**GEOLOGY, GEOTHERMOBAROMETRY AND GEOCHRONOLOGY OF THE
HIGH-*P* GRANULITE-FACIES KRAMANITUAR COMPLEX,
WESTERN CHURCHILL PROVINCE**

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ABSTRACT

The Kramanituar Complex is a window of high- P granulite-facies rocks in the interior of the Archean western Churchill Province. It is dominated by a metagabbroic-anorthositic suite of Paleoproterozoic age, with subordinate paragneiss and charnockite of probable primary Archean age. L-S fabrics throughout the complex are generally strong to mylonitic and are interpreted to reflect progressive deformation during general noncoaxial flow at ca. 1.9 Ga. Near-peak equilibrium metamorphic conditions of 12 ± 1.5 kbar and $810 \pm 70^\circ\text{C}$ are bracketed between ca. 1910 Ma, the age of prograde metamorphic monazite in sillimanite- and kyanite-bearing paragneiss, and 1901 Ma, the cooling ages of titanite and rutile in gabbroic and paragneissic rocks. Rocks with coronitic and symplectitic reaction textures record high- T , near-isothermal decompression to ca. 8 kbar and 800°C . The timing of exhumation is tightly bracketed between magmatic crystallization of gabbroic anorthosite under granulite-facies conditions at 1902 Ma and widespread rutile cooling ages of 1901 Ma. Time-averaged cooling rates of $>450^\circ\text{C Ma}^{-1}$ for gabbroic anorthosite and leucogranite reflect rapid unroofing of the complex at 1901 Ma.

The Kramanituar Complex is surrounded mainly by foliated to gneissic Archean rocks. To the north and east, extensive amphibolite- to greenschist-facies granitoid rocks host gabbroic rocks that record high- P metamorphic conditions, similar to those of the Kramanituar Complex. To the south, amphibolite-facies metasedimentary and plutonic rocks show fabric relationships, compositional variation and isotopic evidence indicative of an older, mainly Neoproterozoic prograde tectonometamorphic history.

The Kramanituar Complex is a window into largely anhydrous, Paleoproterozoic deep-crustal rocks whose metamorphic history reflects magmatic heat superimposed on a regional metamorphism. The complex and country rocks to the north are believed to have been tectonically juxtaposed with rocks that preserve older fabrics and metamorphic assemblages, exposed to the south. Tectonic unroofing via extensional faulting is indicated by rapid cooling of the complex at ca. 1.9 Ga, and by juxtaposition of the Kramanituar Complex with country rocks to the south that appear unaffected by ca. 1.9 Ga high-grade tectonometamorphism.

The Kramanituar Complex may represent a segment of the Archean Snowbird tectonic zone that was reactivated during Paleoproterozoic time, possibly in response to the ca. 1.97-1.9 Ga Thelon Orogen.

ACKNOWLEDGMENTS

This research results from six months of field work and extensive laboratory work that was supported and funded by the Geological Survey of Canada. It was carried out at Carleton University and the Geological Survey of Canada under the supervision of Simon Hanmer, Sharon Carr and Rob Berman. The project was partially funded by Natural Sciences and Engineering Research Council of Canada (NSERC) Operating Research Grants to Simon Hanmer and Sharon Carr, and by Ontario Graduate Study (OGS) postgraduate scholarships, a Geological Society of America (GSA) research grant and financial awards from Carleton University to Mary Sanborn-Barrie. I thank Simon Hanmer for his support, assistance and keen interest at every stage of this study, particularly for thought-provoking discussions of western Churchill geology in the field and office. His constructive criticism has been invaluable to the formulation of ideas during this research and has led to significant improvements in the content, style and format of this thesis. Sharon Carr instructed me in all stages of U-Pb geochronology and provided support and helpful advice in this research and in all aspects of graduate research over the course of this study. Others whose assistance in geochronological aspects of this study I gratefully acknowledge include fellow Carleton colleague Jim Crowley, and G.S.C. scientists Bill Davis, Vicki McNicoll, Reg Thériault and Klaus Santowski. I thank Rob Berman for introducing me to his approach to geothermobarometry, and for many stimulating discussions regarding mineral equilibria, timing constraints on tectonometamorphism in the interior of the western Churchill Province, and the elusive link between P - T estimates and time. Rob also arranged access to microprobe and scanning electron facilities at the Geological Survey of Canada. I am grateful to Mike Williams for sharing his insights into the value of detailed compositional maps in P - T - t studies, his guidance and hospitality in obtaining maps of thin sections for this study, and his constructive comments regarding the data and interpretations presented in this thesis. Annette Briggs and Mike Jackson prepared thin sections; John Stirling, Gordon Pringle, Katherine Venance, Pat Hunt and Peter Bélanger provided assistance with sample analyses using G.S.C. analytical facilities; and Rachelle Lacroix, Sarah McMullen, Catherine Haines and Rob Vanderkam very kindly provided assistance in the final stages of presentation of this thesis. In the field, capable geological assistance and appreciated companionship was provided by Peter Schaub in 1992 and by Julie Watson and Ralph Kownak in 1993. Logistical support during fieldwork was provided by Continental Polar Shelf, Boris and Liz Kotelewitz and family, Peter Tapatai and many other residents of Baker Lake. My thanks to Jim Crowley, Tucker Barrie, Richard Brown, Bill Davis, Tom Skulski, David Corrigan, Lizzie Ann Dunworth and many others for sharing their observations, experiences and for discussion. To Tucker and Hannah Barrie, and our families, my gratitude for their constant encouragement and support.

ORIGINAL CONTRIBUTION

This research focusses on further understanding the crustal-scale Snowbird tectonic zone through a multi-disciplinary investigation into the geology, structure and P - T - t history of the Kramanituar Complex, western Churchill Province. Research involved six months of bedrock mapping during 1992 and 1993, during which time I carried out lithological and structural mapping as well as sampling of an area encompassed by 1:50 000 Rio Island sheet 56D/1 and parts of adjacent sheets 56D/2, 56C/4, 55M/16 and 55N/13. The area was previously mapped by Mikkel Schau of the Geological Survey of Canada (Schau and Ashton, 1980; Schau, 1983).

Petrographic and geothermobarometric data from the Kramanituar Complex and its country rocks, presented here, provide the first quantitative P - T constraints on high-grade tectonometamorphism within the Snowbird tectonic zone. This involved the analysis of mineral assemblages from twenty samples on the Geological Survey of Canada microprobe. These analyses formed the basis for estimates of P - T conditions, calculated using the TWEEQU (Thermobarometry With Estimation of EQUilibrium state) software (Berman, 1991), in consultation with Rob Berman.

Nine samples were selected for geochronological studies, in order to provide constraints on a decades-long debate regarding the timing and extent of tectonometamorphic reworking of the Churchill Province. Of these samples, seven were analysed using the U-Pb isotopic system and two garnet-bearing samples were analysed using the Sm-Nd system. I performed all stages of sample preparation for these samples including crushing and grinding, mineral separation and selection. I assisted Sharon Carr, who performed chemistry for all zircon and monazite fractions. Chemistry on rutile, titanite for U-Pb analyses and on garnet, plagioclase and clinopyroxene for Sm-Nd analysis was done by scientists of the Geological Survey of Canada. Mass spectrometry was performed by Sharon Carr at Carleton University and by Klaus Santowski and Reg Thériault at the Geological Survey of Canada.

An investigation of the geochemical character of selected rocks from the study area was designed to characterize the Kramanituar gabbroic suite and country rocks gabbros, and to investigate a genetic link between Kramanituar gabbro and gabbroic anorthosite: aspects of the study that could not be resolved by other methods. Major and trace element analyses on selected samples were done at the Geological Survey of Canada. Treatment and interpretation of the geochemical data presented in Chapter 4 was done in consultation with Tucker Barrie.

I compiled and wrote each chapter of this thesis and received comments regarding content, style and format from Simon Hanmer, Sharon Carr and Rob Berman. The thesis generally incorporates critical comments on content as suggested by these advisors.

METHOD OF PRESENTATION

This thesis takes a form which includes three major papers (Chapters 1, 2, and 3) that will be published as journal articles. This format is intended to facilitate the publication of this research, however, it introduces some unavoidable redundancies for which I apologize in advance. My intent when submitting these chapters to scientific journals is to be first author on all three papers in co-authorship with Simon Hanmer on Chapter 1, with Rob Berman on Chapter 2, and with Sharon Carr and Reg Thériault on Chapter 3. Chapter 5 may form the basis for a synthesis paper co-authored with Simon Hanmer. Chapter 4 has been published in Geological Survey of Canada Paper 1999-E. Two preliminary progress reports on field relationships in the study area were published in Geological Survey of Canada Paper 93-1C and 94-1C, and are included here as Appendix I and II.

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CHAPTER 1

Geology of the Kramanituar Complex: Paleoproterozoic deep-crustal rocks in the Archean western Churchill Province, Canada

ABSTRACT

The Kramanituar Complex coincides with the northeastern part of the ~2000 km long Snowbird tectonic zone, a geophysically-defined structure whose role in the tectonic evolution of the western Churchill Province remains controversial. The complex is dominated by a high-*P* granulite-facies metagabbroic suite of Paleoproterozoic age, with minor supracrustal rocks and charnockite of probable primary Archean age. It is surrounded mainly by Archean amphibolite-facies rocks, however, discontinuously exposed mafic rocks to the north also record Paleoproterozoic high-*P* conditions.

Mafic metaplutonic rocks within the complex preserve two generations of fabrics that reflect progressive deformation during general noncoaxial flow at ca. 1.9 Ga. Northwest-striking *S*₁ foliation, best preserved in the central part of the complex, is intensified in panels of penetratively recrystallized rock with strongly developed *L*₂/*S*₂ fabrics. Steeply dipping, granulite facies ribbon mylonites of the Northern shear zone form two fault segments along the northern and eastern margins of the complex and mainly record sinistral, general noncoaxial flow. In contrast, foliated to gneissic country rocks display a moderately developed, northeast-striking foliation that may be mainly Archean. This is demonstrated south of the complex by supracrustal rocks whose northeast-striking regional foliation is cut by the dextral, oblique-slip Southern shear zone, dated at ca. 2550 Ma.

The Kramanituar Complex is interpreted to represent a tectonic lens or crustal-scale boudin of mainly Paleoproterozoic rocks, contained within a larger block of Archean crust. It preserves a record of intrusion of mantle-derived mafic magmas at deep-crustal levels during noncoaxial shortening, tectonic exhumation to mid-crustal levels and very rapid cooling at ca. 1.9 Ga. Paradoxically, the two major shear zones within the study area did not accommodate the major displacements responsible for exhumation of the complex. However, a break between penetrative Paleoproterozoic tectonometamorphic fabrics within the Kramanituar Complex and those of Archean age to the south requires a cryptic south-dipping normal fault that likely was related to ca. 1.9 Ga exhumation.

Rocks along the surface trace of the Snowbird tectonic zone have similarities to the Kramanituar Complex in terms of lithology, structural style, and metamorphic grade. However, latest penetrative tectonometamorphism at ca. 2.6 Ga within its regional

northeast-striking segment contrasts with penetrative ca. 1.9 Ga activity in its regional east-striking segment, as documented at the Kramanituvar Complex, indicating that these regional segments were separate structural entities during Paleoproterozoic time.

INTRODUCTION

Prominent high-amplitude linear anomalies in horizontal gravity gradient (Fig. 1-1a) and magnetic anomaly maps of Canada (Goodacre et al., 1987a; GSC, 1987) extend from central Alberta to the northwest shore of Hudson Bay. These correspond to discontinuous occurrences of granulite-facies tectonites of similar lithology and structural style and collectively define the Snowbird tectonic zone (Fig. 1-2; Hoffman, 1988). There are two end-member views regarding the geological nature of the Snowbird tectonic zone within the western Churchill Province: 1) that it represents a Paleoproterozoic suture zone which welds two Archean continents, the Rae to the northwest and the Hearne to the southeast (Walcott and Boyd, 1971; Thomas et al., 1988; Hoffman, 1988; Ross et al., 1991; Ross et al., 1995); and, 2) that it reflects a predominantly intracontinental structure that was active during Archean (Hanmer et al., 1994; Hanmer, 1997a) or Paleoproterozoic (Lewry and Sibbald, 1980; Thomas and Gibb, 1985) time.

The Kramanituvar Complex is a 850 km² exposure of variably strained, granulite-facies, metaplutonic and metasedimentary rocks. It coincides with one of the highest-amplitude anomalies in the horizontal gravity gradient along the trace of the Snowbird tectonic zone (Fig. 1-1a), and is underlain by an extensive positive (+60 mGal) Bouguer gravity anomaly (Fig. 1-1b; Goodacre et al., 1987b). The Kramanituvar Complex is located at the junction of the regional northeast- and east-striking segments of the Snowbird tectonic zone (Fig. 1-2), a position which is pivotal to establishing the relationship between these segments, and to addressing the question of whether the Snowbird tectonic zone is a continuous trans-Churchill structure.

This paper is part of an integrated geologic, thermobarometric and geochronologic study that provides constraints on the magmatic, deformational and meta-

morphic history of the Kramanituar Complex and surrounding rocks. It presents a comprehensive description of deep-crustal units exposed in the interior of an Archean craton, and a structural analysis of their geometrical and kinematic aspects. An important conclusion drawn from this analysis is that shear zones within the study area, including a spectacular granulite-facies mylonite belt, did not accommodate major uplift and exhumation. First-order geological relationships presented in this paper provide the framework on which geothermobarometry (Chapter 2) and U-Pb and Sm-Nd geochronology (Chapter 3) are based, and from which inferences are drawn regarding the relationship of the Paleoproterozoic Kramanituar Complex to the Snowbird tectonic zone, and to the evolution of the Archean western Churchill Province.

REGIONAL GEOLOGICAL SETTING

The western Churchill Province (Fig. 1-2) is largely underlain by Archean crust, including Paleo- to Neoarchean (ca. 3.5 to 2.6 Ga) foliated to gneissic metaplutonic rocks, and linear belts of Neoarchean (ca. 2.9 to 2.6 Ga) metavolcanic and meta-sedimentary rocks (Aspler and Chiarenzelli, 1996). Paleoproterozoic rocks of the province include erosional remnants of a ca. 2.45-2.1 Ga intracontinental platformal clastic sequence represented by the Hurwitz (Hearne) and possibly Amer (Rae) groups (Hoffman, 1988; Aspler et al., 1994); ca. 1.83 Ga alkalic volcanic rocks (Christopher Island Formation) and dykes (Peterson, 1994); ca. 1.83 Ga granites; and an extensive suite of ca. 1.78-1.74 Ga granites (Nueltin suite) and related rhyolites (Wright, 1967; LeCheminant et al., 1981; Loveridge et al., 1988).

Two major Paleoproterozoic orogenic events are known to have affected the western Churchill Province (Hoffman, 1988). The Thelon Orogen is the product of eastward subduction beneath the Churchill Province at ca. 2.0-1.97 Ga (van Breemen et al., 1990; van Breemen et al., 1992) which led to collision with the Archean Slave craton at ca. 1.97 Ga (Tirrul and Grotzinger, 1990; Hanmer et al., 1992), with further indentation

until ca. 1.84 Ga, partly attributed to collisional events along the western margin of the Slave Province (Hoffman and Bowring, 1984). The Trans-Hudson Orogen is the product of subduction beneath the western Churchill Province at 1.885 Ga, which led to terminal collision with the Archean Superior craton at ca. 1.83 Ga and post-collisional shortening until ca. 1.7 Ga (Stern et al., 1995; Lucas et al., 1996). The Kramanituvar Complex is situated in the interior of the western Churchill Province, nearly equidistant from the Thelon and Trans-Hudson orogens (Fig. 1-2).

Previous Work

The name "Kramanituvar complex" (Schau et al., 1982; Schau, 1983) was given to granulite facies rocks, including a gabbro-anorthosite suite, exposed near the east end of Baker Lake to distinguish them from regionally extensive amphibolite facies rocks. The name is derived from "kramaniktuark", the Inuit word for "big lake", referring specifically to Baker Lake. The Baker Lake region was initially mapped by Wright (1967) who identified an anorthosite body characterized by massive, coarse-grained textures in its western part, and well-banded, shear-related structures throughout its eastern part. Detailed mapping of these rocks was undertaken by Schau and Hulbert (1977) and Schau and Ashton (1979), who described a central gabbro-norite-anorthosite suite with adjacent well-layered, granulite-facies gneiss. Recognition of rare cumulate features in gabbroic anorthosite, and of magmatic plagioclase of labradoritic composition, led these workers to interpret the rocks as a sill complex, intruded at shallow crustal levels with subsequent tectonic burial to mid-crustal (medium-*P*) levels followed by exhumation. Gneiss associated with the mafic suite has been interpreted as either metasomatized paragneiss, on account of its persistent layering (Schau and Hulbert, 1977), or orthogneiss (Jak, 1977). The complex was interpreted to be faulted against, and intruded by, plutonic rocks to the north and east, and in fault contact with supracrustal rocks to the south (Schau and Hulbert, 1977; Schau and Ashton, 1979).

Both the Kramanituar Complex and its country rocks were interpreted by Schau (1980, 1982) to be Archean, based on U-Pb (zircon) ages of $2573 \pm 28/-5$ Ma from charnockite gneiss from the center of the complex, and $2675 \pm 33/-11$ Ma from granodiorite gneiss north of the complex (located in Fig. 1-3). A K-Ar age (biotite) of 1895 ± 44 Ma was also reported for the granodiorite north of the complex (Schau, 1980). These dates are included in a summary of age data for this area (Table 1-1), obtained largely during this study.

DEEP-CRUSTAL ROCKS OF THE KRAMANITUAR COMPLEX

The Kramanituar Complex is a window of deep crustal rocks, dominated by a Paleoproterozoic gabbroic-anorthositic suite that intruded felsic to intermediate metaplutonic rocks and paragneiss of probable Archean primary age. Penetrative tectonometamorphism of the complex at ca. 1910-1900 Ma, synchronous with magmatic emplacement, has resulted in highly strained, migmatitic rocks with gradational tectonized contacts within, and between, major units.

Metagabbroic Suite

The Kramanituar metagabbroic suite (Fig. 1-3) is dominated by gabbroic anorthosite, dated at 1902 ± 1.6 Ma (Table 1-1; Chapter 3), which occupies the central part of the complex. Gabbroic anorthosite $\pm$ norite is composed of sodic labradorite, orthopyroxene (up to 8 cm locally) and clinopyroxene, with minor pargasitic amphibole $\pm$ garnet and accessory magnetite $\pm$ ilmenite. In their least strained state, these rocks are coarse-grained with up to 60% phenocrysts of plagioclase (Plate 1-1a). Steeply dipping, centimetre- to metre-scale primary magmatic layering, defined by variable proportions of mafic minerals is present locally (Plate 1-1b), although typically these rocks are thoroughly recrystallized and show moderate to strong amounts of strain (Plates 1-1c,d).

Gabbro and leucogabbro consist of plagioclase, clinopyroxene and up to 20% garnet, with minor orthopyroxene, magnetite and quartz (Fig. 1-3, Plate 1-1e). These rocks can be shown to be cogenetic with ca. 1902 Ma gabbroic anorthosite on the basis of their trace element geochemistry (Chapter 4). Gabbro and leucogabbro typically display subtle metre-scale compositional layering which, in part, may be magmatic. Layering also results from concordant, centimetre-scale veinlets or intrusive sheets of granite $\pm$ charnockite (described below) which cut these rocks. Layers of gabbro, 1 to 100 cm wide, within gabbroic anorthosite (Plate 1-1f) are interpreted as intrusive sheets on the basis of their discrete contacts, uniform composition, and locally, by low-angle crosscutting relationships between gabbroic layers and magmatic layering (and/or foliation) in host gabbroic anorthosite.

Rocks of true anorthosite composition (colour index <10) comprise <5% of the suite. They mainly occur as continuous, tabular, 10 m wide concordant horizons in gabbro and paragneiss immediately north of gabbroic anorthosite and probably represent sills. Subangular, metre-scale blocks of fine-grained anorthosite within gabbroic anorthosite also occur (Fig. 1-3, locality A).

Paragneiss

Paragneiss comprises two elongate units north and south of gabbroic anorthosite, and occurs as narrow, steeply dipping lenses in the northeastern part of the complex (Fig. 1-3). The primary depositional age of these rocks is unknown, however, they yield ca. 1910-1908 Ma monazite (Table 1-1), an age interpreted as the time the rocks experienced prograde metamorphism (Chapter 3). These rocks consist of fine-grained, rusty weathering, centimetre- to metre-wide layers and lenses that alternate with white-weathering quartzo-feldspathic layers, 1-30 cm wide (Plate 1-2a). Rusty-weathering layers are composed mainly of plagioclase, quartz (5-15%), lavender garnet ($\text{Alm}_{48-60}\text{Prp}_{30-46}\text{Grs}_{5-12}\text{Sps}_{1-2}$) and minor biotite. They commonly contain graphite, sillimanite, K-feldspar,

pyrite and rutile, and rarely orthopyroxene. The occurrence of kyanite is restricted to narrow lenses of paragneiss in the northeastern part of the complex (locality **B**; Fig. 1-3), whereas, sapphirine occurs in pelite only in the southeastern part (Schau, 1982; see Fig. 1-4). White-weathering leucocratic layers, typically forming 2% to 20% (up to 60%) of paragneiss consist of plagioclase, quartz (20-30%), up to 10% K-feldspar and garnet (<10%) $\pm$ sillimanite $\pm$ kyanite with minor graphite.

Evidence for primary sedimentary structures, such as grading, is absent or ambiguous in these rocks, which generally show moderate to very high strain (Plate 1-2*b,c*). However, mineralogical and textural criteria indicate that these rocks are paragneisses that have experienced variable degrees of partial melting. Mineralogical criteria include the widespread occurrence of graphite and sillimanite in these rocks, and local occurrence of kyanite and sapphirine. Partial melting is reflected by the close association (centimetre-scale) between aluminous (sillimanite or kyanite) leucocratic layers (leucosome) and intervening rusty-weathering, graphite- and aluminosilicate-bearing rocks which are mineralogically consistent with a compositionally immature (quartz-poor) feldspathic wacke protolith (paleosome). Locally, layers containing up to 50% kyanite and 25% garnet, interpreted as restite, may alternate with kyanite-bearing leucosome (Plate 1-2*c*), in accordance with *in situ* partial melting. More commonly, metawacke and metagabbro alternate with cm-scale layers of garnet- rutile- graphite-bearing leucogranodiorite which may represent injected partial melt material mobilized from its source region.

Felsic to Intermediate Plutonic Rocks

Granitic Rocks

Granitic rocks are a minor, but widespread, component of the complex. They occur as centimetre- and metre-scale veins and sheets that cut gabbro and paragneiss. These veins are mainly concordant to tectonic foliation, however, a second set of ca. 1902 Ma (Table 1-1) granitic veins cut the dominant tectonic foliation at a high angle,

and are interpreted to be late-tectonic (Chapter 3). Both sets of veins are leucogranitic to leucogranodioritic in composition with up to 10% biotite, 5% garnet, minor magnetite $\pm$ pyrite.

Granitic/charnockitic rocks of probable Archean age occur sporadically throughout the complex and also form coherent bodies along the east shore of Baker Lake (locality C on Fig. 1-3). An Archean age for these rocks is inferred from charnockite gneiss from the center of the complex (Fig. 1-3) whose age of ca. 2573 Ma (Schau, 1980) indicates the presence of Archean metaplutonic rocks into which the Kramanituur Complex gabbroic suite was intruded. On the east shore of Baker Lake, granitic rocks are cut by 5-20 m wide moderately foliated dykes of magnetite-bearing gabbro and leucogabbro, which limit the exposure of granitoids to narrow panels less than 3 m wide. Given their close spatial proximity, the mafic dykes are likely related to the Kramanituur gabbroic suite, however, this has not been proven.

Mixed Gneiss, Diatexite and Diatexite Mylonite

Pale buff- to rust-weathering quartzo-feldspathic rock designated as "mixed gneiss" (Fig. 1-3) is a composite layered rock consisting of two units described above: 1) gabbro ($\pm$ leucogabbro $\pm$ quartz diorite), interlayered at a cm-scale with 2) granite ($\pm$ charnockite) (Plate 1-3a). The charnockitic component may be similar to that previously dated at 2573 Ma (Schau, 1980). A strain gradient is observed from readily identifiable mixed gneiss to increasingly homogeneous, subtly layered, migmatitic, garnet-bearing quartzo-feldspathic rocks (Plates 1-3b,c), designated as diatexite (Brown, 1973). Equivalent rock types with quartz and feldspar ribbons are named diatexite mylonite.

Diatexite and diatexite mylonite are compositionally similar to paragneiss but they do not contain sillimanite, kyanite, graphite or sapphirine. Where strongly deformed, these rocks are further distinguished from paragneiss by their medium to coarse grain size, and pitted appearance created by weathering of round garnet porphyroblasts

(1mm to 2cm), draped by sigmoidal ribbons of quartz and feldspar. Contacts between the diatexite, and paragneiss or gabbro, are typically subtle and diffuse.

COUNTRY ROCKS TO THE KRAMANITUAR COMPLEX

North and east of the Kramanitar Complex, quartzo-feldspathic plutonic rocks and minor gabbroic rocks are exposed. Although these rocks may be mainly Archean in age, *P-T-t* constraints from the gabbros within this granitoid terrain demonstrate that, like the complex, this region was at deep-crustal conditions at ca. 1.9 Ga. South of the Kramanitar Complex, Archean-age supracrustal rocks, cut by granites and gabbros, preserve older tectonometamorphic fabrics and assemblages, and lack penetrative ca. 1.9 Ga reworking. Sub-greenschist facies clastic sedimentary rocks with a minimum age of ca. 1832 Ma unconformably overlie the complex at its western margin, and attest to the rapid exhumation experienced by the complex, from deep-crustal conditions at ca. 1.9 Ga.

Metaplutonic rocks

Quartzo-feldspathic metaplutonic rocks of intermediate to felsic composition are widespread north and east of the complex, and are dated at ca. 2675 Ma (Schau, 1980; Table 1-1) at one locality (Fig. 1-3). Hornblende $\pm$ biotite diorite is most common and is typically cut by centimetre-scale veins and metre-scale sheets of granodiorite. These rocks are weakly to moderately foliated, or may possess an irregular gneissic texture (Plate 1-4a) defined by curvilinear mafic schlieren and variably-oriented granitoid veins. Diorite gneiss is garnet-bearing and displays leucocratic segregations indicative of partial melt textures within a 1 km wide by 10 km long zone adjacent to the eastern margin of the complex (Fig. 1-4). These features are interpreted to represent a contact thermal aureole imposed on the dioritic country rocks during intrusion of the Kramanitar gabbroic suite at 1902 Ma.

Gabbroic rocks within the granitoid terrain form discrete intrusive stocks (Fig. 1-3) of pre- and post-exhumation vintage. These include a ca. 2200 Ma (Chapter 3) coronitic metagabbro north of the complex (locality **D**, Fig. 1-3), which records Paleoproterozoic high-*P* conditions (Chapter 2), and a prehnite-bearing gabbro (locality **E**, Fig. 1-3) that may be part of the ca. 1825 Ma sub-greenschist facies alkalic suite (Peterson, 1994). Along the northern margin of the Kramanituar Complex, gabbroic rocks with up to 12 wt% MgO and 5% sapphirine (Fig. 1-4) may represent peraluminous high-Mg dykes or sills, several tens of metres wide.

Granitic Rocks

Northwest of the complex, foliated K-feldspar porphyritic syenogranite (Fig. 1-3) is cut by quartz veins with associated pyrite $\pm$ chalcopyrite $\pm$ fluorite. South of the complex between North and South inlets, moderately strained granite and charnockitic leucogranite are interlayered with dioritic to quartz dioritic rocks which they cut and are, in turn, cut by magnetite-bearing mafic dykes.

Quartz Diorite

Magnetite-bearing leucocratic quartz diorite ($\pm$ diorite) exposed along the North Channel of Chesterfield Inlet (Fig. 1-3) consists of alternating plagioclase-rich and mafic-dominated (orthopyroxene, amphibole) layers which are granoblastic. These rocks occur as straight (Plate 1-4b) and irregular orthogneisses, intruded by gabbro and leucomonzogranite veins. Granoblastic textures are characteristic of Archean-age granitoid gneiss and mylonite gneiss south of the complex (*see* Plate 1-8b), but are absent in mafic metaplutonic rocks of the Kramanituar Complex. Accordingly, magnetite-bearing quartz diorite ($\pm$ diorite) gneiss is not interpreted as part of the Kramanituar metagabbroic suite, but is suspected to represent part of a plutonic complex of probable Archean age.

Anorthosite

Restricted exposures of white, recrystallized anorthosite occur south of Chesterfield Inlet (Fig. 1-3). These rocks are interpreted as Archean country rocks, rather than part of the Kramanituur gabbroic suite because they carry the regional Neoarchean foliation ($S_{1(\text{External})}$; described below) which characterizes country rocks south of the complex. In addition, xenoliths of recrystallized anorthosite are hosted by diorite gneiss, itself interpreted to represent Archean country rocks (described above).

Supracrustal rocks

Amphibolite facies metasedimentary rocks are widespread south of the Kramanituur Complex (Fig. 1-3) and consist of moderately to strongly foliated semi-pelite and pelite. A U-Pb (zircon) magmatic crystallization age of ca. 2550 Ma (Table 1-1) from syntectonic syenogranite pegmatite that cuts the clastic rocks provides a minimum age of deposition for them. They are composed of biotite, plagioclase, quartz (<20%) and up to 50% garnet ($\text{Alm}_{72-77}\text{Prp}_{17-20}\text{Grs}_4\text{Sps}_1$) with sillimanite as coarse-grained porphyroblasts or fine-grained prismatic crystals. Grunerite-bearing magnetite ironstone and magnetite-bearing metawacke occur as 1-100 cm wide units at several localities. These clastic metasedimentary rocks commonly show features consistent with incipient anatexis, such as the pervasive presence of quartzo-feldspathic aggregates and discontinuous layers (<1 cm wide) rimmed by biotite foliae (Plate 1-4c). Mafic dykes, 2 to 200 m wide, cut these rocks and contain irregular plagioclase-rich patches and veins, which locally contain coarse (1-2 cm) clinopyroxene crystals and garnet porphyroblasts. Concentrations of leucocratic material at interboudin necks of mafic dykes (Plate 1-4d) are consistent with migmatization of metasedimentary rocks synchronous with their deformation ($D_{2(\text{External})}$).

Metavolcanic rocks are exposed in a restricted zone along the South Channel of Chesterfield Inlet (Fig. 1-3) where they are tightly infolded with metasedimentary rocks.

These include garnet-bearing amphibolitic rocks with dark anastomosing layers, 1-2 cm wide, that are interpreted as relict selvages in a pillowed mafic flow; massive and brecciated (flow top?) mafic flows; and intermediate pyroclastic breccia. A rhyodacite fragment from dacitic pyroclastic breccia (Plate 1-4e) yielded a U-Pb (zircon) age between ca. 2655 Ma and ca. 2691 Ma (Davis and Sanborn-Barrie, unpublished data, 1998). Fine-grained, layered, tuffaceous rocks of andesitic to dacitic composition appear to be transitional upwards with epiclastic/clastic rocks.

Sub-greenschist facies conglomerates and sandstones of the Baker Lake Group (Gall et al., 1992) unconformably overlie the Kramanituur Complex (Figures 1-2, 1-3). The lowermost part of this succession includes the basal, clast-supported South Channel conglomerate which contains clasts of weakly- to strongly-foliated gabbroic anorthosite, gabbro and mixed gneiss, apparently derived from the Kramanituur Complex. A maximum age of deposition of this conglomerate is 1901 Ma, the time at which the western part of the complex cooled through ca. 400°C (Table 1-1). Its minimum age is 1832 ± 28 Ma, a Pb-Pb (monazite-bearing apatite) isochron age (MacRae et al., 1996) of intrusive ultrapotassic rocks of the overlying Christopher Island Formation (Peterson, 1994).

METAMORPHISM OF THE KRAMANITUUR COMPLEX AND COUNTRY ROCKS

Within the Kramanituur Complex, igneous mineralogy has largely been pervasively recrystallized and overprinted by Paleoproterozoic high-*P* mineral assemblages (Chapter 2) that are well-preserved and generally remain unaffected by subsequent retrograde effects. High-*P* assemblages include clinopyroxene-garnet-plagioclase-quartz in mafic metaplutonic rocks and sillimanite-kyanite-garnet in paragneiss (Fig. 1-4). The highest grade rocks presently identified within the complex are mafic granulites from its north and east parts which yield *P-T* conditions of ca. 12 ± 1.5 kbar and $810 \pm 70^\circ\text{C}$ (Chapter 2). Rocks from the central part of the complex preserve disequilibrium reaction

textures from which a near-isothermal (850°-750°C) decompression history from ca. 13 kbar to 8 ± 1 kbar (Chapter 2) is estimated.

Metaplutonic rocks north and east of the Kramanituur Complex are generally characterized by non-diagnostic mineral assemblages (ie., quartz-plagioclase-amphibole-biotite $\pm$ K-feldspar $\pm$ epidote) compatible with amphibolite- to greenschist-facies conditions. However, isolated gabbros occurring within this granitoid terrain, both north (locality **D**, Fig. 1-3) and east of the complex, yield *P-T* conditions similar to estimated peak conditions from within the complex. These data (Chapter 2) and an age constraint on metamorphism north of the complex (Chapter 3) indicate that country rocks north of the Kramanituur Complex were also at deep-crustal conditions during Paleoproterozoic time.

Supracrustal rocks south of the complex are interpreted to preserve an older tectonometamorphic history. Garnet-bearing pelitic rocks record a two-stage prograde metamorphic path with garnet core conditions estimated at ca. 6.5 kbar and 700°C, and garnet rim conditions estimated at 8 ± 1 kbar and 750°C. These garnet crystals are wrapped by planar biotite-sillimanite-quartz-feldspar fabrics of Archean age (described below), and accordingly, are interpreted as part of a Neoarchean metamorphic assemblage (Chapter 2). Mafic dykes that cut semi-pelite locally contain Amp-Cpx-Grt-Pl-Qtz assemblages (abbreviations after Kretz, 1983). Conditions of 12 kbar and 870°C obtained from a gabbroic unit indicate local attainment of granulite facies conditions prior to ca. 2326 Ma, the titanite age of this rock (Chapter 3). Details of the petrology and thermobarometry of these rocks are provided in Chapter 2.

STRUCTURAL GEOLOGY

Tectonic fabric elements of both Archean and Paleoproterozoic age are recognized across the study area (Fig. 1-5). Within the Kramanituur Complex, two generations of foliations, one set of folds and a granulite-facies shear zone (Northern shear zone), are identified in rocks that, in part, crystallized at ca. 1902 Ma (Table 1-1). These structures

are annotated with the subscript_(KC). Fabric elements outside of the complex, annotated with the subscript_(External), are thought to be mainly Archean. South of the complex, these include the regional foliation and an overprinting ca. 2550 Ma dextral shear zone (Southern shear zone); north of the complex, they may include coplanar fabrics in granitoid rocks.

Fabric elements within the Kramanituur Complex

Planar fabrics

Relatively unstrained gabbroic anorthosite in the central part of the Kramanituur Complex contain a weak to moderately developed foliation (Plate 1-1*a,b*) defined by aligned fine-grained mafic mineral grains, and a weakly developed preferred shape fabric in porphyritic phases. This first generation fabric, designated $S_{1(KC)}$, strikes northwest and generally dips steeply (Fig. 1-5*a*). Where modal layering (S_0) is recognized in gabbroic anorthosite, $S_{1(KC)}$ is parallel to it, or lies at a small (2-15°) counterclockwise angle (Plate 1-1*b*). Near the eastern margin of the complex, shallow-dipping foliations may be related to a right-stepping discontinuity in the Northern shear zone (described below). In gabbro, leucogabbro and mixed gneiss, $S_{1(KC)}$ is defined by aligned mafic mineral grains and a preferred orientation of recrystallized aggregates of pyroxene, feldspar and quartz (Plate 1-1*e*). Granitic veins that cut these rocks are generally concordant to $S_{1(KC)}$, and possess a parallel internal shape fabric. (Plate 1-5*d*).

The dominant penetrative planar fabric within the complex is a strongly to intensely developed, west- to northwest-striking foliation which affects all rock types (Plates 1-1*c,d*; 1-2*b*; 1-3*b*; 1-5*a*). This foliation is designated $S_{2(KC)}$ on the basis of overprinting relationships. These include rare, but widespread, occurrences of mesoscopic folds defined by $S_{1(KC)}$ to which $S_{2(KC)}$ is axial planar (Plate 1-5*b*); panels of folded $S_{1(KC)}$ layering bounded by zones of straight, transposed gneisses where the transposition foliation is parallel to the axial planar $S_{2(KC)}$ (Plate 1-5*c*); and, less com-

monly, obliquity between $S_{2(KC)}$ and $S_{1(KC)}$ (Plate 1-5d). In general, however, a composite $S_{1-2(KC)}$ transposition fabric is widespread.

The style and intensity of $S_{2(KC)}$ ($\pm S_{1(KC)}$) varies according to lithologic unit and this is interpreted to have been controlled largely by the mineralogy and relative competency of the material in which it is developed. In gabbroic anorthosite, $S_{2(KC)}$ is a streaky, discontinuous foliation formed by recrystallized aggregates of feldspar and pyroxene with garnet (Plate 1-1c,d). In gabbro and leucogabbro, it is typically manifested as elongate orthopyroxene and/or clinopyroxene aggregates consisting of a central, monocrystalline core surrounded by an elongate polycrystalline mantle that resembles core-and-mantle structure (White, 1976); both are enveloped by discontinuous, mm-wide layers of garnet and recrystallized plagioclase $\pm$ quartz (Plate 1-5a). In paragneiss, $S_{2(KC)}$ is defined by straight, continuous, alternating layers of feldspathic wacke and leucosome (Plate 1-2b), and by ribbons of quartz and aligned mafic minerals within these layers. $S_{2(KC)}$ is near-vertical to steeply south-dipping throughout the complex, except in the southeast where it is moderately ($38-57^\circ$) north-dipping (Fig. 1-5a).

Linear fabrics

Linear fabric elements include aggregates of feldspar and quartz, aligned metamorphic minerals such as sillimanite and kyanite, streaks of graphite, and tight to isoclinal minor F_2 fold axes. These show a consistent shallowly to moderately ($5-35^\circ$) westward plunging orientation (Fig. 1-5b). This extension lineation is observed on $S_{2(KC)}$ and is interpreted as $L_{2(KC)}$, although rarely shallow west-plunging lineations are also developed on $S_{1(KC)}$. Discrete domains of east-plunging lineations are recognized in the northwestern, north-central, and southeastern parts of the complex (Fig. 1-5b), and are interpreted to reflect transpressional strain associated with shearing (described below).

Northern shear zone

Granulite facies mylonitic and ultramylonitic rocks of gabbroic, gabbroic anorthosite and quartzo-feldspathic (paragneiss and diatexite) composition (Plate 1-6) delineate the Northern shear zone, an anastomosing 3-6 km wide zone that can be traced for a strike length of approximately 60 km (Fig. 1-6). A Paleoproterozoic age for movement on this zone (Chapter 3) is based on a U-Pb magmatic crystallization ages of 1902 ± 1.6 Ma for gabbroic anorthosite cut by the shear zone, and 1902.6 ± 1 Ma for a leucogranite dyke that cuts the shear zone, but is itself deformed (ie., late-tectonic). This steeply dipping shear zone forms two fault segments and is geometrically complex. From James Point it strikes easterly across the northern part of the complex, and then bends southeast to southerly around the eastern margin of the Kramanitu Complex. The bend in the northeast part of the shear zone (locality *b* in Fig. 1-6) is interpreted as a primary bend owing to the continuity of mylonitic fabrics, with no evidence of later modification of a once straight zone. The east-striking southeast segment of the shear zone is separated from the rest of the zone by a right-stepping discontinuity (locality *d* in Fig. 1-6) where low strain conditions are consistent with minimal shearing-related displacement.

Strain gradients and optical microstructures indicate that fine-grained rocks with straight, mm- to cm-scale layering within the Northern shear zone are the products of intense progressive $D_{2(KC)}$ deformation within the temperature-sensitive regime dominated by intracrystalline plasticity and mylonitization (Sibson, 1977, 1982). Gabbroic anorthosite with a moderately-developed $S_{2(KC)}$ foliation (Plate 1-1*c*) shows a progressive decrease in grain size attributed to increasing degrees of dynamic recrystallization of plagioclase and clinopyroxene (Plate 1-1*d*), and to cataclastic refinement of garnet. Optical microstructures in these rocks include a seriate texture to plagioclase, quartz and clinopyroxene (Plate 1-7*a*), which form core and mantle structures (Plate 1-7*b*) and mosaics, characterized by a spectrum of coarse to very fine subgrains (Plate 1-7*c*). These textures suggest that rotational recrystallization (Guillope and Poirier, 1979) occurred. Lobate to

amoeboidal grain boundaries of these phases, some of which are strain free, appear to reflect migrational dynamic recrystallization. Confirmation of these deformation mechanisms and thorough assessment of the degree of crystal plastic slip versus cataclastic flow, for the pyroxene and plagioclase, would require examination by transmission electron microscopy (cf. Tullis and Yund, 1987). Grain size reduction is accompanied by strain-induced segregation of minerals into monomineralic bands (cf. Jordan, 1987) leading to the ideal low strength configuration whereby layered microdomains of dynamically recrystallized plagioclase alternate with equally fine-grained clinopyroxene and garnet (Plates 1-6a, 1-7c). Monocrystalline ribbons of quartz and orthopyroxene, with aspect ratios of up to 20:1, are locally present (Plate 1-7d) and may reflect strain localization in the weakest phase of these rocks. In orthopyroxene, elongation appears to have been accomplished by slip on the system (100), [001].

Gabbroic rocks and paragneiss show a similar progression from a discontinuous $S_{2(KC)}$ foliation (Plates 1-2b, 1-5a) to continuous, straight, millimetre-scale ribbon mylonites (Plates 1-6b,c,d). Concordant granite veins that cut these rocks are attenuated and contain ribbons of quartz and feldspar. Relatively unstrained mixed gneiss outside the shear zone, which is compositionally and texturally heterogeneous, can be traced progressively into homogeneous, subtly layered rocks with a strong rodding lineation and intense foliation defined by ribbons of quartz, feldspar and orthopyroxene, and studded by porphyroclasts of garnet and minor orthopyroxene (Plates 1-3b,c).

