponds to the age of Neoarchean granite plutonism that is widespread throughout the Rae domain and also occurs in the northern Hearne domain (Figure 1.1; Davis et al., 2000a). Although the cause of ca. 2.6 Ga plutonism in the WCP remains poorly constrained, their occurrence in the northern Hearne domain is associated with, and locally constrains the timing of displacement along predominantly discrete strike-slip shear zones (Davis et al., 2000a; Aspler et al., 2000; MacLachlan et al., 2000). Hanmer (1997) documents similar 2.62-2.60 Ga relationships within the Striding-Athabasca segment of the STZ. This study provides evidence for metamorphism within the UC at ca. 2.62-2.60 Ga (see section 5.5.2). This may mark the original age of the Southern bounding shear zone, but there are no observations to directly link geochronological data to fabric development.

The second thermal event evident from the current data set (ca. 2.50-2.53 Ga) corresponds to a recently recognized regional metamorphic and deformation event (D2;

Berman et al., 2000). As is the case for the ca. 2.6 Ga granites, 2.55-2.50 Ga thermotectonism is also best documented in the northern Hearne domain (e.g. Berman et al., 2000; MacLachlan et al., 2000; Ryan et al., 2000), suggesting the possibility of a genetic relationship between magmatism, deformation and metamorphism (Davis et al., 2000a). Results of recent teleseismic studies indicate a significant difference in conductivity properties of the Rae crust relative to Hearne crust (Jones et al., 2000). Isotopic data suggest that Neoarchean granitoids of the southern Hearne domain developed in an intraoceanic setting, whereas some of those from the northern Hearne (rocks of the Cross Bay Complex, in particular) exhibit evidence of having intruded older crustal material (Sandeman et al., 2000). Recent magnetotelluric data has been interpreted to indicate the presence of a south-dipping interface that roughly coincides with Chesterfield Inlet (Jones et al., 2000), just south of the UC. A possible model that ties together these threads of evidence involves Neoarchean juxtaposition of the Rae and Hearne domains, with the latter thrust atop the former resulting in crustal thickening along the Rae-Hearne boundary (Jones et al., 2000; Hanmer et al., 2000; Davis et al., 2000a). In this scenario, initial juxtaposition would have occurred by ca. 2.62 Ga and extensive tectonothermal reworking at 2.55-2.50 Ga would reflect terminal collisional processes focused within the STZ (Davis et al., 2000a). The Rae domain is characterized by widespread ca. 2.6 Ga plutonism relative to the minor component evident in the northern Hearne. This is compatible, however, with initial north-directed subduction. This apparent discrepancy might be explained by subduction reversal from north- to south-directed, producing the south-dipping interface described by Jones et al. (2000).

Rocks of the UC record M1 metamorphic conditions of ~8.4-8.8 kbar and ca. 725-730°C, interpreted to reflect ca. 2.53-2.50 Ga metamorphism. The lack of chemical or thermobarometric evidence for ca. 2.6 Ga metamorphism may be related to thickened crust remaining at depth until exhumation occurred following ca. 2.5 Ga end-stage collision. Evidence for ca. 2.5 Ga granulite-facies mylonitization at the KC and the Big lake shear zone and the UC suggests that these deep crustal remnants may have a shared tectonothermal history. Possibly, they may represent the mid- to lower crustal extensions of Neoarchean faults and these faults may have accommodated significant displacement in Neoarchean times. Ryan et al. (2000b) interpret the orientation of S_2 and the Big lake shear zone to have been originally shallow. Possibly, mylonitization at Big lake, the KC and the UC were all related to ca. 2.5 Ga thickening and metamorphism.

If the Rae and Hearne domains had been juxtaposed by ca. 2.6 Ga, with ca. 2.5 Ga marking terminal collision processes, then ca. 1.9 Ga crustal thickening must have occurred in an intracontinental setting. The interpreted timing of M2 correlates well with the approximate age of the Thelon-Taltson orogen, which marks the timing of collision between the Slave province and the Rae domain of the WCP (ca. 1.97-1.91 Ga). Ca. 1.9 Ga tectonothermal activity within the WCP may reflect the transmission of these far field plate boundary stresses (i.e. Slave-WCP collision) into the upper plate interior (Sanborn-Barrie, 1999; Berman et al., 2000), preferentially reactivating old structures (e.g. possibly the Southern shear zone south of the KC, the Big lake shear zone south of the Cross Bay complex and the Southern bounding shear zone south of the UC). According to this scenario, the bulk of thickening occurred at ca. 2.5 Ga, but collisional forces along the Thelon-Taltson orogen resulted in further thickening at ca. 1.9 Ga.

M3 metamorphism, described above, may be related to a ca. 1.83 Ga magmatic event, consistent with the dominantly post-tectonic manifestation of M3 at mid-amphibolite facies (~5.8-6.0 kbar and 625-695°C). Samples from the Uvauk complex collected for $^{39}\text{Ar}/^{40}\text{Ar}$ dating indicate that cooling through ~530°C (hornblende closure temperature; MacDougall and Harrison, 1988) occurred at ca. 1.8 Ga (Sandeman, 2000, unpublished data), consistent with the above interpretation of ca. 1.83 Ga resetting. Ca. 1780-1830 Ma rutile ages from the anorthosite suite (Roddick and Tella, 1993, unpublished data; see Table 6.1) suggest cooling through ~400°C (rutile closure temperature; Heaman and Parrish, 1991). The 1.85-1.81 Ga Hudson Suite monzonites may be related to terminal collision in the Trans-Hudson orogen or to the indentation of the Slave Province into the WCP (Peterson et al., 2000).

Ca. 1.77-1.75 Ga Nuelin monzonite to granite plutonism occurred *after* much of Laurentia had been assembled (Hoffman, 1989). The anorogenic magmatism may be related to intracontinental rifting (Hoffman, 1989), large-scale mantle upwelling (Hoffman, 1989), or the removal of a thickened continental root (Corrigan and Hanmer, 1997). The ca. 1650 $\pm$ 350 Ma age of rutile (sample 98TXM298b; section 5.5.3) reflects cooling through 400°C possibly due to a thermal pulse related to the regionally extensive Nuelin granites.

6.4 Questions and suggestions for future work

- 1) Does the UC comprise Archean *and* Paleoproterozoic anorthosite rocks? Is the older or younger anorthosite more voluminous? Further geochronology targeting undeformed gabbroic anorthosite and gabbro dykes from the eastern part of the UC and anorthosite dykes within the Southern bounding shear zone may provide insight.

Sm/Nd analyses of the 2711 Ma leucogabbro and other rocks that may or may not be Archean in age should elucidate any inherent source differences and can be used to calculate T_{DM} ages for comparison.

- 2) Are the anorthositic suites and/or mafic magmas of the KC and UC petrogenetically related? Comparative geochemical and isotopic studies may shed light on petrochemical relationships between rocks of the Aqxarneq gneisses.
- 3) What was the effect of the 1.9 Ga thermotectonic event on rocks of the Southern Bounding shear zone (SBSZ)? A monazite lower intercept of ca. 1920 Ma for the quartzofeldspathic gneiss (98TXM352) indicates a thermal event at ca. 1.9 Ga. SHRIMP analysis of the quartzofeldspathic gneiss and tonalite mylonite (98TXM298b) zircons to determine the age of zircons and their overgrowths (bright rims imaged by SEM) may clarify age-morphology relations and test the hypothesis that the bright zircon rims represent Paleoproterozoic zircon growth. In situ analyses of zircons and monazites would illuminate age differences between grains included in low CaO garnet cores, high-CaO garnet rims, and unarmoured matrix grains.
- 4) Why do the southernmost thermobarometric samples not exhibit any evidence for 1.9 Ga overprinting (i.e. lack high-CaO garnet rims)? In-situ study of rocks south of the UC may reveal 1.9 Ga mineral growth, implying that the rocks did not contain a mineralogy conducive to garnet growth at that time. Alternatively, the lack of ca. 1.9 Ga effects would support conjecture of a cryptic fault between the granulite-facies UC rocks and the amphibolite-facies rocks to the south (98TXM339a and b). An east-west trending fault between localities 98TXM339 and 98TXM201 could extend west across Chesterfield Inlet, merging with the Akunak Bay shear zone (Tella et al.,

2000b). Petrographic and thermobarometric studies of samples from either side of the proposed fault/shear zone on both sides of Chesterfield Inlet would clarify this conundrum.