Assemblages within mylonites of the Northern shear zone indicate granulite facies (Chapter 2). They exhibit well-preserved high-temperature metamorphic mineral assemblages and optical microstructures, reflecting the general absence of retrogression and post-tectonic recovery in these rocks. Minimal relaxation is also expressed by patchy undulose extinction within subgrains of plagioclase and quartz, indicative of stored strain energy (Plate 1-7d), and by the preferred crystallographic orientations of quartz (de-

scribed below). Preservation of these features is attributed to the exceedingly rapid rate of cooling of these relatively anhydrous, high-temperature mylonites (Chapter 3).

Within mylonitic and ultramylonitic rocks of the shear zone, foliations define C planes (from the French *cisaillement*, meaning shear; Berthé et al., 1979) which lie parallel to the flow plane of the progressive deformation. Near-vertical mylonitic foliations occur throughout the shear zone except in its southeast sector, where foliations dip moderately to steeply (40-88°) to the north. Most mylonitic rocks are LS and L>S tectonites, with extension lineations that are parallel to $L_{2(KC)}$ elsewhere in the complex. Shallowly to moderately west-plunging lineations are dominant; however, discrete domains of eastward-plunging lineations occur in the northwest (10-55°E) and north-central (5-20°E) parts of the zone. The similar attitude and metamorphic grade of $D_{2(KC)}$ fabric elements throughout the complex and the Northern shear zone, suggest that shearing is an integral part of the $D_{2(KC)}$ deformation and metamorphism, rather than an unrelated, superimposed event.

Straight layered ribbon mylonites of the Northern shear zone generally show a marked absence of porphyroclasts, folds and boudins regardless of their protolith. The uniform nature of ribbon mylonites of gabbroic, gabbroic anorthosite and anorthositic protolith (Plate 1-6c), and of paragneissic and granitic origin, is a feature common to high-grade tectonites (Passchier et al., 1990) and reflects an absence of rheological contrasts in the deforming material at all scales. In spite of this, rare shear-sense indicators identified at a dozen localities along the length of the shear zone generally show a sinistral shear sense in horizontal exposures. These consist of mantled porphyroclasts of δ -type (Plate 1-6e) and rarely σ -type winged inclusions with stair-step geometry, sillimanite-fish, and shear band cleavages corresponding mainly to C and C' fabrics (Platt and Vissers, 1980; Gapais and White, 1982) and rarely to C and S fabrics (Berthé et al., 1979).

Petrofabric data for quartz-bearing rocks from the shear zone are consistent with sinistral, non-coaxial strain operating at high-temperature conditions (Carter, 1976; Passchier and Trouw, 1996). Sinistral non-coaxial strain is best displayed by great circle girdle distributions of quartz *c*-axes that are asymmetric with respect to the finite strain reference frame (samples *K*-290, *K*-432, Fig. 1-6). Patterns of higher symmetry, represented by small circle distributions (*K*-400, *K*-440, Fig. 1-6) reflect a greater component of flattening strain, however, well-developed maxima in the northwest quadrants of these stereograms reflect non-coaxiality and sinistral vorticity of the flow. In comparison, crystallographic preferred orientations of quartz *c*-axes from a less-strained lithon (*K*-80, Fig. 1-6) display diffuse, approximately symmetrical maxima, consistent with lower degrees of coaxial finite strain. High temperature deformation within the Northern shear zone, indicated by its granulite facies mylonites, is reflected by the numerous slip systems active during deformation. Operation of several slip systems within the same sample is indicated in samples *K*-440 and *K*-290 (Fig. 1-6), where *c*-axis orientations from the periphery to the centre of the girdle result from basal, rhomb and prism $\langle a \rangle$ slip, respectively. Samples dominated by a single slip system, include *K*-440 (basal slip) and *K*-400 (prism $\langle c \rangle$ slip). The one sample analysed from the southeast segment of the shear zone (*K*-356; Fig. 1-6) exhibits a crossed girdle distribution of *c*-axes which may reflect a component of dextral shear, or alternatively, basal slip during sinistral noncoaxial strain with a significant population of unfavorably oriented high-strain grains. The kinematic significance of these data requires additional petrofabric work on this, and other samples from this fault segment.

General sinistral vorticity across the northern segment of the shear zone where moderately to shallowly plunging westerly extension is dominant suggests that the shear zone may have accommodated largely sinistral, south-side-up (Kramanituur Complex-side-up) displacement. Accordingly, the bend in the northeast part of the fault trace (locality *b* in Fig. 1-6) would have been a restraining bend, consistent with the steeper pitch

of lineations in this area. Near-horizontal foliations proximal to the right-stepping (anti-dilational) offset of the sinistral oblique-slip Northern shear zone may have developed during $D_{2(KC)}$ due to constrictional strain in this area (Segall and Pollard, 1980).

Although rocks of the Northern shear zone generally display L/S fabrics, two sectors of the zone are characterized by more oblate finite strains. The northwestern sector (James Point area, Fig. 1-6) consists of granulite facies ultramylonitic gabbro, gabbroic anorthosite and anorthosite with moderately ($20-50^\circ$) east-plunging mineral lineations. A greater component of coaxial shortening across the flow plane in this sector is indicated by the $S \gg L$ fabrics of these rocks (Plate 1-6c, f), their petrofabric data (K-440, Fig. 1-6) and the presence of garnet porphyroblasts with strain shadow geometry indicative of both clockwise and anticlockwise rotation. The southeast extremity of the Northern shear zone (Fig. 1-6), consists of moderate to steeply north-dipping, east-striking mylonitic rocks with east-plunging mineral lineations and similar oblate, $S \gg L$ fabrics. Shear-sense indicators are rare. Rather, megascopic structural features consistent with coaxial strain include symmetrical, tight to isoclinal folds (Plate 1-6g) with extension-parallel axes and symmetrical boudins. Accordingly, flattening strains are recorded across the northwest and southeast sectors of the Northern shear zone.

Semi-brittle shear zones and kink folds

Spaced, metre-scale structures including ductile and semi-brittle shear zones and kink folds cut penetrative fabrics within the complex (*see* Fig. 8 in Sanborn-Barrie, 1994, Appendix II) and, in particular, mylonitic layering of the northwestern sector of the Northern shear zone. These steeply-dipping structures generally occur at a high angle to the fabric they overprint and offset, and are designated $D_{3(KC)}$. Preservation of granulite facies mineral assemblages within $D_{3(KC)}$ structures reflect minimal retrogression suggesting that water activity remained low during their formation. The apparent sense of metre-scale displacement across these structures, determined from deflection of the folia-

tion they cut, is both sinistral and dextral consistent with layer-parallel and layer-normal extensional strains. $D_{3(KC)}$ structures are locally filled with granitic material which carries a weakly to strongly developed foliation consistent with late synkinematic injection. These $D_{3(KC)}$ structures may represent accommodation structures, required if the open curvature of the northwest part of the Northern shear zone was caused by bending. The relative absence of these structures in the northeast part of the complex is consistent with a primary bend in the Northern shear zone, described above.

Fabric elements external to the Kramanituar Complex

Rocks south of the Kramanituar Complex carry structural fabric elements that predate those within the complex. Metasedimentary rocks generally have a single, moderately developed foliation defined by aligned biotite and a weak shape fabric in quartz and feldspar. This foliation strikes east-northeast and dips $40-60^\circ$ to the northwest defining a moderately northward-dipping homocline (Fig. 1-5a). Weakly developed extension lineations plunge $15-35^\circ$ to the west and northwest. This fabric is the earliest recognized in the study area and is here designated $S_{1(External)}$. However, in supracrustal rocks ca. 50 km southwest of the study area, this northeast-striking, dominantly north-dipping schistosity overprints an earlier S_1 foliation, is axial planar to F_2 folds, and is cut by 2.19 Ga mafic dykes (Tella et al., 1997a).

Quartz diorite $\pm$ diorite gneiss south of the complex carry a north- to northeast striking granoblastic gneissosity (Plate 1-8a,c) which may correlate with $S_{1(External)}$ in metasedimentary rocks. The absence of a shape fabric parallel to compositional layering in these rocks suggests static recrystallization occurred after layering was formed. In this way these rocks are texturally distinct from rocks of the Kramanituar Complex, in which annealed, granoblastic textures are largely absent. Quartz diorite gneiss carries a superimposed preferred shape fabric which occurs at a small angle to gneissosity, and may correlate with $S_{2(External)}$, described below.

Foliated and gneissic plutonic rocks north of the complex possess a moderately dipping tectonic foliation that strikes northeast to southeast, defining broadly curvilinear trajectories (Fig. 1-5a). These rocks are L/S tectonites, and rarely L-tectonites, with extension lineations that plunge moderately ($10-35^\circ$) to the west (Fig. 1-5b). This fabric is of similar style, intensity and geometry to that of $S_{1(\text{External})}$ south of the complex, and may have developed contemporaneously with it. In contrast, within 1 km of the north margin of the complex, metaplutonic rocks have penetrative, east-striking steeply-dipping L/S fabrics which are parallel to $S_2/L_{2(\text{KC})}$ and are correlated with them.

East of the complex, compositional layering in dioritic orthogneiss strikes northerly and dips steeply to the west (Fig. 1-5a). Extension lineations plunge southwest at $50-80^\circ$, and represent the steepest linear shape fabrics observed in the map area.

Southern shear zone

The Southern shear zone consists of a series of east-striking, right-stepping, ductile fault segments that cut granitoid and metasedimentary rocks south of the Kramanitar Complex (Fig. 1-6). This shear zone was active at ca. 2550 Ma, the U-Pb magmatic crystallization age of a syntectonic syenogranite pegmatite dyke that cuts it (Table 1-1; see Chapter 3). Within this zone, progressive transposition of $S_{1(\text{External})}$ fabrics and grain size refinement in orthogneiss and schistose metasedimentary rocks, accompanied by reduction in the thickness of compositional layering from 20 cm to 0.1-1 cm (Plate 1-8a, c), resulted in straight gneisses ($S_{2(\text{External})}$) with isolated porphyroclasts (Plate 1-8b). Metasedimentary rocks within the zone have shear band cleavages of C- and C'-type (Plate 1-8d), distinguishing them from their less-strained equivalents. Based on the syntectonic mineral assemblage of biotite-garnet-plagioclase-quartz $\pm$ sillimanite, mylonitization took place at mid-amphibolite facies conditions.

$S_{2(\text{External})}$ fabrics within the Southern shear zone are west-striking ($250-270^\circ$) and north-dipping ($55-65^\circ$). Strongly developed extension lineations defined by quartz

and feldspar ribbons in granitoid gneiss and sillimanite lineations in semi-pelite plunge 25-50° to the west and northwest, or are locally down-dip (65°N). Within the shear zone, hinge lines of tight to isoclinal Z-folds (*see* Fig. 7*d* in Sanborn-Barrie, 1993, Appendix I) are generally coaxial with the extension lineations, indicating widespread rotation into the flow direction. At one locality, folds in metasedimentary rocks have hinges that are curved about the interpreted direction of finite extension, resulting in a large degree of non-cylindricity suggestive of sheath folds (Quinquis et al., 1978). Well-developed shear-sense indicators throughout the Southern shear zone consistently indicate north-side-up and dextral displacement. These include shear bands in metasedimentary rocks (Plate 1-8*d*) and δ -type mantled porphyroblasts of K-feldspar in mylonitized granitoid rocks (Plate 1-8*e*).

GEOLOGICAL SUMMARY AND DISCUSSION

The Kramanituur Complex consists of granulite-facies rocks which are dominated by a Paleoproterozoic (ca. 1900 Ma) gabbroic suite, with lesser charnockite of Archean age ($\geq$ ca. 2573 Ma) and metaplutonic and metasedimentary tectonites of probable primary Archean age. It is distinguished from surrounding amphibolite-facies rocks of intermediate metaplutonic and supracrustal origin by its mafic plutonic lithologic association, granulite facies assemblages and its Paleoproterozoic tectonometamorphic history that is in contrast to the mainly Archean (>2550 Ma) age of its country rocks and the older metamorphic mineral assemblages these country rocks locally preserve.

Within the complex, several generations of fabric elements are identified which are consistent with a bulk deformation path of progressive, general, noncoaxial deformation at ca. 1.9 Ga. Low strain domains in the central part of the complex preserve weakly to moderately developed $S_{1(KC)}$ ($\pm L_1$), however, much of the complex is thoroughly dynamically recrystallized with strongly developed west- and northwest-striking foliations marked mainly by shallow to moderate westward-plunging lineations ($S_{2(KC)}, L_{2(KC)}$).

Strains of highest magnitude are localized within the Northern shear zone, an anastomosing belt of granulite facies ribbon mylonites that coincides with the north and parts of the east margins of the complex. Shear fabrics, parallel to $S_2/L_{2(KC)}$ and attributed to the same deformation, display evidence of general, non-coaxial flow with a sinistral sense of vorticity, consistent with south-(Kramanituar Complex) side-up movement across much of the Northern shear zone. More transpressive finite strain is reflected in domains with $S>L$ fabrics, east-plunging lineations and small circle distributions of quartz c -axes.

Strain fabrics south of the complex have different geometry and timing than those within the complex. A regional northeast-striking, moderately northwest-dipping, northwest-lineated fabric ($S_{1(External)}$) defined by amphibolite-facies assemblages in supracrustal rocks developed prior to ca. 2550 Ma, the age of the Southern shear zone that cuts it (Chapter 3). That these rocks preserve Archean fabric elements, metamorphic assemblages and textures implies that a fundamental tectono-metamorphic discontinuity occurs between the ca. 1.9 Ga Kramanituar Complex, and rocks to the south.

Role of structures in accommodating exhumation of high- P rocks

Rocks of the Kramanituar Complex have P - T - t paths which record near-isothermal decompression from deep-crustal conditions (Chapter 2), followed by rapid cooling (Chapter 3). Such paths require extremely efficient exhumation. Tectonic unroofing allows for efficient exhumation by the removal of overlying material from rapidly uplifted rocks. The presence of granulite facies ribbon mylonites of the Northern shear zone along the north and much of the east margins of the high- P Kramanituar Complex, and the dominant sinistral, south-side-up movement recorded by their fabric elements suggest an apparent connection between shear zone deformation and exhumation of the complex. Although the Northern shear zone was active during exhumation, as indicated by localization of decompressional (garnet breakdown) reaction products in strain shadows (Chapter 2), it is not a master fault which accommodated major displacement of

deep-crustal rocks. The right-stepping jog in the shear zone imposes a requirement of minimal displacement at this locality, and this is expressed by the local preservation of a 1 km wide contact metamorphic aureole imposed by the complex on dioritic country rocks. Although the absence of a thermal aureole north of the complex is attributed to greater magnitudes of differential displacement across this part of the Northern shear zone, the presence of Paleoproterozoic high-*P* mafic rocks north, and possibly east, of the complex (Chapters 2, 3) confirm that the Northern shear zone, although spectacular in terms of macro- and microstructural fabric development (Plates 1-6, 1-7) accommodated only modest displacement (cf. Hanmer et al., 1995*b*). The spatial distribution and complex geometry of mylonites of the Northern shear zone suggest that it developed in response to shear localization within a thermally weakened zone, and that its development was controlled, in part, by the rheological interface between hot gabbroic-anorthositic rocks and quartzo-feldspathic country rocks to the north and east (cf. Hanmer, 1997*b*).

The rapid cooling history of exposed deep-crustal rocks of the Kramanituur Complex requires that an efficient exhumation mechanism, such as tectonic unroofing, played a significant role in the evolution of the complex. The preservation of an older metamorphic history in rocks south of the Kramanituur Complex, in contrast to the penetrative Paleoproterozoic tectonometamorphism within it, allows the presence of an intervening tectonic break, which is a likely candidate for an extensional detachment structure that facilitated exhumation. Although fabrics related to a Paleoproterozoic detachment fault were not recognized south of the gabbroic-anorthositic suite, this inferred structural break may coincide with the northwest-trending, 400 m wide northern channel of Chesterfield Inlet (Fig. 1-6). This is supported by preliminary magnetotelluric-teleseismic data (D. Snyder and A.G. Jones, personal communication, 1999) which highlight differences in density, crustal depth and conductivity across Chesterfield Inlet. These differences are consistent with a steep or shallow south-dipping discontinuity, which may correspond to a cryptic Paleoproterozoic detachment fault.

The Snowbird tectonic zone

The Snowbird tectonic zone was originally defined by Hoffman (1988) as a regional, throughgoing geophysical anomaly (Fig. 1-1a) postulated to coincide with a Paleoproterozoic suture. This investigation of the Kramanituur Complex, and other recent geological studies, collectively establish that the Snowbird tectonic zone consists of several segments that share many features, but have some important differences. These differences mainly relate to the timing of latest penetrative tectonometamorphic activity. A summary of major findings from these segments, provided below, highlights the relationship of the Kramanituur Complex to the rest of the zone.

The southwest extremity of the Snowbird tectonic zone (Fig. 1-2) was interpreted by Ross et al. (1991, 1995) to extend beneath Paleozoic cover rocks in Alberta. Geological relationships established for the subsurface Alberta segment have little in common with those at the Kramanituur Complex. Rock types recovered from drillcore are not representative of the anorthosite-gabbro/norite $\pm$ paragneiss $\pm$ diatexite association, nor are they characterized by high-grade assemblages or ductile, mylonitic fabrics (Villeneuve et al., 1993). Subsurface structures in central Alberta, imaged as seismic reflections, are interpreted by Ross et al. (1995) to have been active at ca. 1856 to 1796 Ma and to have been related to convergence across the Trans-Hudson orogen (Lucas et al., 1997). This is ca. 70 m.y. later than Paleoproterozoic metamorphism, magmatism and deformation at the Kramanituur Complex. Exposed parts of the Alberta segment, such as the Virgin River shear zone and associated granulites (Fig. 1-2), yield a complex set of isotopic data which Crocker et al. (1993) use to suggest latest penetrative tectonometamorphism at ca. 1.9 Ga in this area, with subsequent Hudsonian reactivation (Crocker et al. 1993; Bickford et al., 1995).

The Striding-Athabasca mylonite zone (Fig. 1-2) is a ca. 400 km long, anastomosing, granulite- and amphibolite-facies belt of mid-Archean (ca. 3.13 Ga) annealed mylonites, and late Archean (ca. 2.62-2.60 Ga) ribbon mylonites of plutonic (tonalite,

gabbro, anorthosite, granite) and sedimentary (semi-pelitic) origin (Hanmer, 1994; Hanmer et al., 1994, 1995b; Snoeyenbos et al., 1995; Williams et al., 1995). The ribbon mylonites are interpreted as a dextral, transpressive shear zone that initiated adjacent to stiff, crustal-scale boudins in an intracontinental setting at ca. 2.6 Ga (Hanmer et al., 1995a). The Kramanitu Complex and the Striding-Athabasca mylonite zone show strong similarities in terms of lithologic associations, metamorphic grade and structural style. At both, the anorthosite-pyroxenite-gabbro/norite association is volumetrically important; paragneiss into which these mafic suites intruded is subordinate; and felsic intrusive rocks and diatexite are widespread. Granulite-facies mineral assemblages prevail throughout both regions, and steeply dipping high strain zones form major mylonite belts that display spectacular fabrics but record modest movements. At both localities, granulite-facies shear zones have not had an important role in the uplift and exhumation of deep-crustal rocks. A fundamental difference between these areas, however, is the timing of latest penetrative magmato-tectonometamorphism. Within the Striding-Athabasca mylonite zone latest high-grade deformation is essentially Neoarchean (ca. 2.6 Ga) in age; whereas Paleoproterozoic activity is identified as weakly developed with localized plastic deformation, and relatively minor granitoid intrusion (Hanmer et al., 1994; Hanmer et al., 1995a; Hanmer, 1997a). This is in contrast to the Kramanitu Complex which records penetrative Paleoproterozoic metamorphism and ductile deformation.

The Tulemalu fault zone segment (Fig. 1-2) contains several geological elements that invite comparison to other segments of the Snowbird tectonic zone. Xenoliths containing high-*P* mafic granulite assemblages, similar to those recognized in most other segments, are contained in a foliated granitoid pluton that occupies a northeast-striking, steeply west-dipping fault zone (Tella and Eade, 1986). A strand of the fault zone exposed on eastern Angikuni Lake is a near-vertically dipping, northeast-striking, strike-lineated dextral shear zone (Aspler et al., 1998; Aspler et al., 1999) dated at ca. 2.61 Ga, a preliminary age of a syntectonic granite (W. Davis, unpublished data, 1997). A small

exposure (4 km²) of anorthosite, characteristic of the Snowbird tectonic zone, occurs northeast of Tulemalu Lake (Wright, 1967) on strike with the Tulemalu fault zone.

The east-striking Chesterfield segment of the Snowbird tectonic zone includes, in addition to the Kramanituvar Complex, the Uvauk and Daly Bay complexes (Fig. 1-1, 1-2). Similar to Kramanituvar, these complexes contain the gabbroic anorthosite-gabbro/norite-paragneiss-diatexite association; moderately to intensely developed planar and linear fabric elements which, although variable, are oriented predominantly east-west; and marginal belts of granulite-facies mylonites (Tella et al., 1993; Gordon, 1988; Gordon and Lawton, 1995; Mills et al., 1999). Existing data from these complexes suggest that Paleoproterozoic tectonometamorphism, was superposed on previously metamorphosed rocks. Although the relative extent of Archean versus Paleoproterozoic magmatism and tectono-metamorphism are not well understood for these complexes, limited geochronological data bracket the timing of latest high-grade metamorphism at Daly Bay between ca. 2067 Ma and 1951 Ma (Gordon, 1988), and at the Uvauk complex between ca. 1940-1920 Ma (Tella et al., 1993). Combining temporal constraints from the Kramanituvar Complex (Chapter 3) with these preliminary data for the Uvauk and Daly Bay complexes, one can speculate that the Chesterfield segment of the Snowbird tectonic zone shares a common, though apparently diachronous, history with respect to Paleoproterozoic magmatism, metamorphism and exhumation.

CONCLUSIONS

The Kramanituvar Complex is interpreted to represent a tectonic lens of deep-crustal rocks of mainly Paleoproterozoic age contained within a larger block of Archean crust. Intrusion of mantle-derived (Chapter 4) mafic magmas at deep-crustal levels during noncoaxial flow, tectonic exhumation of this crustal block to mid-crustal levels (Chapter 2) and subsequent rapid cooling (Chapter 3) were the processes involved in the formation and preservation of these rocks. Paradoxically, the two major shear zones that

bound the Kramanituar Complex did not accommodate major uplift and exhumation of high-*P* rocks. The geometry of the Paleoproterozoic granulite-facies Northern shear zone imposes constraints on the magnitude of displacement it could have accommodated; whereas, the Neoarchean age of the Southern shear zone discounts it as a candidate for Proterozoic exhumation. However, a fundamental interface appears to exist between ca. 1.9 Ga fabrics and metamorphic assemblages within the complex and those of Neoarchean age to the south. This interface may be a discrete, south-dipping extensional detachment fault that may coincide with northern Chesterfield Inlet.

The Paleoproterozoic timing of latest penetrative magmatic and tectonometamorphic events at the Kramanituar Complex, and elsewhere within the east-striking Chesterfield segment, indicate that this segment was a discrete structural entity which operated separately from other parts of the Snowbird tectonic zone during Paleoproterozoic time.

Table 1-1. Summary of dated geological events for the Kramanituur Complex and surroundings.

PROTEROZOIC EVENTS

Cooling (~700-750°C) of coronitic metagabbro, north of complex	1878±27 Ma ^{SN}
Cooling ~400° (rutile) of paragneiss	ca. 1900 (sil) -1896 (ky) Ma ^R
Cooling ~400° (rutile) of Kramanituur gabbroic anorthosite	1901 Ma ^R
Cooling ~700° (titanite) of Kramanituur gabbroic anorthosite	1901 Ma ^T
Cooling (~700-750°) of mafic granulite from the Kramanituur Complex	1894 ±9 Ma ^{SN}
Crystallization of Kramanituur Complex gabbroic anorthosite	1902 ±1.6 Ma ^Z
Leucogranite dyke, late-tectonic with respect to the Northern shear zone	1902.6 ±1 Ma ^{Z,M}
Prograde metamorphism of paragneiss within the complex ($T \sim 550^\circ\text{C}$)	ca. 1910 (sil) –1908 (ky) Ma ^M
Titanite age in gabbro, south of complex	ca. 2326 Ma ^T

ARCHEAN EVENTS

Syenogranite pegmatite, syntectonic with respect to the Southern shear zone	ca. 2550 Ma ^Z
Charnockite gneiss within the complex	2573 +28/-5 Ma ¹
Foliated granodioritic wallrocks, north of complex	2675 +33/-11 Ma ¹
Crystallization of intermediate metavolcanic rocks, south of complex	>2655 ±12 Ma, <2691 ±10 Ma ²

Notes: U-Pb on Z-zircon, M-monazite, T-titanite, R-rutile ages determined in this study.

SN-samarium-neodymium on garnet-plagioclase±clinopyroxene±whole rock, this study.

¹ Schau, 1980; ² Davis and Sanborn-Barrie, unpublished data, 1998. ky-kyanite, sil-sillimanite.

Figure 1-1. Colour gravity maps of the western Churchill Province. *a*) horizontal gradient of the Bouguer gravity field. *b*) Bouguer free-air gravity anomalies. Canadian Gravity Anomaly Database (Geophysical Data Centre, Geological Survey of Canada).

Figure 1-2. Geology of the western Canadian Shield after Hoffman (1988). Segments of the Snowbird tectonic zone (STZ) from northeast to southwest include the Chesterfield segment which comprises the Daly Bay (DBc), Uvauk (Uc) and Kramanituvar (KC) complexes; the Tulemalu fault zone; the Striding-Athabasca mylonite zone (SAmyz); and the subsurface Alberta segment (eA) bound to the northwest by the Virgin River shear zone (VRsz).

Figure 1-3. General geology of the Kramanituvar Complex and its country rocks, modified from Schau and Ashton (1980), Schau (1983), Sanborn-Barrie (1993) and Sanborn-Barrie (1994). Letters and symbols designate features described in text: **A**-anorthosite blocks; **B**-kyanite-bearing mylonitic paragneiss; **C**-coherent granite body; **D**-coronitic gabbro; **E**-prehnite-bearing gabbro; Filled diamonds are U-Pb localities of Schau (1980; *see* Table 1-1).

Figure 1-4. Metamorphic isograds and metamorphic mineral occurrences. Note extent of contact metamorphic aureole (diagonal lines) imposed by the Kramanituur Complex (shaded) on adjacent intrusive country rocks to the east. Isograds that may be older than those within the complex are shown in gray tones.

Figure 1-5. Planar and linear tectonic elements of the Kramanituvar Complex (shaded) and its country rocks. *a*) Foliation trajectory map; *b*) Mineral lineation data.

Figure 1-6. Distribution of major ductile shear zones. Shaded region delineates the Northern shear zone with restraining bend at *b*, right-stepping discontinuity at *d*. Quartz *c*-axis fabrics are shown contoured on lower hemisphere equal-area (Schmidt) projections. Diagonal lines delineate right-stepping segments of the Southern shear zone. Both ductile shear zones are cut by a sinistral semi-brittle fault (unpatterned).

Plate 1-1. Kramanituvar gabbroic suite. *a)* coarse-grained gabbroic anorthosite, with minor recrystallization of plagioclase shown by partial white rims; *b)* modal layering in gabbroic anorthosite defined by decrease in mafic minerals from base of layer (indicated by scale card) to top of photo. Trace of $S_{1(KC)}$ shape fabric at 082/88, parallel to pencil, forms a small, counterclockwise angle to S_0 ; *c)* trace of $S_{2(KC)}$ in gabbroic anorthosite defined by discontinuous aggregates of pyroxene, garnet $\pm$ amphibole in a recrystallized plagioclase matrix; *d)* penetratively recrystallized gabbroic anorthosite with a well-developed $S_{2(KC)}$ foliation; *e)* gabbro with $S_{1(KC)}$ shape fabric from locality *d* in Figure 1-6; *f)* view to west-northwest of north-verging F_2 folded mafic dyke in gabbroic anorthosite, 23 cm long hammer on fold limb, for scale.

Plate 1-2. Granulite facies paragneiss. *a*) foliated paragneiss, northwest shore of Chesterfield Inlet, with $S_{1(KC)}$ at $102/49^{\circ}S$, parallel to pencil; *b*) sillimanite-bearing paragneiss with straight $S_{2(KC)}$ at $262/77^{\circ}N$, located immediately south of the Northern shear zone; *c*) polished slab of mylonitized paragneiss from the northeast part of the complex showing kyanite-garnet restite (dark grey) and kyanite-garnet-graphite bearing leucosome (white).

Plate 1-3. Mixed gneiss and diatexite. *a*) XZ plane showing mixed gneiss unit consisting of diorite cut by granite veins (ie., below pencil tip) with $S_{2(KC)}$ shape fabric 12° counterclockwise to the intrusive layering; *b*) subtly layered unit transitional between mixed gneiss and diatexite; *c*) garnetiferous diatexite.

Plate 1-4. Country rocks external to the Kramanituvar Complex. *a*) gneissic granodiorite north of the complex; *b*) granoblastic quartz diorite orthogneiss, near North Channel; *c*) semi-pelite (grey) interbedded with pelite south of the complex. Coarse-grained texture of pelite is due to incipient *in situ* partial melting; *d*) mafic dyke intrusive into metasedimentary rocks south of the complex. Leucosome infilling interboudin space is consistent with syntectonic (D₂) metamorphism at ca. 2550 Ma; *e*) intermediate pyroclastic breccia south of the complex.

Plate 1-5. Planar fabrics within the Kramanitaur Complex on surfaces subparallel to XZ plane. *a*) typical nature of $L \geq S$ $S_{2(KC)}$ in gabbro oriented 070/88°S, northwest part of complex; *b*) east-plunging folds with axial surfaces parallel to S_2 at 098/84°S, central part of complex; *c*) panel of folded S_1 layering trending 311/40°NE, enveloped by straight gneiss ($S_{2(KC)}$) at 100/73°S, from northwest part of complex, 23 cm long hammer for scale; *d*) trace of S_2 (125/75°SW), parallel to pencil, defined by shape fabric in quartz and feldspar and by aligned mafic minerals. Note S_1 shape fabric in quartz is parallel to intrusive layering at 145/56°SW.

S_1 S_2

Plate 1-6. Granulite-facies ribbon mylonites from the Northern shear zone, surface subparallel to XZ plane. *a*) anorthosite; *b*) gabbro; *c*) anorthosite mylonite (white) and adjacent gabbro mylonite (grey). Note ribbon texture of both units and the absence of boudins, folds, or porphyroclasts; *d*) sillimanite-bearing paragneiss; *e*) δ -type mantled porphyroclast of orthopyroxene in a pervasively recrystallized matrix of aligned clinopyroxene-plagioclase-garnet $\pm$ quartz defining C planes oriented 120/87°SW. Sinistral shear sense, northeast part of complex; *f*) $S \gg L$ gabbro ultramylonite near James Point, pencil leaning against plane parallel to XZ plane of finite strain ellipsoid. C planes at 262/65°N with stretching lineation at 042/45°SE; *g*) view west of isoclinally folded gabbro and anorthosite from the southeast segment of the Northern shear zone.

Plate 1-7. Microstructures in mylonitic gabbroic anorthosite and gabbro from the Northern shear zone. *a*) fine-grained plagioclase- (clear) and clinopyroxene-bearing (gray) matrix with seriate texture enveloping poikiloblastic garnet porphyroblasts (black). Width of view 5 mm. CPL. *b*) clinopyroxene paleoblasts rimmed by very fine-grained mantles of clinopyroxene neoblasts set in a recrystallized plagioclase-rich matrix of coarse to fine subgrains with seriate grain boundaries. Width of view 1.8 mm. CPL. *c*) layered microdomains of recrystallized plagioclase (Pl) alternate with fine-grained clinopyroxene (Cpx) and garnet (Grt). Width of view 1.8 mm. CPL.; *d*) monocrystalline ribbon of orthopyroxene with 20:1 aspect ratio. Width of view 5 mm. CPL..

Plate 1-8. Mylonitic rocks from the Southern shear zone. *a*) strain gradient from folded diorite gneiss to straight mylonite gneiss; *b*) mylonitized granitoid rocks at Shoal Point (located in Fig. 1-3); *c*) low-strain panels within mylonite preserving nested folds, Shoal Point; *d*) shear bands of C-C'-type in schistose sillimanite-garnet pelite, pencil parallel to C'. Sigmoidal lenses of leucocratic material form a 20°-counterclockwise angle to C planes which bound them, consistent with back-rotated boudins in a dextral shear zone; *e*) mantled K-feldspar δ -type porphyroclast with asymmetrical, in-plane wings indicating dextral shear.

CHAPTER 2

Geothermobarometry of the Kramanituar Complex and its country rocks: petrological evidence of rapid exhumation of deep-crustal rocks in the Archean western Churchill Province at 1.9 Ga

ABSTRACT

The Kramanituar Complex is a window of Paleoproterozoic, high-*P* granulite-facies rocks in the interior of the Archean western Churchill Province, Canada. The complex is dominated by a metagabbroic-anorthositic suite with subordinate kyanite- and sillimanite-bearing paragneiss, and minor charnockite. Near-peak metamorphic conditions of 12 ± 1.5 kbar and $810 \pm 70^\circ\text{C}$ are estimated for rocks with textures and mineral compositions suggestive of a close approach to equilibrium. Those with coronitic and symplectitic reaction textures record evidence of high-*T* decompression from similar near-peak conditions to 8 ± 1 kbar and $800 \pm 50^\circ\text{C}$. Prograde metamorphism of the complex is bracketed between 1910-1908 Ma, the U-Pb age of concordant monazite in the paragneisses, and 1902 ± 1.6 Ma, the time of intrusion and granulite facies metamorphic recrystallization of the Kramanituar gabbroic-anorthositic suite. Exhumation to mid-crustal levels led to very rapid cooling of the complex below ca. 400°C (U-Pb rutile) by 1901-1896 Ma.

North and east of the complex, extensive amphibolite- to greenschist facies granitoid rocks host isolated gabbroic rocks that record high-*P* conditions, similar to those of the Kramanituar Complex. In contrast, rocks south of the complex show fabric relationships, compositional variation and isotopic evidence that highlight a prograde metamorphic history that predates the ca. 1.9 Ga tectonometamorphism recorded by the Kramanituar Complex. For example, Bt-Grt-Pl-Sil-Qtz assemblages in semi-pelite define pre- and syn- ca. 2550 Ma fabrics that wrap garnet porphyroblasts, indicating a minimum Neoarchean age for garnet growth at 6.5 kbar and 700°C (cores) to 8 ± 1 kbar and 750°C (rims). Gabbroic rocks similarly record prograde conditions (to ca. 12 kbar and 870°C) that predate ca. 2326 Ma, the U-Pb titanite age of this unit.

The Kramanituar Complex is interpreted as an exposed window of largely anhydrous, Paleoproterozoic deep-crustal rocks whose thermal budget was a product of magmatic heat superimposed on a regional metamorphism. The complex and country rocks to the north are believed to have been tectonically juxtaposed with rocks that preserve older fabrics and metamorphic assemblages, exposed south of the complex. Efficient exhumation via extensional faulting is suggested by the rapid cooling required for the complex at ca. 1.9 Ga, and by juxtaposition of the Kramanituar Complex against country rocks to the south that are unaffected by high-grade ca. 1.9 Ga tectonometamorphism.

INTRODUCTION

The western Churchill Province of North America (Fig. 2-1*a*) has long been recognized as having experienced protracted polymetamorphism that involved an extensive Archean tectonothermal history and variable degrees of reworking during Paleoproterozoic time (Davidson, 1972; Lewry et al., 1985). Its high-grade metamorphic rocks, which occur as extensive regions (7000-27000 km²) throughout the northern part of the province and as discrete complexes that lie within a corridor across its interior (Fig. 2-1*b*), are poorly understood. The *P-T* conditions, timing of metamorphism and tectonic setting for these rocks are not well established, limiting our understanding of the nature and timing of tectonometamorphic activity in the western Churchill Province throughout its protracted history.

The Kramanituar Complex is one of several exposed high-grade complexes in the interior of the western Churchill Province. These complexes are dominated by granulite-facies plutonic rocks, which include a gabbro-norite-anorthosite suite, and minor supracrustal rocks. They record moderate to intense degrees of ductile strain, typically manifested as kilometre-scale belts of granulite-facies mylonites, and coincide with the surface expression of a regional-scale geophysical anomaly, the Snowbird tectonic zone (Hoffman, 1988). This geophysical anomaly extends across the interior of the western Churchill Province more than 2000 km from the foothills of the Rocky Mountains to Hudson Bay, and is well expressed as a series of high amplitude anomalies in the horizontal gravity gradient field (*see* Fig. 1-1*a*; Goodacre et al., 1987*a*), and by low amplitude anomalies in the regional magnetic field (Geological Survey of Canada, 1987). Interpretations of the tectonic significance of the Snowbird tectonic zone have included: a Proterozoic intracontinental rift system (Wallis 1970); a Paleoproterozoic collisional suture (Gibb and Walcott, 1971; Thomas and Gibb, 1985; Hoffman, 1988); a Neoarchean intracontinental transform fault (Hanmer et al., 1995*b*) and an indentation-related fault network generated during continental collisional activity to the northwest or southeast

(Lewry and Sibbald, 1977; Lewry et al., 1985; Ross et al., 1995). Understanding the role of the Snowbird tectonic zone in the evolution of the western Churchill Province is critically dependent on integrated structural, metamorphic and isotopic studies of the high-grade complexes along it.

This paper presents petrographic and geothermobarometric results from the Kramanituar Complex and its surroundings, drawing from structural (Chapter 1) and geochronological (Chapter 3) constraints. We show that the complex consists of anhydrous mafic metaplutonic rocks that not only retain granulite-facies temperatures (800-950°C), but also record high pressures (12 ± 1.5 kbar). The attainment of these conditions is interpreted to be the product of magmatic heat superimposed on regional metamorphism in the deep crust. Their preservation is attributed both to exhumation-driven cooling shortly after metamorphism at 1.9 Ga (Chapter 3) and the anhydrous nature of the complex. Thermobarometry of surrounding rocks to the north and east reveals that high-*P* rocks occur beyond the high-strain boundaries of the Kramanituar Complex, but that these rocks record lower near-peak temperatures (650-825°C). In contrast, amphibolite-facies rocks south of the complex record pre- to syn-ca. 2550 Ma metamorphic assemblages and fabrics, highlighting the longevity of tectono-metamorphism in this part of the interior of the western Churchill Province.

GEOLOGICAL SETTING

The western Churchill Province of northwest Laurentia is largely underlain by Archean metaplutonic and supracrustal rocks with a substantial Proterozoic lithologic component (Fig. 2-1). Quantitative constraints on *P-T* conditions and the timing of metamorphism within the province indicate that granulite-facies metamorphism affected dominantly metaplutonic rocks at ca. 3.13 Ga (Crocker et al., 1993; Hanmer et al., 1994) and ca. 2.68 to 2.59 Ga (Fraser, 1988; Hanmer et al., 1994; Tella et al., 1994; Snoeyenbos et al., 1995). Amphibolite- to greenschist-facies metamorphism of supracrustal rocks is

recorded at ca. 2.68 to 2.65 Ga (Park and Ralser, 1992) and ca. 2.6-2.5 Ga (Ashton, 1988; Tella et al., 1997b; Berman and Stern, in press).

Paleoproterozoic tectonometamorphism of the province is more cryptic in nature. Although a high-grade metamorphic event at ca. 2.3-2.2 Ga is proposed for the southwest part of the province (Crocker et al., 1993; Bickford et al., 1994; M. Williams, pers. comm., 1999), Paleoproterozoic tectonometamorphism is generally considered to have involved localized greenschist-facies reworking; manifested by dextral shearing and plutonic emplacement in the Striding-Athabasca mylonite zone (Hanmer et al., 1994), and by cataclastic reworking of the MacQuoid-Gibson supracrustal belt (Tella et al., 1997a). High-grade Paleoproterozoic tectonometamorphism is restricted to a narrow strand of ca. 1.9 Ga mylonites in the northeastern Striding-Athabasca mylonite zone (Hanmer, 1997a) but appears to have penetratively affected the east-striking Chesterfield segment of the Snowbird tectonic zone. This is well demonstrated by geothermobarometric (this chapter) and geochronologic (Chapter 3) data for the Kramanituvar Complex which highlight penetrative high-*P* tectonometamorphism at 1910-1901 Ma. High-grade Paleoproterozoic tectonometamorphism is postulated for the Uvauk complex at ca. 1940-1920 Ma (Tella et al., 1994) and for the Daly Bay complex between ca. 2067 Ma and ca. 1951 Ma (Gordon, 1988), based on limited thermobarometric and geochronologic data.

The Kramanituvar Complex

The Kramanituvar Complex is dominated by a granulite-facies metagabbro-gabbroic anorthosite-norite suite dated at 1902 ± 1.6 Ma (Chapter 3), intrusive into charnockite with an inferred metamorphic age of $2573 +28/-5$ Ma (Schau, 1980) plus panels of sillimanite- and kyanite-bearing paragneiss and diatexite of probable primary Archean age. These rock types generally occur as 1 to 5 km wide, west- and northwest-trending panels, that together form a coherent, 850 km² elliptical exposure of high-grade rocks (Fig. 2-2). The complex is in contact with mainly Archean rocks. North of the complex,

these include foliated to gneissic quartzo-feldspathic plutonic rocks, with a U-Pb zircon age of $2675 \pm 33/-11$ Ma from one locality (Schau, 1980), which host isolated mafic plutonic rocks (Fig. 2-2). South of the complex, quartz diorite orthogneiss, pelitic to semi-pelitic metasedimentary rocks and minor ca. 2.68 Ga metavolcanic rocks (Davis and Sanborn-Barrie, unpublished data, 1998) are cut by gabbro and syntectonic pegmatite dated at ca. 2550 Ma (Fig. 2-2). The western margin of the Kramanituur Complex is unconformably overlain by sub-greenschist facies sedimentary rocks of the Baker Lake Group (Figures 2-1a, 2-2; Gall et al., 1992), whose basal polymictic conglomerate contains locally derived rocks, including gabbroic anorthosite. A maximum age of deposition for this transgressive sequence is ca. 1901 Ma, the time at which its basement, the western Kramanituur Complex, cooled through ca. 400°C (Chapter 3). Its minimum age is 1832 ± 28 Ma, the age of associated overlying intrusive ultrapotassic rocks (MacRae et al., 1996).