Table 6.1: Summary of events in the Uvauk complex and surrounding rocks

Age	Method	Dated by	
2730 ± 2 Ma	U-Pb, zircon	Roddick and Tella, 1993, unpublished data	Magmatic age of K-spar augen granite north of the anorthosite suite
2711 ± 5/-4 Ma	U-Pb, zircon	Millis, this study (upper intercept - zircon)	Magmatic age of leucogabbro; possible component of the anorthosite suite
ca. 2620-2600 Ma	U-Pb, zircon, monazite	Roddick and Tella, 1993, unpublished data;	Earliest metamorphism of the SBs; age of regional D1 and associated mm?
	U-Pb, zircon, monazite	Millis, this study (upper intercept of monazite, zircon data; see Table 6.1)	Age of abundant granite plutonism in the northern Hearne and Rae; age of Rae-Hearne collision?
	U-Pb, zircon	Millis, this study (upper intercept of zircon overgrowths; Figure 5.11b)	Possible age of low-CaO M1 garnet cores indicating granulite-facies mm at 7-9 kb and 700-770°C (this study); possible age of S ₄ at UC; probable age of regional D2 event (S2); final collisional stage between Rae-Hearne
1936 ± 8 Ma	U-Pb, zircon	Tella et al., 1994a	Possible age of M2 high-CaO garnet indicating granulite-facies mm at 10.8-14.3 kb, 700-865°C; Uvauk gabbro dykes intrude, possible anorthosite intrusion; age of F2?; age of S _m in NBs; age of regional D3?; possible age of high-T decompression (indicated by plagioclase-hornblende symplectite surrounding garnet); age of clockwise P-T path, indicative of crustal thickening followed by exhumation.
1933 ± 86 Ma	U-Pb, zircon	lower intercepts zircon data; sample 98TXM334b	
1921 ± 13 Ma	U-Pb, monazite	lower intercepts monazite data; sample 98TXM352	
1903 ± 2 Ma	U-Pb, titanite	Tella et al., 1994a	cooling below titanite closure temperature (660-700°C)
1.78-1.83 Ga	U-Pb, rutile	Roddick and Tella, 1993, unpublished data	Probable age of M3, this study; post-tectonic retrograde metamorphism as indicated by pseudomorph structures produced by garnet breakdown (98TXM339a)
ca. 1800 Ma	Ar-Ar, hornblende	Sandeman, 2000, unpublished data	thermal activity may be related to Hudson suite granites; age of F3/F4 folds?
ca. 1650 +278/-341 Ma	U-Pb, rutile	Millis, this study	Cooling below hornblende closure T (~400°C); possibly related to greenschist-facies hydrothermal event associated with ca. 1.75 Nuellin suite plutonism.
Note: zircon = zircon; monazite = monazite; titanite = titanite; hornblende = hornblende; rutile = rutile; NBs = northern Hearne and Rae; mm = metamorphism			

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APPENDIX I: Geological Maps and Legends

Legend - Area A of Figure 1.3

Anorthositic Suite

- **Ano:** White, medium- to very coarse-grained, white anorthosite (60%); and black, fine-grained, mafic granulite (40%). Both are variably foliated and lineated.
- **AnGb:** White with black, medium- to very coarse-grained, variably foliated and lineated anorthositic gabbro, gabbroic anorthosite, leucogabbro, and black, fine-grained, mafic granulite.
- **GtAn:** White to grey, medium- to fine-grained, well banded, garnet-bearing anorthosite, and mafic granulite. Both are well-foliated and lineated.

Surrounding Geology (relative ages unknown)

- **Tmgx:** White to buff, medium-grained, moderately foliated, well lineated, tonalite with abundant green to grey, fine-grained, inclusions of mafic granulite and pyroxenite; minor white to pink, medium-grained, well-lineated, leuconorite to charnockite.
- **Tamp:** Grey, medium-grained, well banded, recrystallized, biotite-hornblende-tonalite; minor pink augen granodiorite; black amphibolite; tan to brown, biotite-schist; all are moderately foliated.
- **Dio:** Green to grey, fine-grained, well foliated and lineated dioritic straight gneiss; subordinate garnet-quartzofeldspathic gneiss, tonalite gneiss, and amphibolite; all are proto- to ultramylonitic.
- **QFgn:** Buff to tan, fine-grained, well foliated and lineated garnet-quartzofeldspathic gneiss; minor well foliated tonalite gneiss and recrystallized amphibolite.
- **Tstr:** White to grey, medium-grained, well foliated and lineated, hornblende-biotite-tonalite orthogneiss to straight gneiss; minor dark green, medium-grained clinopyroxenite; minor mixed paragneiss.
- **Togn:** Grey to white, medium-grained, moderately foliated and locally lineated, biotite-tonalite with concordant foliated veins of pink monzogranitic pegmatite; subordinate white, coarse-grained, weakly foliated hornblende-tonalite; amphibolite.
- **Gsp:** Buff to rust, medium-grained, moderately foliated, plagioclase-porphyroblastic, garnet-semipelitic schist; dark grey, granoblastic amphibolite; minor tonalite.

Symbols

- Lithological contact
- Limit of mapping
- Shear zone rocks: proto- to ultramylonites.
- Foliation (generation unknown)
- Lineation (generation unknown)
- Sample location

Note: True north is 0°13' counterclockwise from grid north.

Note: True north is 0°13' counterclockwise from grid north.

Legend - Area B of Figure 1.3

- Tmgx: White to buff, medium-grained, moderately foliated, well lineated, tonalite with abundant green to grey, fine-grained inclusions of mafic granulite and pyroxenite; minor white to pink, medium-grained, well-lineated, leuconorite to charnockite.
- Glco: Coarse-grained, white with black, variably foliated, generally well lineated leucogabbro to quartz diorite with polycrystalline hornblende aggregates ~1 cm in diameter.
- Togn: White to grey, medium-grained, hornblende-biotite tonalite orthogneiss and minor brown to black discontinuous lenses of melanogabbro to pyroxenite.
- Grd: Buff to cream, medium-grained, weakly foliated biotite-granodiorite with salmon-coloured K-feldspar and trace magnetite.
- Amph: Dominantly recrystallized amphibolite (+/- biotite), minor medium-grained, grey, foliated biotite-tonalite and monzogranite pegmatite.
- Trm: Grey to buff, well foliated and lineated, biotite-tonalite and amphibolite, interpreted to be recrystallized mylonite..
- Gton: White to cream, massive to weakly foliated, granite to tonalite with abundant tectonized amphibolite inclusions at the eastern margin.

Symbols

- Lithological contact
- Limit of mapping
- Shear zone rocks: proto- to ultramylonites.
- Foliation (generation unknown)
- Lineation (generation unknown)
- Sample location

Legend - Area C of Figure 1.3

Proterozoic Rocks

Monzogranite: Massive to poorly foliated, equigranular, medium to coarse grained pink monzogranite.

Anorthositic Suite

Anorthosite: Massive to well foliated and lineated, medium to very coarse grained, white anorthosite, gabbroic anorthosite; moderately to well foliated, fine grained mafic granulite; rare coarse grained gabbro and leucogabbro.

Moderately to well foliated, fine grained, black mafic granulite.

Archean Rocks

Metasedimentary rocks: Well foliated, commonly lineated, buff- to rust-weathering garnet-cummingtonite-biotite schist and paragneiss.

Well-layered to banded, hornblende-biotite tonalitic to granodioritic orthogneiss.

Symbols

- Lithological contact
- Limit of mapping
- Metamorphic isograd (symbols on high-grade side)
- Shear zone rocks: proto- to ultramylonites.
- Foliation (generation unknown)
- Lineation (generation unknown)
- Fold axis (generation unknown)
- Sample location

APPENDIX II: Mineral Chemistry used for P-T Calculations

98TXM561b - Garnet

Spot I.D.	G14B	G34E	G14'A
SiO ₂	37.67	37.46	37.51
Al ₂ O ₃	20.77	20.63	20.11
TiO ₂	0.17	0.25	0.29
Cr ₂ O ₃	0.05	0.03	0.05
FeO	26.87	23.01	19.76
MnO	1.51	0.43	0.29
MgO	3.45	3.72	2.48
CaO	9.65	13.85	18.54
Na ₂ O	0.00	0.00	0.00
K ₂ O	0.00	0.00	0.00
Total	100.14	99.39	99.03
Number of cations based on n oxygens			
n	12	12	12
Si	2.98	2.96	2.97
Al iv	0.02	0.04	0.03
Al vi	1.92	1.88	1.85
Ti	0.01	0.02	0.02
Cr	0.00	0.00	0.00
Fe	1.78	1.52	1.31
Mn	0.10	0.03	0.02
Mg	0.41	0.44	0.29
Ca	0.82	1.17	1.57
%Alm	57.28	48.12	40.97
%Sps	3.27	0.90	0.60
%Prp	13.10	13.87	9.18
%Grs	26.35	37.12	49.25