The Kramanituur Complex contains Paleoproterozoic tectonic fabrics which are interpreted to reflect progressive deformation during general noncoaxial flow. Low strain rocks in the interior carry a weakly to moderately developed, steeply dipping, northwest-trending L/S fabric ($S_{1(KC)}$) locally marked by moderate (30-50°) northwest-plunging mineral lineations. Much of the complex is characterized by penetratively recrystallized rock with strongly developed L/S fabrics ($S_2/L_{2(KC)}$). Intensely strained rocks occur along the northern and much of the eastern margins of the complex forming the Northern shear zone, a discontinuous, anastomosing mylonite belt 60 km long and 3 to 6 km wide (Fig. 2-2 inset). West- to northwest-striking, steeply dipping granulite-facies ribbon mylonites ($S_{2(KC)}$) with a predominance of shallow to moderate westward plunging lineations record evidence of progressive, general, noncoaxial flow with a sinistral sense of vorticity (Chapter 1). A minimum age of ductile movement on the Northern shear zone is 1902.6 ± 1 Ma, the age of a leucogranite dyke that cuts it (Chapter 3).

Foliated to gneissic plutonic rocks north and east of the complex display a moderately developed northeast-striking, variably dipping foliation. Metasedimentary rocks south of the complex contain a regional northeast-striking, moderately northwest-dipping foliation ($S_{1(\text{External})}$) which likely correlates with the coplanar regional foliation of the MacQuoid-Gibson supracrustal belt, 50 km southwest. Near MacQuoid Lake, this regional foliation is superimposed on an earlier foliation, is axial planar to northeast-trending folds (F_2) and is cut by ca. 2.19 Ga mafic dykes (Tella et al., 1997a). Near the Kramanituur Complex, the regional northeast-striking foliation is cut by east-striking, north-dipping shear foliations ($S_{2(\text{External})}$) of the Southern shear zone (Fig. 2-2 inset), dated at ca. 2550 Ma, the age of a syntectonic syenogranite pegmatite (Chapter 3). Accordingly, regional northeast-striking fabrics south of the complex are not Paleoproterozoic, but are interpreted to pre-date ca. 2550 Ma. Coplanar fabrics north of the complex may be Archean as well.

The timing of metamorphism for the Kramanituur Complex and its surroundings is well constrained by U-Pb and Sm-Nd data. These data, presented in Chapter 3, establish Paleoproterozoic metamorphism at 1910-1901 Ma throughout the complex and to the north, but indicate an older, partly Neoproterozoic, history to the south.

METHODOLOGY

P-T conditions within and surrounding the Kramanituur Complex were assessed on the basis of detailed petrography, mineral composition and geothermobarometry. Data from fourteen samples are presented: eight of these are from the Kramanituur Complex and include well-equilibrated samples (6) and those preserving high-temperature reaction textures (2); six samples are external to it. Element distribution images were produced at the University of Massachusetts using a Cameca SX50 electron microprobe with four wavelength-dispersive spectrometers, an accelerating voltage set to 15 kV, a beam current of 100 nA, a spot size of 3 μm and a counting time of 50 to 100 ms per pixel

point. These were generated at an early stage of the study and guided our understanding of relative compositional variations prior to quantitative analysis. Mineral analyses were carried out at the Geological Survey of Canada on a Cameca Camabax electron microprobe using a wavelength-dispersive analysis system with a beam current of 10 nA or 25 nA depending on phase and sample size, an accelerating potential of 15 kV and a counting time of 10-20 s. Pure oxides and natural minerals were used as standards in the analysis procedure. Data were reduced with a PAP correction program (Pouchou and Pichoir, 1985). A summary of mineral analyses used in geothermobarometric calculations is given in Tables 2-1, -2, -3, -4. For each sample, at least two to four domains containing the required assemblage of minerals were analysed and used for geothermobarometric calculations to ensure consistency at a thin section scale. The quoted range of P - T values for each sample reflects the variation in the results between different domains within a thin section, as well as the variation resulting from use of different equilibria, discussed below (*see* Table 2-5). It does not take into account probe error or uncertainty in thermodynamic parameters.

Pressures (P) and temperatures (T) were calculated using the TWEEQU (Thermobarometry With Estimation of EQUilibrium state) software of Berman (1991). This program was designed to provide thermobarometric consistency by calculating the location of reaction equilibria in P - T space using internally consistent thermodynamic data. The data used in this study are: end-member phases (Berman, 1988, 1991), ideal solution models for the pyroxenes, and the solution models of Berman (1990) and Fuhrman and Lindsley (1988) for garnet and plagioclase, respectively. Thermodynamic data for amphibole (Amp) end-members and mixing properties were taken from Mader et al. (1994), and are consistent with the thermodynamic database. For a specified set of end-member phases and their chemical compositional data, the TWEEQU program calculates all possible stable and metastable equilibria applicable to that assemblage. In this study we utilized the following independent set of equilibria, selected as most robust on the basis

of reaction properties and calibration systematics. Reactions are written with the higher- T assemblage on the right, using mineral abbreviations of Kretz (1983).

Barometers

- 1) $\text{Grs} + 2 \text{ Ky/Sil} + \text{Qtz} = 3 \text{ An}$
- 2) $\text{Prp} + 2 \text{ Grs} + 3 \text{ Qtz} = 3 \text{ An} + 3 \text{ Di}$
- 3) $2 \text{ Prp} + \text{Grs} + 3 \text{ Qtz} = 3 \text{ An} + 3 \text{ En}$
- 4) $3 \text{ Ts} + 12 \text{ Qtz} + 2 \text{ Prp} + 4 \text{ Grs} = 12 \text{ An} + 3 \text{ Tr}$

Thermometers

- 5) $\text{Alm} + 3 \text{ Di} = \text{Prp} + 3 \text{ Hd}$
- 6) $5 \text{ Alm} + 3 \text{ Tr} = 3 \text{ FeTr} + 5 \text{ Prp}$
- 7) $\text{Alm} = 3 \text{ Fsl} + \text{alOpx}$
- 8) $\text{Alm} + \text{Phl} = \text{Prp} + \text{Ann}$

With the exception of equilibrium #7, all calculations were performed with TWEEQU version 1.02, because only this version includes amphibole components that appear in equilibria #4 and #6. In order to compute equilibrium #7, that includes an Al_2O_3 component in Opx (alOpx), TWEEQU version 2.02 was used. This inconsistency introduced negligible error, because the differences between version 1.02 and 2.02 databases for γ_{Alm} and γ_{Fsl} , and for the quantity $G^{\text{P},\text{T}}(\text{Alm}) - 3G^{\text{P},\text{T}}(\text{Fsl})$ are small. Fe^{3+} values in clinopyroxene and garnet were estimated assuming charge balance according to the method of Neumann (1976), and in hornblende according to the method of Mader et al. (1994), which is based on that of Spear and Kimball. For consistency with TWEEQU version 1.02 calibration of thermodynamic properties for biotite, all FeO in biotite was calculated as Fe^{2+} .

Data for rocks from within the complex are presented first, beginning with well-equilibrated samples from which peak conditions are estimated, followed by those with reaction textures that record high- T decompression. Data from country rocks are presented next: those located north and east of the complex record high- P conditions comparable to those of the complex; while those south of the complex are interpreted to retain

an older, mainly Neoarchean metamorphic history. Implications of these data for the tectonic setting of the Kramanituar Complex and its surroundings are then discussed.

WELL-EQUILIBRATED MAFIC ROCKS AND PARAGNEISS FROM THE KRAMANITUAR COMPLEX

Mafic Rocks

Petrography and Mineral Compositions

Gabbro and gabbroic anorthosite within the Kramanituar Complex consist of plagioclase, orthopyroxene, clinopyroxene $\pm$ pargasite, garnet, quartz and K-feldspar with accessory ilmenite, apatite, magnetite, titanite, rutile and zircon. Orthopyroxene-bearing assemblages (Plate 2-1a) are prevalent in the low-strain interior of the complex, whereas clinopyroxene-garnet-plagioclase assemblages (Plate 2-1b,c) are widespread throughout its higher-strained north, northeast and eastern margins. The latter rocks generally have uniform mineral compositions and petrographic features suggestive of metamorphic textural equilibrium, and are considered most likely to have preserved peak metamorphic conditions. Representative samples analysed include moderately strained gabbro (K-1053), strongly strained gabbro gneiss (K-596) and a mylonitized gabbro (K-104) from the Northern shear zone (Fig. 2-2). The metamorphic assemblage of one of these (K-1053) was dated at 1894 ± 8 Ma using the Sm-Nd system, based on a Grt-Pl-Cpx-whole rock isochron. This is interpreted as a minimum age of high-grade metamorphism within the complex, bracketed by other isotopic data at 1910-1901 Ma (Chapter 3).

Grt-Cpx-Pl rocks contain 10-50% garnet which varies from elongate porphyroblasts that may be highly embayed (Fig. 2-3a; Plate 2-1c) to subhedral or subrounded equant grains (Plate 2-2). Garnet commonly includes irregular to spherical quartz grains, but may also contain inclusions of plagioclase, clinopyroxene $\pm$ titanite and rutile (Plate 2-2b,c). Subhedral garnet crystals containing optically strain-free inclusions (eg., K-596) are interpreted to be pre-tectonic, whereas, elongate embayed garnet (eg., K-104) which

locally enclose quartz ribbons, are interpreted as syn-tectonic. Garnet composition is typically uniform in these rocks at ca. $\text{Alm}_{47-53}\text{Prp}_{25-35}\text{Grs}_{17-21}\text{Sps}_{1.5}$. In *K-596*, higher-Ca ($X_{\text{Grs}} \leq 0.35$) garnet occur within leucocratic, plagioclase-rich bands (Plate 2-2*a,c*) which are themselves compositionally distinct (described below). A small increase in $\text{Fe}/(\text{Fe}+\text{Mg})$ from garnet core to rim occurs in some crystals, although typically this is only where garnet is in direct contact with other mafic phases, such as clinopyroxene (Fig. 2-3*b*).

Clinopyroxene within these rocks is a fabric-forming mineral and occurs as elongate aggregates in low strain rocks and as straight or undulating polycrystalline ribbons in highly strained samples (Plate 2-1*b,c*). It takes two forms. Cloudy, medium brown-green grains have an average diameter of 1 mm, and commonly contain fine Ti-rich exsolution lamellae; they are interpreted as paleoblasts on the basis on their grain size and preserved lamellae. Clear, pale-green equigranular grains, 100-200 μm in size, comprise polycrystalline fabric-forming ribbons and mantles around paleoblasts, and are interpreted as neoblasts. Although these texturally distinct populations have compositions of $\text{Wo}_{41-51}\text{En}_{30-38}\text{Fs}_{14-24}$, placing them in the salite field (Morimoto, 1988), they exhibit consistent compositional variations between paleoblast cores, rims and neoblasts. These involve a progressive decrease in $\text{Al}^{[6]}$ and Ti from cores of paleoblasts (0.1-0.13, 0.02-0.05 cations per formula unit (cfu), respectively), through clear rims, to neoblasts (0.05-0.09, 0.009-0.02 cfu, respectively). A decrease in Na is observed from paleoblast core (0.07-0.1 cfu) to clear rim (0.04-0.06 cfu), however, the Na content of clear rims and clear neoblasts is uniform. Ca and Mg commonly show a slight increase from core (0.75-0.90, 0.56-0.69 cfu) through rim to the neoblasts (0.92-0.94, 0.62-0.74 cfu). Clinopyroxene inclusions in garnet commonly have higher Ti and $\text{Al}^{[6]}$ contents than matrix grains, and are unzoned.

Plagioclase is a major fabric-forming component (Fig. 2-3*a,b*; Plate 2-1*b,c*;) in these rocks, and like clinopyroxene, occurs both as μm -scale, recrystallized matrix subgrains, and as mm-scale paleoblasts which typically are twinned. Together these form

discontinuous aggregates (moderate strain) or continuous mm-wide layers (high strain) that define the penetrative L-S fabric. Most matrix plagioclase throughout these well-equilibrated rocks exhibit a fairly uniform composition of An_{31-38} . Leucocratic, plagioclase-rich layers associated with high-Ca garnet in *K-596* (Plate 2-2) are An_{50-56} . Ternary feldspars comprise one third of the matrix feldspar of mylonitized gabbro *K-104*. These are both antiperthite, with μm -scale stringlets of ca. $An_{25-31}Or_{18-34}$ in a plagioclase host of ca. $An_{31-34}Or_{2-8}$, and perthite with stringlets of $Or_{55-88}An_{2-15}$ in a host of ca. $Or_{83-92}An_{1-5}$. Feldspar inclusions in garnet are essentially the same composition as matrix feldspar (An_{31-38}). In *K-596*, inclusions of An_{57-68} , more calcic than the feldspars in plagioclase-rich layers, also occur in garnet (Plate 2-2c). In one sample (*K-1053*), plagioclase of An_{48-55} occurs as isolated grains and partial moats, $\sim 30 \mu m$ wide, at the interface between garnet and Ti-oxides (Ilm + Mt $\pm$ Tnt $\pm$ Rt). This more calcic plagioclase tends to occur at, but is not restricted to, strain shadow regions, a relationship which appears to be governed by concentration of Ti-oxides at these sites. This feature suggests localized garnet-breakdown according to a reaction that can be represented by the following idealized end-member equilibria:

Two textural varieties of calcic amphibole occur. Mm-scale grains of ferroan pargasite (nomenclature of Leake, 1978) with 2.0-3.4 wt.% TiO_2 , and $0.49 \leq Mg/(Fe+Mg) \leq 0.64$ are in direct contact with the main matrix phases, and are interpreted to be in textural equilibrium with these. A very small amount of magnesio-hastingsite with lower TiO_2 (<2 wt%), and $Mg/(Fe+Mg) > 0.7$ occurs as fringes on clinopyroxene, and is interpreted to be of replacement origin. Both forms are Cl-poor (<0.25 wt% Cl). In *K-596*, a 2-3 mm wide layer consisting of a monominerallic mosaic of amphibole includes 1 mm grains of ferroan pargasite $\pm$ magnesian hastingsite with μm -scale grains of

ferroan pargasitic hornblende. Trace amounts of biotite ($X_{\text{Phl}}=0.6-0.7$) with 1.5-4.0 wt.% TiO_2 , and chlorite occur only rarely in these rocks.

P-T Results for Well-Equilibrated Mafic Rocks

Geothermobarometric results for mafic granulites, described above, show excellent convergence of up to five independent equilibria for different domains of the same thin section, generally within ± 1.2 kbar and $\pm 50^\circ\text{C}$. Results are summarized in Table 2-5 and are shown for K-1053 in Fig. 2-4. *P-T* estimates for garnet rim compositions in contact with compositionally homogeneous matrix plagioclase and adjacent neoblastic clinopyroxene (separated from garnet by either quartz or feldspar) yield 12 ± 1.5 kbar and $810 \pm 70^\circ\text{C}$. *P-T* conditions of 14 kbar and 850°C for garnet growth conditions in sample K-596 are calculated from composite Pl-Cpx $\pm$ Amp inclusions in garnet (Plate 2-2). Near-peak conditions of 12 ± 1 kbar and $870 \pm 10^\circ\text{C}$ are calculated from core compositions of all porphyroblastic phases in K-1053. Localized garnet breakdown in K-1053 occurred at 9 ± 1 kbar and 760°C (Table 2-5), based on compositions of partial calcic plagioclase moats, adjacent garnet rims and closest matrix neoblastic clinopyroxene in the presence of ilmenite.

Paragneiss

Petrography and Mineral Compositions

Sillimanite-bearing paragneiss (K-440) from the northwest part of the complex, and kyanite-bearing paragneiss mylonite (K-276 and K-290) from the northeast part (Fig. 2-2) contain as their modally abundant mineral phases: quartz, K-feldspar, garnet and sillimanite or kyanite, with accessory biotite, plagioclase, rutile, graphite, monazite, zircon and pyrite. Monazite ages of 1910-1908 Ma for these rocks are interpreted to date prograde regional metamorphism ($\sim 550^\circ\text{C}$) of these rocks in advance of intrusion of the 1902 Ma mafic plutonic complex (Chapter 3, p.149-150). Garnet are typically elongate,

1 to 5 mm in length, anhedral and embayed (Fig. 2-3c,d). Garnet from kyanite-bearing rocks contains inclusions of quartz, kyanite and perthite with trace plagioclase and K-feldspar. Inclusions in garnet from sillimanite-bearing rocks are slender prismatic needles of sillimanite, quartz, kyanite? and rutile. Ilmenite is absent.

Garnet in these rocks is typically zoned with similar core compositions. In kyanite-bearing paragneiss, cores are ca. $\text{Alm}_{50-60}\text{Prp}_{31-43}\text{Grs}_{6-11}\text{Sps}_{<2}$ and may show a small (up to 2.5 mol.%) increase in Ca from core to rim (Figures 2-4c). Where Ca zonation is concentric and is truncated by embayments in garnet (e.g., Fig. 2-3c), it is interpreted to reflect primary growth zonation with increasing P . In some places, a steep decrease in Ca (i.e., 5 mol %) occurs where the outer rim of garnet is touching small isolated matrix plagioclase grains (e.g., arrows in Plate 2-3). This is consistent with limited garnet breakdown during decompression via the GASP reaction (#1). Cores of garnet in sillimanite-bearing paragneiss are $\text{Alm}_{48-52}\text{Prp}_{44-46}\text{Grs}_5\text{Sps}_{<1}$ and show a rimward decrease in Ca to $X_{\text{Grs}}=0.024$. A characteristic feature of garnet in both kyanite- and sillimanite-bearing paragneiss is a Fe/(Fe+Mg) gradient, up to 0.05 in magnitude, perpendicular to the long dimension of elongate grains (Fig. 2-3d, Plate 2-3). The absence of other ferromagnesian minerals proximal to the garnet, and their paucity in the paragneisses in general, indicate that retrograde Fe-Mg exchange likely did not influence the distribution of Fe-Mg in these garnets. Rather, garnet may have crystallized within a significant bulk composition gradient which has not been removed by diffusional reequilibration. Rarely where garnet is in contact with matrix biotite, Fe/(Fe+Mg) increases steeply at garnet rims, consistent with localized retrograde diffusion.

Plagioclase is rare in paragneiss, occurring as isolated, 50-100 μm grains. In kyanite-bearing rocks, plagioclase of An_{34-38} occurs adjacent to lower-Ca garnet rims (described above). In sillimanite-bearing rocks, plagioclase grains of ca. An_{24-30} occur throughout the matrix. String perthite consisting of K-feldspar host with up to 20% 1-5 μm wide lamellae of plagioclase form a dominant component of the matrix of kyanite-

bearing samples. These have a re-integrated composition of $\text{Ab}_{20-25}\text{An}_{5-8}\text{Or}_{69-73}$. Ternary feldspars of K-440 have a composition of $\text{Ab}_{60-62}\text{An}_{25-27}\text{Or}_{11-15}$ after reintegration.

Red-brown biotite, with $X_{\text{FeMg}}=0.27$ and up to 5-6 wt.% TiO_2 , comprises < 3% of sillimanite-bearing paragneiss, and occurs in trace amounts in kyanite-bearing rocks. In K-440, some biotite is weakly zoned with decreasing Ti from core (0.6-0.71 cfu) to rim (0.6-0.49 cfu). Other accessory phases include: 2% graphite; slender 200-300 μm long prismatic rutile as inclusions in garnet and throughout the matrix; and clear, pale yellow, subhedral to subrounded monazite, 80-100 μm long, which occurs as a matrix phase of the paragneisses. Rare, subrounded monazite grains, <10 μm , occur within garnet.

P-T Results for Paragneiss

Samples of paragneiss yielded a less well-constrained thermobarometric data set than the mafic rocks described above, with limitations arising both from an absence of relevant inclusion phases within garnet and from the low abundance of an Fe-Mg phase, other than garnet, in these rocks. The latter condition limits the general applicability of Fe-Mg exchange thermometry. The intersection of the GASP equilibrium (#1) for rim compositions with temperature estimates based on closest matrix plagioclase coexisting with reintegrated perthitic feldspar (Fuhrman and Lindsley, 1988) and, in the case of K-440 on Fe-Mg exchange in biotite, yields minimum "peak" *P-T* conditions for the three paragneiss samples between 9.5 to 12 kbar and 780 to 875°C (Table 2-5), similar to estimated peak conditions for adjacent mafic rocks. Garnet and biotite rim compositions in sillimanite-bearing paragneiss yielded the lowest pressure estimates obtained from rocks from the Kramanituur Complex of 5-6 kbar at 685-740°C, consistent with high temperature decompression of this sample to mid-crustal conditions.

SAMPLES WITH REACTION TEXTURES FROM THE COMPLEX

In contrast to the well-equilibrated rocks described above, samples from the central parts of the complex preserve reaction textures that record peak conditions (as above)

and a high- T decompressional path. Data from two samples are presented here: those of a meta-norite with coronitic texture (K -619), and those of a 2-pyroxene gabbro with symplectitic texture (K -1005). The results suggest that near-isothermal decompression of the Kramanituur Complex took place both passively (K -619) and syn-kinematically (K -1005), due to strain partitioning.

Coronite

Petrography and Mineral Compositions

Weakly foliated (090/80°S) orthopyroxene-porphyritic meta-norite (K -619) is exposed in the southeast part of the complex (Fig. 2-2). This rock is interpreted to be comagmatic with gabbroic anorthosite dated at 1902 ± 1.6 Ma (Chapter 3), located 11 km northwest of K -619 (Fig. 2-2), due to the similar composition and strain state of these samples, and their proximity to each other. The meta-norite consists of a plagioclase matrix with phenocrysts of orthopyroxene averaging 3 cm by 1 cm, but occurring up to 13 cm by 6 cm. Layered mono- and polymineralic subdomains, up to 800 μ m wide, occur along the interface between primary orthopyroxene and matrix plagioclase (Fig. 2-5; Plate 2-4a), and contain garnet, clinopyroxene, second generation orthopyroxene and plagioclase, magnetite, quartz $\pm$ biotite $\pm$ K-feldspar. These layered coronitic reaction textures are interpreted to be multistage, initially involving reaction between orthopyroxene and plagioclase to form a metamorphic assemblage of Grt-Cpx-Qtz (Plate 2-4b). The generalized reaction:

has a moderate, positive dP/dT slope (Hansen, 1981) suggesting that garnet growth may have occurred during re-equilibration at lower T , higher P , or both. Subsequent garnet breakdown appears to have initially involved consumption of coronitic clinopyroxene via back-reaction of (#11) which led to the production of a secondary metamorphic assemblage of orthopyroxene and plagioclase of An_{40-51} , 100 μ m wide, at the interface between

clinopyroxene and garnet (Plate 2-4c IIa). Locally, isolated K-feldspar grains, <30 μm in diameter, form part of this secondary assemblage. Continued garnet breakdown via a reaction like equilibrium (#3) is recorded by a compositionally and texturally distinctive moat of plagioclase of An_{50-58} (up to An_{75}) which contains isolated 50 μm grains of orthopyroxene (Plate 2-4c IIb). This latter moat, in contact with "inboard" garnet, is interpreted as the final stage of the preserved reaction series.

Compositional maps of Ca (Fig. 2-5a) and Mg (Fig. 2-5b) illustrate systematic μm -scale variations in mineral compositions and preservation of elemental zonation across this coronitic structure. Variation in plagioclase composition, highlighted by Ca (Fig. 2-5a), extends from compositions of ca. An_{45-54} for magmatic plagioclase distal to garnet (*see* Plate 2-4a), to a 300 μm wide zone of uniform $\text{An}_{42-45}\text{Or}_2$ adjacent to (outboard of) the garnet corona, interpreted as the equilibrium composition during garnet growth. Second generation plagioclase inboard of garnet shows pronounced compositional variations in that earlier-formed grains adjacent to clinopyroxene are An_{40-51} , whereas later-formed plagioclase adjacent to garnet is An_{50-58} (locally up to An_{75}). These chemical variations are interpreted to track P variations during garnet breakdown. Garnet are zoned, with higher Ca cores up to $X_{\text{Grs}} = 0.23-0.25$ and rims of $X_{\text{Grs}} = 0.11-0.12$ (Fig. 2-5a). They show a corresponding decrease in $\text{Fe}/(\text{Fe}+\text{Mg})$ (Fig. 2-5b) from cores of 0.57 to rims of 0.54.

Clinopyroxene composition is uniformly $\text{Wo}_{44-49}\text{En}_{39-42}\text{Fs}_{8-14}$. Second generation orthopyroxene may be less aluminous ($\Sigma\text{Al} < 0.09$ cfu) and slightly more Fe-rich ($\text{Fe}/(\text{Fe}+\text{Mg}) = 0.34-0.36$) than primary magmatic orthopyroxene ($\Sigma\text{Al} > 0.1$ cfu; $\text{Fe}/(\text{Fe}+\text{Mg}) = 0.33-0.34$).

P-T Results for Coronite

Most reliable conditions are those of final garnet breakdown, at 9 ± 0.5 kbar and $780 \pm 40^\circ\text{C}$, as these are based on compositions of inboard garnet rims in contact with

symplectitic plagioclase of $An_{>50}$ and associated isolated grains of orthopyroxene. Conditions of initial garnet formation, estimated at 12.5 ± 1.5 kbar and $850 \pm 40^\circ\text{C}$ for high-Ca garnets (Fig. 2-5a) are based on core compositions of clinopyroxene and garnet, matrix plagioclase that forms a compositionally uniform moat outboard of garnet, and reactant orthopyroxene. These are also considered to be valid, on account of their reproducibility for comparable mineral assemblages throughout the thin section, and the fact that they compare well to estimated peak conditions of well-equilibrated mafic rocks, described above. Appropriate caution must be exercised in deriving P - T conditions intermediate between initial garnet formation and final garnet breakdown, because a phase of the local intermediate equilibrium assemblage has been consumed. Nevertheless, estimated conditions for final garnet formation at 8.75 ± 1.25 kbar and $740 \pm 40^\circ\text{C}$ are based on rim compositions of clinopyroxene and unreacted outboard garnet, with adjacent outboard plagioclase in contact with garnet. Initial garnet breakdown conditions at 8.7 ± 1 kbar and $740 \pm 15^\circ\text{C}$ are based on compositions of unreacted garnet rims with first-formed symplectitic plagioclase of $An_{<50}$ and associated second generation orthopyroxene. In summary, the P - T path recorded by the coronitic structure extends from its initial development at ca. 12.5 kbar and 850°C to the final recorded stage of garnet breakdown at ca. 9 kbar and 780°C , with intermediate conditions that support a path of progressive high- T decompression.

Symplectite

Petrography and Mineral Compositions

Moderately strained 2-pyroxene gabbro (K-1005, Fig. 2-2) with an assemblage of Opx-Cpx-Grt-Amp-Pl contains 10% garnet, set in a matrix of equal parts of clinopyroxene and orthopyroxene with minor amphibole. Symplectitic intergrowths of $Pl + Opx \pm Amp \pm Spl$ (Fig. 2-6) occur preferentially in strain shadows of garnet, and fill fractures within garnet. Localization of the products of garnet-breakdown in strain shadows, and

their asymmetrical disposition about garnet in mylonitized gabbroic anorthosite from the Northern shear zone (Plate 2-5) are interpreted to indicate syn-kinematic garnet breakdown in these rocks.

In sample K-1005, subrounded to elongate porphyroblasts of garnet, up to 3 mm long, have core compositions of ca. $\text{Alm}_{41-43}\text{Prp}_{36-37}\text{Grs}_{18-22}\text{Sps}_1$. Rim compositions along the long edge of garnet, where there is no textural indication of garnet breakdown, show a decrease in Ca and Mg and an increase in Fe to an average composition of $\text{Alm}_{50-55}\text{Prp}_{27-32}\text{Grs}_{15-17}\text{Sps}_2$. Garnet rims at the ends of the grains, proximal to Opx-Pl symplectite are $\text{Alm}_{51-56}\text{Prp}_{24-28}\text{Grs}_{17-20}\text{Sps}_2$, that is, lowest in Mg content, and intermediate in Ca content between cores and unreacted rim compositions.

Plagioclase is texturally and compositionally variable. Matrix plagioclase is An_{50-61} (up to An_{78} in contact with amphibole), but has a more uniform composition of An_{51-53} adjacent to unreacted (symplectite-absent) garnet rims. Systematic compositional variations characterize symplectitic plagioclase, and are well illustrated by the detailed compositional map of Ca (Fig. 2-6a). Symplectitic plagioclase furthest from garnet, and presumed to be first-formed, is An_{80-84} ; whereas that in contact with garnet, presumed to be last-formed, is An_{86-87} .

First generation orthopyroxene ($\text{En}_{58}\text{Fs}_{40}$) occurs as unaltered, evenly distributed, subhedral matrix grains, ≤ 1 mm in size. Second generation orthopyroxene occurs as elongate, 100-300 μm long grains of $\text{En}_{58-62}\text{Fs}_{37-39}$ forming symplectite (Fig. 2-6a). Both have $\text{Al}^{[6]} < 0.07$ cfu and $\text{Ti} < 0.003$ cfu. Clinopyroxene occurs throughout the matrix as 1-3 mm grains with a relatively uniform composition of $\text{Wo}_{44-46}\text{En}_{37-40}\text{Fs}_{16-20}$. Some zonation is observed with paleoblast cores of En_{40} and $\text{Al}^{[6]} = 0.08$; and rims of En_{36-38} and $\text{Al}^{[6]} = 0.05-0.08$. Small clinopyroxene grains (neoblasts?) are less aluminous with $0.04 \leq \text{Al}^{[6]} \leq 0.06$ cfu.

P-T Results for Symplectite

Poor convergence of Cpx-Amp-Grt-Pl-Qtz equilibria (#2,3,4,5,6) based on core compositions for all phases of the 2-pyroxene gabbro (K-1005) suggest peak metamorphic ($\pm$ magmatic?) conditions of 13.5 ± 1.5 kbar and $900 \pm 50^\circ\text{C}$ (Table 2-5). More importantly, Opx + Pl_(An₈₅₋₈₇) symplectite proximal to garnet in this sample yields *P-T* conditions of 8.5 ± 0.5 kbar and $800 \pm 50^\circ\text{C}$, interpreted to represent conditions of syn-kinematic garnet breakdown. Initial garnet breakdown at $P > 10$ kbar is suggested by the lower anorthite component (An₈₀) of symplectitic plagioclase more distal to garnet (Fig. 2-6a).

ROCKS EXTERNAL TO THE KRAMANITUAR COMPLEX

North and east of the Kramanituar Complex, moderately foliated to gneissic intermediate metaplutonic rocks (Fig. 2-2) are generally characterized by garnet-absent assemblages (Qtz-Pl-Bt-Amp-Kfs $\pm$ Ep), consistent with greenschist- to amphibolite-facies conditions. Isolated mafic plutonic rocks within this granitoid terrain locally contain Cpx-Grt in addition to ubiquitous Amp-Pl-Qtz, thereby providing appropriate assemblages for thermobarometry. South of the Kramanituar Complex, data from metasedimentary rocks with Grt-Bt-Pl-Qtz $\pm$ Sil assemblages and garnet-bearing metagabbro place constraints on metamorphic conditions that pre-date those for the complex.

North of the Complex

Petrography and Mineral Compositions

Plagioclase-porphyritic coronitic gabbro (K-691), located 7 km north of the Kramanituar Complex (Fig. 2-2), is hosted by quartzo-feldspathic metaplutonic rocks. Contacts between these rock types were not observed. The gabbro is weakly foliated, and is dominated by a magmatic assemblage of Cpx-Pl-Opx-Mt $\pm$ Bt with metamorphic coronae of amphibole-garnet. The gabbro has a maximum age of 2.2 Ga, its maximum Nd model age, and its Amp-Grt coronae are dated (Sm-Nd) at 1878 ± 27 Ma, a two-point (Grt-Pl)

isochron age (Chapter 3). In these rocks, clinopyroxene occurs as euhedral to subhedral grains, 1-2 mm, that exhibit hour-glass sector zoning produced by variable abundances of minute Ilm/Rt inclusions aligned parallel to three crystallographic orientations. Clinopyroxene shows increasing Fe/(Fe+Mg) from core to rim (0.25-0.50) which may reflect a primary magmatic trend of increasing Fe during crustal growth. Inclusion-rich sectors are $\text{Wo}_{44-49}\text{En}_{35-40}\text{Fs}_{14-18}$ with $\Sigma\text{Al}=0.14$ cfu and $\text{Ti}=0.023$ cfu, whereas sectors that are less included or inclusion-free are $\text{Wo}_{23-32}\text{En}_{37-41}\text{Fs}_{30-36}$ with $\Sigma\text{Al}=0.077$ cfu and $\text{Ti}=0.006$ cfu. Optically clear clinopyroxene rims adjacent to coronitic phases are $\text{Wo}_{40-50}\text{En}_{36-38}\text{Fs}_{12-23}$ with Al and Ti similar to less included sectors. Other accessory phases that are interpreted to be magmatic include strongly pleochroic orthopyroxene of $\text{Wo}_{40-50}\text{En}_{36-38}\text{Fs}_{12-23}$ and biotite with 3.0-6.2 wt.% TiO_2 and $\text{Fe}/(\text{Fe}+\text{Mg})=0.3-0.4$.

Amphibole and garnet form 50-150 μm wide coronae in this rock (Fig. 2-7). Amphibole occurs as discontinuous coronae, 80-100 μm wide, which mantle clinopyroxene and vary in composition from ferroan-pargasitic hornblende to magnesian-hastingsitic hornblende with 0.09-0.19 cfu Ti, $\text{Al}^{[6]}=0.37-0.50$ cfu and $\text{Fe}/(\text{Fe}+\text{Mg})=0.38-0.50$. Euhedral to subhedral garnet forms partial coronae, 50 μm wide, between matrix plagioclase and coronitic amphibole (Fig. 2-7); between matrix plagioclase and clinopyroxene (Fig. 2-7); and between matrix plagioclase and biotite-ilmenite aggregates. Garnet is typically zoned with respect to Ca, with cores of $X_{\text{Grs}}=0.25-0.34$ and rims of $X_{\text{Grs}}=0.18-0.24$, whereas X_{Sps} and $\text{Fe}/(\text{Fe}+\text{Mg})$ are generally constant at 0.02 and 0.80-0.82, respectively. Locally, garnet has elevated $\text{Fe}/(\text{Fe}+\text{Mg})$, up to 0.84, where rims of innermost garnet are proximal to, but not touching, clinopyroxene.

Plagioclase is texturally and compositionally diverse. Equant, primary magmatic phenocrysts, up to 1 cm^2 , comprise 2% of the rock and have a uniform composition of An_{45} , except for <400 μm wide rims of An_{47-48} . Matrix plagioclase occurs as aligned subhedral laths, 1 mm in length, with simple albite twinning, and as equant subgrains, 30-50 μm , defining linear recrystallized subdomains within twinned laths. It varies in com-

position from An₅₁, distal to garnet coronae, to An₃₃₋₄₄ immediately outboard of garnet (Fig. 2-7). Plagioclase of An₂₀₋₂₆ occurs inboard of garnet coronae, forming a discontinuous interface between garnet ± amphibole and clinopyroxene (Fig. 2-7). Optical continuity between outboard and inboard plagioclase is observed at gaps in garnet coronae, indicating that garnet may have overgrown and/or replaced plagioclase of ca. An₂₇₋₃₅ composition. An interpretation of garnet growth by replacement is further supported by plagioclase of An₃₁ near the gap in the garnet corona.

Amphibole-garnet coronae in K-691 appear to represent a single-stage reaction such as the generalized reaction:

This reaction has a steep positive dP/dT slope implying that re-equilibration of the magmatic assemblage under hydrous conditions at lower T could have produced both garnet and amphibole in these clinopyroxene-bearing rocks. Amphibole-absent garnet coronae adjacent to clinopyroxene may have formed via a reaction like equilibrium (#2), which has a moderate to shallow positive slope in P - T space, consistent with Grt-Qtz corona formation at lower T , higher P or both. Alternatively, two-stage development of the coronitic assemblage may have occurred by initial hydration of clinopyroxene to form continuous amphibole coronae, followed by a subsequent dehydration event involving growth of garnet from amphibole and plagioclase represented by the generalized reaction:

P-T Results from North of the Complex

P - T conditions of garnet corona formation in plagioclase-porphyritic gabbro (K-691) north of the complex are estimated at 8.25 ± 0.75 kbar and $700 \pm 50^\circ\text{C}$ (Table 2-5). These results represent an average, based on mineral compositions of the inboard rim of garnet, in contact with inboard plagioclase and adjacent clinopyroxene, and those of the outboard rim of garnet, in contact with outboard plagioclase (Fig. 2-7). P - T conditions of

8.5-12 kbar and $700 \pm 50^\circ\text{C}$ (Table 2-5) for initial garnet formation are calculated from garnet core compositions for the full range of plagioclase compositions (An_{20-43}) presently in contact with garnet.

East of the Complex

Petrography and Mineral Compositions

East of the complex, foliated to gneissic plutonic rocks of dioritic composition contain garnet within 1 km of the eastern limit of the Kramanituvar anorthositic suite (Fig. 2-2 inset, Grt-in isograd). These garnet-bearing rocks typically contain leucocratic aggregates rimmed by mafic minerals, consistent with incipient partial melting. Together these features are interpreted to reflect the effects of contact metamorphism imposed by the Kramanituvar Complex on dioritic country rocks to the east. Metamorphic constraints in this region are provided by garnet-porphyroblastic gabbro (K-272) which occurs within this 1 km wide aureole (Fig. 2-2).

Gabbro K-272 contains up to 35% garnet as spherical porphyroblasts, up to 4 cm across (Plate 2-6), set in an equigranular, foliated matrix of Pl-Di-Amp $\pm$ Qtz with minor Opx $\pm$ Chl. Low strain mineral aggregates within strain shadows of garnet (Plate 2-6) suggest that garnet growth occurred early in the deformational history of this rock. Garnet composition is relatively uniform (ca. $\text{Alm}_{32}\text{Prp}_{42-44}\text{Grs}_{23}\text{Sps}_1$) although where in direct contact with amphibole $\pm$ chlorite, X_{Alm} increases to 0.42-0.44. Clinopyroxene forms a principal component of the matrix and also occurs as inclusions within garnet. Matrix clinopyroxene is a high Al-diopside ($\text{Wo}_{49-52}\text{En}_{42-44}\text{Fs}_{3-8}$) with 5-8.5 wt.% Al_2O_3 and 0.5-0.8 wt.% Na_2O , and decreasing Al_2O_3 from core to rim. Clinopyroxene inclusions have up to 10 wt.% Al_2O_3 . Amphibole occurs throughout the matrix as 100 μm grains of magnesio-hornblende $\pm$ tschermakitic hornblende ($\text{Na}(\text{M4})=0.13-0.34$ cfu) that appear to be in textural equilibrium with other matrix phases. Minor (~2%) patchy tschermakitic hornblende partially replaces clinopyroxene.

Matrix plagioclase in K-272 is typically An_{45-56} , however, more calcic compositions (An_{69-79}) occur randomly throughout the recrystallized matrix and dominate within strain-shadows of garnet porphyroblasts. Because strain shadow regions are less affected by deformation and recrystallization (Plate 2-6), this more calcic plagioclase may represent a preserved early composition. Within garnet, plagioclase occurs commonly as twinned inclusions of $An_{50-55}Or_{\sim 1.5}$ or as symplectitic intergrowths (An_{78-88}) with orthopyroxene adjacent to clinopyroxene inclusions. The latter suggests localized garnet breakdown via reaction #11.

P-T Results from East of the Complex

East of the complex, near-peak equilibration at 13 ± 1 kbar and $790 \pm 35^\circ\text{C}$ is estimated for K-272 from core compositions of all matrix phases with garnet. Similar conditions of 12 kbar and 700°C are derived from compositions of included clinopyroxene and plagioclase within a garnet core. Localized garnet breakdown in the presence of clinopyroxene and quartz to a symplectite of second generation Pl+Opx is estimated to have occurred at 8.5 ± 0.5 kbar and $720 \pm 30^\circ\text{C}$, consistent with near-isothermal decompression of ca. 4 kbar (ca. 15 km) for this sample.

Country Rocks South of the Complex

Metasedimentary Rocks

Petrography and Mineral Compositions

Metasedimentary rocks south of the complex (Fig. 2-2, Plate 2-7) have as their main phases garnet, biotite, quartz, and plagioclase with minor sillimanite and accessory rutile, graphite and monazite. Ilmenite is absent. These rocks contain up to 50% garnet, with an average size of 2 by 3 mm, but up to 1 cm in diameter. The garnet are texturally and chemically distinctive, commonly consisting of poikiloblastic cores mantled by inclusion-free garnet rims (Plate 2-7). Compositional profiles across these garnet are con-

sistent with two-stage prograde growth zonation. Sample K-556 is a biotitic pelite containing the penetrative pre-2550 Ma northeast-striking $S_{1(\text{external})}$ foliation (Plate 2-7a), whereas sample K-386 is a semi-pelite (Plate 2-7b) containing the ca. 2550 Ma east-striking shear foliation of the Southern shear zone (*see* Fig. 3-11c). Garnet porphyroblasts within both samples deflect enveloping fabrics implying these garnets formed at, or prior to, ca. 2550 Ma.

These garnet crystals are dominated by cores which have numerous inclusions of biotite with $X_{\text{Ann}} = 0.299$ to 0.343 , $.225$ -. 300 c.p.u. Ti and $.79$ -. 85 c.p.u. Al^[6], and plagioclase (An_{27-31}). Poikiloblastic garnet cores show relatively uniform compositions, in contrast to 1 mm wide, compositionally variable inclusion-free garnet rims (Plate 2-7a,b). In K-556, average Ca contents of $X_{\text{Grs}} = 0.03$ across garnet cores are elevated to $X_{\text{Grs}} = .046$ -. 05 at the interface between the poikiloblastic and inclusion-free parts of the garnet and at the outermost garnet edge (Plate 2-7a). In K-386, highest Ca compositions, up to $X_{\text{Grs}} = .12$, also occur within the inclusion-free rim, generally close to the interface between poikiloblastic and inclusion-free garnet (Plate 2-7b). Mn content of poikiloblastic cores averages $X_{\text{Sps}} = .0075$ -. 015 , with a decrease of 0.5 - 1.0 mol% locally occurring from poikiloblastic core to poikiloblastic rim. Mn may be uniform or increase slightly (by 0.5 mol% Sps) across inclusion-free rims suggesting that a progressively depleted Mn reservoir was replenished at a late-stage of garnet growth. Mn-enriched rims are commonly attributed to limited diffusion of Mn during resorption of garnet (Yardley, 1977), however, resorption is not evident in this case. In K-556, $\text{Fe}/(\text{Fe}+\text{Mg})$ is nearly constant (0.79 - 0.82) across the garnet, regardless of textural domains, except for a gradual increase to 0.860 - 0.875 , ca. 1 mm from garnet edges where they are in contact with biotite (Plate 2-7a). This increase is interpreted to reflect retrograde Mg-Fe diffusional exchange with matrix biotite. In K-386, $\text{Fe}/(\text{Fe}+\text{Mg})$ shows more variation. It is elevated at the interface between the poikiloblastic and inclusion-free domains (Plate 2-7b), possibly reflecting retrograde diffusion at this stage of garnet growth. Inclusion-free rims show

decreasing $X_{\text{Fe}/(\text{Fe}+\text{Mg})}$ contents, consistent with prograde conditions during final garnet growth. Elevated $\text{Fe}/(\text{Fe}+\text{Mg})$ at the outermost edge of garnet is consistent with a narrow 200 μm wide zone that experienced retrograde diffusional exchange.

Matrix phases of these rocks include quartz, sillimanite, plagioclase of An_{22-26} composition and biotite, which is $X_{\text{Ann}}=0.399$ to 0.404 with higher Ti ($.275-.320$ c.p.u.) and lower Al^[6] ($.735-.777$ c.p.u.) than biotite inclusions within garnet. Staurolite is not recognized in these rocks.

P-T Results for Semi-pelite K-386 South of the Complex

Metasedimentary rocks cut by, or contained within, the ca. 2550 Ma Southern shear zone display fabric relationships with respect to garnet porphyroblasts and compositional variations that are consistent with Archean prograde metamorphism. *P-T* conditions of 6.5 kbar and 700°C are estimated for sample K-386 garnet cores based on compositions of composite Pl-Bt inclusions and adjacent garnet, whereas, inclusion-free garnet rims in contact with matrix plagioclase and quartz yield conditions estimated at 8 ± 1 kbar and 750°C.

Gabbro

Petrography and Mineral Compositions

Foliated gabbro dykes cut metasedimentary rocks south of the complex. These dykes are 3-6 m wide on average and typically have an assemblage of Amp-Pl $\pm$ Cpx $\pm$ Tnt. Locally, they contain irregular plagioclase-rich patches and veins, and coarse (1-2 cm) Cpx and Grt porphyroblasts. Where mafic dykes are boudinaged, leucocratic material may occur at interboudin necks (*see* Plate 1-4d), consistent with migmatization of the host pelitic to semi-pelitic country rocks during ca. 2550 Ma $D_{2(\text{External})}$ shear deformation.

A unit of medium-grained equigranular gabbro, 200 m wide, occurs south of the complex at the interface between quartz diorite orthogneiss and metasedimentary rocks (Fig. 2-2). Samples from this unit (K-704, K-1057) are garnet-bearing with 15-35% amphibole, 8-15% quartz, 5-20% plagioclase, up to 25% clinopyroxene, and contain minor ilmenite and titanite. A U-Pb (titanite) upper intercept age of ca. 2326 Ma for sample K-704 indicates that this unit retains a metamorphic record older than the ca. 1.9 Ga Kramanitar Complex (Chapter 3). These equigranular (1-2 mm) mafic rocks are characterized by an interlocking assemblage of calcic amphibole which varies in $(\text{Na}+\text{K})_{\text{A}}$ to include magnesio- and ferro-hornblende, ferroan pargasite and magnesian hastingsitic hornblende; garnet with an average composition of $\text{Alm}_{55-60}\text{Grs}_{25-30}\text{Prp}_{10-14}\text{Sps}_{1.5-2}$; plagioclase (An_{48-54}); quartz; $\pm$ clinopyroxene ($\text{Wo}_{46-47}\text{En}_{26-27}\text{Fs}_{26-27}$). Microprobe data for rare garnet porphyroblasts, up to 7 mm across, indicate they are zoned from core to rim with increasing X_{Grs} , decreasing $X_{(\text{Fe}/\text{Fe}+\text{Mg})}$ and decreasing X_{Sps} (Fig. 2-8), suggesting garnet porphyroblast growth under prograde conditions involving an increase in T and P . Narrow selvages of plagioclase of composition An_{56-59} which occur at the interface between matrix amphibole and garnet may have formed at an early stage in the decompression path for these rocks. A later stage of the decompression is recorded by plagioclase of An_{83-86} composition forming symplectite with magnesian-hastingsitic hornblende adjacent to garnet, and plagioclase of An_{81-84} forming symplectite with clinopyroxene adjacent to ilmenite-titanite.

P-T Results for Gabbro from South of the Complex

Geothermobarometric results from gabbro south of the complex, at the interface between quartz diorite gneiss and metasedimentary country rocks, are consistent with prograde metamorphism which, in this case, is older than the U-Pb titanite age of 2326 Ma for this unit. Garnet rim and near-rim compositions and adjacent Pl-Cpx-Amp-Qtz yield estimated P - T conditions of 12 ± 2 kbar and $870 \pm 100^\circ\text{C}$. Growth of garnet cores

appears to have initiated at lower-grade conditions than the rims because garnet cores have lower X_{GrS} and higher $X_{\text{Fe}/(\text{Fe}+\text{Mg})}$. Localized garnet breakdown to Am-Pl and Am-Cpx symplectite took place at 7 ± 1 kbar and $700 \pm 30^\circ\text{C}$.

DISCUSSION

Petrographic and geothermobarometric data from the Kramanituar Complex and its surroundings, presented here, provide the first detailed quantitative P - T constraints on high-grade Paleoproterozoic tectonometamorphism within the Snowbird tectonic zone. Data from rocks south of the complex document older, partly Neorchean, metamorphic conditions.

High- P Granulite-facies Tectonometamorphism at 1.9 Ga

Kramanituar Complex

The dominant component of the Kramanituar Complex is its mafic metaplutonic suite from which near-peak conditions of 12 ± 1.5 kbar and $810 \pm 70^\circ\text{C}$ are estimated from well-equilibrated, extensively recrystallized Cpx-Grt-Pl-Qtz $\pm$ Prg metamorphic mineral assemblages. Mafic rocks still preserving magmatic textures are dominated by orthopyroxene and plagioclase, indicating that the generalized reaction:

was important in the metamorphic transformation of these rocks. This reaction has a moderate positive slope (Hansen, 1981) such that garnet is predicted to form during increasing P or decreasing T . However in these mafic granulites, variations in mineral composition from garnet core to rim are generally consistent with an estimated 1-2 kbar decrease in P , with only a small decrease in temperature during garnet growth (Fig. 2-9a; Table 2-5). Garnet growth during decompression and slight cooling suggests that the garnet-forming reaction was overstepped during prograde metamorphism, or that some mafic magmas crystallized within the stability field of garnet (Green and Ringwood,

1967). Trace element data, presented in Chapter 4, supports limited garnet fractionation during crystallization of gabbroic anorthosite, consistent with fractional crystallization of mantle-derived melts at deep-crustal conditions corresponding to 12-15 kbar.

Paragneiss samples within the Kramanituur Complex yield metamorphic conditions of 11 ± 1.5 kbar and $810 \pm 30^\circ\text{C}$, similar to peak conditions for adjacent mafic rocks (Fig. 2-9a,b). Given the absence of robust thermometers in these rocks and the potential for diffusional re-equilibration, these likely represent minimum conditions only. Garnet in paragneiss appear to preserve subtle growth zoning, defined by an increase in Ca from core to rim (Fig. 2-3c; Plate 2-3). This is interpreted to reflect increasing pressure during prograde metamorphism, dated at ca. 1910 Ma, and possibly as early as 1917 Ma, by monazite in these rocks (Chapter 3). Few other examples of preserved growth zoning in high-grade garnet are known (eg., Indares, 1995; Chen et al., 1998), because during granulite-facies metamorphism (750-900°C or higher) growth zoning in garnet is usually erased by volume diffusion. As such, little information about the prograde P - T path prior to such high-temperature conditions is generally retained. Experiments have shown that garnet will usually be homogenized between 600-750°C, depending on the duration of high- T conditions and grain size (Spear, 1991). The presence of growth zoning in garnets from paragneiss suggests that the residence time of the garnet-bearing assemblage at peak granulite conditions was very short, and that these rocks cooled rapidly shortly after peak metamorphism. Exhumation to mid-crustal levels is recorded by sillimanite-bearing paragneiss (K-440) whose high- T decompressional history to 5-6 kbar and 700-740°C is also reflected by retrograde replacement of peak kyanite by sillimanite.

Coronitic and symplectitic reaction textures within mafic granulites from the Kramanituur Complex record maximum P - T conditions of 13 ± 2 kbar and $880 \pm 70^\circ\text{C}$, consistent with near-peak conditions for well-equilibrated mafic rocks from the complex (Fig. 2-9a,c), followed by high- T decompression ($800 \pm 50^\circ\text{C}$) to 8 ± 1 kbar. Localization of Opx-Pl symplectites within strain shadows in some samples is interpreted to reflect

syn-kinematic decompression of these rocks, while other samples (i.e., K-619) are interpreted to record conditions during passive exhumation, presumably due to strain partitioning.

Prograde metamorphism of the complex is bracketed between ca. 1910 Ma, the U-Pb age of concordant monazite grains from sillimanite- and kyanite-bearing paragneiss, and 1902 ± 1.6 Ma, the time of intrusion and granulite-facies recrystallization of the Kramanituvar gabbroic anorthosite suite (Chapter 3). Exhumation was initiated prior to 1901 Ma, the age at which which most rocks had cooled below ca. 400°C , the closure temperature of rutile (Mezger et al., 1991). The time interval of only ca. 1 m.y. between magmatic crystallization of gabbroic anorthosite and its cooling through ca. 400°C , and the high- T decompression history of garnet in these mafic granulites, are consistent with intrusion of mafic magmas at deep-crustal levels (i.e., $P=12\text{--}15$ kbar) and subsequent garnet growth during rapid exhumation. If a large overstepping of the equilibrium boundary occurred, the driving force for the garnet-forming reaction would have been large. Consequently reaction rates, possibly catalysed by penetrative deformation (Rutter and Brodie, 1995), would have been relatively fast.

High-P Country Rocks

Mafic rocks north and east of the complex recorded conditions of 9-14 kbar and $650\text{--}825^{\circ}\text{C}$ during garnet growth. These pressure estimates are comparable to those experienced by the Kramanituvar Complex, indicating that deep-crustal mafic rocks occur beyond its boundaries. The timing of high- P metamorphism north of the complex is bracketed between ca. 2.2 Ga, the maximum Nd model age for gabbro K-691, and 1878 ± 27 Ma the isochron age of a Grt-Pl Sm-Nd isochron for this rock (Chapter 3), and is likely coeval with that of the ca. 1.9 Ga complex. Conditions during garnet formation in coronitic gabbro, 7 km north of the complex, are consistent with a depth-averaged geothermal gradient of $23^{\circ}\text{C km}^{-1}$, based on the P - T estimates of 8.25 kbar (ca. 30 km) and

700°C for Grt-Amp coronae formation. This supports an interpretation of a regional metamorphic event at ca. 1.9 Ga, however, more regional-scale data is required before the relative importance of magmatic heat and regional metamorphism is understood. A temperature differential of 100-200°C between the Kramanituur Complex, where peak T of $840 \pm 100^\circ\text{C}$ are estimated from mafic granulites, and high- P country rocks to the north is interpreted to reflect the additive influence of magmatic heat to the thermal budget of the Kramanituur Complex.

Although only isolated examples of deep-crustal country rocks were examined north of the complex during this study, the regional extent of high- P country rocks at, or near, the present-day erosion level may correlate with a broad, 300 km (E-W) by 100 km (N-S), positive Bouguer gravity anomaly (+60 mGal) that coincides with the Chesterfield segment of the Snowbird tectonic zone (Goodacre et al., 1987b; *see* Fig. 1-1b). Investigations into the P - T - t history of rocks from this region will further constrain the extent of Paleoproterozoic metamorphism, and the relative importance of regional and magmatic heat.

Older Tectonometamorphic Constraints

Supracrustal and gabbroic rocks south of the Kramanituur Complex preserve a tectonometamorphic history older than that of the complex and rocks to its north, described above. Pelitic to semi-pelitic rocks are interpreted to preserve a two-stage prograde metamorphic record that culminated in upper amphibolite conditions of 8 ± 1 kbar and 750°C (Fig. 2-9d). Prograde metamorphism took place prior to, and likely during, ca. 2550 Ma, the time of fabric development within the Southern shear zone, as indicated by localization of leucosome in the necks of D_2 deformed gabbro boudins. Similarly, a prograde metamorphic history is recorded by a gabbro unit south of the complex where higher grade conditions of 12 kbar and 870°C have a minimum age of ca. 2326 Ma, the U-Pb titanite ($T_c \sim 650$ -750°C) age for this unit (Fig. 2-9d). It is not known whether the

granulite-facies metamorphic assemblage of the gabbro reflects: 1) Neoarchean metamorphism as recorded by adjacent supracrustal rocks; 2) conditions of mafic magmatic crystallization and cooling; or 3) a separate ca. 2.3 Ga high-grade metamorphic event, as suggested elsewhere in the western Churchill Province (Crocker et al., 1993; Bickford et al., 1994; M. Williams, pers. comm., 1999). Additional analytical work on this and other samples from south of the complex is required to further investigate these three possibilities. Regardless, a conclusion that can be drawn from the available data is that neither supracrustal rocks or metaplutonic rocks south of the Kramanituar Complex appear to have been penetratively affected by ca. 1.9 Ga high-grade tectonometamorphism, as recorded by the complex. Post-tectonic monazite dated at 1901 Ma in semi-pelite (K-386) is interpreted to have formed during intermediate-grade fluid fluxing (*see* Chapter 3), likely related to exhumation of the Kramanituar Complex at 1.9 Ga. Fluid-assisted isotope transport is thought by some workers to be more effective than thermally induced diffusion (Hansen and Friderichsen, 1989; Villa, 1998), and may account for: 1) new monazite growth; 2) up to 40% Pb loss in zircon from volcanic rocks south of the complex; and 3) a general pattern of Pb loss from minerals in rocks of the western Churchill Province, as indicated by widespread lower intercept ages of ca. 1.9 Ga (W. Davis, personal communication, 1999).

Preservation of metamorphic assemblages, mineral compositions and fabrics dated at ≥ 2550 Ma and > 2326 Ma in country rocks south of the Kramanituar Complex requires that a tectonic break separates the country rocks from the complex. This break may represent a south-dipping normal fault across which structurally higher, colder Archean hangingwall rocks were juxtaposed against hot deep-crustal footwall rocks of the complex. Alternatively, south-vergent thrusting of the Kramanituar Complex onto country rocks to the south could have led to exhumation of the complex, if assisted by tectonic unroofing. However, this second scenario seems unlikely, because an inverted metamorphic gradient (i.e., of ca. 1.9 Ga age) is not recorded by the country rocks on which hot,

deep-crustal rocks would have been thrust. Fabrics related to a Paleoproterozoic detachment fault were not recognized south of the Kramanituar gabbroic suite and therefore the exact location of a break is not known. However, a break could coincide with the north-west-trending, 400 m wide northern channel of Chesterfield Inlet (Fig. 2-2) across which differences in rock density, crustal depth and conductivity are interpreted, based on preliminary magnetotelluric-teleseismic data (D. Snyder and A.G. Jones, personal communication, 1999). These differences are consistent with a steep or shallow south-dipping discontinuity, which may correspond to a cryptic Paleoproterozoic detachment fault.

Geodynamic Implications

England and Richardson (1977) and Thompson and England (1984) have demonstrated that a clockwise P - T - t loop is a likely consequence of crustal thickening during collision followed by uplift and erosion. Estimated peak pressures reported here require that this part of the interior of the western Churchill Province attained crustal thicknesses in excess of ca. 45 km. The preserved assemblages, the textural record of decompression, quantitative P - T constraints on near-isothermal decompression, and the requirement of thick crust provide speculative, yet compelling evidence of a clockwise P - T path which may be explained by collision tectonics followed by uplift and collapse of the thickened crust. Although constraints on the geodynamic setting are uncertain, the complex may represent a deep crustal level of the hinterland to the ca. 1.97 Ga Thelon-Taltson Orogen, for which granulite-facies conditions persisted in northeast Alberta until 1933-1913 Ma (Grover et al., 1997). Exhumation during extension may have been initiated by instability of an overthickened, upper plate hinterland to the orogen.

CONCLUSIONS

Metamorphic, petrologic and isotopic data indicate that the Kramanituar Complex represents part of a deep-crustal block that was exhumed at ca. 1.9 Ga, under conditions

that largely inhibited re-equilibration and thermal relaxation. The complex is distinct because of its predominantly mafic metaplutonic character, which includes gabbroic anorthosite; its related magnetic and gravity expression; its high pressures corresponding to deep-crustal conditions of 12 ± 1.5 kbar; and, its high temperatures (750-950°C) attained by the addition of magmatic heat to a thick crustal section, and preserved due to the anhydrous character of these rocks and their rapid exhumation to a level where tectonic unroofing led to very rapid cooling.

In contrast to many other high-*P* rocks worldwide, mafic rocks within the Kramanituar Complex record a single-cycle tectonic history. Rather than remain buried in the lower to middle crust, these rocks crystallized, were metamorphosed to high-grade conditions at ca. 1910-1901 Ma, and were then exhumed, cooled, and unconformably overlain by coarse clastic rocks before ca. 1832 Ma. The heterogeneous crustal succession exposed today consists of the Paleoproterozoic high-*P* granulite-facies Kramanituar Complex in contact with Archean-age upper amphibolite-facies mid-crust to the south. This juxtaposition is considered to have been caused by extensional faulting that led to the rapid exhumation of these deep-crustal rocks.

Table 2-1. Selected mineral analyses from well-equilibrated mafic metaplutonic rocks, Kramanituar Complex.

TWQ/CMP	GARNET														
	104m1e	m2c	m1g	m4b	1053mbb	m1a	m3c	596m1j	1k3	m1e	m2b	m2c			
Sample no:	104	104	104	104	1053	1053	1053	596	596	596	596	596			
Comment:	C	R	C	R	C	C	R @ moat	C	C-R	R	C	R			
Analysis:	2-1	2-3	4-2	4-3	g14-6b	g1-1	g3-3	1-11	1-9b	g1-4	g10-1	g10-2			
SiO2	38.29	38.05	37.81	37.42	38.69	38.99	38.01	38.22	38.15	38.59	38.15	38.57	37.84		
Al2O3	21.71	21.76	21.72	21.34	20.86	20.61	21.10	20.83	21.21	21.57	21.12	21.08	21.45		
TiO2	.21	.22	.21	.18	.17	.08	.15	.16	.22	.21	.21	.21	.29		
Cr2O3	.00	.00	.00	.00	.00	.00	.00	.03	.00	.04	.00	.06	.00		
MgO	9.37	9.28	9.36	8.06	6.07	5.82	5.45	5.56	5.68	5.40	5.22	7.30	6.38		
FeO	23.23	23.46	22.92	24.94	25.28	25.04	25.55	26.33	21.10	22.22	21.46	24.77	25.64		
MnO	.58	.66	.54	.74	.52	.60	.74	.75	.84	.92	.51	.69	.64		
CaO	6.57	6.23	6.49	6.27	7.49	7.63	7.13	7.26	12.40	11.92	12.28	7.09	7.19		
K2O	.02	.02	.01	.01	.03	.02	.00	.01	.01	.02	.00	.00	.00		
Total	99.98	99.68	99.09	99.00	99.11	98.79	98.13	99.15	99.65	100.89	98.95	99.77	98.90		
Fe/(Fe+Mg)	.582	.586	.579	.634	.700	.707	.724	.727	.676	.698	.697	.656	.693		
Alm	47.46	48.22	47.29	51.86	54.69	54.69	56.58	56.88	44.01	46.24	45.64	52.09	54.71		
Prp	34.13	34.01	34.44	29.89	23.41	22.64	21.52	21.40	21.10	20.03	19.80	27.35	24.27		
Grs	17.19	16.40	17.15	16.70	20.76	21.34	20.25	20.09	33.12	31.79	33.46	19.11	19.65		
Sps	1.21	1.37	1.13	1.55	1.14	1.33	1.65	1.64	1.76	1.94	1.10	1.46	1.38		

TWQ/CMP	PLAGIOCLASE														
	104m1e	m2c	m1g	m4b	1053mbb	m1a	m3c	m3f	596m1j	1k3	m1e	m2bb	m2b	m2c	
Sample no:	104	104	104	104	1053	1053	1053	1053	596	596	596	596	596	596	596
Comment:	I @ Grt R	M	I	M				moat	C I	C-R I	R	ss	C-R I	C-R I	
Analysis:	2-8	2-4	4-3	4-5	p14-5	p1-2	3-7	3-8	1-24	1-21	p1-5	1-18	10-9	10-10	p10-1
SiO2	59.44	59.31	60.67	59.60	59.47	59.65	58.81	54.91	52.69	53.40	53.22	49.01	59.16	61.19	57.75
Al2O3	24.51	24.58	24.28	24.96	25.37	25.33	25.28	27.71	30.35	29.90	28.72	33.40	26.00	24.87	25.86
Na2O	7.61	7.80	7.57	7.28	7.11	7.06	7.26	5.56	3.99	4.39	4.85	2.41	6.79	7.54	7.08
CaO	6.41	6.73	6.15	6.74	7.49	7.41	7.63	10.65	13.60	12.66	11.56	16.39	8.16	6.89	8.02
K2O	.43	.15	.70	.33	.17	.18	.21	.10	.05	.10	.07	.06	.03	.07	.02
FeO	.25	.18	.21	.17	.25	.09	.27	.54	.21	.21	.10	.08	.19	.03	.11
Total	98.65	98.75	99.58	99.08	99.88	99.72	99.46	99.47	100.89	100.66	98.52	101.35	100.33	100.59	98.84
Ab	66.56	67.17	66.25	64.86	62.53	62.61	62.51	48.31	34.57	38.32	43.00	20.92	59.96	66.15	61.44
An	30.97	31.99	29.73	33.20	36.43	36.31	36.33	51.12	65.18	61.11	56.63	78.75	39.83	33.44	38.46
Or	2.47	.84	4.01	1.94	.99	1.08	1.16	.57	.25	.58	.37	.32	.20	.41	.10

TWQ/CMP	PYROXENE														
	104m1e	m2c	m1g	m4b	1053mbb	m1a	m3c	596m1j	m1k	m1e	m2b	m2c			
Sample no:	104	104	104	104	1053	1053	1053	596	596	596	596	596			
Comments:	C,P	R,P	M	R,I	N	C	C	M subgr	I	C-R I	R	I	R, P		
Analysis:	1-1	2-4	c2-2b	4-2	4-4a	c14-5b	1-2	c3-5	1-3	1-10	c1-3	c10-1	c10-2		
SiO2	48.51	51.51	50.13	48.35	49.47	49.89	49.13	50.86	47.17	48.07	47.75	50.33	50.45		
Al2O3	6.65	3.21	4.91	6.18	5.03	3.93	3.36	3.99	7.84	7.27	6.39	4.70	4.09		
MgO	10.68	13.21	11.75	11.52	11.90	11.57	11.76	11.03	10.11	10.41	10.29	12.46	12.28		
TiO2	.99	.41	.72	1.23	.66	.44	.30	.40	1.09	.87	.98	.77	.67		
FeO	10.44	9.05	8.33	8.95	9.26	11.29	11.20	11.13	9.49	9.37	10.09	8.55	9.10		
CaO	20.89	22.85	20.94	20.76	21.81	21.35	21.59	21.57	22.43	23.04	22.61	21.59	21.91		
Na2O	1.31	0.62	.98	1.60	.88	.69	.52	.55	1.03	.74	1.05	.96	.87		
MnO					.17	na	.13				.02	.05	.15		
TOTAL	99.47	100.86	99.13	98.59	99.01	99.38	100.56	99.66	99.16	99.77	99.16	99.57	99.71		
Fe/(Fe+Mg)	.354	.278	.311	.303	.304	.363	.348	.361	.345	.336	.345	.278	.293		
Fe3+	.09	.08	.07	.15	.09	.08	.05	.06	.13	.09	.13	.07	.08		
Wo	47.61	47.32	47.82	47.45	47.84	47.34	48.98	47.32	51.06	51.38	51.06	49.32	49.60		
En	33.84	38.05	37.34	36.61	36.31	33.75	37.13	33.66	32.05	32.28	32.05	39.55	38.70		
Fs	18.55	14.63	14.84	15.95	15.85	18.90	13.89	19.02	16.89	16.34	16.89	11.13	11.69		

TWQ/CMP	AMPHIBOLE						
	104m1e	m2c	m4b	1053mbb	m1a	596m1k	m2c
Sample no:	104	104	104	1053	1053	596	596
Comments:	M	M	M	C	C	I	M
Analysis:	2-1	2-3b	4-1	14-3b	1-2	1-1	10-3
SiO2	41.00	40.88	41.03	40.59	39.89	38.52	45.32
Al2O3	11.54	11.44	11.05	12.90	12.68	14.38	10.57
TiO2	2.70	3.23	2.73	2.17	1.84	2.39	.90
CR2O3	.00	.00	.00	.01	.07	.00	.02
FeO	13.48	13.87	13.50	17.25	17.39	16.70	13.80
MNO	.01	.09	.03	.04	.11	.10	.16
MGO	11.97	11.59	12.05	9.12	9.64	8.76	12.41
CAO	12.05	12.12	12.07	11.29	11.35	11.98	11.60
NA2O	1.39	1.43	1.41	2.40	2.31	1.24	1.33
K2O	2.08	2.24	1.92	.73	.83	2.61	.26
Total	96.21	96.89	95.79	96.77	96.61	96.80	96.93

Composition expressed as wt.% oxides.
C-core, R-rim, M-matrix, P-paleoblast, N-neoblast, I-inclusion, ss-strain shadow.
Fe3+ calculated assuming charge balance according to Mader et al. (1994).

Table 2-2. Selected mineral analyses from paragneiss, Kramanituur Complex.

TWQ/CMP	440m5d	m5b	m5c	m1c	m2b	GARNET		m3cc	m1a	m2a
						276m3c	276			
Sample no:	440	440	440	440	440	276	276	276	290	290
Comment:	C,M	C	R	C	R	C	inner R	R	C	R
Analysis:	5-2	5-6	5-4	6-2	6-1	g3-6	g3-1	3-8	g1-8	1-2
SiO ₂	38.41	39.03	38.70	38.77	39.18	38.76	38.91	39.06	39.61	38.88
Al ₂ O ₃	22.23	22.13	21.74	22.27	22.09	21.99	21.70	22.36	22.11	22.65
TiO ₂	.09	.05	.05	.10	.09	.07	.09	.04	.11	.12
Cr ₂ O ₃	.05	.04	.07	.06	.07	.00	.03	.07	.03	.06
MgO	11.76	12.37	11.28	12.44	12.37	9.31	9.10	10.66	10.71	11.63
FeO	24.21	23.01	25.40	22.70	23.77	24.18	24.04	24.80	23.40	23.56
MnO	.30	.21	.31	.23	.26	.83	.91	.85	.25	.28
CaO	1.68	1.84	.91	1.82	.90	4.18	4.12	2.35	2.38	3.33
K ₂ O	.01	.02	.01	.01	.02	.00	.00	.03	.03	.00
Total	98.74	98.70	98.52	98.40	98.75	99.32	98.90	100.22	98.63	100.51
Fe/(Fe+Mg)	.536	.511	.558	.506	.519	.593	.597	.566	.551	.551
Alm	50.84	48.30	54.05	47.86	50.31	51.47	51.75	52.01	51.10	48.25
Prp	44.01	46.29	42.79	46.74	46.68	35.33	34.90	39.86	41.68	42.44
Grs	4.51	4.96	2.49	4.91	2.45	11.42	11.37	6.31	6.67	8.72
Sps	.64	.44	.68	.49	.56	1.78	1.98	1.81	.55	.58

TWQ/CMP	440m5d	m5c	m2b&m1c	PLAGIOCLASE		290m1a	m2a
				276m3bb/m3c	m3cc		
Sample no:	440	440	440	276	276	290	290
Analysis:	5-4	5-1	6-4	3-1b	3-2b	1-3	2-2
Comment:	M	R	M	R	R	M	M
SiO ₂	60.61	60.27	59.29	58.68	59.32	58.87	59.96
Al ₂ O ₃	23.66	24.83	24.42	25.63	25.81	25.76	25.21
Na ₂ O	8.26	8.05	8.13	7.00	7.34	7.59	7.32
CaO	5.76	6.34	6.14	7.16	7.42	7.64	6.94
K ₂ O	.22	.15	.14	.16	.16	.08	.19
FeO	.03	.24	.05	.33	.26	.19	.10
Total	98.54	99.88	98.17	98.96	100.31	100.13	99.72
Ab	71.28	69.09	69.99	63.28	63.63	63.97	64.88
An	27.49	30.07	29.21	35.76	35.49	35.59	34.02
Or	1.23	.84	.81	.96	.88	.44	1.10

TWQ/CMP	440m5d	m5b	m5c	m1c	m2b	290
Sample no:	440	440	440	440	440	290
Analysis no:	5-1	5-4	5-5	6-3	6-4	1-4
Comments:	C,M	C	R	C	R	only grain
SiO ₂	37.22	36.72	36.42	36.65	37.44	39.52
Al ₂ O ₃	15.52	15.24	15.89	15.36	15.61	13.10
TiO ₂	5.57	5.59	5.64	6.34	4.36	3.53
Cr ₂ O ₃	.07	.05	.06	.08	.07	.02
FeO	9.93	9.97	9.84	10.55	9.46	4.76
MnO	.00	.01	.00	.00	.00	.00
MgO	15.90	16.06	16.27	15.44	16.74	22.85
CaO	.00	.00	.02	.00	.00	.01
Na ₂ O	.11	.12	.12	.10	.12	.03
K ₂ O	9.74	10.00	9.56	10.04	9.85	10.24
Total	94.06	93.80	93.80	94.57	93.73	94.06
Fe/(Fe+Mg)	.260	.258	.253	.220	.248	.105

Table 2-3. Selected mineral analyses from coronitic and symplectitic mafic samples, Kramanituur Complex.

TWQ/CMP	GARNET						
	619m9a	m9b	m9c	m9d	1005m4c	m3f	m21
Sample no:	619	619	619	619	1005	1005	1005
Comment:	C	outb. R	outb. R	inb. R	C	R	R
Analysis:	g1-13	g1-10	g1-14	g1-18	2-6	2-3	g1-2
SiO ₂	38.81	38.79	39.18	37.92	39.75	39.15	38.78
Al ₂ O ₃	22.19	22.52	22.18	21.89	22.50	22.41	21.43
TiO ₂	.10	.07	.01	.07	.08	.07	.14
Cr ₂ O ₃	.05	.02	.01	.11	.04	.04	.01
MgO	9.03	10.97	10.74	8.71	9.84	8.84	7.04
FeO	21.00	22.48	23.01	24.76	20.37	22.39	23.50
MnO	.51	.47	.52	.70	.42	.46	.83
CaO	8.65	4.11	4.40	5.22	8.26	7.81	6.51
K ₂ O	.00	.04	.01	.00	.00	.00	.01
Total	100.34	99.47	100.06	99.38	101.26	100.90	98.25
Fe/(Fe+Mg)	.566	.535	.546	.615	.537	.587	.652
Fe ³⁺							
Alm	43.13	47.05	47.62	51.96	41.65	46.03	51.95
Prp	33.07	40.93	39.62	32.56	35.85	32.43	27.73
Grs	22.75	11.02	11.67	14.02	21.63	20.59	18.44
Sps	1.05	1.00	1.09	1.47	.87	.96	1.88

TWQ/CMP	PLAGIOCLASE						
	619m9a	m9b	m9c	m9d	1005m4c	m3f	m21
Sample no:	619	619	619	619	1005	1005	1005
Analysis:	p1-10	p1-9	p1-18	p1-16	2-1	2-11	p1-2
Comment:	outb. M	outb. M in cnt w/Gr	inb. in cnt w/Cpx	inb. Grn medial			
SiO ₂	57.18	56.36	56.17	54.24	54.58	46.42	47.58
Al ₂ O ₃	27.54	27.97	27.97	29.25	28.68	33.61	32.61
Na ₂ O	6.28	6.10	6.18	5.27	5.00	1.44	2.02
CaO	9.18	9.56	9.65	11.29	11.64	18.20	16.24
K ₂ O	.31	.14	.11	.14	.17	.00	.02
FeO	.09	.24	.46	.18	.05	.43	.26
Total	100.58	100.37	100.54	100.37	100.12	100.10	98.73
Ab	54.36	53.15	53.33	45.40	43.31	12.55	18.37
An	43.86	46.07	46.03	53.79	55.75	87.45	81.50
Or	1.77	.78	.65	.81	.94	.00	.13

TWQ/CMP	PYROXENE						
	619m9a	m9a,9b	m9b	m9c	m9c	m9d	1005m4c
Sample no:	619	619	619	619	619	619	1005
Comments:	C, P	R, P	M	R, I	M	R, I	C Cpx
Analysis:	c1-1	o1-9	c1-4	c1-2	o1-7	o1-4	2-4
SiO ₂	50.50	51.20	50.76	50.08	51.73	50.71	49.74
Al ₂ O ₃	4.06	2.15	3.17	3.33	1.54	2.90	4.49
MgO	13.00	23.30	13.73	13.50	22.65	22.53	12.58
TiO ₂	.36	.15	.72	.44	.13	.07	.60
FeO	7.84	21.06	8.12	7.92	22.19	21.51	10.28
CaO	21.99	.50	21.92	21.98	.41	.43	21.59
Na ₂ O	.76	0.00	.62	.63	.00	.02	.63
MnO	.12	.25	.11	.11	.24	.15	na
TOTAL	99.11	99.01	99.13	98.64	99.28	98.60	99.91
Fe/(Fe+Mg)	.253	.336	.249	.247	.355	.303	.315
Fe ³⁺	.07	.07	.08	.10	.04	.05	.07
Wo	49.33	1.05	48.19	49.31	.83	47.45	45.82
En	40.60	67.85	41.99	42.13	65.41	36.61	37.14
Fs	10.07	31.10	9.82	8.56	33.76	15.95	17.04

TWQ/CMP	AMPHIBOLE	
	1005m4c	
Sample no:	1005	
Comments:	C	
Analysis:	2-1	
SiO ₂	41.78	
Al ₂ O ₃	13.41	
TiO ₂	1.93	
Cr ₂ O ₃	.09	
FeO	14.13	
MnO	.12	
MgO	11.83	
CaO	11.37	
Na ₂ O	2.26	
K ₂ O	.77	
Total	97.69	

Compositions expressed as wt.% oxides.

C-core, R-rim, M-matrix, P-paleoblast, N-neoblast, I-inclusion, outb.-outboard, inb.-inboard.

Fe³⁺ estimated by charge balance according to Mader et al. (1994).

Table 2-4. Selected mineral analyses from country rocks external to the complex.

GARNET														
TWQ/CMP	Mafic metaplutonic rocks										Semipelite			
	272m3a	m2b	m2f		691m2gg	m2hh	m2l	m2d	m2c	m2dd	704m3b	386m6e	m7d	
Sample no:	272	272	272		691	691	691	691	691	691	704	386	386	
Comment:				R	R inb.	R outb.	C	R inb	R outb.	C	R	R/M	R	
Analysis:	3-6	2-9	2-12	4-3	g2-1b	g2-5c	g2-2b	g2-11b	g2-8b	g2-9b	g1-15	g6-6b	g7-2	
SiO2	38.33	39.05	39.14	37.42	37.74	36.82	37.04	37.60	37.58	37.99	37.82	SiO2	36.52	37.16
Al2O3	22.32	22.67	22.96	21.34	21.02	20.97	21.85	21.25	21.82	21.47	20.71	Al2O3	21.35	21.44
TiO2	.06	.05	.07	.18	.07	.09	.05	.09	.06	.09	.13	TiO2	.01	.03
Cr2O3	.05	.03	.01	.00	.08	.05	.02	.04	.03	.00	.06	Cr2O3	.07	.00
MgO	11.55	11.28	11.12	8.08	3.45	3.66	3.16	2.88	3.68	3.23	3.44	MgO	4.95	5.29
FeO	17.13	17.39	17.57	24.94	27.89	28.40	24.98	30.68	29.88	28.69	27.85	FeO	33.20	34.85
MnO	.49	.44	.55	.74	.99	.94	.65	1.33	.91	1.05	.93	MnO	.33	.47
CaO	8.01	8.70	8.61	6.27	8.77	7.76	12.64	7.64	7.41	9.19	8.82	CaO	1.84	1.81
K2O	.00	.02	.02	.01	.01	.00	.00	.00	.02	.01	.01	K2O	.00	.01
Total	97.94	99.63	100.05	99.00	99.80	98.69	100.17	101.51	101.39	101.72	99.77	Total	98.27	101.05
Fe/(Fe+Mg)	.424	.464	.470	.634	.818	.813	.816	.857	.820	.833	.820	Fe/(Fe+Mg)	.79	
Alm	35.35	35.44	35.85	51.86	60.11	61.99	52.60	65.37	63.76	60.68	60.25	Alm	74.25	74.04
Pyr	42.46	40.95	40.47	29.89	13.34	14.23	11.88	10.92	14.01	12.17	13.27	Pyr	19.72	20.03
Grs	21.18	22.71	22.53	16.70	24.39	21.70	34.13	20.84	20.26	24.90	24.45	Gr	5.27	4.92
Sps	1.02	0.91	1.14	1.55	2.16	2.08	1.39	2.86	1.97	2.25	2.04	Sps	0.75	1.01

PLAGIOCLASE												
TWQ/CMP	Mafic metaplutonic rocks									Semipelite		
	272m3a	m2f	m2b	691m2gg/m2l	m2hh	m2d/m2dd	m2c	704m3b		386m6e	m7d	
Sample no:	272	272	272	691	691	691	691	704		386	386	
Analysis:	2-1	2-18	2-16	2-4	2-9	2-15	2-14	1-15		p6-2	7-4	
Comment:	C-R	S	S	inb.	outb.	inb.	outb.	R		R/M	clear R	
SiO2	54.02	47.87	45.47	62.33	57.57	61.46	59.60	57.39		SiO2	61.45	63.28
Al2O3	28.22	31.04	32.85	23.67	27.04	24.28	24.95	26.89		Al2O3	23.95	23.23
Na2O	4.80	2.43	1.35	8.47	6.77	8.62	7.51	6.18		Na2O	8.73	8.60
CaO	11.08	15.76	17.64	5.15	9.05	5.50	7.21	9.16		CaO	5.23	4.83
K2O	.21	.05	.01	.16	.09	.12	.12	.08		K2O	0.12	0.12
FeO	.20	.36	.36	.51	.21	.75	.31	.15		FeO	0.27	0.05
Total	98.53	97.51	97.68	100.29	100.73	100.73	99.70	99.85		Total	99.8	100.22
Ab	43.43	21.78	12.18	74.18	57.25	73.44	64.87	54.72		Ab	74.57	75.78
An	55.32	77.95	87.75	24.93	42.26	25.86	34.44	44.82		An	24.68	23.53
Or	1.25	.27	.07	.89	.49	.69	.70	.46		Or	0.67	0.70

PYROXENE							
TWQ/CMP	Mafic metaplutonic rocks						
	272m3a	m2b	m2f	691m2gg	m2gg	m2hh	704m3b
Sample no:	272	272	272	691	691	691	704
Comments:	I	S	S	R,P	R inb.	R inb.	R M
Analysis:	c3-1	o2-8	o2-6	c2-6	c2-1b	o2-1	c1-8
SiO2	48.82	51.73	51.89	52.20	52.02	50.28	50.04
Al2O3	7.70	2.28	1.80	1.38	1.76	.65	1.33
MgO	13.00	24.25	24.71	12.68	12.12	15.57	11.87
TiO2	.29	.11	.04	.14	.18	.04	.18
FeO	4.08	20.62	19.81	14.47	11.21	32.54	11.08
CaO	22.91	.44	.37	19.26	21.85	.51	22.30
Na2O	.76	.00	0.03	.53	.52	.04	.56
MnO	.10	na	na	na	.17	.37	.14
Cr2O3	.03	na	na	na	.03	.03	.01
NiO	.01	na	na	na	.04	na	.04
Total	97.70	99.43	98.65	100.64	99.71	100.06	99.17
Fe/(Fe+Mg)	.150	.323	.310	.391	.54	.39	.39
Fe3+	.05			.06	.02	.04	.05
Wo	51.86	.90	.75	40.00	1.08		46.69
En	40.93	67.08	68.46	36.56	46.51		32.50
Fs	7.21	32.02	30.79	23.44	52.41		20.81

AMPHIBOLE					
TWQ/CMP	Mafic metaplutonic rocks				
	272	272	272	691	704
Sample no:	272	272	272	691	704
Comments:	I	S	S	R	R
Analysis:	3-1	2-8	2-6	am2-1	a1-3
SiO2	48.82	51.73	51.89	42.59	43.12
Al2O3	7.70	2.28	1.80	11.67	10.27
MgO	13.00	24.25	24.71	8.96	9.21
TiO2	.29	.11	.04	1.53	1.47
FeO	4.08	20.62	19.81	19.14	14.9
CaO	22.91	.44	.37	11.78	11.45
Na2O	.76	.00	0.03	1.65	1.49
MnO	.10	na	na	.06	.17
Cr2O3	.03	na	na	.07	.05
K2O	.01	na	na	1.32	.74
NiO				.03	na
Total	97.70	99.43	98.65	98.80	97.21
Fe/(Fe+Mg)	.150	.323	.310	.504	.476

BIOTITE		
TWQ/CMP	Semipelite	
	386m6e	m7d
Sample no:	386	386
Comments:	R/M	clear R
Analysis:	b6-1b	b7-3
SiO2	34.91	34.26
Al2O3	18.95	18.94
MgO	10.46	10.15
TiO2	2.10	2.70
FeO	18.51	17.92
CaO	.05	.05
Na2O	.09	.22
MnO	.02	na
Cr2O3	.10	na
K2O	8.37	9.04
NiO	na	na
Total	93.56	93.28
Fe/(Fe+Mg)	.498	.498

Compositions expressed as wt.% oxides.