98TXM561b - Hornblende

Spot I.D.	H4B	H6B	H12E
SiO ₂	39.61	39.06	42.17
Al ₂ O ₃	16.87	17.38	12.23
TiO ₂	0.43	0.46	0.49
Cr ₂ O ₃	0.00	0.00	0.00
FeO	18.08	17.60	22.99
MnO	0.22	0.16	0.14
K ₂ O	0.38	0.40	0.20
MgO	7.60	7.31	5.68
CaO	11.33	11.38	11.83
Na ₂ O	1.90	1.87	1.24
Cl	0.24	0.20	0.04
F	0.00	0.00	0.00
NiO	0.00	0.03	0.00
Total	96.80	96.14	97.41
Number of cations based on n oxygens			
n	23	23	23
Si	6.05	6.00	6.53
Al iv	1.95	2.00	1.47
Al vi	1.08	1.15	0.76
Ti	0.05	0.05	0.06
Cr	0.00	0.00	0.00
Fe	2.31	2.26	2.98
Mn	0.03	0.02	0.02
K	0.07	0.08	0.04
Mg	1.73	1.67	1.31
Ca	1.85	1.87	1.96
Na	0.28	0.28	0.19
Cl	0.02	0.02	0.00
F	0.00	0.00	0.00

98TXM561b - Plagioclase

Spot I.D.	P16B	P34E	P4'A
SiO ₂	56.04	47.67	51.52
Al ₂ O ₃	27.36	33.27	29.30
K ₂ O	0.06	0.01	0.04
FeO	0.10	0.23	0.21
CaO	9.70	16.54	12.40
Na ₂ O	5.75	1.98	4.94
SrO	0.00	0.00	0.09
BaO	0.00	0.00	0.15
Total	99.01	99.70	98.63
Number of cations based on n oxygens			
n	8	8	8
Si	2.54	2.19	2.38
Al iv	1.46	1.81	1.62
Al vi	0.00	-0.01	-0.02
K	0.00	0.00	0.00
Fe	0.00	0.01	0.01
Ca	0.47	0.81	0.61
Na	0.51	0.18	0.44
%An	48.08	82.15	57.99
%Ab	51.57	17.80	41.77
%Or	0.35	0.06	0.24

Microdomain indicated by last letter of spot I.D.

98TXM304 - Garnet

Spot I.D.	G10A	G26A	G29A
SiO ₂	37.56	38.53	38.51
Al ₂ O ₃	20.75	21.35	20.89
TiO ₂	0.16	0.20	0.15
Cr ₂ O ₃	0.03	0.08	0.00
FeO	25.52	23.01	25.49
MnO	0.74	0.34	0.51
MgO	3.50	5.06	5.23
CaO	10.85	11.55	9.10
Na ₂ O	0.00	0.00	0.00
K ₂ O	0.02	0.02	0.02
Total	99.11	100.12	99.88
<u>Number of cations based on n oxygens</u>			
n	12	12	12
Si	2.99	2.99	3.01
Al iv	0.01	0.01	-0.01
Al vi	1.93	1.94	1.94
Ti	0.01	0.01	0.01
Cr	0.00	0.00	0.00
Fe	1.70	1.49	1.67
Mn	0.05	0.02	0.03
Mg	0.41	0.58	0.61
Ca	0.92	0.96	0.76
%Alm	54.99	48.79	54.23
%Sps	1.62	0.73	1.10
%Prp	13.43	19.11	19.85
%Grs	29.96	31.37	24.81

98TXM304 - Hornblende

Spot I.D.	H7A	H15A	H16A
SiO ₂	40.52	40.69	41.62
Al ₂ O ₃	14.32	13.53	13.08
TiO ₂	0.76	0.80	0.69
Cr ₂ O ₃	0.03	0.05	0.02
FeO	20.80	21.13	20.83
MnO	0.16	0.28	0.28
K ₂ O	0.99	0.82	0.79
MgO	5.92	6.19	6.39
CaO	11.56	11.42	11.59
Na ₂ O	1.12	1.15	1.07
Cl	0.03	0.04	0.07
F	0.00	0.19	0.47
NiO	0.00	0.01	0.02
Total	96.50	96.46	97.16
<u>Number of cations based on n oxygens</u>			
n	23	23	23
Si	6.30	6.34	6.44
Al iv	1.70	1.66	1.56
Al vi	0.92	0.83	0.83
Ti	0.09	0.09	0.08
Cr	0.00	0.01	0.00
Fe	2.70	2.75	2.70
Mn	0.02	0.04	0.04
K	0.20	0.16	0.16
Mg	1.37	1.44	1.48
Ca	1.92	1.91	1.92
Na	0.17	0.17	0.16
Cl	0.00	0.00	0.01
F	0.00	0.10	0.23

98TXM304 - Plagioclase

Spot I.D.	P13A	P22A	P25A
SiO ₂	57.55	47.82	57.75
Al ₂ O ₃	26.46	33.05	26.30
K ₂ O	0.11	0.06	0.13
FeO	0.09	0.44	0.13
CaO	8.47	15.97	8.29
Na ₂ O	6.26	1.92	6.28
SrO	0.00	0.00	0.00
BaO	0.00	0.12	0.06
Total	98.94	99.38	98.98
<u>Number of cations based on n oxygens</u>			
n	8	8	8
Si	2.60	2.21	2.61
Al iv	1.40	1.79	1.39
Al vi	0.01	0.00	0.01
K	0.01	0.00	0.01
Fe	0.00	0.02	0.00
Ca	0.41	0.79	0.40
Na	0.55	0.17	0.55
%An	42.49	81.81	41.83
%Ab	56.85	17.80	57.40
%Or	0.67	0.38	0.78

Microdomain indicated by last letter of spot I.D.

98TXM564a - Garnet

Spot I.D.	g8b	g1b
SiO ₂	38.68	38.53
Al ₂ O ₃	22.35	21.92
TiO ₂	0.07	0.03
Cr ₂ O ₃	0.01	0.05
FeO	17.94	19.84
MnO	0.37	0.39
MgO	11.02	9.71
CaO	8.05	7.47
Na ₂ O	0.00	0.00
K ₂ O	0.00	0.00
Total	98.49	97.95
Number of cations based on n O		
n	12	12
Si	2.95	2.98
Al iv	0.05	0.02
Al vi	1.96	1.97
Ti	0.00	0.00
Cr	0.00	0.00
Fe	1.14	1.28
Mn	0.02	0.03
Mg	1.25	1.12
Ca	0.66	0.62
%Alm	37.17	42.10
%Sps	0.78	0.85
%Prp	40.69	36.73
%Grs	21.36	20.32

98TXM564a - Plagioclase

Spot I.D.	p3b	p16b
SiO ₂	53.78	57.23
Al ₂ O ₃	27.57	26.73
K ₂ O	0.10	0.12
FeO	0.63	0.12
CaO	10.57	8.59
Na ₂ O	5.07	6.57
SrO	0.00	0.01
BaO	0.00	0.00
Total	97.73	99.37
Number of cations based on n O		
n	8	8
Si	2.49	2.58
Al iv	1.51	1.42
Al vi	-0.01	0.00
K	0.00	0.00
Fe	0.02	0.00
Ca	0.27	0.23
Na	0.23	0.29
%An	54.12	43.75
%Ab	45.27	55.58
%Or	0.61	0.67

98TXM564a- Orthopyroxene

Spot I.D.	o2b	o1b
SiO ₂	51.53	51.74
Al ₂ O ₃	2.74	2.62
TiO ₂	0.11	0.06
Cr ₂ O ₃	0.03	0.01
FeO	19.25	20.04
MnO	0.18	0.16
K ₂ O	0.00	0.01
MgO	23.75	23.87
CaO	0.47	0.28
Na ₂ O	0.00	0.00
Total	98.07	98.79
Number of cations based on n O		
n	6	6
Si	1.93	1.93
Al	0.12	0.12
Ti	0.00	0.00
Cr	0.00	0.00
Fe	0.60	0.62
Mn	0.01	0.01
K	0.00	0.00
Mg	1.33	1.33
Ca	0.02	0.01
Na	0.00	0.00

98TXM564a - Clinopyroxene

Spot I.D.	clb	c3b
SiO ₂	52.38	50.88
Al ₂ O ₃	5.40	4.30
TiO ₂	0.27	0.39
Cr ₂ O ₃	0.04	0.01
FeO	5.51	7.05
MnO	0.02	0.02
K ₂ O	0.01	0.02
MgO	13.93	13.59
CaO	22.07	21.58
Na ₂ O	0.95	0.77
Total	99.63	97.86
Number of cations based on n O		
n	6	6
Si	1.91	1.91
Al	0.23	0.19
Ti	0.01	0.01
Cr	0.00	0.00
Fe	0.17	0.22
Mn	0.00	0.00
K	0.00	0.00
Mg	0.76	0.76
Ca	0.86	0.87
Na	0.07	0.06

Microdomain indicated by last letter of spot I.D.