C-core, P-paleoblast, R-rim, M-matrix, I-inclusion, S-symplectite, sg-subgrain, inb.-inboard, outb.-outboard.

Fe3+ calculated according to Mader et al. (1994)

Table 2-5. Thermobarometric results from the Kramanituar Complex and its country rocks.**Grt-Cpx-Hbl-Pl-Qtz thermobarometry of well-equilibrated mafic metaplutonic rocks, KC**

Sample	Grt-Cpx-Pl-Qtz			Grt-Hbl			Grt-Pl-Rt-Ilm-Qtz	
	P ₍₅₎	T ₍₂₎	P@850°C (avg T ₂ , T ₄)	P ₍₆₎	T ₍₄₎	P@850°C (avg T ₂ , T ₄)	P	T
K-1053								
1053mbb ¹	11.80	830	11.85	13.60	910	12.50		
1053mla ¹	11.90	850	11.90	14.30	950	12.50		
1053m3f ⁵							8.00	730
K-596								
596mlj ¹	14.00	850	14.00					
596mlk3 ^{1,2}	12.00	830	12.10	14.10	970	12.40		
K-104								
104mlc ¹	13.30	890	13.00	13.95	910	13.00		
104m2cc ^{2,3}	12.10	810	12.60	12.60	890	11.90		
104m4b ^{2,3}	11.30	810	11.65	10.40	810	11.15		

Grt-Ky/Sil-Qtz-Pl-Kf thermobarometry of paragneiss, KC

Sample	Grt-Ky/Sil-Qtz-Pl-Kf				Grt-Bt
	P ₍₈₎	P _(F&L)	P@825°C (avg of T ₁ , T _{F&L})	T _(F&L)	T ₍₁₎
K-440					
440m5b ¹	10.8-12.25		10.8-12.25		825
440m5d ³		9.70	10.50		780
440mlc ¹	11.75		11.00		875
440m5c ⁵	5.00	3.85	6.80	584	685
440m2b ⁵	6.00	4.25	7.25	583	740
K-276					
276m3bb ¹		13.00	12.50	856	
276m3c ²		11.70	11.15	856	
276m3cc ²		10.90	10.50	856	
K-290					
290mla ²		12.25	11.85	836	
290m2a ²		11.40	12.10	788	

Grt-Cpx-Opx-Hbl-Pl-Qtz thermobarometry of coronitic and symplectitic mafic rocks, KC

Sample	Grt-Cpx-Pl-Qtz			Grt-Hbl			Grt-Opx-Pl-Qtz	
	P ₍₅₎	T ₍₂₎	P@970°C (avg T ₂ , T ₄)	P ₍₆₎	T ₍₄₎	P@970°C (avg T ₂ , T ₄)	P ₍₇₎	T ₍₃₎
K-1005								
1005m1d ¹	12.00	950	12.1	14.00	980	13.9	13.60	1080
1005m4b ¹	13.15	930	13.1	16.75	1020	15.0	14.00	1060
1005mla ¹							10.00	850
1005m2f ⁵							8.05	825
1005m3f ⁵							8.60	870
1005m7d ⁵				7.90	720		6.80	775
K-619								
619m9a ^{4, initial}	12.65	820					11.60	850
619m9b ^{4, final}		750					9.00	730
619m9c ^{5, initial}	7.80	725					9.40	725
619m9d ^{5, final}							8.20	750

Grt-Cpx-Opx-Hbl-Pl-Qtz thermobarometry of country rocks, external to the KC

Sample	Grt-Cpx-Pl-Qtz			Grt-Hbl			Grt-Opx-Pl-Qtz	
	P ₍₅₎	T ₍₂₎	P@740°C (avg T ₂ , T ₄)	P ₍₆₎	T ₍₄₎	P@740°C (avg T ₂ , T ₄)	P ₍₇₎	T ₍₃₎
K-272								
272m3a ¹	11.80	700						
272mla ³	11.80	600	12.85	13.8	820	12.35		
272m7a ³	13.50	725	13.65	14.4	820	13.00		
272m2e ⁵							8.70	680
272m2d ⁵							8.80	750
K-691								
691m2dd ^{4, initial for An₂₁₋₂}				9.40-11.20	750			
691m2c ^{4, final, outboard}				8.60	760			
691m2d ^{4, final, inboard}				9.60	700			
691m2i ^{4, initial for An₂₁₋₂}	10.6-12.5	750						
691m2hh ^{4, final, outboard}	7.00	675						
691m2gg ^{4, final, inboard}	10.75	725					9.50	735
K-704			P@770°C (avg T ₂ , T ₄ , excl. 704m3f)			P@770°C (avg T ₂ , T ₄ , excl. 704m3f)		
704mla ¹				8.25	770	8.30		
704mlb ¹				9.30	780	9.25		
704m3b ²	8.50	720	9.00	8.80	780	9.00		
704m3c ²	11.50	770	11.50	12.50	810	12.00		
704m3f ²	12.80	900	11.50	12.30	980	9.40		

Notes: ¹core, ²rim, ³matrix, ⁴garnet formation, ⁵garnet breakdown; P₍₅₎ P₍₆₎ P₍₇₎ P₍₈₎ are pressure values computed at intersection with thermometers corresponding to equilibria 5, 6, 7, 8 (p.66), respectively; T₍₂₎ T₍₄₎ T₍₃₎ T₍₁₎ are temperature values computed at intersection with barometers corresponding to equilibria 2, 4, 3, 1 (p.66), respectively; T_(F&L) are temperatures calculated according to Fuhrman and Lindsley (1988); P_(F&L) are pressures calculated at T_(F&L); P@x°C are calculated pressure values at a reference T (i.e., average of thermometers, as indicated) for a specific part of the P-T path.

Figure 2-1. Geology of the western Canadian Shield after Hoffman (1988). *a)* segments of the proposed Snowbird tectonic zone (STZ) transect the western Churchill Province (unornamented) and include: the Virgin River shear zone (VRsz) and subsurface east Alberta orogen (eAo), the Striding-Athabasca mylonite zone (SAmz), the Tulemalu fault zone, and the Chesterfield segment which includes the Kramanituvar (KC), Uvauk (Uc) and Daly Bay (DBc) complexes. Other abbreviations are sz-shear zone, mz-magmatic zone. *b)* distribution of high-grade metamorphic rocks of the western Churchill Province.

Figure 2-2. Geology of the Kramanituvar Complex and its wall rocks, modified from Schau and Ashton (1980), Schau (1983) and Sanborn-Barrie (1993, 1994). Sample sites for this study with geothermobarometric results (maximum/minimum) are shown, as is the location of dated gabbroic anorthosite discussed in text. Inset shows regional metamorphic isograds and major ductile shear zones, the Northern shear zone (NSZ) and Southern shear zone (SSZ).

Figure 2-3. Element distribution maps for well-equilibrated mafic (*a,b*) and paragneiss (*c,d*) from the Kramanituur Complex with variation in elemental abundance indicated by colour scale increasing from cool blue to hot red tones. *a*) for mafic sample *K-104*, Ca distribution in 3 mm garnet (flat blue-green), matrix Cpx (orange-yellow tones) and Pl (green-blue tones). Qtz is black.; *b*) Mg distribution for area shown in *a*). Note lower Mg (corresponding to higher Fe) is restricted to garnet in direct contact with Cpx; *c*) for paragneiss *K-290*, Ca distribution in 4 mm garnet (blue ~Grs_{6.7}; orange ~Grs_{9.3}) with surrounding matrix of Qtz (black) and perthite (purple-red); *d*) Mg distribution for area shown in *c*). Note Mg:Fe gradient perpendicular to long axis of garnet.

Figure 2-4. Equilibria used to determine estimated peak conditions of 12.5 ± 1 kbar and $875 \pm 35^\circ\text{C}$ for mafic granulite K-1053 as plotted by TWEEQU software of Berman (1991). Abbreviations after Kretz (1983).

Figure 2-5. Element distribution maps for coronite K-619. *a*) Ca distribution across layered coronitic texture. From right to left is matrix plagioclase from An₅₁ (yellow) to An₄₄ (pale blue) immediately *outboard* of garnet; garnet with concentric zonation from X_{Grs}= (pale yellow-blue) to X_{Grs}= (dark blue); *inboard* plagioclase (last-formed is An (yellow-orange), first-formed is An (uniform green); second generation Opx (black); Cpx (uniform red); rim of magmatic Opx phenocryst (black with Cpx lamellae). *b*) Mg distribution for area shown in *a*). Note concentric Mg zonation in garnet (pale blue to dark blue).

Figure 2-6. Element distribution maps for symplectite in 2-pyroxene gabbro *K*-1005. *a*) Ca distribution across symplectite, from garnet (purple-black); last-formed Pl closest to Grt (orange); first-formed Pl (yellow to green); matrix Amp (dark blue); Opx (black). *b*) Fe distribution for area shown in *a*).

Figure 2-7. Sketch of coronitic amphibole (Amp) and garnet (Grt) in plagioclase porphyritic gabbro *K-691*, located 7 km north of the complex.

Figure 2-8. Profile through garnet porphyroblast from metagabbro *K*-704.

Figure 2-9. Summary of P - T conditions and possible P - T paths for samples from the study area; *a*) well-equilibrated mafic granulites from the Kramanituvar Complex; *b*) paragneiss from the Kramanituvar Complex. *c*) samples with symplectitic and coronitic reaction textures from the complex; *d*) country rocks located south of the Kramanituvar Complex.

Plate 2-1. Mafic metaplutonic rocks of the Kramanituur Complex. *a*) relatively unstrained gabbroic anorthosite. *b*) Cpx-Pl-Grt mylonite, located 3 km east of *K*-1053 (Fig. 2.2); *c*) photomicrograph of well-equilibrated mafic granulite *K*-104. Pbl-paleoblast, nbl-neoblast, discussed in text.

Plate 2-2. Mafic gneiss (*K-596*) with compositional layering. *a*) polished slab showing mesocratic Cpx-Grt layers that alternate with leucocratic garnetiferous layers; *b*) representative compositions of Grt (Grs mol%) and Pl (An mol%) from mesocratic layers (ci=80); *c*) representative compositions of Grt (Grs mol%) and Pl (An mol%) from leucocratic layers.

Plate 2-3. Backscatter electron image of 3 mm garnet (light grey) from kyanite-bearing paragneiss, sample *K-276*. Also shown is K-feldspar (medium grey), quartz (darkest contrast). Ca (mol% Grs) contoured. Arrows indicate low Ca garnet adjacent to matrix plagioclase (discussed in text). Profiles of Sps and Fe/(Fe+Mg) illustrate asymmetric Fe:Mg zonation perpendicular to garnet length, a pattern also present in garnet of Fig. 2-4*d*.

Plate 2-4. Coronitic texture in metanorite *K-619*. *a*) photomicrograph (crossed nicols) of Opx phenocryst (left) and matrix Pl (right) with intervening two-stage corona described in text; *b*) schematic illustration of *a*) showing initial stage of Grt-Cpx-Qtz corona formation; *c*) schematic detail showing second stage of corona formation due to garnet breakdown initially involving Cpx to produce a continuous moat of second generation Opx and Pl (An₄₀₋₅₁), and then late-stage breakdown with Qtz to produce more calcic Pl (An₅₀₋₇₆) and isolated blebs of Opx.

a)

b) GARNET CORONA FORMATION

c) GARNET BREAKDOWN

Plate 2-5. Photomicrograph of mylonitized garnet-bearing gabbroic anorthosite. Garnet is mantled by Opx-Pl symplectite interpreted to reflect synkinematic garnet breakdown during decompression.

Plate 2-6. Polished hand sample of garnet porphyroblastic gabbro *K-272*, located 1 km east of the complex. Low strain mineral aggregates within strain shadows of garnets are indicated by white ellipses.

Plate 2-7. Element zonation profiles for garnets from metasedimentary country rocks located south of the complex. Note garnet porphyroblasts are characterized by poikiloblastic cores and inclusion-free rims. *a*) pelite (*K*-556) with regional northeast-striking foliation ($S_{1(\text{External})}$) wrapping garnet porphyroblasts. *b*) semi-pelite (*K*-386) with ($S_{2(\text{External})}$) wrapping garnet, from the ca. 2550 Ma Southern shear zone.

CHAPTER 3

Geochronological constraints on Paleoproterozoic high-*P* metamorphism, magmatism and rapid exhumation of deep-crustal rocks, Kramanituvar Complex, western Churchill Province, Canada.

ABSTRACT

U-Pb and Sm-Nd geochronology establishes an important Paleoproterozoic (ca. 1.9 Ga) history for the Kramanituvar Complex, located in the interior of the Archean western Churchill Province, and provides further insight into the Snowbird tectonic zone, a regional-scale (>2000 km) feature whose role in the Precambrian remains controversial. The Kramanituvar Complex represents a window of deep-crustal rocks, dominated by a granulite-facies metagabbroic suite of ca. 1902 Ma age, with minor supracrustal rocks and charnockite of probable primary Archean age. These yield peak equilibrium conditions of 12-15 kbar and 850-900°C, which are bracketed between ca. 1910 Ma, the age of prograde metamorphic monazite in sillimanite- and kyanite-bearing paragneiss, and 1901 Ma, the cooling ages of titanite and rutile in gabbroic and paragneissic rocks. Although mineral assemblages reflecting peak metamorphic conditions are widespread, some rocks record near-isothermal decompression to ca. 8 kbar and 800°C. The timing of uplift and exhumation is tightly bracketed between magmatic crystallization of gabbroic anorthosite under granulite-facies conditions at 1902 Ma and widespread rutile cooling ages of 1901 Ma. Very rapid time-averaged cooling rates of $>450^{\circ}\text{C Ma}^{-1}$ are estimated for gabbroic anorthosite and leucogranite.

Rocks surrounding the Kramanituvar Complex are mainly amphibolite-facies Archean plutonic and supracrustal rocks, however, those to the north were at deep-crustal conditions coeval with the complex. Accordingly, an intervening crustal-scale belt of granulite-facies mylonites (Northern shear zone), which was active during ca. 1901 Ma exhumation of the complex, is not a master fault that accommodated major displacements. The southern boundary of the complex may coincide with a cryptic tectonic break that separates the Paleoproterozoic Kramanituvar Complex from rocks that preserve older, mainly Neoarchean tectonometamorphic assemblages and fabrics. This proposed break is inferred to be a normal fault which may, in part, coincide with a ca. 2550 Ma amphibolite-facies shear zone which is unrelated to ca. 1900 Ma exhumation.

The geochronological data highlight important differences in the timing of events along the geophysically-defined Snowbird tectonic zone. Latest significant magmatism and tectonometamorphism along the southwest-trending Striding-Athabasca segment at

ca. 2600 Ma, and interpreted ca. 1800 Ma activity in the subsurface Alberta segment, contrast with penetrative ca. 1900 Ma activity in the east-trending Chesterfield segment, as documented at the Kramanituar Complex.

INTRODUCTION

Archean provinces are generally regarded as having been stable, with respect to tectonism and metamorphism since the Neoproterozoic, and only peripherally affected by Proterozoic igneous, metamorphic and deformation events. Examples include the Slave, Superior, Nain and Wyoming provinces of North America (Hoffman, 1989), the Zimbabwe and Kaapvaal cratons of South Africa (Boyd et al., 1985), and the Yilgarn (Gee et al., 1981; Nemchin et al., 1994) and Pilbara (Hickman, 1981) of western Australia. This stability is attributed, in part, to the presence of an ancient, cold, chemically-distinct (FeO-poor), high velocity lithospheric keel, or tectosphere, that stabilized to depths of ~200 km in the first two billion years of earth's history (Jordan, 1975; Anderson and Bass, 1984; Polet and Anderson, 1995). However, the western Churchill Province of North America (Fig. 3-1) has not retained such structural integrity. In addition to an extensive Archean history, considerable Proterozoic magmatism (Peterson, 1994) and sedimentation (Bell, 1970; Eade and Chandler, 1975; Aspler et al., 1994), as well as variable degrees of metamorphism and deformation are recognized, both marginal and internal to the province (see Fig. 3-1, Appendix III). Although Proterozoic tectonometamorphic reworking of Archean western Churchill crust has been recognized for almost three decades (Davidson, 1970; Lewry et al., 1985), its nature and extent are not well understood. This is particularly true in the interior of the province, far removed from peripheral Paleoproterozoic orogenic belts. Mechanisms to account for tectonometamorphic reworking of the province are not well established, nor are boundary conditions which influenced Proterozoic deformation in its interior.

This contribution documents Paleoproterozoic metamorphism, magmatism and deformation in the interior of the Archean western Churchill Province. It focusses on the

Kramanituar Complex (KC in Figs. 3-1, 3-2; Fig. 3-3), one of several high-grade complexes exposed within the interior of the province that coincide with the surface expression of a regional-scale geophysical anomaly, the Snowbird tectonic zone (Fig. 3-1). This geophysical anomaly extends more than 2000 km, from the foothills of the Rocky Mountains to Hudson Bay, and is well-expressed as a series of high amplitude linear anomalies in the horizontal gradient of the gravity field (Goodacre et al., 1987a; *see* Fig. 1-1a). Its northeastern Chesterfield segment which includes the Kramanituar Complex (Fig. 3-1), is underlain by an extensive positive (+60 mGal) Bouguer gravity anomaly (Goodacre et al., 1987b; *see* Fig. 1-1b). Interpretations of the tectonic significance of the Snowbird tectonic zone that have sought to explain Paleoproterozoic intra-Churchill tectonometamorphism have included a Proterozoic intracontinental rift system (Wallis, 1970), a Paleoproterozoic collisional suture (Gibb and Walcott, 1971; Thomas and Gibb, 1985; Hoffman, 1988), and an indentation-related fault network generated during continental collisional activity from the northwest and/or from the southeast (Lewry and Sibbald, 1977; Lewry et al., 1985; Ross et al., 1995). However, detailed geological and geochronological studies along a 400 km length of this feature (SAMZ, Fig. 3-1) have shown that latest, significant tectonometamorphic activity occurred at ca. 2600 Ma (Hanmer et al., 1994), a time when significant portions of this zone are interpreted as intracontinental (Carolan and Collerson, 1989; Hanmer et al., 1995a). If significant parts of the western Churchill Province were largely intracontinental by ca. 2600 Ma, the nature and significance of Paleoproterozoic magmatic and tectonometamorphic events in its interior remain an outstanding geological problem.

This paper presents U-Pb and Sm-Nd geochronological data that establish timing constraints on: 1) voluminous mafic magmatism, which accounts for 75% of the Kramanituar Complex and which is responsible for its magnetic and gravity expression; 2) high-*P* granulite-facies metamorphism which is preserved in at least 60% of the complex; and, 3) ductile deformation, localized within mylonite zones that are marginal to the complex,

and south of it. From these data, a chronology of events for the study area is reconstructed and considered in the context of the western Churchill Province and the Snowbird tectonic zone.

GEOLOGICAL SETTING

The Kramanituar Complex is situated in the interior of the western Churchill Province, nearly equidistant from two Paleoproterozoic orogenic belts (Fig. 3-1). The Thelon Orogen marks the northwest margin of the province and is the product of east-dipping subduction-related magmatism between ca. 2000-1960 Ma (Frith and van Breemen, 1990; van Breemen et al., 1992) which led to continental collision with the Archean Slave craton at ca. 1970-1960 Ma (Tirru and Grotzinger, 1990; Hanmer et al., 1992) and continued post-collisional shortening until ca. 1900 Ma. The Trans-Hudson Orogen at its southern margin is the product of subduction beneath the western Churchill Province at 1885 Ma, which led to terminal collision with the Archean Superior craton at ca. 1830 Ma and post-collisional shortening until ca. 1700 Ma (Stern et al., 1995; Lucas et al., 1996).

The Kramanituar Complex

The Kramanituar Complex is dominated by high-*P* granulite-facies plutonic rocks, including a metagabbro-norite-gabbroic anorthosite suite with minor charnockite (Fig. 3-3). Rocks of supracrustal origin are mainly sillimanite-bearing paragneiss, with kyanite-bearing paragneiss occurring as narrow units in the extreme northeast part of the complex (locality A on Fig. 3-3). All rock types generally occur as 1 to 5 km wide, west- to north-west-trending, steeply dipping panels which together comprise a coherent, elliptical 850 km² exposure of high-grade rocks (Fig. 3-3). The complex is mainly distinguished from surrounding rocks on the basis of lithology and, as will be shown here, by age. North and east of the complex, foliated to gneissic quartzo-feldspathic plutonic rocks with minor mafic metaplutonic rocks are exposed. To the south, Archean-age (this study) supracrustal

rocks are cut by granitoid rocks and gabbro (Fig. 3-3). An intrusive relationship between the Kramanituar Complex and metaplutonic country rocks to the east is indicated by partial melt textures in garnet-bearing diorite within a 1 km-wide aureole adjacent to the eastern margin of the complex, but is obscured along the north margin due to shear deformation (Chapter 1). To the south, an inferred tectonic break separates rocks of the complex that were penetratively affected by Paleoproterozoic tectonometamorphism from those recording older fabrics and metamorphic assemblages to the south (Chapter 2). Sub-greenschist facies conglomerates and sandstones of the Baker Lake Group (Gall et al., 1992) unconformably overlie the western part of the complex (Fig. 3-3), and contain clasts of variably strained metaplutonic rocks, including gabbroic anorthosite, that appear to be derived from the complex.

Several generations of fabric elements identified within the Kramanituar Complex indicate a bulk deformation path of progressive, general, noncoaxial deformation. Low strain domains in the center of the complex (Plate 3-1a) preserve a weak to moderately developed, northwest-striking shape fabric ($S_{1(KC)}$). These low strain rocks are transitional to rocks characterized by moderate to high strain and thorough dynamic recrystallization (Plate 3-1b). Strains of highest magnitude are localized within the Northern shear zone (Fig. 3-4), an anastomosing, granulite-facies ductile shear zone, 60 km in length and 3 to 6 km wide, that forms two segments along the north and east margins of the complex. East-striking, near-vertically dipping ($S_{2(KC)}$) granulite facies ribbon mylonites (Plate 3-1c) of gabbroic, anorthositic and paragneissic origin within the shear zone record evidence of general, noncoaxial flow with a sinistral sense of vorticity (Chapter 1).

Rocks south of the complex preserve tectonic fabric elements that predate those within the complex. Semi-pelitic and metaplutonic rocks contain a moderate to strong foliation which strikes northeast and dips moderately to steeply ($55-65^\circ$) to the northwest. This fabric, designated $S_{1(External)}$ is the earliest fabric recognized in the study area (Chapter 1). However, in the MacQuoid-Gibson supracrustal belt to the south (MGB in

Fig. 3-2), this regional northeast-striking, dominantly northwest-dipping foliation overprints an earlier foliation (S_1), is axial planar to F_2 , and is cut by 2.19 Ga mafic dykes (Tella et al., 1997a). $S_{1(\text{External})}$ south of the Kramanituar Complex is deflected into the east-striking, moderately north-dipping amphibolite-facies Southern shear zone ($S_{2(\text{External})}$; Fig. 3-4) that records oblique, north-side-up and dextral movement (Chapter 1). North of the complex, northeast-striking fabrics in granitoid rocks are coplanar to $S_{1(\text{External})}$ and may be contemporaneous.

Thermobarometric determinations of metamorphic conditions for the Kramanituar Complex indicate peak equilibrium conditions of 12 ± 1.5 kbar and $810 \pm 70^\circ\text{C}$ for mafic metaplutonic rocks and paragneiss, with near-isothermal decompression to ca. 8 kbar and 800°C (Chapter 2). North and east of the complex, isolated gabbroic rocks (e.g. K-691, this study) within an extensive granitoid terrain yield P - T conditions of 8-13 kbar and $760 \pm 60^\circ\text{C}$ (Grt-Cpx-Pl-Amp), indicating the presence of high- P rocks beyond the north and east boundaries of the Kramanituar Complex. South of the complex, a prograde metamorphic history is recorded by semi-pelite (K-386, this study) which attained conditions of 8.5-9 kbar and 750°C , and gabbro (K-704, this study) which attained conditions up to 12 kbar and $870 \pm 100^\circ\text{C}$.

PREVIOUS GEOCHRONOLOGY

Two U-Pb age determinations for rocks within the study area (Fig. 3-3; Appendix III) were available at the start of this investigation. Within the complex, charnockitic gneiss previously considered to be comagmatic with gabbroic anorthosite and anorthositic rocks, yielded an inferred metamorphic age of $2573 \pm 28/-5$ Ma (Schau, 1980). North of the complex, foliated to gneissic granodiorite is dated at $2675 \pm 33/-11$ Ma (Schau, 1980). These two ages had been interpreted to indicate a minimum Neoproterozoic age for the Kramanituar Complex and its wallrocks (Schau, 1980; Schau et al., 1982). A K-Ar age (biotite) of 1895 ± 44 Ma was reported for the granodiorite north of the complex (Schau,

1980). A minimum age of deposition of the coarse clastic sequence that unconformably overlies the complex is 1832 ± 28 Ma, based on a Pb-Pb (monazite-bearing apatite) isochron age of associated overlying ultrapotassic plutonic rocks (MacRae et al., 1996).

ANALYTICAL PROCEDURES

U-Pb geochronology studies using zircon and monazite, and involving sample preparation, chemical procedures and mass spectrometry, were performed in the Geochronology Laboratory of the Earth Science Department of Carleton University. Those using titanite and rutile were carried out at Carleton University and the Geochronology Laboratory of the Geological Survey of Canada. Samples were crushed in a steel jaw crusher and pulverized in a grinder. A heavy mineral concentrate, obtained using a Roger'sTM table, was further processed by conventional mineral separation techniques including sieving, and use of heavy liquids (methylene iodide) and a FrantzTM isodynamic separator. Mineral grains chosen for analyses were selected using a binocular microscope. All zircon fractions analyzed were abraded (3-16 hours) following the technique described by Krogh (1982). Abraded fractions were washed in warm 3 N HNO₃. Prior to dissolution, all fractions were cleaned with a hot 3 N HNO₃ bath and rinsed with ultrapure acetone. The fractions were weighed on disposable aluminum boats using a microbalance.

Zircon and monazite fractions were spiked with a mixed ²⁰⁵Pb/²³⁵U tracer (Parrish and Krogh, 1987) and dissolved in acid in TeflonTM micro-capsules (Parrish, 1987): zircon for 30 hours with 16 N HF at 245°C, and monazite for 60 hours with 12N HCl at 210°C. Chemical extraction of U and Pb using anion exchange resin followed procedures modified from those described by Krogh (1973). Additional purification was achieved for rutile and titanite fractions by using 1N HBr for Pb and 8 N HNO₃ and 6.2N HCL for U (Parrish et al., 1991).

Pb and U were loaded separately onto outgassed Re filaments using Si-gel. Analyses at Carleton University utilized a Finnigan-Mat 261 thermal ionization mass

spectrometer outfitted with multicollector Faraday cups and a secondary electron multiplier detector. A mass fractionation correction for Pb of 0.09% per atomic mass unit was used. U fractionation was calculated for each sample by comparison of measured ^{233}U with known concentrations of ^{233}U in the mixed spike. Analytical blanks were generally 18 pg or less for ^{206}Pb and 5 pg or less for ^{238}U . Isotopic data for titanite and rutile collected at the Geological Survey of Canada followed the procedures of Parrish et al. (1987) and Davis et al. (1997).

For Sm-Nd analyses, mineral separates were prepared at Carleton University and chemistry and mass spectrometry were done at the Geochronology Laboratory of the Geological Survey of Canada following a chemistry procedure modified from Richard et al. (1976). Whole rock powder and mineral separates, spiked with a mixed ^{148}Sm - ^{149}Nd solution, were dissolved in an HF-HNO₃ mixture. Separation and purification of Sm and Nd was carried out by cation exchange and HDEHP (Di[2-ethylexyl] orthophosphoric acid)-teflon powder chromatography methods. Total procedural blanks were approximately 200 pg for Nd and 100 pg for Sm. Mass spectrometry was carried out on a Finnegan-MAT 261 thermal ionization mass spectrometer in static multicollection mode. Nd isotopic compositions were normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$, and corrected to LaJolla $^{143}\text{Nd}/^{144}\text{Nd} = 0.511860$ for fractionation. Measurements of LaJolla standard yielded $^{143}\text{Nd}/^{144}\text{Nd} = 0.511842 \pm 10$ (2-sigma error).

Results of U-Pb isotopic analyses for the seven samples dated are given in Table 3-1. All error ellipses are shown at the two sigma confidence level and linear regressions were calculated using the method of Davis (1982). Age errors are quoted at the 95% confidence level. Sm-Nd isotopic data for two garnet-bearing samples are given in Table 3-2.

SAMPLE DESCRIPTIONS AND RESULTS

Samples from within the Kramanituur Complex

1. *Kramanituur Complex gabbroic anorthosite (K-676)*

A 100 kg sample of least strained and minimally recrystallized gabbroic anorthosite (K-676) from the central part of the Kramanituur Complex (Figs. 3-3, 3-4a), was collected to date the age of voluminous gabbroic-anorthositic magmatism. The sample is coarse-grained with an assemblage of plagioclase-orthopyroxene-clinopyroxene-pargasitic amphibole-magnetite-ilmenite and trace amounts of zircon, titanite, rutile $\pm$ apatite. Garnet occurs rarely as fine (150 μm) grains within plagioclase, or as partial coronae around magnetite and ilmenite. The unaltered nature of this rock is reflected by the brilliant blue-green schiller of labradorite phenocrysts, and fresh clinopyroxene, pargasite and orthopyroxene, the latter preserving platelets and lamellae of exsolved ilmenite. Recrystallization is limited to minor development of subgrains along grain boundaries of some plagioclase. This rock contains a moderate L=S foliation (Plate 3-1a) which corresponds to $S_{1(KC)}$. Elsewhere in the complex, however, rocks of the gabbroic suite are intensely deformed, and a strain gradient between sample K-676 and mylonitic gabbroic anorthosite within the Northern shear zone is observed (Plate 3-1; Chapter 1). An age of magmatic crystallization for this suite therefore places constraints on mafic magmatism and on movement along the Northern shear zone, the major zone of localized strain in the study area.

The gabbroic anorthosite yielded a moderate amount of inclusion-free, colourless to very pale yellow-pink zircon of variable morphology. Most abundant are euhedral and subhedral prismatic crystals (Plate 3-2a), some with dipyrarnidal symmetry, that are 100-200 μm wide, with length to width ratios of 1:2 and euhedral internal zonation (Plate 3-2b). These crystals possess attributes that are consistent with zircon that have crystallized from a magmatic melt (Speer, 1980). Slender prismatic crystals with aspect ratios up to 6 are also interpreted as magmatic. A small population of rounded, equant, multi-faceted grains (Plate 3-2c), 50-150 μm in diameter, are morphologically similar to zircons com-

monly described as synmetamorphic. Two single-grain fractions of euhedral prismatic zircon (Z1 and Z3 in Plate 3-2a) and one multigrain fraction of subrounded equant grains (indicated by arrows in Plate 3-2c) were analysed. Analyses of two fractions are concordant (Fig. 3-5a, Table 3-1), the larger single prismatic crystal (Z1) at 1899 ± 5 Ma and the multigrain fraction (Z2) at 1901 ± 3 Ma. The third single-grain fraction (Z3) is 17.3% discordant and when regressed (Davis, 1982) with the other two zircon fractions, defines a chord from the origin of the concordia diagram to an upper intercept age of 1901 ± 2 Ma (Fig. 3-5a).

Two fractions of titanite from this sample are concordant with ages of 1901 ± 4 Ma and 1905 ± 3 Ma (Fig. 3-5a). These ages are indistinguishable from the zircon age, such that the weighted average of the titanite fractions and the two concordant zircon fractions is 1902 ± 1.6 (MSWD=1.43). Regardless of whether the closure temperature of titanite is ca. 700°C (Cherniak, 1993; Scott and St-Onge, 1995; Zhang and Scharer, 1996) or ca. 600°C (Heaman and Parrish, 1991; Spear and Parrish, 1996), these data imply rapid cooling of gabbroic anorthosite from at least ca. 800°C, the estimated minimum closure temperature of zircon (Black et al., 1986; Maboko et al., 1991).

Two fractions of rutile were analysed. One fraction (R2) is 0.7% discordant with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1899 ± 3 Ma, and the other (R1) is 1.5% discordant with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1890 ± 5 Ma (Fig. 3-5a). A regression of these two fractions (York, 1969) yields an upper intercept of $1906 +90/-9$ Ma with a ca. 993 Ma lower intercept. Rutile is interpreted to have grown above the closure temperature of ca. 400°C (Mezger et al., 1991) since minimum T conditions for all rocks within the complex exceeds 600°C (Chapter 2). Accordingly, the rutile data are interpreted to reflect a cooling age which overlaps that of titanite and zircon from this sample. In summary, U-Pb isotopic results from zircon, titanite, and rutile from gabbroic anorthosite are interpreted to record magmatic crystallization of gabbroic anorthosite at 1902 ± 1.6 Ma, followed by extremely rapid cooling below ca. 400°C.

2. *Syntectonic leucogranite intrusion (K-198)*

Fine-grained, biotite-bearing leucogranite (K-198; Fig. 3-3) occurs within a 20 cm wide brittle-ductile shear zone oriented $194/64^{\circ}\text{W}$, that cuts the penetrative, east-striking, near-vertically dipping mylonitic foliation ($S_{2(KC)}$) of the Northern shear zone (Plate 3-3). Crosscutting relations between the leucogranite and the mylonitic foliation of the host diatexite clearly establish that the leucogranite postdated mylonitization. The leucogranite is interpreted as late-tectonic, since it carries a strongly to intensely developed planar shape fabric ($120/67^{\circ}\text{SW}$). It has not imposed a retrograde mineral assemblage on the mylonitic rocks it cuts, suggesting intrusion of water-undersaturated leucogranite melt. Zircon and monazite from this sample were analysed using the U-Pb system to obtain a minimum age for ductile mylonitization along the Northern shear zone and date localized late-stage reworking within it.

The leucogranite yielded abundant slender-prismatic, dipyrarnidal zircon, 70-100 μm long, with aspect ratios ≥ 4 and subrounded tips. Smaller populations of zircon with flat, subhedral morphology ($< 62\text{-}74\ \mu\text{m}$) and with anhedral, equant morphology (average 80 μm diameter) also occur. All morphological types are clear, colourless to very pale yellow, with no discernable fractures, cores or inclusions. These qualities, together with their variable U contents (Table 3-1) suggest these zircon represent a single igneous population (i.e., not homogenized during subsequent high-grade metamorphism). Analyses from the fraction comprising the largest (equant) grains (Z1) is concordant at 1902 ± 2 Ma (Fig. 3-5b). Four other analyses, including the only single-grain fraction (Z5) and two fractions of slender-prismatic crystals (Z3, Z4), are 1.4-3.5% discordant, and if regressed with the concordant fraction define an upper intercept age of $1903.6 \pm 5/-4$ Ma (MSWD=0.73). The lower intercept age defined by the discordia line is $1019 \pm 69/-65$ Ma, indicating partial disturbance of the four discordant fractions during Mesoproterozoic time. Given that all zircon fractions except one have experienced partial Pb loss, the U-Pb

zircon age of this leucogranite dyke is taken to be 1902 ± 2 Ma, the age of the concordant fraction.

Two single grains of clear, medium to pale yellow monazite, with a subrounded to disc-shaped morphology of 100 μm diameter, were analysed from the leucogranite. Data from both fractions are slightly discordant (Fig. 3-5b): M1 is 0.7% discordant with $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1902.6 ± 1.2 Ma and M2 is 1.6% with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1903.3 ± 2.3 Ma. The chord defined by these two monazite data ellipses goes through the origin, consistent with a small degree of recent Pb loss, and intersects concordia at $1902.3 +2/-1.3$ Ma. The weighted average of the concordant zircon fraction and the $^{207}\text{Pb}/^{206}\text{Pb}$ ages of the two monazite fractions is 1902.6 ± 1 Ma (MSWD=0.36, POF=69%), which is taken as the age of the leucogranite.

3. & 4. *Diatexite of metasedimentary origin*

Two samples of granulite-facies paragneiss were dated: kyanite-bearing paragneiss mylonite (K-290) from the northeast part of the complex, and sillimanite-bearing paragneiss mylonite (K-440) from the northwest part (Fig. 3-3). In addition to their major constituents: plagioclase, K-feldspar, quartz, garnet $\pm$ sillimanite $\pm$ kyanite (Plate 3-4a), these samples contain zircon, monazite, and rutile from which independent constraints on the timing of metamorphism (monazite) and cooling (rutile) of the complex are obtained. Both samples contain only trace amounts of hydrous phases, mainly represented by high-Ti (3.5-6 wt%) biotite. Geothermobarometry from these rocks yields near-peak conditions of 11 ± 1.5 kbar and $810 \pm 30^\circ\text{C}$, with high- T decompression to 5-6 kbar and $720 \pm 20^\circ\text{C}$ recorded by the sillimanite-bearing sample (Chapter 2). Since high-grade paragneisses from the complex generally contain *either* kyanite or sillimanite, samples dominated by each were selected for U-Pb analysis in order to compare their thermal histories.

Kyanite-bearing mylonitic paragneiss (K-290) yielded clear, pale yellow monazite crystals, 50-100 μm diameter, with a subhedral to anhedral, equant morphology. Only

subhedral monazite crystals smaller than 25 μm , were identified in the matrix using backscatter SEM imaging. Single-grain fraction M1 is concordant at 1910.5 ± 2 Ma and another single-grain fraction M3 is 0.4% discordant with an identical $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1910 ± 2 Ma (Fig. 3-5c). A third multi-grain fraction (M2) is 0.8% discordant with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1907 ± 11 Ma. The concordant age of 1910 ± 2 Ma is taken to represent the age of monazite closure for the rock, with a minimal amount of subsequent Pb loss from the other fractions. Two rutile fractions analysed from this sample were concordant with ages of 1896 ± 1.8 Ma (R1) and 1895 ± 6 Ma (R2).

Sillimanite-bearing mylonitic paragneiss (K-440) yielded a moderate amount of clear, pale yellow monazite grains, 50-170 μm long, with subrounded, irregular, equant shapes. Several monazite grains with a subhedral morphology have visible cores. Monazite observed using backscatter SEM images occurs as a widely dispersed matrix phase, generally <50 μm and rarely up to 90 μm . Rare inclusions of monazite <10 μm long occur within garnet. A single-grain concordant fraction (M1) has an age of 1908 ± 2 Ma, and a 2-grain fraction (M2) is 1.3% discordant with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1910 ± 3 Ma (Table 3-1; Fig. 3-5d). These data are the same, within error, as monazite ages from the kyanite-bearing paragneiss described above, and indicate monazite closure at ca. 1910-1908 Ma. A third single-grain fraction (M3) is 1.1% discordant with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1917 ± 6 Ma, within error of M2 and slightly older than the concordant M1 fraction (Fig. 3-5d). The slightly older age for M3 may reflect a component of inherited Pb, possibly reflected by core material in other grains, or may indicate closure of some monazite grains as early as ca. 1917 Ma.

Four multigrain fractions of rutile were analysed from this sample. Fraction R3 is concordant with an age of 1901 ± 6 Ma. Other fractions are 0.5% to 1.7% discordant and lie along a chord (MSWD=2.58) with a lower intercept age of $1254 +143/-130$ Ma (Fig. 3-5d), interpreted as a time of Pb loss.

5. *Granulite-facies gabbro (K-1053)*

Granulite-facies metagabbro (K-1053) was collected from the eastern part of the Kramanitar Complex (Fig. 3-3) to date high-grade metamorphism of the mafic plutonic suite. This sample is ideally suited for Sm-Nd geochronology. Firstly, analyses of garnet, clinopyroxene and plagioclase separates and whole rock powder ensure a reasonable spread in Sm/Nd ratios, required for a precise isochron age (Faure, 1986). Secondly, these minerals are texturally consistent (Plate 3-4b) with an unannealed equilibrium assemblage produced during a single granulite-facies event. High grain boundary energy is reflected by lobate subgrain boundaries in matrix quartz ribbons and by the irregular grain boundary network of plagioclase and pyroxene in which 120° triple junctions are absent. Attainment of equilibrium is suggested by uniform compositions of coexisting mineral phases (Chapter 2) and the absence of coronitic or symplectitic textures in these rocks. Near-peak *P-T* conditions of ca. 11 kbar and 820°C are estimated for this sample, and for other mafic granulites throughout the northeastern part of the complex (Chapter 2). A single metamorphic event is consistent with Ca and Mg growth profiles through garnet which show gradual and systematic variation (Chapter 2). Finally, quartz-only inclusions in garnet ensured that potential problems associated with mixing or contamination due to mineral inclusions were avoided. The randomly distributed pattern of quartz inclusions is consistent with single-stage garnet growth.

Garnet, clinopyroxene and plagioclase mineral separates and whole rock powder define an isochron (MSWD=0.96) with a slope equivalent age of 1894 ± 9 Ma (Fig. 3-6). The excellent statistical fit to the regression line confirms a lack of diffusion of Sm and Nd. An initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.51032 for the gabbro corresponds to a positive $\epsilon_{\text{Nd}}(1902 \text{ Ma})$ value of +2.8 (Table 3-2).

Samples External to the Kramanitar Complex

6. *Syntectonic syenogranite pegmatite (K-574)*

To constrain the timing of ductile deformation south of the Kramanituur Complex, a 50 kg sample of syntectonic syenogranite pegmatite (K-574) was collected from the Southern shear zone. The sample is from one dyke of a swarm of biotite-bearing syenogranite pegmatite dykes that occur within the shear zone and are concordant with the tectonic foliation of mylonitic metasedimentary rocks they cut. The dykes display a range of deformation states: from weakly foliated planar dykes (Plate 3-5a) to mylonitic porphyroclastic rocks with well-developed pinch and swell structure (Plate 3-5b). The variable intensity of strain is interpreted to reflect syntectonic timing of syenogranite magmatism with respect to shearing along the Southern shear zone. Early-intruded dykes, subject to a longer interval of deformation, perhaps at a higher strain rate, show a correspondingly higher degree of accumulated finite strain than later-intruded dykes of this suite. Differences in finite strain state are not attributed to variable degrees of strain partitioning between metasedimentary host and dykes, because boudinage of least foliated dykes is not observed. The pegmatite sampled contains a strong planar and linear shape fabric, parallel to that of mylonitic wallrocks, but is not itself mylonitized. Shear sense indicators within mylonitized syenogranite pegmatite elsewhere reveal oblique, north-side-up, dextral shearing, consistent with that observed throughout the Southern shear zone (Sanborn-Barrie, 1993, Chapter 1).

The pegmatite sample yielded large (~200 μm), euhedral, prismatic zircon of a uniform medium brown colour with a length to width ratio of 4:1 to 1:1. The central portions of a few crystals contain very fine inclusions, defining irregular wispy cores. Four single-grain zircon fractions were analysed: three clear, strongly abraded crystals with no discernable core material; and one euhedral tip from a large crystal with a visible core. The data are 0.5 to 1% discordant, with $^{207}\text{Pb}/^{206}\text{Pb}$ ages ranging from 2524 ± 2 Ma to 2536 ± 1 Ma (Table 3-1). These data do not define a single colinear array (Fig. 3-7) and can be interpreted in several ways. If the zircon analysed did not incorporate older inherited zircon, but experienced minor recent Pb loss, they could represent a system that crys-

tallized slowly from 2536 ± 1 Ma to 2524 ± 1 Ma, the age range recorded by the $^{207}\text{Pb}/^{206}\text{Pb}$ ages (Table 3-1). If they had incorporated an inherited component, the pegmatite may be as young as 2524 ± 7 Ma, the $^{207}\text{Pb}/^{206}\text{Pb}$ age of the single abraded euhedral tip. Alternatively, a more complex Pb-loss history may be reflected by the data. Indications of Pb loss at ca. 1.9 Ga from south of the complex come from a volcanic rock whose zircons record up to 40% Pb loss at 1.89 Ga (Davis and Sanborn-Barrie, unpublished data, 1998). Ca. 1.9 Ga Pb loss in zircons from the syenogranite pegmatite would imply a maximum age of magmatic crystallization for the syenogranite pegmatite sample of 2549 ± 10 Ma, the upper intercept age of a chord through the most concordant zircon fraction (Z1) for a fixed lower intercept of 1901 ± 2 Ma (Fig. 3-5a). Recent Pb loss, in this case, has mainly affected fraction Z4, the smallest zircon which was abraded for 7 hours (in comparison to 30 hours for Z2, 16 hours for Z1 and 13 hours for tip Z3).

Although this sample lacked accessory monazite, titanite or rutile, preventing further insight into its thermal history subsequent to magmatic crystallization, two very important constraints are derived from the zircon data. First, Archean-age rocks occur south of the Kramanituar Complex. These not only include the pegmatite suite, but also metasedimentary rocks they cut (see sample 7 below). Second, syntectonic emplacement of the syenogranite pegmatite swarm provides a Neoproterozoic age of ca. 2524-2550 Ma for oblique, ductile, north-side-up and dextral movement on the Southern shear zone.

7. *Metatexite from the Southern shear zone (K-386)*

Sillimanite-bearing semi-pelite (K-386) from the ca. 2550 Ma Southern shear zone was examined to further investigate the metamorphic history of Archean rocks south of the complex. These rocks, and their less-strained equivalents to the south, contain garnet porphyroblasts which record a two-stage prograde metamorphic history (Chapter 2). An Archean age for garnet growth is demonstrated by the fact that within the Southern shear zone garnets are wrapped by ca. 2550 Ma shear band foliations (Plate 3-5c), while south

of the shear zone they are wrapped by pre-2550 Ma $S_{1(\text{External})}$ foliation. GSC SHRIMP II ion microprobe analyses of zircon and monazite inclusions from chemically similar garnet in pelite from the MacQuoid Lake area, 50 km south, also indicate ca. 2560 to 2500 Ma garnet growth (Berman and Stern, in press).

Monazite crystals analysed are pale to medium yellow, anhedral to subhedral with an average size of 180 μm long by 120 μm diameter. Optical microscopy and backscatter SEM images show that monazite of this shape and size occurs exclusively as a matrix phase (Plate 3-6a), and are most commonly overgrown on biotite, which appears to have partly controlled monazite crystal form (Plate 3-6b). Three single-grain fractions of monazite yielded slightly discordant (0.3-0.5%) data whose $^{207}\text{Pb}/^{206}\text{Pb}$ ages overlap within error, defining an age of 1901 ± 1 Ma (Fig. 3-8).

Two multi-grain (100) fractions of amber, slender prismatic rutile were analysed from this sample (Table 3-1), however, small sample size and low abundance of U (7 to 9 ppm) made it impossible to obtain ages of useful precision.

8. *Metagabbro (K-704)*

A tabular mafic unit, 200 m wide, south of the Kramanituar Complex (Fig. 3-3) comprises medium-grained, equigranular, indurated rock with an assemblage of clinopyroxene-amphibole-plagioclase-quartz-garnet-titanite. In spite of its recrystallized nature, this unit is interpreted to be intrusive on account of its grain size and lack of volcanic textures. A 5 kg sample (K-704) was collected from a locality which is on strike with the Southern shear zone (Fig. 3-4). At this locality, the mafic rocks do not display a high degree of finite shear strain, but rather a moderately developed foliation which is obscured, in part, by superposed randomly oriented subhedral hornblende. These textures may reflect intrusion of this gabbro after movement on the Southern shear zone, and subsequent hydrothermal activity.

Medium honey-brown titanite recovered from this sample occurs as clear, subangular fragments, typically 200 μm long and rarely as subrounded disk-shaped grains. In the sample, titanite is a matrix phase associated with amphibole, garnet and clinopyroxene, and occurs as inclusions in garnet. Three multigrain fractions (40 grains each) with Th/U ratios of 1.16 to 1.41, are 0.6-4% discordant with $^{207}\text{Pb}/^{206}\text{Pb}$ ages ranging from 2273 to 2317 Ma (Table 3-1, Fig. 3-9). These analyses do not lie on a single mixing line, indicating complex isotopic systematics possibly involving: 1) components of inherited Pb; 2) more than one period of Pb loss or crystal growth; or 3) both of these. If the data are interpreted as representing one population, as suggested by the uniform character of the 40 grains from each of the three analysed fractions, a titanite closure date of 2326 ± 3 Ma is derived from the upper intercept of a chord (Fig. 3-9, inset) through the most concordant fraction and ca. 1.9 Ga, the time of titanite growth and Pb loss in zircons south of the complex (Davis and Sanborn-Barrie, unpublished data, 1998). In this scenario, all fractions would have lost Pb at ca. 1.9 Ga in the order of 5% (T3), 12% (T1) and 20% (T2), followed by a minimal amount of recent Pb loss from T2 (0.5% loss) and T3 (3% loss). If the data reflect more than one population of titanite, for which there is no textural or chemical evidence, they could represent a $2300 +5/-4$ Ma population, the upper intercept age of a regression line through T1 and T2, with an older inherited population contained within fraction T3. Although interpretation of these data is not straightforward, derivation of a slightly (0.6%) discordant analysis from a 40-grain fraction with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2294 ± 2 Ma does not support significant disturbance or growth of titanite in this rock at ca. 1.9 Ma.

9. *Coronitic gabbro (K-691)*

Foliated to gneissic plutonic country rocks north of the Kramanituur Complex are, in part, of Archean age (ca. 2675 Ma; Schau, 1980). Metamorphism of this region was dated using the Sm-Nd technique on a garnet-bearing gabbro (K-691) located 7 km north

of the complex (Fig. 3-3). The sample is a weakly foliated (056/38°SE), plagioclase porphyritic gabbro with an assemblage of plagioclase-clinopyroxene-amphibole-biotite-orthopyroxene-garnet-ilmenite-quartz. It is dominated by a magmatic texture, however, metamorphic amphibole occurs as continuous coronitic rims around clinopyroxene, and garnet forms partial rims at the interface between coronitic amphibole and matrix plagioclase. These microtextures are consistent with single-stage production of amphibole and garnet due to breakdown of clinopyroxene and plagioclase in the presence of water, or alternatively, two-stage production by partial hydration of clinopyroxene to produce amphibole coronae, followed by prograde metamorphism of amphibole and plagioclase to produce garnet. Garnet-amphibole equilibria yield estimated temperatures of 620-750°C and decompression from ca. 11 kbar to 8 kbar (Chapter 2).

Analysed garnet and plagioclase separates (Table 3-2) form a two-point isochron (Fig. 3-10) with a slope equivalent age of 1878 ± 27 Ma, which is interpreted to date the time of garnet formation ($\pm$ cooling). The igneous age of the gabbro is bracketed between 2.2 Ga, its maximum Nd model age, and 1878 ± 27 Ma, the time of metamorphism ($\pm$ cooling). This suggests it may be correlative with 2.19 Ga mafic dykes (Tulemalu swarm) that cut the MacQuoid-Gibson supracrustal belt (Tella et al., 1997b).

DISCUSSION

U-Pb and Sm-Nd geochronological data presented here for the Kramanituur Complex and its surroundings highlight the importance of Paleoproterozoic magmatic, metamorphic and tectonic activity in the interior of the Archean western Churchill Province. They also place constraints on the timing of older, mainly Neoarchean, metamorphism and deformation. The chronology of events, summarized below, provides a well-constrained template for this area, to which other parts of the Snowbird tectonic zone are compared.

Chronology of Events

Archean

The widespread occurrence of Archean plutonic rocks in the study area, previously indicated by ca. 2675 Ma granodiorite north of the complex and $\geq$ ca. 2573 Ma charnockite from the complex (Schau, 1980), is supported by an age of ca. 2550 Ma for syenogranite pegmatite south of the complex. In addition, a substantial component of Neoproterozoic supracrustal rock is represented by pre- ca. 2550 Ma metasedimentary rock into which the pegmatite intruded, and by minor metavolcanic rocks (*see* Fig. 3, Sanborn-Barrie, 1994 in Appendix II) from which a dacite fragment yielded a U-Pb zircon age between ca. 2655 and 2690 Ma (Davis and Sanborn-Barrie, unpublished data, 1998). These supracrustal rocks are interpreted to represent the northern extension of the MacQuoid-Gibson supracrustal belt (Fig. 3-2).

Neoproterozoic deformation is established at ca. 2550 Ma, based on the age of the syntectonic syenogranite pegmatite in the Southern shear zone. Neoproterozoic deformation south of the complex places important constraints on the timing of metamorphism in this region. Because garnet porphyroblasts in metasedimentary rocks are wrapped either by ca. 2550 Ma mylonitic foliations of the Southern shear zone, or by an earlier $S_{1(\text{external})}$ foliation, metamorphism of these rocks is interpreted to have occurred during the Archean. Preservation of Neoproterozoic tectonometamorphic assemblages and tectonic fabrics in rocks south of the Kramanitar Complex suggests that these country rocks were insulated from penetrative Paleoproterozoic high-grade tectonometamorphism, as recorded by the complex. Given the rapid cooling history of the complex, which requires extremely efficient exhumation, it seems likely that country rocks to the south were insulated from events within the complex by a structural break, that was likely active during exhumation (discussed below).

Amphibolite-facies dextral mylonitization at ca. 2550 Ma within the Southern shear zone is similar in age to ca. 2.6 Ga granulite-facies dextral transpression in the

Striding-Athabasca mylonite zone (Hanmer et al., 1994) and Angikuni Lake area (Aspler et al., 1998, 1999) to the southwest, and possible ca. 2.59 Ga amphibolite- to granulite-facies mylonitization in the Uvauk Complex to the east (Tella et al., 1994).

Paleoproterozoic

Metagabbro that cuts Archean supracrustal rocks immediately south of the complex contains titanite with an interpreted upper intercept age of ca. 2326 Ma (Fig. 3-9). Upper amphibolite- to granulite-facies assemblages in this rock (Chapter 2) indicate this is likely a metamorphic and/or cooling age. These data may reflect: 1) slow cooling of a Neoarchean metamorphic event; 2) magmatic crystallization and cooling in the deep crust; or 3) a separate ca. 2.3 Ga high-grade metamorphic event, interpreted elsewhere in the western Churchill Province (Crocker et al., 1993; Bickford et al., 1994; M. Williams, pers. comm., 1999). Most importantly, these titanite data are inconsistent with penetrative ca. 1.9 Ga high-grade metamorphism in country rocks south of the complex, thereby supporting the presence of a tectonic break that accommodated large displacements north of this sample, at the southern margin of the Kramanituur Complex.

Monazite ages of 1910 ± 3 Ma from kyanite-bearing mylonitic diatexite and 1908 ± 3 Ma from sillimanite-bearing paragneiss are interpreted to reflect Paleoproterozoic metamorphism, that culminated in granulite-facies conditions. In these paragneisses, metamorphic monazite likely crystallized at amphibolite-facies conditions of 500-550°C as a breakdown product of allanite (Smith and Barreiro, 1990), apatite and/or LREE-oxides (Kingsbury et al., 1993). The monazite ages are not interpreted to date peak metamorphism, or cooling, because slightly younger granulite-facies assemblages are contained within adjacent gabbro and gabbroic anorthosite. A magmatic crystallization age of 1902 ± 1.6 Ma for gabbroic anorthosite (discussed below) indicates that granulite-facies conditions prevailed ca. 8 m.y. after monazite closure. Although two high-grade events (i.e., at 1910-1908 Ma and ca. 1902 Ma) cannot be ruled out, geochronological data

together with textural and compositional observations are consistent with single-stage metamorphism from prograde conditions at ca. 1910 Ma leading to peak granulite-facies conditions by ca. 1901 Ma. Monazite data from paragneisses plot close to concordia (Fig. 3-5c,d), suggesting little disturbance of the U-Pb isotopic system after 1908 Ma. This is notable given subsequent granulite-facies temperatures up to, and exceeding, 850°C. Inheritance in monazite has been described in granitic rocks for which calculated melt temperatures are $710 \pm 30^\circ\text{C}$ (Copeland et al., 1988) and $>740^\circ\text{C}$ (Kingsbury et al., 1993), and also in rocks that have been metamorphosed to T in excess of 650°C (Copeland et al., 1988; Parrish, 1990; Kingsbury et al., 1993). Although the observation that inheritance in the U-Pb systematics of monazite is relatively rare suggests that monazite is less retentive of Pb than zircon, it would appear from these data that in some cases monazite remains closed to Pb diffusion to T up to 850°C . The restitic (ie., dry) nature of these samples with only trace amounts of biotite, may account for closure of monazite under anhydrous conditions up to, and possibly exceeding, 850°C . A $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1917 ± 6 Ma for a single monazite from sillimanite-bearing paragneiss may indicate prograde metamorphic conditions as early as ca. 1917 Ma.

A magmatic crystallization age of 1902 ± 1.6 Ma for gabbroic anorthosite is based on the euhedral to subhedral, prismatic morphology and internal euhedral zonation of many zircons, and also on their low uranium contents (28 to 86 ppm U) and relatively high Th/U ratios (0.46 to 0.60), irrespective of morphology. These values are consistent with magmatic zircon from Archean and Paleoproterozoic rocks of anorthositic and gabbro anorthositic composition (Sultan et al., 1994) but are inconsistent with those typically reported for zircon of metamorphic origin (Percival and Krogh, 1983; Corfu, 1988; Kamo and Davis, 1994). Two additional reasons why these data (Fig. 3-5a) are interpreted to date magmatic crystallization, rather than the effects of metamorphism superimposed on an older plutonic rock, are the absence of any indication of inherited Pb, and absence of evidence of high- T Pb loss from a zircon population that had crystallized prior to ca. 1902

Ma. For example, high temperature Pb loss in an "open" system may result in an *en echelon*, non-colinear data array produced via diachronous closure of zircon from an initial magmatic population.

A magmatic age of 1902 ± 1.6 Ma for gabbroic anorthosite is consistent with an $\epsilon_{\text{Nd}}(1902 \text{ Ma})$ value of +2.8 for gabbro (sample 5) which is comagmatic with the gabbroic anorthosite (Chapter 4). This positive ϵ_{Nd} value is similar to, but slightly less than, depleted mantle at ca. 1.9 Ga (ca. +4; DePaolo, 1981a; Goldstein et al., 1984; Liew and McCulloch, 1985) and is interpreted to reflect derivation from an undepleted (E-type MORB) mantle source (Chapter 4). While acknowledging that mantle-derived melts contribute heat to the deep crust (Furlong and Fountain, 1986; Bohlen and Mezger, 1989; Voshage et al., 1990; Wilshire, 1990), magmatic crystallization of gabbroic anorthosite under granulite-facies conditions some 6-8 m.y. *after* formation of metamorphic monazite in adjacent paragneiss suggests that mantle-derived mafic magmatism was attendant with high-grade metamorphism, but was not the cause of it. Relating metamorphic monazite growth to earlier thermal pulses related to a hitherto unrecognized older (i.e., 1917-1910 Ma) suite of mafic rocks can be tested by future geological mapping and geochronological studies.

In gabbroic anorthosite, gradients from low strain to mylonitic rocks, and from primary magmatic (orthopyroxene-plagioclase) to peak metamorphic (clinopyroxene-garnet-plagioclase $\pm$ quartz) assemblages (*see* Plate 3-1), establish that ductile deformation and metamorphism outlasted magmatic crystallization of gabbroic anorthosite at 1902 ± 1.6 Ma. Metamorphic conditions of 12-15 kbar and 850°C prior to and/or synchronous with ductile deformation is indicated by peak metamorphic assemblages that define the penetrative $S_{2(\text{KC})}$ fabric in both paragneiss and gabbroic rocks (Plate 3-4). Ductile deformation persisted during high- T decompression of these rocks. This is indicated by P - T estimates of 8-10 kbar and ca. 800°C from symplectitic reaction (garnet-breakdown) products localized within strain shadows of garnet porphyroblasts (Chapter 2). A mini-

imum age of ductile deformation at 1902.6 ± 1 Ma is provided by zircon and monazite data from a fine-grained leucogranite dyke that cuts the Northern shear zone. This age is indistinguishable from the crystallization age of gabbroic anorthosite, itself ductilely deformed within the Northern shear zone (Plate 3-1c), suggesting that heat from mantle-derived mafic magmas may have triggered crustal melting which led to contemporaneous production of late-tectonic leucogranite melts in this region. Overlap in the timing of high-grade metamorphism, mafic and felsic magmatism and deformation reveals the dynamic nature of this part of the western Churchill Province at 1910-1900 Ma.

Extremely rapid cooling of the Kramanituvar Complex by ca. 1901 Ma is indicated by overlap of zircon, titanite and rutile U-Pb ages from gabbroic anorthosite, overlap of zircon and monazite ages from leucogranite, and a rutile age of 1901 Ma for sillimanite-bearing paragneiss. A slightly younger rutile age of ca. 1896 Ma for kyanite-bearing paragneiss suggests a slower rate of cooling, and/or a slightly deeper crustal level for this part of the complex.

A Sm-Nd (Grt-Cpx-Pl-whole rock) isochron age of 1894 ± 9 Ma from metagabbro from the eastern part of the complex is a minimum age of Paleoproterozoic metamorphism. Because isotopic ratios of the garnet fraction largely control the slope of the isochron, the calculated age is highly dependent on the closure temperature of the Sm-Nd system in garnet. This has been estimated for garnet radii of 1-2 mm in anhydrous mafic rock at ca. 650°C (Burton et al., 1995) for slow-cooling conditions, and at >700-750°C (Cohen et al., 1988; Jagoutz, 1988; Henson and Zhou, 1995) for rapidly cooled terranes. Crystallization of garnet, plagioclase and clinopyroxene in the Kramanituvar metagabbroic suite at temperatures well above the closure temperature of garnet (for Sm-Nd) indicates that 1894 ± 9 Ma is a cooling age, or minimum age of metamorphism. The Sm-Nd cooling age is indistinguishable from the U-Pb cooling ages, because uncertainties in the decay constant for the Sm-Nd system are in the order of 30 m.y. at 1.9 Ga, for comparisons between Sm-Nd and U-Pb data (Renne et al., 1998).

The metamorphic history of rocks north of the complex appears to be coeval with that of the complex. For instance, high-*P* metamorphism is bracketed between 2.2 Ga, the maximum Nd model age of coronitic gabbro, and 1878 ± 27 Ma, its metamorphic cooling age (Sm-Nd Grt-Pl). Similarly, cooling is indicated by a K-Ar biotite ($T_c \sim 300^\circ\text{C}$) age of 1895 ± 44 Ma from Neoarchean granodiorite (Schau, 1980).

Semi-pelite south of the complex yielded monazite ages of 1901 ± 1 Ma, indicating an event in these rocks some 600 m.y. *after* the main stage(s) of their metamorphic mineral growth as recorded by pre- to syn- 2550 Ma garnet-biotite $\pm$ sillimanite assemblages (discussed above). The three single-grain monazite fractions analysed (*see* Fig. 3-8) do not lie on a mixing line between Archean and Paleoproterozoic metamorphic events, as might be expected for monazite that had experienced variable Pb loss due to a significant 1.9 Ga thermal overprint. A thermal pulse was probably not the principal mechanism causing growth of new monazites because the meta-pelites had already experienced anatectic melting and extraction of melt fluids during Neoarchean upper amphibolite-facies metamorphism (Chapters 1, 2). Further anatectic mobilization would not be expected unless the ca. 1.9 Ga thermal pulse was even higher than the Neoarchean event, and this is contradicted by the titanite age of ca. 2326 Ma in adjacent gabbroic rock. Furthermore, it is unlikely that these data reflect advective heat input from the Kramanitar gabbroic complex, given other isotopic evidence requiring a tectonic break between the complex and country rocks to the south. Rather, these data may reflect fluid-induced growth of new monazite associated with lower grade (lower amphibolite?) fluid fluxing (*cf.* Vavra and Schaltegger, 1999), possibly related to exhumation-related faulting at ca. 1.9 Ga. A post-tectonic hydrothermal origin for the monazite is supported by the subhedral morphology of monazite crystals, by its overgrowth on biotite (Plate 3-6*b*) and by the relative uniformity of U contents and low Th/U ratios for the three monazite grains analysed (Table 3-1). More analyses of texturally or chemically variable monazite may provide further constraints on the Archean metamorphic history of this sample. A titanite age of 1900 ± 2

Ma from an Archean volcanic rock 7 km south of the complex (Davis and Sanborn-Barrie, unpublished data, 1998) supports a hydrothermal/thermal event at this time.

Sometime after ca. 1901 Ma, the western Kramanituur Complex was exposed and unconformably overlain by clastic sedimentary rocks of the Baker Lake Group (Fig. 3-1). A maximum age of deposition of the basal South Channel conglomerate is 1901 Ma, the time at which the western Kramanituur Complex had cooled through ca. 400°C. This clastic succession was subsequently overlain by ultrapotassic volcanic rocks, whose intrusive equivalents are dated at 1832 ± 28 Ma (MacRae et al., 1996).

Time-averaged Cooling Rates

Time-averaged cooling rates for the Kramanituur Complex can be estimated using data from the gabbroic anorthosite sample. Closure of zircon in the order of ca. 850°C, of titanite (ca. 650-750°C) at 1902 ± 1.6 Ma, and of rutile (ca. 400°C) at the same time portrays a telescoped thermal history which corresponds to a time-averaged cooling rate of $>450^\circ\text{C Ma}^{-1}$. Very rapid cooling is consistent with overlapping zircon and monazite ages from the leucogranite dyke, and with the excellent preservation of granulite-facies mineral assemblages and microstructures that characterize much of the complex. This rate greatly exceeds that estimated for many regional granulite terranes (Harley, 1989) where cooling rates of $\leq 10^\circ\text{C Ma}^{-1}$ are attributed to isobaric re-equilibration. Very fast cooling rates, on the order of $150\text{-}350^\circ\text{C Ma}^{-1}$, can be driven by tectonic unroofing (Platt and Vissers, 1989) in cases where rapid uplift leads to uplifted isotherms and increased thermal gradients, followed shortly after by exhumation. For example, tectonic unroofing during the final stages of nappe emplacement and orogenic development in the Betic Cordilleras, Spain (Zeck et al., 1992; Andriessen and Zeck, 1996) is interpreted to account for cooling rates on the order of $250\text{-}350^\circ\text{C Ma}^{-1}$. Tectonic unroofing of the Valhalla metamorphic complex of North America, is thought to be accommodated by thrusting of the metamorphic complex onto colder basement rocks and coeval extensional faulting (Carr et al.,

1987). Locally this is postulated to have resulted in potential cooling rates of up to $1000^{\circ}\text{C Ma}^{-1}$ (Spear and Parrish, 1996).

Regional Implications

Marginal Paleoproterozoic Orogens

Historically, Paleoproterozoic reworking of Archean rocks of the Churchill Province has been referred to as the Hudsonian thermotectonic overprint, in part implying a genetic link with the Trans-Hudson orogeny (Davidson, 1972; Lewry and Sibbald, 1980; Lewry et al., 1985; Lucas et al., 1997). However, U-Pb and Sm-Nd data presented here for the Kramanituur Complex indicate that Paleoproterozoic tectonometamorphic and magmatic activity in this part of the western Churchill Province took place some 70 Ma *prior* to orogen-scale continental collisional activity associated with the Trans-Hudson Orogen, and is not linked to it.

Events at the Kramanituur Complex are synchronous with post-collisional processes along the northwest margin of the western Churchill Province where east-dipping subduction led to continental collision with the Slave Province at ca. 1.97-1.96 Ga (van Breemen et al., 1990; van Breemen et al., 1992; Tirrul and Grotzinger, 1990; Hanmer et al., 1992; Berman and Bostock, 1997) and subsequent collisional-related shortening. Further indentation of the western Churchill Province until 1.84 Ga is partly attributed to collisional events along the western margin of the Slave Province (Hoffman and Bowring, 1984).

The Snowbird tectonic zone

Penetrative ca. 1.9 Ga tectonometamorphic activity within the Kramanituur Complex highlights a major difference between the east-striking Chesterfield segment of the Snowbird tectonic zone and its northeast-striking segments (Fig. 3-11), where latest penetrative activity is mainly Archean. For instance, a strand of the Tulemalu fault zone

(Tella and Eade, 1986; Fig. 3-1) on Angikuni Lake is a steeply-dipping, strike-lineated dextral shear zone (Aspler et al., 1998; Aspler et al., 1999), which was active at ca. 2.61 Ga, the age of syn-tectonic granite (W. Davis, unpublished data, 1997). The 400 km long, northeast-striking Striding-Athabasca segment (Fig. 3-1) comprises a voluminous tonalitic complex with mafic-anorthositic inclusions and mafic dykes, postulated to have crystallized prior to ca. 3100 Ma (Hanmer et al., 1994). Penetrative, high-grade, mid- and late-Archean deformation events are documented at ca. 3120 Ma and ca. 2600 Ma (Hanmer et al., 1994), the latter involving northwest-directed shortening across a bulk dextral shear plane (Hanmer, 1997a). These rocks were subsequently affected by variable degrees of static recrystallization and retrogression to amphibolite facies. In contrast to the Kramanituar Complex, Paleoproterozoic tectonometamorphism in the Tulemalu and Striding-Athabasca segments is interpreted to consist of localized low-grade reworking and minor magmatic activity (Hanmer et al., 1994; Hanmer, 1997a), with high-grade reactivation restricted to a narrow (<500 m wide) ca. 1907 Ma dextral granulite facies mylonite belt, 10 km long, at Wholdaia Lake (Hanmer, 1997a).

Exposed parts of the southwest Snowbird tectonic zone include the Virgin River shear zone (*VRsz* in Fig. 3-1) and adjacent granulite-facies gneisses. The gneissic rocks are interpreted to record a penetrative Mesoarchean history, overprinted by high-grade events at ca. 2.3-2.2 Ga and ca. 2.0-1.9 Ga, with low-grade brittle deformation and retrogression at ca. 1.85 Ga, the time of greenschist-facies mylonitization along the Virgin River shear zone (Carolan and Collerson, 1989; Crocker et al., 1993; Bickford et al., 1994). The subsurface Alberta segment of the Snowbird tectonic zone (*eA* in Fig. 3-1) is interpreted by Ross et al. (1991, 1995) as a ca. 1856 to 1796 Ma east-vergent collisional orogen associated with the Trans-Hudson Orogen (Lucas et al., 1996).

The east-striking Chesterfield segment of the Snowbird tectonic zone records widespread evidence of high-grade Paleoproterozoic tectonometamorphism. East of the Kramanituar Complex, limited isotopic data allow only broad constraints on this activity.

At the Uvuak Complex (Fig. 3-1), gabbroic anorthosite has been postulated to be mid-Archean (2.9 and 2.77 Ga) age, based on a Sm-Nd model age (Tella et al., 1994). However, least-discordant, single-grain zircon fractions described as having magmatic morphology, have $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 1917 Ma and 1920 Ma (J.C. Roddick, unpublished data, 1992). These data leave open the possibility of a Paleoproterozoic age (with inheritance) for the Uvuak gabbroic anorthosite, possibly slightly older than the Kramanituak gabbroic suite. Within the Uvuak Complex, two penetrative syntectonic granulite-facies events at 2600-2560 Ma and ca. 1940 Ma have been proposed (Tella et al., 1994). Further east, granulite-facies metamorphism at Daly Bay is bracketed between ca. 2067 Ma and 1951 ± 26 Ma (Gordon, 1988). Collectively, these data point to diachronous, high-grade Paleoproterozoic magmato-tectonometamorphism within the east-striking Chesterfield segment at ca. 2067-1951 Ma for Daly Bay in the east, ca. 1940-1917 Ma for the Uvuak Complex, and at ca. 1910-1900 Ma for the Kramanituak Complex in the west. Recognition of the Chesterfield segment as a discrete ca. 1.9 Ga structural entity raises the following questions. Was the Chesterfield segment a locus of mantle-derived mafic magma, which was subsequently uplifted? Or, is the lithologic association observed at the Kramanituak Complex and elsewhere in the Chesterfield segment typical of the deep crust everywhere, such that this segment is remarkable only in its exhumation history at ca. 1.9 Ga?

The presence of intermediate- to high-grade dextral mylonites of Neoarchean age throughout the northeast-striking (Striding-Athabasca and Tulemalu) and east-striking (Kramanituak, Uvuak $\pm$ Daly Bay?) segments of the Snowbird tectonic zone, supports the presence of a throughgoing belt of Neoarchean mylonites across the western Churchill Province (Fig. 3-11a). If the geophysically-defined Snowbird tectonic zone was an ancestral, Neoarchean dextral transpressive fault zone, it may have acted as a crustal-scale heterogeneity which focussed magmatic and tectonic activity along favourably-oriented segments, such as the east-striking Chesterfield segment, during Paleoproterozoic time (Fig. 3-11b).

CONCLUSIONS

U-Pb and Sm-Nd geochronology presented here demonstrate that, in contrast to many other Archean cratons, Paleoproterozoic (ca. 1910-1901 Ma) magmatism, deformation and metamorphism penetratively affected part of the interior of the Archean western Churchill Province. This involved: 1) the existence, at ca. 1915-1902 Ma of largely Archean-age, thickened ($>40\text{-}55$ km depth) crust that sustained syn-tectonic granulite-facies metamorphism; 2) deep-seated intrusion of mantle-derived mafic magmas and minor leucocratic granitic melts, and their crystallization at ca. 1902 Ma under granulite-facies conditions estimated at 13.25 ± 1.75 kbar and $850 \pm 100^\circ\text{C}$; 3) penetrative deformation and near-isothermal decompression of these rocks before ca. 1901 Ma, to crustal levels of ca. 30 km (8 kbar); and 4) extremely rapid cooling ($>450^\circ\text{C}/\text{Ma}$) by 1901-1896 Ma, with exhumation to surface before ca. 1832 Ma to be unconformably overlain by a coarse clastic succession, in part derived from local high-grade source rocks.

The isotopic data presented here clearly establish the importance of ca. 1.9 Ga high-grade tectonometamorphic activity along the east-striking segment of the geophysically-defined Snowbird tectonic zone. This contrasts with other parts of the Snowbird tectonic zone where a largely Mesoarchean to Neoarchean history is recorded by the northeast-trending Tulemalu and Striding-Athabasca segments, and ca. 1.85-1.7 Ga reactivation is interpreted for the subsurface Alberta segment. Isotopic, thermobarometric and structural data from the Kramanituar Complex and elsewhere suggest that high-grade tectonites of the Chesterfield segment may share a common, though possibly diachronous, history involving 2.0 to 1.9 Ga metamorphism, magmatism and deformation. This activity may reflect far-field, intracontinental effects of collision between the Slave and Churchill cratons (Thelon Orogen), localized on a favourably-oriented segment of a Neoarchean lithospheric fault.

Table 3-1. U-Pb analytical data for the Kramanitar Complex and country rocks

Fraction ^a	Wt (ug) ^b	U (ppm) ^c	Pb ^d (ppm) ^e	Pb (pg) ^f	²⁰⁶ Pb/ ²³⁸ U	Th	²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²⁰⁶ Pb	Age (Ma) ^g	Corr Coeff	²⁰⁷ Pb/ ²⁰⁶ Pb	Age (Ma) ^g	% Disc.
1. GABBROIC ANORTHOITE (HVA K-93-676)														
Z1 [s]eu	17	28	11	961	11	12.9	0.569	0.3431 ± 0.22%	0.11619 ± 0.18%	1901.8 ± 7.3	0.79	1898.5 ± 6.4	1900.2 ± 5.0	-0.2
Z2 [3]sb	25	86	32	7978	6	10.6	0.456	0.3430 ± 0.13%	0.11636 ± 0.06%	1901.1 ± 4.4	0.89	1901.1 ± 2.3	1901.1 ± 2.3	0
Z3 [s]eu	24	53	17	1853	12	13.6	0.605	0.2832 ± 0.18%	0.11622 ± 0.11%	1607.5 ± 5.2	0.88	1898.9 ± 3.8	1738.0 ± 3.7	17.3
T1 [40]pl-y,fr-d	146	38	29	1296	90	56.2	4.94	0.3442 ± 0.10%	0.11657 ± 0.08%	1907.0 ± 3.3	0.83	1904.2 ± 2.8	1905.7 ± 2.4	-0.2
T5 [33]pl-y,an	100	31	25	560	111	60.2	5.84	0.3423 ± 0.13%	0.11639 ± 0.17%	1898.0 ± 4.4	0.72	1901.5 ± 6.2	1899.6 ± 4.1	0.2
R4 [50]r-b,fr	162	31	10	491	204	0.9	0.0352	0.3355 ± 0.17%	0.11563 ± 0.24%	1864.8 ± 5.6	0.67	1899.7 ± 8.5	1876.6 ± 5.4	1.5
R5 [73]r-b,fr	180	37	12	1101	121	0.7	0.0287	0.3401 ± 0.13%	0.11622 ± 0.11%	1887.4 ± 4.4	0.77	1898.8 ± 4.0	1892.8 ± 3.0	0.7
2. LEUCOGRANITE DYKE (HVA K-92-198)														
Z1 [4]an,eq	5	360	121	2314	16	2.2	0.0874	0.3432 ± 0.16%	0.11644 ± 0.05%	1901.9 ± 5.2	0.95	1902.3 ± 2.0	1902.1 ± 3.0	0
Z2 [6]sb	6	735	236	2121	42	4.1	0.165	0.3218 ± 0.11%	0.11359 ± 0.07%	1798.7 ± 3.3	0.86	1857.6 ± 2.5	1826.1 ± 2.3	3.6
Z3 [45]eu	18	336	109	1602	81	3.5	0.138	0.3256 ± 0.11%	0.11417 ± 0.11%	1817.1 ± 3.4	0.84	1866.8 ± 4.1	1840.4 ± 2.8	3
Z4 [15]eu	3	849	279	510	107	4.9	0.196	0.3266 ± 0.26%	0.11464 ± 0.30%	1821.8 ± 8.1	0.69	1874.2 ± 10.7	1846.3 ± 7.0	3.2
Z5 [s]sb	1	754	256	1289	19	4.8	0.194	0.3366 ± 0.23%	0.11581 ± 0.13%	1870.5 ± 7.4	0.87	1892.5 ± 4.8	1881.0 ± 4.6	1.3
M1 [s]y,an	2	3047	3619	9552	14	72.3	10	0.3410 ± 0.09%	0.11646 ± 0.03%	1891.5 ± 3.0	0.95	1902.6 ± 1.2	1896.8 ± 1.8	0.7
M2 [s]pl-y,sb	4	1056	1787	5384	18	80.7	16.1	0.3381 ± 0.12%	0.11651 ± 0.06%	1877.5 ± 3.9	0.86	1903.3 ± 2.3	1889.8 ± 2.1	1.6
3. KYANITE-BEARING PARAGNEISS MYLONITE (HVA K-92-290)														
M1 [s]pl-y,an	2	901	4468	3048	13	93.3	53.4	0.3450 ± 0.12%	0.11697 ± 0.06%	1910.8 ± 4.1	0.91	1910.5 ± 2.1	1910.6 ± 2.5	0
M2 [s]pl-y,an	3	1487	7657	5337	17	93.6	56.2	0.3414 ± 0.31%	0.11673 ± 0.30%	1893.3 ± 10.1	0.88	1906.8 ± 10.7	1899.8 ± 7.1	0.8
M3 [s]y,sb	3	568	3185	6193	6	94.1	61.2	0.3434 ± 0.11%	0.11693 ± 0.05%	1902.8 ± 3.5	0.77	1909.7 ± 1.9	1906.1 ± 1.9	0.4
R1 [67]br,an,fr	738	31	10	2383	194	0.1	0.003	0.3411 ± 0.09%	0.11602 ± 0.05%	1892.0 ± 2.9	0.89	1895.8 ± 1.8	1893.8 ± 1.9	0.2
R2 [100]r-br	413	30	10	649	389	0.5	0.0192	0.3425 ± 0.09%	0.11596 ± 0.16%	1898.5 ± 3.1	0.69	1894.8 ± 5.7	1896.7 ± 3.6	-0.2
4. SILLIMANITE-BEARING PARAGNEISS MYLONITE (HVA K-93-440)														
M1 [s]sb	21	942	1369	31227	14	77	12.9	0.3453 ± 0.12%	0.11683 ± 0.06%	1911.9 ± 4.1	0.89	1908.2 ± 2.1	1910.1 ± 2.1	-0.2
M2 [2]pl-y,an	6	476	4727	3112	21	96.7	112	0.3403 ± 0.39%	0.11694 ± 0.08%	1888.2 ± 12.8	0.98	1909.9 ± 2.8	1898.6 ± 6.9	1.3
M3 [s]pl-y,sb	2	2518	14534	6822	16	94.3	63.3	0.3424 ± 0.16%	0.11742 ± 0.17%	1898.3 ± 5.2	0.62	1917.4 ± 6.0	1907.4 ± 3.6	1.1
R1 [30]br,fr	315	31	10	1740	110	0.5	0.0175	0.3323 ± 0.08%	0.11479 ± 0.06%	1849.5 ± 2.7	0.86	1876.5 ± 2.2	1862.3 ± 2.0	1.7
R2 [30]br,fr	309	35	11	2279	95	0.2	0.0068	0.3409 ± 0.08%	0.11628 ± 0.05%	1891.2 ± 2.7	0.95	1899.8 ± 1.8	1895.3 ± 1.9	0.5
R3 [31]br,fr	379	30	10	560	426	0.5	0.0184	0.3425 ± 0.12%	0.11635 ± 0.16%	1898.9 ± 3.9	0.71	1900.9 ± 5.7	1899.8 ± 3.8	0.1
R4 [100]br,eu	258	25	8	792	166	1	0.0382	0.3400 ± 0.10%	0.11641 ± 0.13%	1886.5 ± 3.3	0.73	1901.8 ± 4.6	1893.8 ± 3.1	0.9
5. SYENOGRAHITE PEGMATITE (HVA K-93-574)														
Z1 [s]br,eu	20	1153	581	51263	13	4.7	0.206	0.4781 ± 0.08%	0.16723 ± 0.03%	2519.0 ± 3.5	0.96	2530.1 ± 1.0	2525.2 ± 1.8	0.5
Z2 [s]br,eu	60	527	267	103966	9	5.1	0.223	0.4791 ± 0.09%	0.16784 ± 0.03%	2523.2 ± 3.7	0.97	2536.2 ± 0.9	2530.4 ± 1.9	0.6
Z3 [s]br,lp	30	972	478	106409	8	3.4	0.147	0.4732 ± 0.29%	0.16665 ± 0.03%	2497.6 ± 12.2	0.99	2524.3 ± 0.9	2512.3 ± 5.5	1.2
Z4 [s]br,eu	11	1227	617	47364	8	4.8	0.213	0.4765 ± 0.09%	0.16757 ± 0.03%	2512.1 ± 3.5	0.96	2533.5 ± 0.9	2524 ± 1.8	1
6. SEMI-PELITE, SOUTH (HVA K-93-386)														
M1 [s]y,an	6	6786	5415	60835	15	58.7	5.48	0.3417 ± 0.10%	0.11645 ± 0.03%	1894.8 ± 3.3	0.97	1902.4 ± 1.0	1898.5 ± 1.9	0.4
M2 [s]pl-y,an	5	9792	5879	47384	21	45.1	3.17	0.3418 ± 0.08%	0.11633 ± 0.03%	1895.4 ± 2.8	0.96	1900.6 ± 1.0	1897.9 ± 1.7	0.3
M3 [s]pl-y,sb	7	3714	4111	33644	16	70.2	9.06	0.3421 ± 0.09%	0.11639 ± 0.03%	1896.6 ± 2.8	0.96	1901.5 ± 1.0	1898.9 ± 1.7	0.3
R1 [100]br,eu	111	7	3	64	352	5.6	0.229	0.3574 ± 0.59%	0.11738 ± 2.0%	1970.0 ± 20.1	0.7	1916.7 ± 71.1	1944.2 ± 40.6	-3.2
R2 [100]br-gn,eu	95	9	3	167	104	3	0.12	0.3306 ± 0.45%	0.11908 ± 0.76%	1841.2 ± 14.4	0.62	1942.5 ± 27.1	1889.2 ± 16.2	5.9
7. HORNFELTIC GABBRO, SOUTH (HVA K-93-704)														
T1 [40]pl-br,fr	291	92	52	2124	305	25.9	1.42	0.4247 ± 0.13%	0.14552 ± 0.05%	2282.0 ± 5.0	0.94	2293.9 ± 1.7	2288.3 ± 2.7	0.6
T2 [37]pl-br,fr	270	106	58	2267	294	26.6	1.46	0.4120 ± 0.30%	0.14379 ± 0.05%	2224.2 ± 11.3	0.97	2273.4 ± 1.7	2249.9 ± 5.5	2.5
T3 [36]pl-br,fr	342	96	53	1946	399	25.9	1.41	0.4153 ± 0.22%	0.14747 ± 0.06%	2239.1 ± 8.5	0.99	2316.7 ± 2.1	2279.9 ± 4.4	3.9

^a Z, zircon; M, monazite; T, titanite; R, rutile. Unless otherwise stated all zircons are clear, transparent grains, from least paramagnetic fractions of Frantz separates, free of cracks and inclusions, and are abraded (Krogh, 1982). [s], single crystal; [3], number of grains in fraction; eu, euhedral; sb, subhedral; an, anhedral; lp, crystal lip; br, brown; pl, pale; y, yellow; r, red.