98TXM333 - Garnet

Spot I.D.	G1B	G8B
SiO ₂	38.39	38.04
Al ₂ O ₃	21.57	21.40
TiO ₂	0.07	0.07
Cr ₂ O ₃	0.00	0.05
FeO	25.80	27.51
MnO	0.81	0.94
MgO	6.83	5.67
CaO	6.47	6.48
Na ₂ O	0.00	0.01
K ₂ O	0.00	0.01
Total	99.95	100.17
<u>Number of cations based on n O</u>		
n	12	12
Si	2.98	2.98
Al iv	0.02	0.02
Al vi	1.96	1.95
Ti	0.00	0.00
Cr	0.00	0.00
Fe	1.68	1.80
Mn	0.05	0.06
Mg	0.79	0.66
Ca	0.54	0.54
%Alm	54.79	58.69
%Sps	1.74	2.03
%Prp	25.86	21.57
%Grs	17.60	17.71

98TXM333 - Plagioclase

Spot I.D.	P1B	P4B
SiO ₂	57.21	58.08
Al ₂ O ₃	27.23	27.19
K ₂ O	0.12	0.13
FeO	0.16	0.19
CaO	8.83	8.92
Na ₂ O	6.43	6.51
SrO	0.00	0.06
BaO	0.00	0.06
Total	99.97	101.14
<u>Number of cations based on n O</u>		
n	8	8
Si	2.56	2.58
Al iv	1.44	1.42
Al vi	0.00	0.00
K	0.01	0.01
Fe	0.01	0.01
Ca	0.42	0.42
Na	0.56	0.56
%An	42.85	42.77
%Ab	56.46	56.49
%Or	0.69	0.74

98TXM333- Orthopyroxene

Spot I.D.	O3B	O2B
SiO ₂	50.30	50.67
Al ₂ O ₃	2.00	1.77
TiO ₂	0.11	0.07
Cr ₂ O ₃	0.05	0.00
FeO	29.66	30.99
MnO	0.38	0.48
K ₂ O	0.01	0.00
MgO	17.07	16.17
CaO	0.66	0.56
Na ₂ O	0.00	0.00
Total	100.23	100.70
<u>Number of cations based on n O</u>		
n	6	6
Si	1.94	1.95
Al	0.09	0.08
Ti	0.00	0.00
Cr	0.00	0.00
Fe	0.96	1.00
Mn	0.01	0.02
K	0.00	0.00
Mg	0.98	0.93
Ca	0.03	0.02
Na	0.00	0.00

98TXM333 - Clinopyroxene

Spot I.D.	C1B	C3B
SiO ₂	54.46	51.34
Al ₂ O ₃	3.52	3.46
TiO ₂	0.09	0.39
Cr ₂ O ₃	0.01	0.05
FeO	15.78	11.06
MnO	0.13	0.06
K ₂ O	0.16	0.01
MgO	16.00	12.40
CaO	9.04	21.03
Na ₂ O	0.30	0.36
Total	99.48	100.16
<u>Number of cations based on n O</u>		
n	6	6
Si	2.01	1.92
Al	0.15	0.15
Ti	0.00	0.01
Cr	0.00	0.00
Fe	0.49	0.35
Mn	0.00	0.00
K	0.01	0.00
Mg	0.88	0.69
Ca	0.36	0.84
Na	0.02	0.03

Microdomain indicated by last letter of spot I.D.

98TXM342- Garnet

Spot I.D.	G1A	G8A
SiO ₂	38.42	38.30
Al ₂ O ₃	21.76	21.29
TiO ₂	0.14	0.05
Cr ₂ O ₃	0.04	0.11
FeO	21.54	25.51
MnO	0.48	0.77
MgO	7.81	7.06
CaO	8.52	6.14
Na ₂ O	0.00	0.00
K ₂ O	0.00	0.01
Total	98.70	99.24
Number of cations based on n O		
n	12	12
Si	2.98	2.99
Al iv	0.02	0.01
Al vi	1.97	1.95
Ti	0.01	0.00
Cr	0.00	0.01
Fe	1.40	1.67
Mn	0.03	0.05
Mg	0.90	0.82
Ca	0.71	0.51
%Alm	45.96	54.57
%Sps	1.04	1.67
%Prp	29.71	26.93
%Grs	23.29	16.83

98TXM342 - Plagioclase

Spot I.D.	P2A	P7A
SiO ₂	53.35	57.92
Al ₂ O ₃	28.21	26.69
K ₂ O	0.06	0.09
FeO	1.02	0.13
CaO	10.87	8.43
Na ₂ O	4.49	6.38
SrO	0.00	0.00
BaO	0.08	0.02
Total	98.09	99.67
Number of cations based on n O		
n	8	8
Si	2.46	2.60
Al iv	1.54	1.40
Al vi	-0.01	0.01
K	0.00	0.01
Fe	0.04	0.00
Ca	0.54	0.40
Na	0.40	0.55
%An	57.01	41.98
%Ab	42.61	57.49
%Or	0.37	0.53

98TXM342 Orthopyroxene

Spot I.D.	O3A	O4A
SiO ₂	50.67	51.83
Al ₂ O ₃	1.52	1.57
TiO ₂	0.08	0.23
Cr ₂ O ₃	0.07	0.00
FeO	26.75	26.82
MnO	0.29	0.34
K ₂ O	0.00	0.00
MgO	18.88	18.92
CaO	0.31	0.49
Na ₂ O	0.00	0.00
Total	98.56	100.19
Number of cations based on n O		
n	6	6
Si	1.96	1.97
Al	0.07	0.07
Ti	0.00	0.01
Cr	0.00	0.00
Fe	0.86	0.85
Mn	0.01	0.01
K	0.00	0.00
Mg	1.09	1.07
Ca	0.01	0.02
Na	0.00	0.00

98TXM342- Clinopyroxene

Spot I.D.	C1A	C6A
SiO ₂	52.01	50.44
Al ₂ O ₃	2.18	2.83
TiO ₂	0.25	0.27
Cr ₂ O ₃	0.04	0.10
FeO	8.19	9.51
MnO	0.09	0.08
K ₂ O	0.01	0.00
MgO	13.46	12.92
CaO	22.10	21.02
Na ₂ O	0.20	0.47
Total	98.53	97.63
Number of cations based on n O		
n	6	6
Si	1.96	1.93
Al	0.10	0.13
Ti	0.01	0.01
Cr	0.00	0.00
Fe	0.26	0.30
Mn	0.00	0.00
K	0.00	0.00
Mg	0.76	0.74
Ca	0.89	0.86
Na	0.01	0.03

Microdomain indicated by last letter of spot I.D.

98TXM178d - Garnet

Spot I.D.	G1C	G6C
SiO ₂	38.82	38.66
Al ₂ O ₃	21.88	21.83
TiO ₂	0.05	0.07
Cr ₂ O ₃	0.12	0.13
FeO	28.05	27.94
MnO	0.34	0.46
MgO	9.41	8.11
CaO	1.87	3.80
Na ₂ O	0.00	0.00
K ₂ O	0.02	0.01
Total	100.56	101.00
Number of cations based on n O		
n	12	12
Si	2.98	2.97
Al iv	0.02	0.03
Al vi	1.96	1.95
Ti	0.00	0.00
Cr	0.01	0.01
Fe	1.80	1.80
Mn	0.02	0.03
Mg	1.08	0.93
Ca	0.15	0.31
%Alm	58.97	58.53
%Sps	0.72	0.98
%Prp	35.27	30.29
%Grs	5.04	10.20

98TXM178d - Plagioclase

Spot I.D.	P7C	P9C
SiO ₂	62.09	62.01
Al ₂ O ₃	24.16	24.20
K ₂ O	0.16	0.12
FeO	0.21	0.20
CaO	5.56	5.43
Na ₂ O	8.34	8.44
SrO	0.00	0.07
BaO	0.00	0.15
Total	100.52	100.64
Number of cations based on n O		
n	8	8
Si	2.74	2.74
Al iv	1.26	1.26
Al vi	0.00	0.00
K	0.01	0.01
Fe	0.01	0.01
Ca	0.26	0.26
Na	0.71	0.72
%An	26.68	26.05
%Ab	72.41	73.27
%Or	0.91	0.69

98TXM178d - Orthopyroxene

Spot I.D.	O3C	O5C
SiO ₂	53.40	51.97
Al ₂ O ₃	4.14	2.63
TiO ₂	0.11	0.08
Cr ₂ O ₃	0.06	0.08
FeO	22.04	22.81
MnO	0.13	0.08
K ₂ O	0.00	0.02
MgO	23.15	22.30
CaO	0.27	0.13
Na ₂ O	0.00	0.00
Total	103.30	100.09
Number of cations based on n O		
n	6	6
Si	1.91	1.93
Al	0.17	0.12
Ti	0.00	0.00
Cr	0.00	0.00
Fe	0.66	0.71
Mn	0.00	0.00
K	0.00	0.00
Mg	1.23	1.24
Ca	0.01	0.01
Na	0.00	0.00

Microdomain indicated by last letter of spot I.D.

98TXM350 - Garnet

Spot I.D.	G1A	G9A
SiO ₂	37.37	36.85
Al ₂ O ₃	21.36	21.23
TiO ₂	0.00	0.03
Cr ₂ O ₃	0.06	0.00
FeO	27.94	28.06
MnO	2.00	1.84
MgO	4.57	5.13
CaO	5.82	5.42
Na ₂ O	0.00	0.00
K ₂ O	0.00	0.01
Total	99.13	98.57
Number of cations based on n O		
n	12	12
Si	2.98	2.95
Al iv	0.02	0.05
Al vi	1.98	1.96
Ti	0.00	0.00
Cr	0.00	0.00
Fe	1.86	1.88
Mn	0.13	0.12
Mg	0.54	0.61
Ca	0.50	0.47
%Alm	61.31	60.98
%Sps	4.45	4.05
%Prp	17.88	19.88
%Grs	16.36	15.09

98TXM350 - Plagioclase

Spot I.D.	P5A	P8A
SiO ₂	56.43	53.67
Al ₂ O ₃	27.06	27.55
K ₂ O	0.14	0.13
FeO	0.15	0.35
CaO	9.22	9.92
Na ₂ O	6.03	5.54
SrO	0.43	0.43
BaO	0.00	0.00
Total	99.46	97.58
Number of cations based on n O		8
n	8	8
Si	2.56	2.49
Al iv	1.44	1.51
Al vi	0.00	0.00
K	0.00	0.00
Fe	0.01	0.01
Ca	0.24	0.25
Na	0.26	0.25
%An	47.35	50.08
%Ab	51.86	49.16
%Or	0.79	0.76

98TXM350 - Orthopyroxene

Spot I.D.	O1A	O10A
SiO ₂	50.99	50.43
Al ₂ O ₃	1.69	1.71
TiO ₂	0.14	0.10
Cr ₂ O ₃	0.04	0.06
FeO	22.39	24.20
MnO	0.41	0.42
K ₂ O	0.00	0.01
MgO	21.97	20.56
CaO	0.69	0.40
Na ₂ O	0.00	0.00
Total	98.32	97.88
Number of cations based on n O		6
n	6	6
Si	1.94	1.94
Al	0.08	0.08
Ti	0.00	0.00
Cr	0.00	0.00
Fe	0.71	0.78
Mn	0.01	0.01
K	0.00	0.00
Mg	1.25	1.18
Ca	0.03	0.02
Na	0.00	0.00

98TXM350 - Clinopyroxene

Spot I.D.	C1A	C3A
SiO ₂	50.27	49.44
Al ₂ O ₃	3.87	4.15
TiO ₂	0.43	0.57
Cr ₂ O ₃	0.25	0.19
FeO	10.32	8.95
MnO	0.15	0.10
K ₂ O	0.02	0.37
MgO	13.57	14.03
CaO	21.11	20.06
Na ₂ O	0.54	0.41
Total	100.52	98.27
Number of cations based on n O		6
n	6	6
Si	1.88	1.88
Al	0.17	0.19
Ti	0.01	0.02
Cr	0.01	0.01
Fe	0.32	0.28
Mn	0.00	0.00
K	0.00	0.02
Mg	0.76	0.79
Ca	0.85	0.82
Na	0.04	0.03

Microdomain indicated by last letter of spot I.D.

98TXM201 - Garnet

Spot I.D.	G3G	G1G	G7G
SiO ₂	37.31	38.03	37.25
Al ₂ O ₃	21.03	21.25	20.96
TiO ₂	0.06	0.09	0.03
Cr ₂ O ₃	0.00	0.02	0.00
FeO	24.75	25.96	29.38
MnO	1.77	2.94	2.94
MgO	2.83	2.11	3.10
CaO	12.08	11.04	6.48
Na ₂ O	0.00	0.00	0.00
K ₂ O	0.00	0.01	0.01
Total	99.84	101.44	100.15
Number of cations based on n oxygens			
n	12	12	12
Si	2.96	2.98	2.98
Al iv	0.04	0.02	0.02
Al vi	1.93	1.95	1.95
Ti	0.00	0.01	0.00
Cr	0.00	0.00	0.00
Fe	1.64	1.70	1.96
Mn	0.12	0.20	0.20
Mg	0.33	0.25	0.37
Ca	1.03	0.93	0.55
%Alm	52.59	55.42	63.61
%Sps	3.81	6.36	6.45
%Prp	10.72	8.03	11.97
%Grs	32.88	30.19	17.98

98TXM201 - Hornblende

Spot I.D.	H7A	H15A	H16A
SiO ₂	40.52	40.69	41.62
Al ₂ O ₃	14.32	13.53	13.08
TiO ₂	0.76	0.80	0.69
Cr ₂ O ₃	0.03	0.05	0.02
FeO	20.80	21.13	20.83
MnO	0.16	0.28	0.28
K ₂ O	0.99	0.82	0.79
MgO	5.92	6.19	6.39
CaO	11.56	11.42	11.59
Na ₂ O	1.12	1.15	1.07
Cl	0.03	0.04	0.07
F	0.00	0.19	0.47
NiO	0.00	0.01	0.02
Total	96.50	96.46	97.16
Number of cations based on n oxygens			
n	23	23	23
Si	6.30	6.34	6.44
Al iv	1.70	1.66	1.56
Al vi	0.92	0.83	0.83
Ti	0.09	0.09	0.08
Cr	0.00	0.01	0.00
Fe	2.70	2.75	2.70
Mn	0.02	0.04	0.04
K	0.20	0.16	0.16
Mg	1.37	1.44	1.48
Ca	1.92	1.91	1.92
Na	0.17	0.17	0.16
Cl	0.00	0.00	0.01
F	0.00	0.10	0.23

98TXM201 - Plagioclase

Spot I.D.	P1G	P19G	P25G
SiO ₂	58.53	61.43	61.80
Al ₂ O ₃	26.62	23.75	23.03
K ₂ O	0.14	0.11	0.09
FeO	0.18	0.02	0.02
CaO	8.18	8.60	8.96
Na ₂ O	7.09	5.58	5.09
SrO	0.38	0.39	0.11
BaO	0.00	0.01	0.00
Total	101.13	99.88	99.09
Number of cations based on n oxygens			
n	8	8	8
Si	2.60	2.62	2.61
Al iv	1.40	1.38	1.39
Al vi	0.00	0.01	0.01
K	0.01	0.00	0.01
Fe	0.01	0.01	0.00
Ca	0.39	0.37	0.40
Na	0.61	0.57	0.55
%An	38.6	39.34	41.83
%Ab	60.6	60.19	57.40
%Or	0.8	0.47	0.78

Microdomain indicated by last letter of spot I.D.

98TXM339a - Garnet

Spot I.D.	G9B	G18B
SiO ₂	37.23	36.95
Al ₂ O ₃	20.85	20.90
TiO ₂	0.09	0.04
Cr ₂ O ₃	0.04	0.02
FeO	25.94	27.54
MnO	2.61	4.62
MgO	2.19	1.52
CaO	10.41	7.40
Na ₂ O	0.00	0.00
K ₂ O	0.01	0.01
Total	99.37	98.99
Number of cations based on n oxygens		
n	12	12
Si	2.98	2.99
Al iv	0.02	0.01
Al vi	1.95	1.99
Ti	0.01	0.00
Cr	0.00	0.00
Fe	1.74	1.87
Mn	0.18	0.32
Mg	0.26	0.18
Ca	0.89	0.64
%Alm	56.61	62.01
%Sps	5.77	10.54
%Prp	8.52	6.10
%Grs	29.10	21.35

98TXM339a - Hornblende

Spot I.D.	H8B	H11B
SiO ₂	44.22	43.52
Al ₂ O ₃	12.05	13.07
TiO ₂	0.62	0.50
Cr ₂ O ₃	0.12	0.12
FeO	18.15	18.49
MnO	0.39	0.27
K ₂ O	0.31	0.29
MgO	8.58	8.24
CaO	11.12	11.34
Na ₂ O	0.84	0.90
Cl	0.04	0.03
F	0.00	0.00
NiO	0.00	0.08
Total	96.59	97.14
Number of cations based on n oxygens		
n	23	23
Si	6.69	6.57
Al iv	1.31	1.43
Al vi	0.84	0.90
Ti	0.07	0.06
Cr	0.01	0.01
Fe	2.30	2.34
Mn	0.05	0.03
K	0.06	0.06
Mg	1.93	1.86
Ca	1.80	1.84
Na	0.12	0.13
Cl	0.00	0.00
F	0.00	0.00

98TXM339a - Plagioclase

Spot I.D.	P10B	P13B
SiO ₂	45.49	47.80
Al ₂ O ₃	34.71	33.61
K ₂ O	0.00	0.01
FeO	0.22	0.13
CaO	17.61	16.55
Na ₂ O	1.22	2.04
SrO	0.00	0.00
BaO	0.14	0.00
Total	99.41	100.15
Number of cations based on n oxygens		
n	8	8
Si	2.11	2.19
Al iv	1.89	1.81
Al vi	0.01	0.00
K	0.00	0.00
Fe	0.01	0.00
Ca	0.87	0.81
Na	0.11	0.18
%An	88.86	81.71
%Ab	11.14	18.23
%Or	0.00	0.06

Microdomain indicated by last letter of spot I.D.