^b weighing error is 1 microgram; ^c radiogenic Pb; ^d = Radiogenic Pb; ^e = Radiogenic Pb, corrected for spike and Pb fractionation of 0.09%; ^f = total common Pb in analysis corrected for spike and fractionation;

^g corrected for Pb and U laboratory blank where 208/204:207/204:206/204 are 19.02:15.65:38.23; errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean when expressed in Ma; ^h corrected for common Pb and laboratory blank, errors are 2 standard errors of the mean in Ma.

Table 3-2. Sm-Nd isotopic data for whole rock and mineral separates from gabbroic rocks.

Fraction	wt. (μm)	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Sm}/^{144}\text{Nd}$ (2σ)	$\epsilon_{\text{Nd}(i)}$
5. GRANULITE FACIES METAGABBRO (HVA K-92-1053)					UTM 15 458665E/7109695N	
WR	286.96	3.0033	8.9315	0.20	0.512848 ± 4	+2.8
Pl	80.12	0.0580	0.6315	0.056	0.511001 ± 22	
Cpx	40.58	4.7545	15.9844	0.18	0.512565 ± 10	
Grt	68.79	2.404	1.732	0.84	0.520791 ± 13	
9. CORONITIC METAGABBRO (HVA K-93-691)					UTM 15 444007E/7124677N	
WR		3.35	21.05	0.10	0.511454 ± 7	-0.6
Grt	29	2.88	3.41	0.51	0.516210 ± 21	
Pl						

Figure 3-1. Geology of the western Canadian Shield after Hoffman (1988). Segments of the proposed Snowbird tectonic zone (STZ) within the western Churchill Province (unornamented) comprise, from northeast to southwest: the Chesterfield segment which includes the Daly Bay complex (DBc), Uvauk complex (Uc), and Kramanituvar Complex (KC); the Tulemalu fault zone (Tfz); the Striding-Athabasca mylonite zone (SAMz); and the Virgin River shear zone (VRsz) and subsurface east Alberta segment (eA). Other abbreviations are sz-shear zone, mz-magmatic zone.

Figure 3-2. Archean supracrustal belts of the western Churchill Province subdivided according to age. Solid: ca. 2.9-2.75 Ga; Shaded: ca. 2.71-2.68 Ma. KC denotes Kramanituvar Complex.

Figure 3-3. Geology of the Kramanituvar Complex and surrounding rocks, modified from Schau and Ashton (1980), Schau (1983), Sanborn-Barrie (1993, 1994). Closed diamond symbols indicate geochronologic sample sites for this study. Open diamond symbols with age in Ma indicate U-Pb sample sites of Schau (1980). Letter A indicates lenses of kyanite-bearing paragneiss, discussed in text.

Figure 3-4. Distribution of major ductile shear zones. Dark shading delineates the Northern shear zone with a restraining bend at *b*, and antidilational right-stepping discontinuity at *d*. Diagonal lines delineate the right-stepping Southern shear zone which is offset by a northeast-striking sinistral fault (unpatterned), which also cuts the Northern shear zone. Open diamond symbols indicate location of geochronologic sample sites of this study shown in Figure 3-3.

Figure 3-5. Concordia diagrams for the Kramanituvar Complex. Zircon-unfilled; monazite-light shade; titanite-dark shade; rutile-line fill. *a*) zircon, titanite, rutile data from gabbroic anorthosite (sample 1); *b*) zircon, monazite data from leucogranite dyke (sample 2); *c*) monazite, rutile data from mylonitized kyanite-bearing paragneiss (sample 3); *d*) monazite, rutile data from mylonitized sillimanite-bearing paragneiss (sample 4).

Figure 3-6. Sm-Nd isochron diagram for granulite facies gabbro (sample 5). The garnet (Grt), clinopyroxene (Cpx), plagioclase (Pl) and whole rock (WR) form a well-correlated (MSWD=0.96) isochron age of 1894 ± 9 Ma.

Figure 3-7. Concordia diagram of zircon data from syenogranite pegmatite (sample 6).

Figure 3-8. Concordia diagram of monazite data from amphibolite-facies semi-pelite (sample 7) from the Southern shear zone.

Figure 3-9. Concordia diagram of titanite data from metagabbro (sample 8) exposed south of the complex.

Figure 3-10. Sm-Nd isochron diagram for coronitic gabbro (sample 9) located north of the complex. Two-point (Grt-Pl) isochron has a slope equivalent age of 1878 ± 27 Ma.

Figure 3-11. Sketch of the western Churchill Province highlighting events associated with the Snowbird tectonic zone through time. *a*) recognized occurrences of Neoarchean mylonites (sigmoidal symbol), consistent with a throughgoing dextral transpressive deformation zone represented by the Neoarchean Snowbird tectonic zone (*STZ*). Open arrows denote constraints on local shortening across the zone, consistent with a far-field northwest-directed shortening regime, from the Southern shear zone (*SSZ*), MacQuoid (*MGB*), Angikuni Lake (*ANG*) and Striding-Athabasca (*SA*) areas. *b*) ca. 1.97 Ga collision of the Slave craton with the western Churchill Province, and continued shortening, may have been an influence on ca. 1.9 Ga intracontinental tectonometamorphic activity along the Chesterfield segment of the Neoarchean, ancestral Snowbird tectonic zone.

Plate 3-1. Gabbroic anorthosite from the Kramanituvar Complex. *a)* Low strain, coarse-grained foliated gabbroic anorthosite (sample 1) dated in this study; *b)* Heterogeneously deformed gabbroic anorthosite; *c)* Ultramylonitized gabbroic anorthosite.

Plate 3-2. Zircon from least strained and least recrystallized gabbroic anorthosite (sample 1). *a)* Prismatic zircons with analysed single-crystal fractions Z1 and Z3 indicated. *b)* Scanning electron microprobe image of euhedral, prismatic zircon with internal zoning parallel to euhedral crystal faces. *c)* Anhedral, subrounded zircons with arrows denoting the three grains that comprised fraction Z2.

Plate 3-3. Plan view of 20 cm wide leucogranite dyke (sample 2) oriented 194/64°W, that cuts the penetrative, east-striking, subvertically-dipping mylonitic foliation of the Northern shear zone. Deflection of mylonitic foliation (left of 15 cm long pencil) shows an apparent sense of dextral offset.

Plate 3-4. Photomicrographs of granulite facies rocks within the complex, cut parallel to mineral lineation and perpendicular to foliation. *a)* Kyanite-bearing paragneiss (sample 3) from the Northern shear zone with shear foliation defined by ribbons of quartz (clear), layered subdomains of finely recrystallized feldspar (light gray mosaic texture) and aligned rutile (dark needles). Garnets (medium gray) at lower centre show a shape-preferred orientation parallel to the shear direction. *b)* Metagabbro (sample 5) with a metamorphic assemblage of garnet (Grt), clinopyroxene (Cpx), plagioclase (Pl) and pargasite (Prg) with minor magnetite/ilmenite (black). Some garnets contain round quartz inclusions 20-80 microns across.

Plate 3-5. Country rocks south of the complex. *a)* Variably strained syenogranite pegmatite dykes cutting pelitic and granitic mylonite (site of sample 6). *b)* Strongly deformed syenogranite dyke showing well-developed pinch and swell structure, located 2 m from dykes shown in *a)*. *c)* Garnet porphyroblastic sillimanite-biotite pelite with C'-type shear band cleavage transecting the main C-foliation. General shear direction is parallel to 15 cm long pencil. Dextral shear sense.

Plate 3-6. Backscatter electron images of semi-pelite *K*-386 (sample 7). *a*) 3 mm garnet with poikiloblastic core and inclusion-free rim. Matrix monazite (dated) are indicated by white arrows. *b*) detail of subhedral matrix monazite (M) from semi-pelite *K*-386 associated with biotite (Bt). Monazite morphology is interpreted to be controlled, in part, by biotite crystal faces.

CHAPTER 4

Geochemistry of gabbroic rocks from the 1.9 Ga Kramanituar Complex and surrounding country rocks of the Archean western Churchill Province

ABSTRACT

The Paleoproterozoic Kramanituar Complex is a window of deep-crustal rocks exposed in the interior of the Archean western Churchill Province. It is dominated by high-*P* granulite-facies mafic metaplutonic rocks, including gabbro, gabbroic anorthosite, norite and anorthosite. Major and trace element geochemistry indicate that these rocks, dated at 1902 ± 1.6 Ma (U-Pb zircon), likely represent a cogenetic suite derived from an undepleted mantle source. Gabbroic rocks hosted within Archean country rocks north and east of the Kramanituar Complex, are petrochemically distinct from the Kramanituar suite and appear to have been derived through partial melt processes.

INTRODUCTION AND REGIONAL SETTING

The Kramanituar Complex is one of several high-grade, mafic-dominated, tectono-metamorphic complexes that coincide with the surface trace of the Snowbird tectonic zone, a geophysically-defined feature that extends over 2000 km from the foothills of the Rocky Mountains to Hudson Bay (Fig. 4-1). In order to better understand the role of the Snowbird tectonic zone in the tectono-metamorphic evolution of the western Churchill Province, several recent studies have investigated lithotectonic, geothermobarometric and geochronologic aspects of its associated high-grade complexes (Gordon, 1988; Tella et al., 1994; Snoeyenbos et al., 1995; Hanmer, 1997; Aspler et al., 1999; Sanborn-Barrie and Hanmer, 1999; Sanborn-Barrie et al., 1999; Sanborn-Barrie and Berman, 1999). Collectively, these studies have shown that high-grade complexes along the Snowbird tectonic zone share similarities in terms of lithological association, structural style and metamorphic grade, but that they differ with respect to the timing of latest penetrative tectono-metamorphism and magmatism. Available constraints indicate that the regional northeast-striking segment of the Snowbird tectonic zone (Fig. 4-1) experi-

enced latest penetrative tectono-metamorphism during Neoarchean time (Hanmer et al., 1994; Hanmer, 1997a; Aspler et al., 1999), whereas the regional east-striking Chesterfield segment (Fig. 4-1) experienced high-grade tectonic and magmatic activity during the Paleoproterozoic (Gordon, 1988; Tella et al. 1994; Sanborn-Barrie et al., 1999), subsequent to Neoarchean events.

The presence of ca. 1.9 Ga mafic intrusive rocks within the Chesterfield segment of the Snowbird tectonic zone allows the opportunity to investigate processes of Paleoproterozoic crustal evolution in this part of the Archean western Churchill Province. The geochemistry of gabbroic rocks of the Kramanituar Complex, and of gabbroic country rocks to the complex, are presented and analysed here to address magmatic relationships that could not be resolved by field mapping, petrographic and/or isotopic methods alone. Trace element geochemistry is used to assess whether gabbroic and anorthositic rocks of the Kramanituar Complex are cogenetic with gabbroic anorthosite, dated at 1902 ± 1.6 Ma (Sanborn-Barrie et al., 1999) in order to better constrain the extent of Paleoproterozoic magmatic activity in this area. Similarly, trace element data are used to compare the Kramanituar gabbroic suite with gabbroic country rocks to determine the extent of magmatism of similar chemistry, and perhaps similar age. Lastly, geochemical characterization of the Kramanituar Complex gabbroic suite is considered in the context of mantle source composition, and the tectonic setting within which these rocks evolved.

THE KRAMANITUAR COMPLEX

The Kramanituar Complex is a coherent, 850 km² elliptical exposure of high-grade rocks (Fig. 4-2) that is dominated by granulite-facies metagabbro-gabbroic anorthosite-norite-anorthosite, dated at 1902 ± 1.6 Ma (Sanborn-Barrie et al., 1999) from one locality (*K-676b*, this study). It includes charnockite layers of Archean age (ca. 2573 $\pm$ 28/-5 Ma; Schau, 1980) and panels of sillimanite- and kyanite-bearing paragneiss and diatexite of probable primary Archean age. Penetrative Paleoproterozoic (ca. 1.91-1.90

Ga) tectonometamorphism of the complex has obscured essentially all primary magmatic features of these rocks (Sanborn-Barrie and Hanmer, 1999) and has resulted in the penetrative development of steeply dipping, northwest- to west-striking L/S fabrics defined mainly by granulite facies assemblages (12-15 kbar, 800-850°C; Sanborn-Barrie and Berman, 1999). These fabrics include a steep belt, 3-6 km wide, of granulite facies ribbon mylonites (Northern shear zone; inset Fig. 4-2), which is cut by a leucogranite dyke dated at 1902.6 ± 1 Ma, providing a minimum age of penetrative ductile deformation for the complex (Sanborn-Barrie et al., 1999).

Prograde metamorphism of the complex at 1910-1908 Ma, and possibly as early as 1917 Ma, is recorded by monazite ages from sillimanite- and kyanite-bearing paragneiss. High-grade conditions continued until at least 1902 ± 1.6 Ma, the primary magmatic U-Pb zircon age of gabbroic anorthosite, part of the high-*P* metagabbroic suite. Exhumation of the complex is recorded by rocks with disequilibrium textures, in that these record a near-isothermal (850°C-750°C) decompression history from peak conditions of ca. 13 kbar to 8-9 kbar (Sanborn-Barrie and Berman, 1999). Exhumation to mid-crustal levels led to very rapid cooling ($>450^\circ\text{C}/\text{Ma}$) of the complex between 1902 Ma and 1901 Ma. This is recorded by widespread titanite and rutile ages of 1901-1898 Ma (Sanborn-Barrie et al., 1999) and is reflected by the general absence of textures indicative of re-equilibration or thermal relaxation in these rocks.

COUNTRY ROCKS TO THE KRAMANITUAR COMPLEX

The Kramanituvar Complex is in contact with mainly amphibolite facies Archean rocks. Geothermobarometric and geochronologic constraints on their tectonometamorphic evolution point to a similar Paleoproterozoic history to the north, but indicate an older, mainly Neoarchean history to the south.

North and east of the complex, foliated to gneissic quartzo-feldspathic plutonic rocks, with a U-Pb zircon age of $2675 \pm 33/-11$ Ma from one locality (Schau, 1980) host

hornblende-bearing gabbroic rocks. Two of these gabbroic units yield estimated P - T conditions of 8-13 kbar and $760 \pm 60^\circ\text{C}$ (Grt-Cpx-Pl-Amp), indicating the presence of high- P rocks beyond the north and east boundaries of the Kramanituur Complex.

South of the complex, metasedimentary rocks and minor ca. 2.68 Ga metavolcanic rocks (Davis and Sanborn-Barrie, unpublished data, 1998) contain a regional north-east-striking, moderately northwest-dipping foliation which becomes more intense and wraps into the dextral, right-stepping Southern shear zone (Fig. 4-2 inset). An age of ca. 2550 Ma for syntectonic syenogranite pegmatite within the Southern shear zone dates movement on the shear zone and provides a minimum Neoproterozoic age for structures south of the complex. These Archean-age rocks include a significant component of gabbro. Near Chesterfield Inlet, the country rocks are cut by undated gabbro dykes (eg., K-1029 and K-709, this study) that are shown here to be geochemically similar to the Kramanituur gabbroic suite, and are interpreted to be related to it. Further work on the distribution of these dykes may help constrain the presence and location of a proposed cryptic fault (see Chapters 1, 2, 3) south of the complex. That these dykes are related to the complex does not preclude the presence of a tectonic break, in that they could represent remote mafic apophyses in the upper plate of the proposed detachment fault.

GEOCHEMISTRY

Analytical Techniques

Representative samples of gabbroic units were analysed at the Geological Survey of Canada for major, minor and trace elements (GSC, 1996). Major and minor element oxides were analysed by fused disk wavelength dispersive X-Ray Fluorescence with limits of determination better than 0.5%. Trace elements including Ba, Be, Co, Cr, Cu, Ni, Sc, Sr, V, Zn were analysed by inductively coupled plasma (ICP) emission spectrometry with limits of determination better than ± 20 ppm for Ba, Rb and Sr and better than ± 10 ppm for the other elements listed. All other trace elements, including the rare earth ele-

ments were analysed by ICP mass spectrometry for which determination limits better than 0.1 ppm are obtained for all trace elements except for Mo and Ta (0.2 ppm), Zr (0.5 ppm) and Pb (2 ppm). Major and trace element geochemical data for gabbroic rocks are presented in Table 4-1 and described below.

Geochemistry of Gabbroic Rocks within the Kramanituur Complex

Major element data for gabbroic to anorthositic rocks of the Kramanituur Complex are plotted on AFM and Jensen ternary diagrams in Fig. 4-3. These diagrams are intended for samples representative of liquid compositions, such as the gabbro samples which are mainly from <15 m wide dikes/sills, two of which are interpreted to cut country rocks exposed south of the complex (*K-1029* and *K-709*; Fig. 4-2). Other samples are included on these plots in order to portray the range of rock compositions and highlight the geochemical continuity shown by the samples. The gabbros plot in the high-Fe tholeiite basalt field, with two samples between the tholeiitic andesite and calc-alkalic basalt fields. With decreasing colour index, samples plot progressively closer to the alkali apex on the AFM diagram and toward the Al_2O_3 apex on the Jensen diagram. Mg numbers for the suite range from 37 to 70, except for the most ferroan sample (*K-113a*) which has a Mg number of 19.4 (Table 4-1). Cr and Ni contents for gabbro and gabbroic anorthosite are moderate to low, ranging between 13 and 260 ppm, and 14 and 220 ppm, respectively; both elements are generally below detection limits in anorthosite. Rare earth element (REE) data for rocks of gabbroic composition (Fig. 4-4a) show two main patterns: 1) flat REE patterns at 10 times (*K-679b*) to 30 times (*K-113a*) chondrite values without substantial enrichment or depletion in Eu ($\text{Eu}/\text{Eu}^* = 0.86\text{--}1.08$)¹; or 2) gently sloped patterns ($\text{La}_N/\text{Yb}_N = 2.2\text{--}5.2$) with small negative europium anomalies ($\text{Eu}/\text{Eu}^* = 0.78\text{--}0.87$). Gabbroic anorthosite samples typically have moderately sloped patterns

¹ $\text{Eu}/\text{Eu}^* = \text{Eu}_N / \sqrt{[(\text{Sm}_N)(\text{Gd}_N)]}$ after Taylor and McLennan (1985)

($\text{La}_\text{N}/\text{Yb}_\text{N}=5.1-13$), with LREE from 10 to 45 times chondrite values and HREE from 0.9 to 6 times chondrite values (Fig. 4-4b). Strong positive Eu anomalies ($\text{Eu}/\text{Eu}^*=1.5-5.0$) characterize the REE pattern of gabbroic anorthosite and are consistent with variable cumulate plagioclase in these rocks. One gabbroic anorthosite sample (K-1007a) shows less LREE enrichment ($\text{La}_\text{N}/\text{Yb}_\text{N}=2.5$) with LREE at 4.5 times and HREE at 1.8 times chondrite values. Anorthosite is characterized by LREE enrichment ($\text{La}_\text{N}/\text{Yb}_\text{N}>17.8$) and strong positive Eu anomalies ($\text{Eu}/\text{Eu}^*=2.8-6.8$; Fig. 4-4c).

Incompatible trace element data for the Kramanituar gabbros, those samples most closely approximating liquid compositions, are shown on a multi-element diagram (Fig. 4-5), in order of decreasing incompatibility from left to right. A gentle concave down pattern, interrupted only by Th, Sr $\pm$ Rb troughs characterizes these gabbroic rocks.

Geochemistry of Gabbroic Country Rocks, North and East of the Kramanituar Complex

Country rocks of gabbroic composition, though not demonstrated to represent a suite of petrogenetically coherent rocks, are described here as a group for comparison with those of the Kramanituar Complex. Their major element compositions plot near the tholeiitic/calc-alkalic boundary (open squares in Fig. 4-3a) and lie within the calc-alkalic basalt and komatiitic basalt (K-273a) fields (Fig. 4-3b). This suite has higher Mg numbers, from 58 to 80, Cr contents of 50 to 520 ppm and Ni contents of 75 to 340 ppm, relative to the Kramanituar gabbro samples. REE data show moderately sloped patterns ($\text{La}_\text{N}/\text{Yb}_\text{N}=4.5-46$), with LREE from 8 to 70 times chondrite values and HREE from 1.5 to 3 times chondrite values (Fig. 4-6a). They show either minor enrichment or depletion in Eu ($\text{Eu}/\text{Eu}^*=0.97-1.27$). Normalized incompatible element abundance patterns for gabbroic country rocks show decreasing abundance of less incompatible elements, resulting in a negatively sloped pattern which is punctuated by strong spikes. Negative spikes occur at Th, Nb and Ti, with positive spikes at Rb, K and a lesser positive peak at La-Sr (Fig. 4-6b).

GEOCHEMICAL MODELING OF MAGMATIC PROCESSES

The trace element geochemistry of gabbroic rocks from the study area show several features of petrogenetic interest. Flat REE patterns for gabbroic rocks from the Kramanituur Complex suggest that several of the analysed samples may represent bulk liquid compositions from which cumulate rocks of gabbroic anorthosite composition have been derived. Petrochemical modeling, presented below, investigates the potential for a cogenetic relationship between these suites.

Parameters for Geochemical Modeling

Gabbro compositions characterized by flat REE patterns from 10 to 25 times chondrite values (Fig. 4-4a, Table 4-1) are permissive of bulk liquid compositions for the Kramanituur gabbroic suite. Of these, samples *K-679b* and *K-709* are most appropriate, with 50 to 52 wt% SiO₂, 4.26 to 5.26 wt% MgO, Mg numbers of 42, 27-150 ppm Cr, and 49 ppm Ni, values that are typical of tholeiitic basalts that have fractionated some olivine (Basaltic Volcanism Study Project, 1981). An estimate of the bulk liquid composition for the Kramanituur Complex gabbroic suite (Table 4-2) is therefore taken as an average of samples *K-679b* and *K-709* and is high-Fe tholeiitic basalt with a flat REE pattern ($\text{La}_N/\text{Yb}_N=1.19$, $\text{Eu}/\text{Eu}^*=1.04$) at 14 times chondrite values.

Mineral/melt partition coefficients used for AFC modeling are listed in Table 4-3. The partition coefficients for plagioclase are from the Proterozoic Kiglapait Intrusion of Labrador (Morse, 1988), a layered complex of similar size, modal lithology and An content to the Kramanituur Complex. Kiglapait partition coefficients for plagioclase are similar to those for other granulite facies anorthosite complexes (Phinney and Morrison, 1990), and are an order of magnitude higher than those of McKay (1982) which are appropriate for lower pressure conditions. The modal mineralogy used to determine bulk partition coefficients representative of a gabbroic anorthosite fractionate is: 74% plagioclase, 8% orthopyroxene, 12% clinopyroxene, 3% amphibole, 2% ilmenite and 1% mag-

netite. This is based on petrographic observation of gabbroic anorthosite samples, excluding garnet as a fractionating phase. In a separate series of AFC models, 1% garnet, which strongly sequesters the middle and heavy REE, was included. It was determined that even this low proportion of garnet was inappropriate for most samples with the possible exception of *K-1007a* (discussed below).

Fractionational Crystallization Model

Fractional crystallization describes the crystallization of solid phases from a melt. It is best described by the Rayleigh Law, whereby crystals are removed from their site of formation after crystallization such that the distribution of trace elements is not an equilibrium process. Rayleigh fractionation is modeled according to:

$$C_L/C_O = F^{(D-1)}$$

where C_L is the weight concentration of a trace element in the fractionating liquid, C_O is the weight concentration of a trace element in the parental liquid, F is the fraction of melt remaining, and D is the bulk partition coefficient (based on mode) for the fractionating assemblage (Rollinson, 1993). A simple Rayleigh fractionation model is shown in Fig. 4-7a for cumulates derived from a parent composition represented by the estimated bulk composition of the Kramanituur gabbro suite. The model REE profiles for simple Rayleigh fractionation show a poor fit to the observed gabbroic anorthosite compositions due to the shallower negative slope to the model REE profiles and the restricted compositional range they represent for 10% to 50% fractionation (shaded; Fig. 4-7a).

A separate model investigated derivation of the gabbroic anorthosite suite via Rayleigh fractionation from a more enriched bulk liquid composition (i.e., modeled after sample *K-714c*; Fig. 4-4a). Since even the initial fractionate in equilibrium with such a liquid (Fig. 4-7b) is more enriched in the LREE than any of the observed cumulate rocks (shaded), it is unlikely that the gabbroic anorthosites were derived from a parental melt of this composition.

Assimilation-Fractional Crystallization Model

Assimilation of wallrock during fractional crystallization is a common process driven by the enthalpy of the magma and the latent heat of crystallization. The role of simultaneous assimilation and fractional crystallization (AFC) in generating cumulate rocks was investigated by using a crustal contaminant (Table 4-2) which is an average of sillimanite- (*K-86b*) and kyanite-bearing (*K-289a*) leucosome from the Kramanituur Complex. These are interpreted as partial melt phases derived from paragneiss and are presently in contact with, and occur as partially digested xenoliths within, gabbroic anorthosite. Assimilation is modeled on a 1:1 ratio with fractionation; that is, removal of 10% fractionate is accompanied by assimilation of 10% metasedimentary leucosome. This degree of assimilation is geologically reasonable in an AFC process where derivative mafic magmas can contain up to 50% to 80% contaminant (Longhi et al., 1983; Sparks, 1986). Assimilation of crustal material is enhanced for: 1) more primitive and hotter mafic magmas; 2) where crustal rocks have low fusion temperatures; and, 3) where crustal rocks are already at temperatures close to fusion upon intrusion of magma (Sparks, 1986). The second and third factors pertain to the Kramanituur Complex, where quartzo-feldspathic metasedimentary rocks were at high-grade conditions just prior to intrusion of the mafic magma (Sanborn-Barrie et al., 1999). Furthermore, contamination of tholeiitic magma within the Kramanituur Complex by siliceous material is consistent with the presence of orthopyroxene as a significant cumulus phase in many of these rocks (Stolper, 1980; Ashwal 1993). Results of a simplified AFC process are shown in Fig. 4-7c. The model REE profiles for 10-50% AFC (shaded; Fig. 4-7c) are parallel to the observed profiles for gabbroic anorthositic rocks from the complex, however, the highly restricted range in REE abundances that can be derived by up to 50% fractionation does not match the range in REE abundances observed for the gabbroic anorthosite suite.

Modified Assimilation-Fractional Crystallization Model

The model that best fits the observed REE patterns for gabbroic anorthosite cumulates is one of AFC with increasing degrees of intercumulus trapped liquid (Fig. 4-8). The proportion of intercumulus liquid was increased incrementally during fractionation, from 1% intercumulus liquid at 10% fractionation, to 25% liquid at 50% fractionation. This is consistent with adcumulus and intercumulus textures in anorthositic rocks worldwide (Ashwal, 1993). Furthermore, the cumulates with the lowest REE abundances are also the most primitive (Fig. 4-4b), consistent with the observation that adcumulate textures (<5% intercumulus liquid) are more common in hotter, more primitive cumulates (i.e., Stillwater Complex, Ashwal 1993). Four gabbroic anorthosite samples correspond closely to model REE patterns for 10% to 50% AFC (Fig. 4-8a). The flat HREE pattern which characterizes these samples cannot be modeled if even 1% garnet is included in the fractionate mode. That garnet was not fractionating from the magma implies magmatic crystallization at depths corresponding to $P \leq 15$ kbar (Thompson, 1972). One gabbroic anorthosite sample (K-676b) has LREE which are modeled for 10% AFC but has much lower HREE than predicted by this model. This sample has a low colour index (no normative diopside and 4.18 wt% normative hypersthene) and a mode of 85% plagioclase, 8% orthopyroxene, 3% clinopyroxene, 2% pargasitic amphibole, 1% magnetite and 1% ilmenite and trace zircon. Modified AFC modeling using a bulk partition coefficient based on this mode results in close agreement between sample K-676b and a model REE pattern for 5% fractionate (Fig. 4-8b).

Only one gabbroic anorthosite REE pattern could not be modeled. Sample K-1007a, with a distinctive REE pattern showing much less LREE enrichment than other gabbroic anorthosite samples ($\text{La}_N/\text{Yb}_N=2.5$) has a high colour index (4.31 wt% normative diopside and 32.52 wt% normative hypersthene). A bulk partition coefficient was calculated for the observed mode of 58% plagioclase, 30% orthopyroxene, 7% clinopyroxene, 2% amphibole, 1% magnetite, 1% ilmenite and 1% garnet. Model REE patterns

generated from this are *parallel* to the distinctive REE pattern of K-1007a (Fig. 4-8c), however, even the initial fractionate, with no intercumulus liquid and no assimilated country rock, produces a pattern which is higher by 1.5 times chondrite values. This sample is the only gabbroic anorthosite with HREE that show a marked concave up pattern, a feature generally attributed to garnet $\pm$ hornblende fractionation.

COMPARISON BETWEEN THE KRAMANITUAR GABBROIC SUITE AND GABBROIC COUNTRY ROCKS

Foliated gabbroic country rocks north and east of the Kramanituar Complex are characterized by an incompatible trace element pattern which is distinct from that of the Kramanituar gabbroic suite (Fig. 4-5b). The two most distinctive features of the country rock gabbro samples are the moderate, negative slope of the patterns and depletion of high field strength elements (HFSE: Th, Nb, Ti), neither of which characterize the Kramanituar gabbroic suite. Negatively sloped multi-element patterns are diagnostic of rocks derived through partial melting of a hydrated crust, or of rocks derived from a source which itself was a product of partial melting, since products of partial melting will be enriched in the most incompatible elements and depleted in the least incompatible elements (Wilson, 1989). Depletion of HFSE, especially Nb, indicates decoupling between the HFSE and the LILE. This has been attributed to the presence of a Nb compatible residual Ti-rich phase (e.g. rutile, sphene, ilmenite) in a source region (e.g., convergent plate margin) undergoing partial melting (Nicholls and Ringwood, 1973; Saunders et al., 1980; Smedley, 1986; Green and Pearson, 1987). Alternatively, it may reflect relative enrichment in the LILE, as observed in arc settings. The important point here is that incompatible trace element geochemistry reveals significant differences between gabbroic country rocks and the Kramanituar gabbroic suite. The data are consistent with derivation of gabbroic country rocks by partial melt processes or from metasomatized

mantle lithosphere, where trace elements are decoupled, in contrast to the Kramanituar gabbroic suite, whose geochemistry is consistent with a more pristine mantle source.

DISCUSSION AND CONCLUSIONS

A Cogenetic Suite

Modeling of trace element data for mafic plutonic rocks of the Kramanituar Complex, presented here, is consistent with a cogenetic relationship between gabbroic anorthosite and gabbroic rocks within the complex. Limited isotopic data for these rocks further supports this finding. Granulite facies gabbro (K-1053) from the complex yields an $\epsilon_{\text{Nd}}(1902 \text{ Ma})$ value of +2.8, in accordance with (see further discussion below) extraction from the mantle at 1.9 Ga (DePaulo, 1981a), the U-Pb (zircon) magmatic crystallization age of gabbroic anorthosite (K-676b; see Chapter 3).

Deep-crustal Conditions

Within the Kramanituar Complex, the association of a high- P (12 ± 1.5 kbar) metagabbroic suite (Sanborn-Barrie and Berman, 1999) intrusive into metasedimentary migmatites indicates that the complex was metamorphosed at deep crustal conditions at ca. 1.9 Ga. The cogenetic relationship demonstrated for rocks of gabbroic, gabbroic anorthositic and anorthositic composition indicate that *fractionation* of mantle melts took place at deep-crustal conditions at ca. 1902 Ma, the U-Pb (zircon) age of gabbroic anorthosite. Indications of garnet fractionation in sample K-1007a, as reflected by its HREE profile (Fig. 4-7c) support crystallization of mantle-derived melts in the deep crust at conditions of 12-15 kbar, as indicated by geothermobarometric data presented in Chapter 2. The low $f\text{O}_2$ for the Kramanituar Complex gabbroic suite, reflected by a mineral/melt partition coefficient of close to 1 for Eu in plagioclase as required for the AFC modeling, is more typical of magmatism in the lower crust than in the upper crust (Drake and Weill, 1975). Drake and Weill (1975) have shown that high (~ 1) partition coefficients for plagioclase imply low $f\text{O}_2$ conditions, because a strong correlation exists between oxygen

fugacity (fO_2) and the partition coefficient for Eu in plagioclase: europium forms Eu^{2+} at low oxygen activities and Eu^{3+} at high oxygen activities. Eu^{2+} and Eu^{3+} behave very differently in their partitioning between plagioclase and basaltic melts, with Eu^{2+} much more compatible than Eu^{3+} in plagioclase. In the case of the Kramanituvar Complex gabbroic suite, low fO_2 conditions are consistent with lower crust/mantle conditions, not more oxidized conditions characteristic of the upper crust.

Mantle Source

The trace element data allow preliminary speculation on the composition of the mantle source for the Kramanituvar gabbroic suite, and on aspects of the tectonic setting of this region. The bulk liquid composition of the Kramanituvar gabbroic suite shows moderate enrichment in incompatible trace elements (Fig. 4-5), comparable to that of enriched mid-ocean ridge basalt (E-MORB). The degree of enrichment shown is comparable to basaltic compositions variably referred to as *plume*-type MORB (Sun, 1980), E-type MORB (Wilson, 1989; McKenzie and O'Nions, 1991); or T- (transitional) type MORB (Humphris et al., 1985). These compositions are all enriched relative to N- (normal) type MORB, but are depleted relative to an enriched MORB end-member. The enriched nature of the Kramanituvar bulk liquid is displayed by the enrichment in LREE for the Kramanituvar gabbros relative to N-type MORB (Fig. 4-4a), by the gentle concave down pattern of the incompatible element abundances for the gabbros (Fig. 4-5) which reveals no significant depletion in these elements, and by the virtually flat pattern at 1x produced when the Kramanituvar Complex bulk liquid composition is normalized to E-type MORB (values of Sun and McDonough, 1989). Enrichment is *not* attributed simply to crustal contamination, because attributes of continental crust such as high silica, a pronounced negative Nb anomaly and high La/Nb ratios of 1.5 to 7.0 (Thompson et al., 1984; Campbell and Griffiths, 1993), do not characterize the Kramanituvar gabbroic suite (Table 4-1). As previously discussed, crustal contamination has been important in derivation of gab-

broic anorthosite *from* this bulk liquid. An ϵ_{Nd} value of +2.8 for these gabbroic rocks, slightly lower than the predicted value of +4 for depleted mantle at ca. 1900 Ma (DePaolo, 1981a), is consistent with an enriched (E-type) MORB source for the Kramanituur Complex gabbroic suite (Campbell and Griffiths, 1992). The modified AFC model presented here predicts that more fractionated gabbroic rocks from the Kramanituur Complex would yield a range of significantly lower ϵ_{Nd} values, reflecting the assimilation of crustal rocks of probable Archean age.

Whereas normal or N-type MORB is derived by partial melting of an isotopically homogeneous, depleted mantle reservoir, E-type MORB is thought to be derived by partial melting from an undepleted, variably-enriched reservoir which is isotopically heterogeneous (Allegre et al., 1984; Zindler et al., 1984; Wilson, 1989). Parts of this reservoir may reflect recycled subduction-altered oceanic crust ($\pm$ pelagic and terrigenous sediments) and are generally designated the ocean island basalt (OIB) reservoir (Zindler and Hart, 1988). Other parts are believed to be primordial (undepleted) in composition. It is partial melting of near primordial portions of the mantle that is thought by Sun (1980), Wilson (1989) and Campbell and Griffith (1992) to give rise to E-type MORB geochemical signatures.

The production of tholeiitic magma with an E-type MORB signature is attributed by Campbell and Griffiths (1992) to thermal plumes because they believe that thermal anomalies at the core-mantle boundary are required to cause partial melting of a lower, primordial part of the mantle. This relationship is demonstrated in Iceland where plume-related basalts derived from deeper levels in the mantle have an E-type MORB signature in contrast to more shallowly derived normal or N-type MORB (McKenzie and O'Nions, 1991). Initiation of a thermal plume is considered by some workers to arise as a result of the insulating properties of a stable craton (Anderson, 1994; Hoffman, 1989). Stability of the Churchill craton for 500 m.y., from ca. 2.5 to 2.0 Ga, may have contributed to elevated mantle temperatures.

An undepleted E-type MORB source for the Kramanituur gabbroic suite at 1.9 Ga is similar to that proposed for mafic rocks of similar age presently exposed in this part of Laurentia. These include enriched, plume-related mafic volcanics of the ca. 1.96 Ga Povungnituk Group, Cape Smith belt (Campbell and Griffiths, 1992; Lucas et al., 1992), and ca. 1904 Ma basalts with E-type MORB geochemistry which form the oldest Paleoproterozoic units in the Trans-Hudson Orogen (Stern et al., 1995). Although it is premature to speculate on the presence of a plume in the generation of high-grade mafic-dominated complexes of Paleoproterozoic age along the Chesterfield segment of the Snowbird tectonic zone (Fig. 4-1), migration of a plume-related hotspot could explain the apparent diachroneity of events along this 250 km long segment, where granulite facies metamorphism and mafic magmatism are bracketed between ca. 2067-1950 Ma at the Daly Bay Complex (Gordon, 1988) in the east, ca. 1920-1940 Ma at the Uvauk complex (Tella et al., 1994), and ca. 1910-1902 Ma at the Kramanituur Complex (Sanborn-Barrie et al., 1999) in the west.

In considering the tectonic setting of the western Churchill Province during Paleoproterozoic time, it is notable that the small degree of enrichment of the Kramanituur Complex gabbros does not approach that of continental tholeiites (high K, Sr, Ba; low Nb) or alkali basalts (i.e., $Nb/Y > 1.4$; high K_2O ; higher P_2O_5). These rocks are derived by partial melting of more enriched mantle, and showing significant degrees of crustal contamination (Wilson, 1989; Anderson, 1996). Continental tholeiites and alkali basalts are generally associated with the initial stages of lithospheric extension or rifting. Accordingly, trace element geochemistry for the Kramanituur gabbroic suite does not appear to reflect the onset of intracontinental rifting and alkalic magmatism of the western Churchill Province which is well-expressed by rocks dated at ca. 1.83-1.75 Ga (Peterson, 1994).

Detailed investigation of the petrogenesis of gabbroic country rocks surrounding the Kramanituur Complex is not addressed in this study, however, the geochemical sig-

nature of these rocks (Fig. 4-6) provides preliminary evidence of a supra-subduction environment during their evolution, presumably during Neoarchean time. Their geochemical signature may indirectly link them to Neoarchean subduction-related calc-alkaline volcanic rocks and associated tonalite-granodiorite that dominate this part of the western Churchill Province.

Table 4-1. Geochemical data for representative gabbroic samples from the Kramanituur Complex and surrounding country rocks.