98TXM339b - Garnet

Spot I.D.	G7B	G9B
SiO ₂	36.59	36.12
Al ₂ O ₃	21.26	20.88
TiO ₂	0.04	0.02
Cr ₂ O ₃	0.04	0.05
FeO	35.48	35.72
MnO	0.71	0.81
MgO	2.95	2.80
CaO	2.83	2.36
Na ₂ O	0.00	0.00
K ₂ O	0.01	0.02
Total	99.92	98.77
Number of cations based on n oxygens		
n	12	12
Si	2.96	2.96
Al iv	0.04	0.04
Al vi	1.98	1.97
Ti	0.00	0.00
Cr	0.00	0.00
Fe	2.40	2.45
Mn	0.05	0.06
Mg	0.36	0.34
Ca	0.24	0.21
%Alm	78.69	80.15
%Sps	1.60	1.85
%Prp	11.68	11.21
%Grs	8.04	6.78

98TXM339b - Biotite

Spot I.D.	B5B	B6B
SiO ₂	34.26	34.44
Al ₂ O ₃	19.44	19.47
TiO ₂	1.56	1.67
Cr ₂ O ₃	0.10	0.09
FeO	19.31	19.54
MnO	0.00	0.02
K ₂ O	8.69	8.25
MgO	9.85	9.75
CaO	0.07	0.11
Na ₂ O	0.19	0.28
Cl	0.01	0.00
F	0.00	0.00
Total	94.13	94.27
Number of cations based on n oxygens		
n	12	12
Si	2.90	2.90
Al	1.94	1.93
Ti	0.10	0.11
Cr	0.01	0.01
Fe	1.36	1.38
Mn	0.00	0.00
K	0.47	0.44
Mg	1.24	1.22
Ca	0.01	0.01
Na	0.02	0.02
Cl	0.00	0.00

98TXM339b - Plagioclase

Spot I.D.	P4B	P7B
SiO ₂	56.11	55.17
Al ₂ O ₃	27.94	27.31
K ₂ O	0.07	0.03
FeO	0.90	0.21
CaO	9.44	9.05
Na ₂ O	6.36	6.56
SiO	0.15	0.12
BaO	0.00	0.05
Total	100.97	98.51
Number of cations based on n oxygens		
n	8	8
Si	2.51	2.52
Al iv	1.49	1.48
Al vi	-0.01	0.00
K	0.00	0.00
Fe	0.03	0.01
Ca	0.45	0.44
Na	0.55	0.58
%An	44.88	43.18
%Ab	54.72	56.65
%Or	0.40	0.17

Microdomain indicated by last letter of spot I.D.

98TXM248 - Garnet

Spot I.D.	G6B	G10B
SiO ₂	40.39	39.30
Al ₂ O ₃	20.91	22.17
TiO ₂	0.00	0.01
Cr ₂ O ₃	0.02	0.00
FeO	24.34	27.25
MnO	0.76	1.29
MgO	6.24	6.25
CaO	6.58	6.28
Na ₂ O	0.00	0.00
K ₂ O	0.00	0.03
Total	99.24	102.58
Number of cations based on n O		
n	12	12
Si	3.12	2.99
Al iv	-0.12	0.01
Al vi	2.03	1.98
Ti	0.00	0.00
Cr	0.00	0.00
Fe	1.57	1.73
Mn	0.05	0.08
Mg	0.72	0.71
Ca	0.55	0.51
%Alm	54.49	57.07
%Sps	1.72	2.74
%Prp	24.91	23.34
%Grs	18.87	16.85

98TXM248 - Plagioclase

Spot I.D.	P6B	P10B
SiO ₂	57.04	61.75
Al ₂ O ₃	26.84	24.20
K ₂ O	0.16	0.31
FeO	1.01	0.18
CaO	8.91	5.88
Na ₂ O	6.17	7.93
SrO	0.34	0.38
BaO	0.00	0.11
Total	100.47	100.74
Number of cations based on n O		
n	8	8
Si	2.56	2.74
Al iv	1.44	1.26
Al vi	-0.01	0.00
K	0.01	0.02
Fe	0.04	0.01
Ca	0.43	0.28
Na	0.54	0.68
%An	43.97	28.54
%Ab	55.09	69.66
%Or	0.94	1.79

98TXM248 - Orthopyroxene

Spot I.D.	O1B	O4B
SiO ₂	51.41	52.43
Al ₂ O ₃	1.03	1.23
TiO ₂	0.05	0.04
Cr ₂ O ₃	0.11	0.07
FeO	25.61	25.30
MnO	0.39	0.25
K ₂ O	0.00	0.02
MgO	20.70	19.61
CaO	0.36	0.70
Na ₂ O	0.00	0.01
Total	99.66	99.66
Number of cations based on n O		
n	6	6
Si	1.95	1.98
Al	0.05	0.05
Ti	0.00	0.00
Cr	0.00	0.00
Fe	0.81	0.80
Mn	0.01	0.01
K	0.00	0.00
Mg	1.17	1.11
Ca	0.01	0.03
Na	0.00	0.00

98TXM248 - Clinopyroxene

Spot I.D.	C1B	C2B
SiO ₂	52.39	56.59
Al ₂ O ₃	2.60	2.33
TiO ₂	0.04	0.05
Cr ₂ O ₃	0.07	0.12
FeO	8.97	9.83
MnO	0.14	0.14
K ₂ O	0.00	0.05
MgO	13.49	15.23
CaO	21.50	16.57
Na ₂ O	0.96	0.54
Total	100.17	101.45
Number of cations based on n O		
n	6	6
Si	1.95	2.03
Al	0.11	0.10
Ti	0.00	0.00
Cr	0.00	0.00
Fe	0.28	0.30
Mn	0.00	0.00
K	0.00	0.00
Mg	0.75	0.82
Ca	0.86	0.64
Na	0.07	0.04

Microdomain indicated by last letter of spot I.D.

98TXM320 - Garnet				98TXM320 - Plagioclase				98TXM320 - Biotite				98TXM320 - Hornblende			
Spot I.D.	g3d	g9d		Spot I.D.	p2d	p6d		Spot I.D.	b1d	b4d		Spot I.D.	h1d	h3d	
SiO ₂	36.63	36.26		SiO ₂	58.44	55.73		SiO ₂	34.42	33.68		SiO ₂	39.63	39.54	
Al ₂ O ₃	20.79	20.70		Al ₂ O ₃	25.17	27.33		Al ₂ O ₃	16.87	15.84		Al ₂ O ₃	14.97	14.66	
TiO ₂	0.04	0.05		K ₂ O	0.88	0.04		TiO ₂	2.35	2.42		TiO ₂	0.47	0.43	
Cr ₂ O ₃	0.01	0.01		FeO	1.24	0.05		Cr ₂ O ₃	0.09	0.07		Cr ₂ O ₃	0.09	0.02	
FeO	33.12	34.52		CaO	6.01	9.49		FeO	22.13	23.41		FeO	21.76	22.22	
MnO	0.50	0.58		Na ₂ O	6.35	5.82		MnO	0.04	0.00		MnO	0.10	0.10	
MgO	2.92	2.26		SrO	0.00	0.00		K ₂ O	8.63	8.50		K ₂ O	0.52	0.58	
CaO	5.13	4.37		BaO	0.00	0.16		MgO	8.55	8.29		MgO	5.64	5.77	
Na ₂ O	0.00	0.00		Total	98.10	98.63		CaO	0.07	0.06		CaO	11.10	10.97	
K ₂ O	0.00	0.01		Number of cations based on n O				Na ₂ O	0.14	0.15		Na ₂ O	1.15	1.16	
Total	99.13	98.75		n	8	8		Cl	0.09	0.07		Cl	0.03	0.12	
Number of cations based on n O				Number of cations based on n O				F	0.45	0.52		F	0.26	0.31	
n	12	12		n	8	8		Total	94.44	93.38		Total	95.87	96.14	
Number of cations based on n O				Number of cations based on n O				Number of cations based on n O				Number of cations based on n O			
Si	2.97	2.97		Si	2.66	2.54		Si	2.96	2.96		Si	6.22	6.22	
Al iv	0.03	0.03		Al iv	1.34	1.46		Al	1.71	1.64		Al iv	1.78	1.78	
Al vi	1.95	1.96		Al vi	0.01	0.01		Ti	0.15	0.16		Al vi	0.99	0.93	
Ti	0.00	0.00		K	0.05	0.00		Cr	0.01	0.00		Ti	0.06	0.05	
Cr	0.00	0.00		Fe	0.05	0.00		Fe	1.59	1.72		Cr	0.01	0.00	
Fe	2.24	2.36		Ca	0.29	0.46		Mn	0.00	0.00		Fe	2.86	2.92	
Mn	0.03	0.04		Na	0.56	0.51		K	0.47	0.48		Mn	0.01	0.01	
Mg	0.35	0.28		%An	32.39	47.28		Mg	1.10	1.09		K	0.10	0.12	
Ca	0.45	0.38		%Ab	61.97	52.48		Ca	0.01	0.01		Mg	1.32	1.35	
Number of cations based on n O				%Or	5.64	0.24		Na	0.01	0.01		Ca	1.87	1.85	
%Alm	72.94	77.16		Number of cations based on n O				Cl	0.00	0.00		Na	0.17	0.18	
%Sps	1.12	1.32		n	12	12		Number of cations based on n O				Cl	0.00	0.01	
%Prp	11.48	9.00		Number of cations based on n O				Number of cations based on n O				F	0.13	0.15	
%Grs	14.46	12.52		Number of cations based on n O				Number of cations based on n O				Number of cations based on n O			

Microdomain indicated by last letter of spot I.D.

APPENDIX III: Whole Rock Geochemistry

sample	description	easting	northing	zone	SiO ₂ XRF	TiO ₂ XRF	Al ₂ O ₃ XRF	Fe ₂ O ₃ T XRF	Fe ₂ O ₃ XRF	FeO Titre	FeOT	MnO XRF	MgO XRF
98TXM205	banded anorthosite/gabbro	516209	7081264	15	57.50	0.79	15.80	9.20	2.10	6.4	8.3	0.14	4.09
98TXM332	anorthosite	515351	7081845	15	54.60	0.08	28.50	0.50	0.10	0.4	0.4	0.00	0.21
98TXM333	mafic dike	515888	7082101	15	48.20	1.87	12.80	18.00	5.40	11.3	16.2	0.26	6.24
98TXM334	gabbroic anorthosite	516008	7081429	15	50.00	0.41	19.00	9.50	3.60	5.3	8.5	0.12	6.56
98TXM335	biotite tonalite	516083	7079663	15	68.60	0.34	16.50	3.20	1.00	2.0	2.9	0.04	0.99
98TXM342	mafic dike	517526	7082992	15	49.80	1.02	13.80	13.90	5.40	7.6	12.5	0.22	7.52
98TXM345	gabbroic anorthosite	517870	7081880	15	53.30	0.11	27.20	1.40	0.30	1.0	1.3	0.01	0.90
98TXM346	anorthosite	515697	7082393	15	54.80	0.05	27.80	0.60	0.10	0.4	0.5	0.00	0.25
98TXM347a	mafic dike	515920	7082500	15	49.60	1.08	12.30	16.10	4.50	10.5	14.5	0.25	7.01
98TXM347b	mafic dike	515920	7082500	15	50.50	1.41	11.80	18.60	6.90	10.5	16.7	0.27	6.52
98TXM348	gabbroic anorthosite	515493	7082578	15	52.70	0.08	28.40	1.20	0.30	0.8	1.1	0.01	0.58
98TXM354	anorthosite	512871	7081597	15	53.50	0.05	28.50	0.90	0.40	0.5	0.8	0.01	0.30
98TXM322	anorthosite	513451	7081821	15	47.90	0.02	33.40	0.40	0.00	0.4	0.36	0.00	0.30
98TXM313	mafic dike	515008	7081847	15	50.00	0.89	14.40	14.00	2.10	10.7	12.60	0.21	7.35
98TXM306b	c.g. gt-gabbroic anorthosite	515589	7083260	15	52.50	0.64	25.50	4.80	1.20	3.3	4.32	0.04	1.17
98TXM315	c.g. gabbroic anorthosite	515974	7082184	15	54.00	0.18	23.90	4.90	0.50	3.9	4.41	0.06	3.31
98TXM299	mafic dike	515792	7081179	15	52.40	1.50	13.10	16.60	3.90	11.5	14.94	0.23	5.14
98TXM306a	f.g. gt-gabbroic anorthosite	515589	7082260	15	52.00	0.21	24.20	5.90	1.00	4.4	5.31	0.08	3.01
98TXM242a	gt-anorthositic gabbro	515839	7083443	15	51.00	0.47	27.10	4.60	0.90	3.3	4.14	0.05	1.33
98TXM242b	opx-bearing mafic dyke	515839	7083443	15	45.50	0.54	17.10	10.50	3.80	6.0	9.45	0.18	11.90

note: negative numbers denote below detection; XRF= X-ray fluorescence; Titre=wet chemical titration; ICPEs = Inductively coupled emission spectroscopy; ICPMS = Inductively coupled mass spectrometry; gt= garnet; opx= orthopyroxene; c.g.= coarse grained; f.g.= fine grained.

sample	CaO	Na2O	K2O	P2O5	LOI	Cr	Ni	Co	Sc	V	Cu	Pb	Pb	Zn	Mo
Analysis method	XRF	XRF	XRF	XRF		ICPES	ICPES	ICPES	ICPES	ICPES	ICPES	XRF	ICPMS	ICPES	ICPMS
98TXM205	6.98	4.50	1.05	0.20	0.2	170	87	35	21.0	180	110	10	5.0	82	0.6
98TXM332	11.62	4.80	0.22	0.02	0.2	0.0	0.0	0.0	0.5	0.0	0.0	2	3.0	0.0	0.2
98TXM333	10.81	1.70	0.21	0.07	0.3	120	79	67	45.0	510	300	4	0.0	110	0.4
98TXM334	10.07	3.70	0.49	0.06	0.4	89	82	41	18.0	73	0.0	2	0.0	63	0.3
98TXM335	3.85	4.90	1.38	0.09	0.9	16	0.0	9	3.1	41	10	11	9.0	54	0.4
98TXM342	12.14	2.00	0.08	0.08	0.1	240	100	55	45.0	340	140	1	0.0	85	0.6
98TXM345	11.66	4.50	0.49	0.02	0.5	19	36	6	1.4	0.0	0.0	1	3.0	7	0.3
98TXM346	11.37	4.80	0.25	0.01	0.1	0.0	0.0	0.0	0.6	0.0	0.0	1	2.0	10	0.3
98TXM347a	11.31	2.00	0.27	0.08	0.4	110	88	60	50.0	410	330	3	0.0	120	0.6
98TXM347b	10.69	2.00	0.12	0.08	0.0	73	72	70	49.0	500	280	2	0.0	120	0.5
98TXM348	12.60	4.40	0.17	0.02	0.4	13	13	0.0	1.9	0.0	0.0	0.0	2.0	7	0.3
98TXM354	12.49	4.50	0.34	0.01	0.2	12	0.0	0.0	0.9	0.0	0.0	0.0	0.0	6	0.3
98TXM322	16.24	2.30	0.07	0.01	0.3	0.0	11	0.0	0.9	0.0	15	3	4.0	9	0.5
98TXM313	11.68	2.00	0.11	0.06	0.1	220	89	57	42.0	300	180	4	0.0	86	0.4
98TXM306b	10.90	4.70	0.30	0.05	0.2	0.0	28	20	10.0	130	0.0	4	0.0	25	0.6
98TXM315	9.52	3.90	0.48	0.02	0.3	76	62	22	7.7	42	16	4	3.0	38	0.3
98TXM299	9.72	2.30	0.46	0.13	0.0	39	63	67	44.0	410	190	7	3.0	120	1.0
98TXM306a	10.63	4.10	0.33	0.03	0.3	40	55	25	12.0	48	44	4	3.0	39	0.8
98TXM242a	11.79	4.00	0.33	0.06	0.2	16	42	19	7.7	50	47	4	0.0	32	0.9
98TXM242b	13.51	1.20	0.06	0.02	0.2	570	310	57	37.0	240	67	3	0.0	54	0.5

note: negative numbers denote below detection; XRF= X-ray fluorescence; Titre=wet chemical titration; ICPES = Inductively coupled emission spectroscopy; ICPMS = Inductively coupled mass spectrometry; gl= garnet; opx= orthopyroxene; c.g.= coarse grained; f.g.= fine grained.

sample	Ag	Rb	Rb	Cs	Ba	Sr	Sr	Tl	Ga	Ga	In	Hf	Nb	Ta
Analysis method	ICPMS	XRF	ICPMS	ICPMS	ICPMS	XRF	ICPES	ICPMS	XRF	ICPMS	ICPMS	ICPMS	XRF	ICPMS
98TXM205	0.3	8	5.7	0.57	930	724	730	0.05	18	18	0.12	1.9	7	6.0
98TXM332	0.0	30	30.0	7.90	270	535	540	0.33	18	19	0.06	0.2	4	0.2
98TXM333	0.6	6	3.9	1.60	33	102	98	0.07	19	18	0.07	1.4	6	4.0
98TXM334	0.3	4	2.6	0.69	360	363	360	0.0	16	17	0.0	1.0	3	0.5
98TXM335	0.3	59	63.0	3.90	450	323	330	0.54	20	21	0.0	3.4	5	3.5
98TXM342	0.0	3	1.3	0.48	0.0	74	67	0.04	18	16	0.10	1.4	6	3.4
98TXM345	0.0	7	4.9	2.00	260	472	490	0.05	17	19	0.0	0.5	4	0.2
98TXM346	0.0	15	15.0	4.30	150	460	470	0.14	20	22	0.0	0.2	4	0.2
98TXM347a	0.2	12	12.0	0.45	37	81	80	0.14	17	17	0.08	1.6	5	3.5
98TXM347b	0.2	5	2.8	1.00	55	88	78	0.07	19	19	0.07	1.8	5	3.9
98TXM348	0.0	6	3.6	6.40	140	430	440	0.05	19	21	0.0	0.1	5	1.2
98TXM354	0.0	4	1.3	0.42	110	439	440	0.02	20	22	0.0	0.1	4	0.2
98TXM322	0.1	4	2.8	3.70	85	206	210	0.04	18	21	0.0	0.1	0.0	0.4
98TXM313	0.2	4	2.1	0.74	77	92	88	0.04	17	17	0.07	1.3	5	2.7
98TXM306b	0.0	4	1.6	1.20	120	413	410	0.02	23	24	0.0	0.6	5	1.4
98TXM315	0.0	37	39.0	17.00	160	419	420	0.38	19	21	0.0	0.2	4	0.5
98TXM299	0.3	13	12.0	0.45	170	166	160	0.08	18	20	0.11	3.1	0.0	6.5
98TXM306a	0.0	18	18.0	11.00	150	384	380	0.21	19	21	0.0	0.4	0.0	0.7
98TXM242a	0.0	3	0.7	0.19	130	433	440	0.0	22	22	0.0	0.7	5	2.2
98TXM242b	0.0	5	2.2	1.20	62	118	100	0.06	14	13	0.0	0.6	4	0.3

note: negative numbers denote below detection; XRF= X-ray fluorescence; Titre=wet chemical titration; ICPEES = Inductively coupled emission spectroscopy; ICPMS = Inductively coupled mass spectrometry; gl= garnet; opx= orthopyroxene; c.g.= coarse grained; f.g.= fine grained.

sample	Y		Zr		Th		U		La		Ce		Pr		Nd		Sm		Eu		Gd	
	XRF	ICPMS	XRF	ICPMS	XRF	ICPMS	XRF	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS
98TXM205	18	20.0	73	84	2.2	0.50	0.0	0.08	22.0	47.0	5.60	24.0	4.60	2.10	4.00							
98TXM332	3	0.6	0.0	0.0	0.0	0.03	0.0	0.03	2.9	5.0	0.50	1.9	0.26	0.92	0.16							
98TXM333	23	24.0	50	47	3.3	0.08	1.0	0.03	2.6	7.6	1.20	6.6	2.20	1.10	3.20							
98TXM334	10	8.1	27	32	0.0	0.08	0.0	0.0	4.2	9.4	1.30	6.0	1.50	1.30	1.50							
98TXM335	5	2.7	123	140	0.0	1.40	0.0	0.22	19.0	34.0	3.30	10.0	1.30	0.55	0.85							
98TXM342	22	24.0	55	58	1.4	0.08	0.0	0.04	4.1	11.0	1.70	8.4	2.30	0.91	3.00							
98TXM345	4	0.9	0.0	0.0	0.0	0.06	0.0	0.0	3.0	5.2	0.53	1.9	0.26	0.83	0.19							
98TXM346	3	0.5	0.0	0.0	0.0	0.12	0.0	0.02	2.1	3.7	0.37	1.2	0.14	0.62	0.12							
98TXM347a	24	24.0	55	55	2.6	0.25	0.0	0.12	5.1	13.0	1.80	8.6	2.40	0.96	3.10							
98TXM347b	21	23.0	62	70	3.0	0.20	1.2	0.10	4.0	11.0	1.70	8.5	2.70	1.10	3.70							
98TXM348	4	1.3	0.0	0.0	0.0	0.21	0.0	0.07	2.7	5.0	0.52	2.1	0.32	0.64	0.26							
98TXM354	4	0.6	0.0	0.0	0.0	0.08	0.0	0.0	1.6	2.6	0.28	0.9	0.19	0.71	0.12							
98TXM322	0.0	0.7	0.0	3.1	1	0.08	0.0	0.03	2.6	4.4	0.44	1.5	0.23	0.25	0.16							
98TXM313	20	19.0	43	43.0	1	0.08	0.0	0.02	2.3	7.0	1.20	6.9	2.00	0.77	2.70							
98TXM306b	8	6.0	20	24.0	0.0	0.73	0.0	0.11	7.2	14.0	1.60	6.9	1.20	0.95	1.10							
98TXM315	5	2.3	4	10.0	0.0	0.15	0.0	0.02	3.2	5.7	0.61	2.6	0.47	0.64	0.41							
98TXM299	24	32.0	107	120.0	11	1.10	9.4	0.19	12.0	28.0	3.70	17.0	4.20	1.30	5.00							
98TXM306a	2	4.2	7	16.0	4	0.33	2.8	0.05	3.6	7.1	0.81	3.4	0.67	0.71	0.65							
98TXM242a	10	7.4	22	30.0	0.0	1.30	0.0	0.27	7.9	17.0	1.80	8.3	1.60	0.98	1.40							
98TXM242b	15	13.0	12	16.0	0.0	0.04	0.0	0.0	0.5	2.3	0.51	3.4	1.30	0.51	2.00							

note: negative numbers denote below detection; XRF= X-ray fluorescence; Titre=wet chemical titration; ICPES = Inductively coupled emission spectroscopy; ICPMS = Inductively coupled mass spectrometry; gt= garnet; opx= orthopyroxene; c.g.= coarse grained; f.g.= fine grained.

sample	Tb	Dy	Ho	Er	Tm	Yb	Lu	Be
Analysis method	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS	ICPES
98TXM205	0.66	3.30	1.70	0.70	0.28	1.70	0.26	0.6
98TXM332	0.01	0.10	0.04	0.02	0.01	0.01	0.01	0.0
98TXM333	0.57	3.50	2.30	0.82	0.37	2.40	0.39	0.0
98TXM334	0.23	1.40	0.68	0.30	0.11	0.64	0.11	0.0
98TXM335	0.10	0.44	0.20	0.08	0.03	0.19	0.04	2.3
98TXM342	0.59	3.50	2.10	0.82	0.36	2.30	0.36	0.0
98TXM345	0.03	0.15	0.06	0.03	0.01	0.05	0.01	0.0
98TXM346	0.01	0.06	0.03	0.01	0.01	0.01	0.01	0.9
98TXM347a	0.60	3.70	2.10	0.88	0.37	2.20	0.36	0.0
98TXM347b	0.63	3.90	2.20	0.86	0.35	2.30	0.35	0.0
98TXM348	0.04	0.23	0.12	0.04	0.01	0.10	0.01	0.0
98TXM354	0.02	0.09	0.04	0.02	0.01	0.06	0.01	0.0
98TXM322	0.02	0.12	0.06	0.02	0.01	0.06	0.01	8.6
98TXM313	0.51	3.20	2.00	0.67	0.31	2.10	0.30	0.0
98TXM306b	0.17	0.99	0.58	0.22	0.09	0.52	0.08	0.0
98TXM315	0.06	0.36	0.25	0.08	0.04	0.25	0.04	0.5
98TXM299	0.84	5.40	3.20	1.10	0.45	3.10	0.47	0.5
98TXM306a	0.11	0.63	0.42	0.14	0.06	0.42	0.07	0.0
98TXM242a	0.23	1.30	0.73	0.27	0.11	0.71	0.11	0.0
98TXM242b	0.33	2.30	1.50	0.49	0.21	1.50	0.23	0.0

note: negative numbers denote below detection; XRF= X-ray fluorescence; Titr= wet chemical titration; ICPES = Inductively coupled emission spectroscopy; ICPMS = Inductively coupled mass spectrometry; gt= garnet; opx= orthopyroxene; c.g.= coarse grained; f.g.= fine grained.

REE content of UC rocks (symbols below) vs. KC rocks (grey shading).

- Anorthosites
- ◇ Gabbroic anorthosites
- Gabbro dykes