KRAMANITUUR GABBROIC SUITE																
Anorthosite			Gabbroic Anorthosite			Gabbro										
K-143c	K-714b	K-11a	K-125a	K-605a	K-619c	K-676b	K-1007a	K-113a	K-620b	K-671b	K-679b	K-697b	K-709	K-714c	K-1029a	
SiO ₂	51.60	53.80	51.90	50.60	53.30	51.60	54.60	53.00	48.50	50.70	47.60	50.40	49.50	51.90	50.20	50.50
TiO ₂	0.21	0.08	0.30	1.27	0.31	2.18	0.14	0.18	3.14	1.38	1.05	1.04	1.43	1.25	2.50	1.29
Al ₂ O ₃	28.60	28.30	24.60	22.50	24.80	21.60	26.50	17.30	13.10	14.10	15.60	18.40	14.60	12.10	13.10	17.60
Fe ₂ O ₃	0.00	bd1	1.60	2.90	1.20	1.70	0.40	1.60	7.30	3.40	5.90	2.60	3.10	4.60	5.40	3.20
FeO	0.80	0.60	3.10	4.90	2.80	5.20	1.30	6.80	13.20	9.60	6.40	9.20	10.10	10.20	10.10	6.80
MnO	0.01	0.01	0.06	0.09	0.05	0.09	0.02	0.14	0.24	0.20	0.20	0.18	0.23	0.33	0.24	0.13
MgO	0.41	0.34	2.03	2.25	1.80	3.38	0.93	9.41	2.40	7.24	6.41	4.26	6.93	5.26	4.42	5.27
CaO	11.00	11.50	10.70	10.60	10.10	9.21	10.50	8.40	8.52	10.70	11.50	11.00	10.80	10.70	9.00	9.60
Na ₂ O	3.60	4.10	4.00	3.70	4.40	3.60	4.40	2.50	3.10	2.30	1.60	2.80	2.20	1.20	1.90	3.50
K ₂ O	1.32	0.39	0.30	0.37	0.42	0.83	0.51	0.27	0.19	0.23	1.19	0.13	0.55	0.36	1.09	0.32
H ₂ O	1.00	0.50	0.50	0.50	0.90	0.50	0.60	0.50	0.70	0.60	1.80	0.40	0.80	1.30	1.80	1.10
CO ₂	0.70	0.60	0.20	0.40	0.20	0.20	0.30	0.20	bd1	bd1	1.20	0.20	0.40	bd1	0.20	0.50
P ₂ O ₅	0.01	bd1	bd1	0.07	0.04	0.11	0.02	bd1	0.15	0.11	0.08	0.08	0.10	0.10	0.31	0.39
TOTAL	99.40	100.40	99.30	100.30	100.40	100.40	100.30	100.50	100.80	100.90	100.80	100.80	101.00	100.70	100.40	100.60
Mg'	50.40	52.90	46.99	37.26	47.91	49.89	52.62	69.36	19.40	53.13	52.04	42.26	51.59	42.10	36.94	51.91
Rb	14.00	7.00	0.65	0.93	2.90	6.10	3.30	2.10	1.10	1.10	24.00	0.78	10.00	8.90	40.00	1.30
Sr	340.00	430.00	410.00	360.00	390.00	390.00	440.00	260.00	150.00	270.00	160.00	130.00	130.00	64.00	150.00	430.00
Cs	0.92	0.46	0.07	0.09	0.11	0.13	0.07	0.05	0.06	0.08	0.73	0.03	0.20	0.67	2.40	1.60
Ba	420.00	150.00	130.00	140.00	150.00	400.00	200.00	110.00	80.00	130.00	190.00	50.00	90.00	90.00	210.00	260.00
Y	3.00	1.00	5.70	14.00	6.70	9.60	2.20	2.90	48.00	20.00	26.00	24.00	28.00	29.00	42.00	23.00
Zr	17.00	14.00	32.00	86.00	34.00	80.00	20.00	17.00	bd1	97.00	81.00	67.00	91.00	bd1	bd1	bd1
Hf	0.41	0.30	0.71	2.10	0.81	1.70	0.45	0.40	4.30	2.40	2.00	1.70	2.30	2.70	5.00	3.80
Nb	2.30	1.20	0.59	5.10	2.20	7.60	0.54	0.21	8.60	7.00	3.10	3.60	5.80	5.50	33.00	8.40
Ta	0.50	0.50	bd1	0.50	0.20	0.50	bd1	0.20	0.80	0.40	0.20	0.30	0.50	0.70	2.20	0.50
Th	1.40	0.13	0.43	2.60	1.30	0.59	0.23	0.18	0.42	0.10	0.28	0.33	1.20	0.64	0.87	0.28
U	0.49	0.07	0.08	0.52	0.41	0.13	0.06	0.07	0.11	0.04	0.08	0.09	0.24	0.43	0.71	0.08
La	13.00	2.40	4.20	14.00	6.80	8.80	3.90	1.40	8.00	10.00	4.10	3.80	8.50	5.30	29.00	33.00
Ce	28.00	4.40	7.90	32.00	14.00	18.00	7.00	2.50	22.00	25.00	12.00	10.00	20.00	14.00	68.00	72.00
Pr	3.10	0.46	0.83	3.60	1.60	2.10	0.74	0.29	3.10	3.30	1.80	1.60	2.80	2.10	8.60	8.90
Nd	13.00	1.80	3.50	15.00	6.40	9.40	3.00	1.20	17.00	16.00	9.40	8.00	13.00	11.00	38.00	37.00
Sm	2.20	0.31	0.75	3.00	1.20	2.00	0.54	0.24	5.10	3.80	2.80	2.50	3.70	3.20	8.30	6.70
Eu	1.50	0.58	0.64	1.40	0.99	1.30	0.81	0.44	2.10	1.20	0.89	0.92	1.10	1.30	2.30	1.60
Gd	1.20	0.22	0.83	2.70	1.20	1.90	0.46	0.34	7.00	4.00	3.60	3.30	4.40	4.30	8.00	5.90
Tb	0.13	0.03	0.13	0.39	0.18	0.27	0.06	0.06	1.20	0.56	0.59	0.57	0.72	0.72	1.20	0.77
Dy	0.59	0.15	0.83	2.30	1.00	1.50	0.35	0.42	7.50	3.40	3.90	3.50	4.40	4.50	7.00	4.20
Ho	0.10	0.03	0.17	0.46	0.21	0.30	0.07	0.09	1.60	0.65	0.82	0.79	0.92	0.98	1.40	0.82
Er	0.24	0.08	0.50	1.20	0.58	0.85	0.18	0.32	4.40	1.70	2.30	2.30	2.60	2.80	4.00	2.20
Tm	0.03	bd1	0.09	0.20	0.10	0.13	0.03	0.06	0.75	0.29	0.41	0.39	0.44	0.48	0.65	0.34
Yb	0.20	0.09	0.55	1.20	0.56	0.84	0.20	0.38	4.50	1.60	2.40	2.30	2.60	2.80	3.70	1.90
Lu	0.03	bd1	0.08	0.18	0.09	0.12	0.03	0.06	0.64	0.23	0.35	0.36	0.40	0.43	0.53	0.26
Co	15.00	11.00	34.00	39.00	21.00	29.00	11.00	52.00	66.00	49.00	51.00	39.00	46.00	45.00	39.00	34.00
Ni	bd1	bd1	31.00	34.00	35.00	88.00	19.00	220.00	14.00	57.00	130.00	49.00	100.00	49.00	15.00	60.00
Cr	bd1	10.00	19.00	16.00	23.00	150.00	13.00	210.00	bd1	74.00	260.00	27.00	160.00	150.00	39.00	99.00
Cu	bd1	bd1	17.00	49.00	31.00	17.00	bd1	91.00	97.00	79.00	49.00	110.00	68.00	120.00	79.00	63.00
Zn	bd1	bd1	16.00	37.00	20.00	46.00	bd1	45.00	140.00	72.00	99.00	71.00	93.00	210.00	120.00	74.00
Pb	8.00	3.00	3.00	2.00	6.00	5.00	3.00	bd1	2.00	2.00	2.00	6.00	2.00	bd1	3.00	3.00
Mo	0.30	1.10	0.30	0.70	2.60	2.00	0.20	bd1	2.20	0.30	0.50	1.10	1.10	1.80	1.30	0.40
V	bd1	bd1	53.00	120.00	25.00	110.00	bd1	99.00	450.00	310.00	310.00	270.00	290.00	310.00	370.00	220.00
Sc	1.80	0.80	11.00	16.00	8.30	14.00	2.90	20.00	42.00	36.00	43.00	31.00	36.00	39.00	40.00	25.00
Be	1.80	bd1	2.20	12.00	2.10	16.00	0.80	1.00	31.00	10.00	7.50	7.30	10.00	8.60	17.00	8.50
Ga	31.00	22.00	24.00	24.00	23.00	19.00	24.00	14.00	26.00	20.00	18.00	19.00	19.00	18.00	24.00	22.00
In	bd1	0.07	0.08	bd1	bd1	bd1	bd1	bd1	0.12	0.17	0.07	0.06	0.08	0.17	0.20	0.11
Tl	0.20	0.03	bd1	bd1	0.03	0.02	bd1	bd1	bd1	bd1	0.22	bd1	0.08	0.09	0.19	bd1

GABBROIC COUNTRY ROCKS

K-222 K-273a K-451b K-456a

SiO ₂	50.20	50.90	46.90	50.30
TiO ₂	0.28	0.27	0.17	0.40
Al ₂ O ₃	18.70	12.40	22.20	18.00
Fe ₂ O ₃	1.50	1.90	2.00	3.60
FeO	4.00	4.50	4.90	6.00
MnO	0.08	0.14	0.09	0.14
MgO	7.89	12.00	9.43	6.48
CaO	9.48	14.40	10.60	9.90
Na ₂ O	2.80	1.10	1.80	2.90
K ₂ O	0.83	0.47	0.56	0.35
H ₂ O	2.10	0.80	1.60	1.50
CO ₂	1.30	0.30	0.40	0.30
P ₂ O ₅	0.03	bd1	0.02	0.12
TOTAL	99.80	99.30	100.80	100.20
Mg'	74.51	79.30	73.62	58.16
Rb	6.40	21.00	9.40	5.60
Sr	1300.00	350.00	280.00	490.00
Cs	0.73	0.53	0.21	0.21
Ba	410.00	140.00	200.00	150.00
Y	4.30	8.00	3.70	6.30
Zr	100.00	61.00	24.00	25.00
Hf	2.10	1.50	0.55	0.71
Nb	0.93	0.42	0.58	0.48
Ta	0.20	bd1	0.20	0.20
Th	0.16	0.33	0.37	0.38
U	0.10	0.08	0.08	0.08
La	22.00	4.50	2.70	7.90
Ce	41.00	12.00	5.80	17.00
Pr	4.40	1.60	0.68	2.10
Nd	19.00	8.40	2.90	9.70
Sm	3.00	2.00	0.59	1.90
Eu	0.99	0.60	bd1	0.74
Gd	1.90	1.70	0.65	1.60
Tb	0.18	0.25	0.10	0.20
Dy	0.86	1.40	0.59	1.10
Ho	0.14	0.28	0.12	0.22
Er	0.35	0.70	0.36	0.53
Tm	0.06	0.11	0.06	0.09
Yb	0.35	0.66	0.33	0.48
Lu	0.06	0.09	0.05	0.08
Co	49.00	42.00	51.00	39.00
Ni	260.00	180.00	340.00	75.00
Cr	170.00	520.00	50.00	60.00
Cu	15.00	bd1	22.00	73.00
Zn	50.00	14.00	24.00	53.00
Pb	13.00	bd1	6.00	4.00
Mo	22.00	0.30	0.30	0.60
V	24.00	130.00	bd1	200.00
Sc	11.00	53.00	7.70	23.00
Be	2.60	2.50	1.50	3.60
Ga	16.00	12.00	14.00	19.00
In	bd1	0.14	0.10	0.11
Tl	0.04	0.19	0.08	0.03

Table 4-2. Geochemical data for derived Kramanituar bulk liquid and crustal contaminant compositions.

	Kramanituar Bulk Liquid	Paragneissic Leucosome		
		K-86b	K-289a	Contaminant
SiO ₂	51.15	64.20	70.70	67.45
TiO ₂	1.14	0.89	0.35	0.62
Al ₂ O ₃	15.25	18.40	16.30	17.35
Fe ₂ O ₃	3.60	4.90	1.60	3.25
FeO	9.70	3.60	2.00	2.80
MnO	0.25	0.16	0.04	0.10
MgO	4.76	2.85	1.85	2.35
CaO	10.85	0.91	1.02	0.96
Na ₂ O	2.00	2.10	1.80	1.95
K ₂ O	0.24	1.44	3.51	2.47
H ₂ O	0.85	1.20	0.40	0.80
CO ₂	bd1	bd1	0.40	bd1
P ₂ O ₅	0.09	0.03	0.03	0.03
TOTAL	100.75	100.90	100.10	100.50
Mg'	42			
Rb	4.84	30.00	76.00	53.00
Sr	97.00	170.00	140.00	155.00
Cs	0.35	0.40	0.10	0.25
Ba	70.00	370.00	590.00	480.00
Y	26.50	27.00	29.00	28.00
Zr	67.00	bd1	bd1	bd1
Hf	2.20	4.10	3.20	3.65
Nb	4.55	11.00	7.70	9.35
Ta	0.50	0.50	0.70	0.60
Th	0.48	5.60	0.58	3.09
U	0.26	0.69	0.19	0.44
La	4.55	38.00	29.00	33.50
Ce	12.00	79.00	53.00	68.00
Pr	1.85	7.90	4.80	6.35
Nd	9.50	31.00	17.00	24.00
Sm	2.85	5.40	3.20	4.30
Eu	1.11	1.30	1.20	1.25
Gd	3.80	4.90	3.70	4.30
Tb	0.64	0.74	0.65	0.69
Dy	4.00	4.40	4.30	4.35
Ho	0.88	0.86	0.90	0.88
Er	2.55	2.40	2.60	2.50
Tm	0.43	0.41	0.48	0.44
Yb	2.55	2.40	2.70	2.55
Lu	0.39	0.35	0.39	0.37
Co	42.00	43.00	40.00	41.50
Ni	49.00	29.00	bd1	9.50
Cr	88.50	120.00	92.00	108.00
Cu	115.00	34.00	bd1	12.00
Zn	140.50	82.00	47.00	64.50
Pb	2.00	31.00	28.00	29.50
Mo	1.45	1.80	1.80	1.80
V	290.00	140.00	52.00	96.00
Sc	35.00	23.00	13.00	18.00
Be	7.95	8.60	3.30	5.95
Ga	18.50	23.00	21.00	22.00
In	0.11	0.10	bd1	0.02
Tl	0.03	0.15	0.39	0.27

Table 4-3. Mineral/melt partition coefficients used for AFC modeling of gabbro anorthosite from liquid gabbro composition.

REE	PLAG ¹	OPX ²	CPX ³	AM ⁴	GRT ⁵	MT ⁶	ILM ⁷
La	0.65	0.0067	0.058	0.17	0.01	0.2	0.0064
Ce	0.49	0.0078	0.1	0.26	0.07	0.2	0.0065
Nd	[0.321]	0.011	0.22	0.44	0.26	0.25	0.0085
Sm	0.152	0.019	0.45	0.76	1	0.3	0.014
Eu	0.813	0.022	0.48	0.88	1.58	0.25	0.00655
Gd	0.09	0.033	0.5	0.86	2.16	0.25	0.0065
Tb	0.072	0.05	0.5	0.83	2.74	0.25	0.0062
Dy	[0.068]	0.061	0.5	0.78	3.32	0.25	0.0067
Er	[0.058]	0.094	0.61	0.68	5.95	0.25	0.0062
Yb	0.048	0.14	0.58	0.59	8	0.25	0.0067
Lu	0.047	0.18	0.53	0.51	10	0.25	0.0065

¹ meso-Proterozoic Kiglapait intrusion (Morse, 1988)

^{2,3,4} basalt at 10-20 kb (Irving and Frey, 1984)

⁵ basalt at 30 kb and 1240°C (Nicholls and Harris, 1980)

⁶ andesitic liquids (Gill, 1981)

⁷ mare basalt composition at 1 atm and 1140°C (McKay and Weill, 1976)

Figure 4-1. Geology of the western Canadian Shield after Hoffman (1988). Segments of the Snowbird tectonic zone (STZ), from northeast to southwest, are the Chesterfield segment which includes the Daly Bay complex (DBc), Uvauk complex (Uc) and Kramanituvar Complex (KC); the Tulemalu fault zone (Tfz); the Striding-Athabasca mylonite zone (SAmz); and the Virgin River shear zone (VRsz) and subsurface east Alberta (eA) segment.

Figure 4-2. General geology of the Kramanituvar Complex and its country rocks, modified from Schau and Ashton (1980) and Sanborn-Barrie (1999).

Figure 4-3. Ternary variation diagrams. Symbols correspond to compositions of representative samples from the Kramanituvar Complex including anorthosite-asterisk; gabbroic anorthosite-open circle; gabbro-closed circle. Also shown are gabbroic country rocks (open squares). *a*) AFM diagram after Irvine and Baragar (1971). *b*) Jensen cation plot (Jensen and Pyke, 1982).

Figure 4-4. REE abundances for mafic rocks from the Kramanituvar Complex normalized to chondrite values of Sun (1980). *a*) Gabbro. *b*) Gabbroic anorthosite with Mg number (*see* Table 4-1) for each representative sample shown. *c*) Anorthosite with Mg number.

Figure 4-5. Incompatible trace element plot for gabbros from the Kramanituvar Complex normalized to primitive mantle values of Hofmann (1988). Also shown is the pattern for N-type MORB, E-type MORB and P-type MORB for comparison (values of Sun and McDonough, 1989).

Figure 4-6. Trace element geochemistry of gabbroic country rocks. *a*) REE abundances normalized to chondrite values of Sun (1980). *b*) Spider diagram showing REE and other incompatible trace element abundances normalized to primitive mantle values of Hofmann (1988). Shaded region is the field for Kramanituvar gabbro samples, for comparison (discussed in text).

Figure 4-7. Results of Rayleigh fractionation and simplified AFC modeling for the Kramanituvar gabbroic suite. *a)* Shaded region represents field for cumulates derived by 10-50% Rayleigh fractionation from an estimated bulk liquid composition (shown). Actual abundances for gabbroic anorthosite samples are indicated for comparison. *b)* First fractionate derived via simple Rayleigh fractionation in equilibrium with an enriched bulk liquid composition (*K-714c*). Shaded region represents field for actual gabbroic anorthosite cumulates, for comparison. *c)* Shaded region represents field for cumulates derived by 10-50% simplified AFC from an estimated bulk liquid composition (shown) with a crustal contaminant (also shown). Actual abundances for gabbroic anorthosite samples are indicated for comparison.

Figure 4-8. Results of modified AFC modeling of REE for the Kramanituvar gabbroic suite. Shaded regions represent fields for cumulates derived through AFC $\pm$ trapped interstitial liquid from the estimated bulk liquid composition (shown) with a crustal contaminant (also shown). Actual abundances for Kramanituvar gabbroic anorthosite samples are indicated for comparison. *a*) For 10-50% AFC; *b*) For 5-10% AFC with a mode representative of gabbroic anorthosite sample *K-676b*, shown for comparison. *c*) For <10% AFC for a mode representative of gabbroic anorthosite sample *K-1007a* and including 1% garnet.

CHAPTER 5

Tectonic evolution of Paleoproterozoic deep-crustal rocks of the Kramanituar Complex, western Churchill Province

INTRODUCTION

This study of the Kramanituar Complex, located at the junction of the regional northeast- and east-trending segments of the Snowbird tectonic zone (Fig. 5-1), has documented the formation of high-*P* granulite-facies tectonites through processes that involved intrusion and crystallization of mantle-derived mafic melts (Chapter 4) in the deep crust during ca. 1910-1901 Ma regional metamorphism (Chapter 3); deformation under conditions of non-coaxial shortening strain (Chapter 1); near-isothermal exhumation from deep to mid-crustal levels (Chapter 2); and, subsequent quenching. The presence of these high-*P* rocks at surface before ca. 1832 Ma (MacRae et al., 1996), the minimum age of coarse clastic rocks that unconformably overlie the complex, implies 45-55 km of exhumation took place in the interior of the Archean western Churchill Province at ca. 1.9 Ga. The tectonic significance of the formation and exhumation of the Kramanituar Complex within the context of the western Churchill Province is addressed in this more speculative final chapter.

Geodynamic Setting of the Kramanituar Complex

The Kramanituar Complex is located in the interior of the western Churchill Province, a frontier region with relatively limited geological constraints. This allows only broad speculation regarding its geodynamic setting. The lack of a ca. 1.9 Ga volcanic or magmatic arc in the interior of the western Churchill Province argues against the influence of a subduction zone in proximity to the Chesterfield segment of the Snowbird tectonic zone at ca. 1910-1900 Ma, and against proximal supra-subduction dynamics (cf. Platt, 1986; Hsü, 1991; Hynes et al., 1996; Chemenda et al., 1995) as a driving force for exhumation of the Kramanituar Complex. The *P-T-t* paths of rocks from the Kramanituar

Complex, which either preserve peak high- P , T metamorphic assemblages, or record near-isothermal decompression and subsequent rapid cooling ($450^{\circ}\text{C Ma}^{-1}$; Chapter 3), are inconsistent with a setting dominated by isostasy-driven exhumation. Continental collisional activity related to the Trans-Hudson Orogen, historically attributed to tectonometamorphic reworking of the western Churchill Province (Davidson, 1972; Lewry and Sibbald, 1980; Lewry et al., 1985; Lucas et al., 1997), took place some 70 m.y. *after* events recorded at the Kramanituvar Complex and is not obviously linked to it. However, continental collisional activity along the western margin of the Churchill Province at ca. 1.97-1.90 Ga is synchronous with tectonometamorphism and exhumation of the Kramanituvar Complex, and may have had a role in crustal thickening, heat production and exhumation of this region.

Crustal Thickening

Estimated peak pressure conditions of 12 to 15 kbar, within the Kramanituvar Complex and within rocks to its north and east, require a crustal thickness of 45 to 55 km for this region of the western Churchill Province at ca. 1902 Ma. This is likely a minimum thickness since ultramafic rocks, representative of upper-mantle peridotite, are not recognized.

Available data do not provide unequivocal constraints on the timing of crustal thickening. Country rocks south of the complex record a prograde metamorphic history to conditions of 8 kbar (metapelite) and 12 kbar (gabbro) at ca. 2550-2326 Ma, suggesting that crustal thicknesses of >45-55 km at the Kramanituvar Complex may have been attained during Neoproterozoic (± 200 m.y.) time (Fig. 5-2a). Preservation of these metamorphic assemblages and fabrics, adjacent to ca. 1.9 Ga high- P granulites of the Kramanituvar Complex, suggests that the Archean country rocks were, at least partially, exhumed/cooled after ca. 2.5-2.3 Ga and may have subsequently remained at mid- to upper

crustal levels. The Neoarchean history of the Archean rocks into which the Kramanituvar Complex intruded is not known.

Prograde metamorphism, dated by monazite ages from granulite-facies paragneiss within the Kramanituvar Complex at 1910 Ma, and possibly as early as 1917 Ma, also may have been induced by crustal thickening of the upper plate hinterland to the Slave-Churchill collision (Fig. 5-2*b*); cf. Dewey and Burke, 1973; Thomas and Gibb, 1985; Hoffman, 1987). Crustal thickening (i.e., loading) during Paleoproterozoic metamorphism is indicated in paragneiss by the rimward increase in calcium during garnet growth. If anomalously thick crust in the interior of the western Churchill Province is related to the Slave-Churchill collision, it should be possible to identify structures that accommodated ca. 1.97-1.91 thickening between the Kramanituvar Complex and the Thelon front. Potential structures in the western part of the hinterland include several crustal-scale northeast-striking fault zones that transect the high-grade Queen Maud block and occur along its eastern margin (Heywood, 1961; Hoffman, 1989). Eastward, reverse structural vergence on these faults (Fig. 5-2*b*) is suggested by paired geophysical anomalies that straddle the fault zones. Positive gravity and magnetic anomalies, potentially correlative with exhumed high-grade rocks, occur west of the faults, in contrast to low amplitude anomalies to the east. A link between collisional activity along the northwest margin of the western Churchill Province and thickening events in its interior clearly lies in establishing the presence, timing and kinematic history of fault zones west of the Kramanituvar Complex.

Heat Production

At the Kramanituvar Complex, fractionation of mantle-derived mafic magmas in the deep crust took place at ca. 1902 Ma, the U-Pb zircon age of gabbroic anorthosite. In general, mafic magmatism signifies a thermal pulse, possibly initiated by delamination or convective removal of the subcrustal lithosphere (see discussion below). Mantle-derived

melts are commonly recognized as having been a contributing factor in elevating the geothermal gradient of the lower crust (Fountain, 1989; Mareschal and Bergantz, 1990; Henk et al., 1997). Undoubtedly they contributed heat to this region, where peak temperatures of ca. 850°C correspond to a simplified linear geotherm of about 20°C km⁻¹, nearly twice that implied for stable shield regions today (Thompson, 1992). The addition of magmatic heat at 1902 Ma to rocks already subjected to regional metamorphism is consistent with monazite ages of 1910-1908 Ma, and possibly as early as 1917 Ma, for high-grade paragneiss from the complex. These ages are interpreted to date prograde regional metamorphism (~550°C) of paragneiss in advance of intrusion of the 1902 Ma mafic plutonic complex, which itself is at granulite facies (Chapter 3). Regional metamorphism synchronous with intrusion of the Kramanituag gabbroic suite is supported by data from coronitic gabbro (K-691) located 7 km north of the complex. Here, *P-T* estimates of formation of garnet coronae, dated at 1878 ± 27 Ma (Sm-Nd, Grt-Pl), correspond to a depth-averaged geothermal gradient of ca. 23°C km⁻¹ (Chapter 2).

Exhumation

Rapid uplift and exhumation of thickened crust in this region is recorded by isothermal decompression of rocks from the Kramanituag Complex from peak conditions of 12.5 ± 1.5 kbar and 810 ± 70°C sometime after 1910-1908 Ma, followed by very high time-averaged cooling rates (≥ 450°C Ma⁻¹) to below 400°C by 1901-1898 Ma. Rapid cooling is reflected by the excellent preservation of high-grade metamorphic mineral assemblages and optical microstructures in these rocks. As previously discussed, such *P-T-t* paths are consistent with exhumation via extensional unroofing, as is juxtaposition of blocks with tectonometamorphic histories of different ages.

Extensional unroofing at ca. 1.9 Ga may reflect N-S tectonic escape which occurred in response to remote tectonic forces imposed on the western Churchill Province during ca. 1.97-1.90 Ga collision of the Slave craton (Fig. 5-2b; Hoffman, 1987). The

regional extent of the uplifted and exhumed block is known to exceed that of the Kramanituar Complex, because isolated occurrences of high- P rocks are located at least 7 km north of the complex. Its dimensions may correspond to the 300 km by 100 km positive (+60 mGal) Bouguer gravity anomaly that underlies much of the Chesterfield segment of the Snowbird tectonic zone (*see* Fig. 1-1b). The potential occurrence in this block of:

1) mafic magmatism; 2) the onset of extension; and related 3) regional uplift, is indicative of removal of the lower lithosphere a few m.y. prior to the onset of mantle-derived magmatism (Bird, 1979; England and Houseman, 1988; Meissner and Mooney, 1998). For instance, delamination of the subcrustal lithosphere (Bird, 1979; Bird and Baumgardner, 1981) may have driven diachronous events in the interior of the western Churchill Province at ca. 1950-1900 Ma (Fig. 5-2a), since delamination may have contributed significantly to uplift by triggering an increase in buoyancy of the crust. Uplift, in turn, may have facilitated extensional unroofing by increasing the potential energy of the uplifted block (Bird, 1979; England and Houseman, 1989; Dewey et al., 1993).

Intracontinental exhumation of the Kramanituar Complex some 30 to 50 Ma after the terminal stage of the Slave-Churchill collision is potentially similar to extensional exhumation of some metamorphic core complexes of the North American Cordillera (Coney, 1980; Parrish et al., 1988) and to the Himalayan Orogen, where convergence between India and Asia, resulted in very strong crustal deformation over a broad area some 2000-2500 km wide from the Himalaya to Tien Shan and Mongolia (Molnar and Tapponier, 1975). Continued convergence between India and Asia, some 50 Ma after the start of collision, is driven by continued underthrusting of the Indian craton beneath Asia. By analogy with the Himalayan Orogen, continued shortening and extension in the hinterland of the Thelon Orogen until ca. 1.9 Ga, some 50 Ma after collision started, may have been driven by underthrusting of the Slave craton below the western margin of the Churchill Province until ca. 1917-1910 Ma.

In contrast to the Himalayan and North American settings where extensional unroofing has resulted in broadly orogen-parallel uplift terrains, the Kramanituvar Complex and potentially correlative (though possibly diachronous) complexes along Chesterfield Inlet appear to comprise an orogen-normal block of exhumed rocks (Fig. 5-2b). One explanation for this is that uplift and exhumation in this part of the hinterland may have preferentially initiated and remained focussed along a favorably-oriented segment of an ancestral crustal-scale fault zone. The presence near the Kramanituvar Complex of ca. 2550 Ma dextral mylonites of the Southern shear zone, which are kinematically compatible with a ca. 2.6 Ga strand of the Tulemalu fault zone (Aspler et al., 1999) and ca. 2600 Ma mylonites of the Striding-Athabasca mylonite zone (Hanmer et al., 1994; Hanmer, 1997a) suggest the presence of a throughgoing crustal-scale fault zone, the Snowbird tectonic zone, during Neoarchean time (Fig. 5-2a). The Chesterfield segment of the Snowbird tectonic zone may have subsequently localized exhumation-related deformation during Paleoproterozoic convergence due to its favourable orientation, approximately parallel to the principal shortening direction during Thelon convergence.

Instability of the Archean western Churchill Province at 1.9 Ga

In contrast to many Archean cratons, which have been tectonically and magmatically quiescent since Neoarchean time (de Wit and Ashwal, 1997), the western Churchill Province has undergone significant tectonometamorphic activity in its interior. This activity is particularly well highlighted by the results of this study of the Kramanituvar Complex which document ca. 1.9 Ga mantle-derived magmatism, high-*P* granulite-facies mylonitization, exhumation of rapidly cooled high-grade rocks and unconformable coarse clastic deposition. Lithospheric stability is generally attributed to the presence of an ancient, chemically distinct (FeO-poor), high velocity lithospheric keel, or tectosphere, that stabilized to depths of ~200 km in the first few billion years of earth history (Jordan, 1975; Polet and Anderson, 1995). The hydration state of tectospheric roots to Archean

cratons is believed to be important in maintaining stability, since maintaining strength at high temperatures requires a very dry mantle (Hirth and Kohlsted, 1995; Polet and Anderson, 1995). Instability of the western Churchill Province at ca. 1.9 Ga may have evolved through inefficient devolatilization of part of the North American tectosphere, possibly indicated by an undepleted mantle source of the Kramanituar gabbroic suite (Chapter 4). Alternatively, instability at ca. 1.9 Ga may be an indication that part of the tectosphere did not remain isolated from recycled volatiles. Geodynamic factors leading to destabilization at ca. 1.9 Ga could have been a far-field response to the Slave-Churchill collision, and may have included introduction of subduction-related volatiles (cf. Peterson and Rainbird, 1990) or enhanced mantle temperatures associated with slab detachment.

Figure 5-1. Geology of the western Canadian Shield after Hoffman (1988). Segments of the proposed Snowbird tectonic zone (STZ) transect the western Churchill Province and include: the Virgin River shear zone (VRsz) and subsurface east Alberta segment (eA), the Striding-Athabasca mylonite zone (SAmz), the Tulemalu fault zone (Tfz), and the Chesterfield segment which includes the Kramanituar (KC), Uvauk (Uc) and Daly Bay (DBc) complexes. Other abbreviations are sz-shear zone, mz-magmatic zone.

Figure 5-2. Sketch illustrating possible tectonic setting of the western Churchill Province leading to formation of the Kramanituar Complex. *a*) Plan and cross-sectional view at ca. 2.6-2.5 Ga showing the Snowbird tectonic zone (STZ) as a throughgoing Neoproterozoic structure. Approximate direction of principal shortening based on local strain regimes interpreted for the Striding-Athabasca mylonite zone (SA), Angikuni Lake area (ANG), MacQuoid belt (MGB) and the Southern shear zone (SSZ). *b*) Plan and cross-sectional view at ca. 1.97-1.90 Ga with estimated principal shortening direction during the Slave-Churchill collision (after Hoffman, 1987). High-grade complexes within the Chesterfield segment of the ancestral Snowbird tectonic zone (STZ) include the Kramanituar Complex (KC; this study), Uvauk (Uc; Tella et al., 1994), Hanbury Island (H; Tella and Annesley, 1988) and Daly Bay (DBc; Gordon, 1988) complexes. Potential detachment structures that bound the exhumed Chesterfield block, as defined by the positive Bouguer gravity anomaly (see Fig. 1-1b) are designated *SDF*-Southern detachment fault, and *NDF*-Northern detachment fault.

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Structural investigation of high-grade rocks of the Kramanituar complex, Baker Lake area, Northwest Territories

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Abstract: The Kramanituar complex comprises granulite-grade gabbro, gabbroic anorthosite, and paragneiss and is hosted by amphibolite-grade plutonic and supracrustal wall rock. Its northern boundary coincides with a granulite-grade shear zone, 60 km long and up to 6 km wide, across which sinistral, south-side-up displacement is recorded. Its southern boundary appears to coincide with an amphibolite-grade shear zone, 30 km long and up to 2 km wide, across which dextral, north-side-up movement has occurred. Preliminary observations suggest that the complex represents a high-grade, crustal-scale tectonic lens, or boudin, which was rapidly exhumed during noncoaxial, southwest-directed shortening.

Résumé : Le complexe de Kramanituar est composé de gabbro, d'anorthosite gabbroïque et de paragneiss du faciès des granulites, encaissés dans des roches supracrustales et plutoniques du faciès des amphibolites. La limite nord du complexe coïncide avec une zone de cisaillement du faciès des granulites mesurant 60 km de longueur et jusqu'à 6 km de largeur, qui a été le siège d'un déplacement senestre avec soulèvement du compartiment sud. Sa limite sud semble coïncider avec une zone de cisaillement du faciès des amphibolites, mesurant 30 km de longueur et jusqu'à 2 km de largeur, dans laquelle s'est produit un déplacement dextre avec soulèvement du compartiment nord. Les premières observations indiquent que le complexe représente une lentille tectonique (ou boudin) d'échelle crustale fortement métamorphisée qui a été mise à nu rapidement durant un épisode de raccourcissement non coaxial à vergence sud-ouest.

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INTRODUCTION

The Kramanituar complex comprises granulite-grade gabbroic anorthosite and gabbroic rocks (Schau and Hulbert, 1977; Schau and Ashton, 1980; Schau et al., 1982; Sanborn-Barrie, 1993) and is hosted by amphibolite-grade plutonic and supracrustal wall rock. The complex coincides with a regional geophysical anomaly, designated the Snowbird tectonic zone (Hoffman, 1988), along which discontinuously exposed rocks of similar composition, structural style, and metamorphic grade also occur (Gordon, 1988; Hanmer et al., 1992; Hanmer and Kopf, 1993; Tella and Annesley, 1988; Tella et al., 1993). This study is part of a collaborative effort to unravel the tectono-metamorphic history of these complexes and, in turn, that part of the Churchill Province transected by the Snowbird tectonic zone (Fig. 1). Well-constrained structural, petrological, and geochronological studies of the individual complexes will reveal whether each possesses a distinct history; or alternately, whether their histories are similar and reflect regional-scale crustal processes.

The Kramanituar complex is an ideal setting in which to undertake magmatic, structural, and metamorphic studies. It preserves several generations of fabric elements. It displays strain of variable intensity which, where low, preserves lithological relationships and, where high, results in regional-scale shear zones up to 60 km long and 6 km wide. It is characterized by high-grade, generally unannealed metamorphic mineral assemblages. Two significant structural interpretations that evolved from 1992 mapping were:

1. that the northern boundary of the granulite-grade complex coincides with a steep, curvilinear, regional-scale shear zone across which sinistral, south-side-up displacement has occurred. This shear zone may have had a major role in juxtaposition of granulite-grade rocks with amphibolite-grade rocks to the north.
2. that the southern boundary of the complex coincides with an east-striking, moderately dipping shear zone across which dextral, north-side-up displacement may, similarly, have been responsible for juxtaposition of the granulites with amphibolite-grade rocks to the south.

A primary goal of 1993 fieldwork was to further delineate the geometry, kinematics, and metamorphic character of these two bounding shear zones. Their character and the temporal relationship between them are critical to understanding uplift mechanisms involved in exhuming this granulite-grade complex, and may pertain to others along the Snowbird tectonic zone, and elsewhere. Emphasis was also placed upon documenting metamorphic mineral assemblages across the granulite complex and into its amphibolite-grade wall rocks to establish the relationship between metamorphism and both magmatism and structure.

This report emphasizes structural aspects of the Kramanituar complex based on 1992 and 1993 field observations. A preliminary description of its metamorphic character is also provided. Ongoing studies will integrate the structural and metamorphic aspects of these rocks to gain a better

understanding of their evolution, their relationship with similar complexes, and their role in the evolution of the central Churchill Province.

LITHOLOGICAL UNITS WITHIN THE COMPLEX

Comprehensive descriptions of the principal lithological units of the map area are provided in Sanborn-Barrie (1993). In general, the Kramanituar complex comprises a gabbroic-anorthosite suite with minor interlayered paragneiss, orthogneiss, and granite-charnockite (Fig. 2). Where less deformed, gabbroic anorthosite is composed of phenocrysts

Figure 1. Geology of the western Canadian Shield after Hoffman (1988). Segments of the proposed Snowbird tectonic zone, at the interface between the Rae and Hearne provinces, include the East Athabasca mylonite zone (EAmz), the Tulemalu fault zone, the Kramanituar complex (Kc), the Uvauk complex (Uc) and the Daly Bay complex (DBc).

Figure 2. General geology of the Kramanituur complex and adjacent units. Modified from Schau and Ashton (1980).

of plagioclase±orthopyroxene in a groundmass of plagioclase, orthopyroxene, and clinopyroxene±magnetite. Metre-scale layers of gabbro and leucogabbro commonly occur with gabbroic anorthosite. These generally appear to represent intrusive dykes or sills owing to their uniform composition and abrupt contacts with gabbroic anorthosite. Magmatic layering, defined by systematic decrease in the percentage of mafic minerals across adjacent metre-scale layers is observed locally (Sanborn-Barrie, 1993). Anorthosite comprises <10% of the suite, occurring as continuous, tabular, 10 m wide horizons (sills?) in gabbro and paragneiss immediately north of gabbroic anorthosite. Gabbro and leucogabbro (±quartz diorite) are regionally extensive, occurring north and south of gabbroic anorthosite (Fig. 2). These consist of 25 to 60% mafic minerals, generally with clinopyroxene > garnet > orthopyroxene, with plagioclase and minor magnetite. These rocks typically display subtle metre-scale layering owing mainly to variations in colour index and are cut by parallel, centimetre-scale veinlets or sheets of granitic-charnockitic rock (described below).

Rocks interpreted as paragneiss occur immediately north of gabbroic anorthosite and along the north channel of Chesterfield Inlet (Fig. 2). These well-layered rocks contain 5-25% quartz, with biotite, plagioclase, and garnet±sillimanite±graphite. Two narrow panels of kyanite-bearing paragneiss occur in the northeast part of the complex, and rare sapphirine-bearing panels occur in the southeast (see Fig. 10).

A texturally distinctive unit designated intermediate tectonite occurs northwest and south of gabbroic anorthosite (Fig. 2). This pale buff- to rust-weathering unit, previously described as diatexite (Sanborn-Barrie, 1993), is compositionally similar to paragneiss in that these rocks contain 10-20% quartz, plagioclase, 2-10% garnet, and biotite±orthopyroxene. In contrast to paragneiss, these rocks do not contain sillimanite, kyanite, graphite, or sapphirine. A texture distinctive to this unit is created by garnet porphyroblasts (1 mm to 2 cm) which are draped by sigmoidal ribbons of quartz and feldspar. Layering within intermediate tectonite is much more nebulous than in paragneiss, and observed contacts with paragneiss and gabbroic rocks are subtle and diffuse. The descriptive term intermediate tectonite was assigned to these rocks since it remains unclear from field relationships whether they represent homogeneous diatexite derived through advanced anatexis of paragneiss. Alternate origins may include anatexis of an intermediate plutonic protolith, or single-stage crystallization at granulite-grade conditions of a peraluminous plutonic body of quartz dioritic composition.

Granitic rocks are a minor but widespread component of the complex and occur as centimetre- to metre-scale veins and panels that cut gabbro, paragneiss, and intermediate tectonite. These pink- to white-weathering rocks are of leucogranodioritic composition with 10 to 35% quartz, 10 to 15% K-feldspar, plagioclase, up to 5% garnet, 2-10% mafic minerals, and minor magnetite±pyrite. Mafic minerals are generally orthopyroxene in the southern part of the complex and

clinopyroxene in the north, a distribution which reflects increasing pressure conditions toward the north (see *Metamorphism*).

UNITS EXTERNAL TO THE COMPLEX

North and east of the complex, foliated to gneissic hornblende- and/or biotite-bearing plutonic rocks are widespread (Fig. 2). These vary in composition from gabbro-diorite and quartz diorite with a colour index of 25 to 45, to granodiorite with 25% quartz and a colour index of 5 to 20. Gabbroic to dioritic host rocks are commonly cut by centimetre-scale veins and metre-scale panels of biotite±hornblende granodiorite. Northwest of the complex, foliated K-feldspar porphyritic syenogranite is exposed (Fig. 2).

South of the complex, metasedimentary rocks dominated by lithic wacke with lesser feldspathic wacke, pelites, and rare magnetite-ironstone occur. Amphibolite-facies metamorphism is reflected by the widespread occurrence of biotite+garnet. Sillimanite occurs locally as coarse grained porphyroblasts or fine grained fibrolite (see Fig. 10). In general, metasedimentary rocks display a strong foliation, transposition, and show incipient partial melt textures. In spite of this, primary lithological variations are recognized locally (Sanborn-Barrie, 1993); however, stratigraphic younging or structural facing could not be established.

Metavolcanic rocks are exposed in a restricted zone along the south channel of Chesterfield Inlet (Fig. 2) where they are tightly folded with metasedimentary rocks. Isolated occurrences of volcanics are compositionally and texturally diverse. Garnet-bearing amphibolites display remnant 2 cm wide selvages and are interpreted as pillowed mafic flows. Possible flow top breccia (Fig. 3a) occurs at the interface between pillowed and massive flows. Intermediate pyroclastic breccias with up to 40% buff-weathering rhyodacite fragments in a grey-weathering dacitic groundmass (plagioclase+biotite+garnet) also occur (Figure 3b). Fine grained, layered, tuffaceous rocks of andesitic to dacitic composition appear to be gradational with the metasedimentary rocks.

Early Proterozoic sedimentary and volcanic rocks of the Dubawnt Supergroup unconformably overlie the Kramanitar complex (Fig. 2). The lowermost part of this succession is the Baker Lake Group which includes the basal South Channel conglomerate and red-weathering Kazan sandstone.

STRUCTURE

Planar and linear elements within the complex

Most localities within the Kramanitar complex display a single, strong to intensely developed planar fabric with an associated strong linear fabric. A number of localities, however, display two generations of planar fabrics, and at these localities, relative timing relationships (i.e., S_1 , S_2) can be discerned on the basis of overprinting relationships. In exposures where only one fabric is observed, it is then generally

feasible to identify it as S_1 or S_2 on the basis of its character and attitude (described below). Resultant patterns of S_1 and S_2 are shown in the foliation trajectory map of Figure 4a.

The character of S_1 is dependent to some extent upon rock type. In relatively unstrained gabbroic anorthosite in the central part of the complex, S_1 occurs as a weak to moderately developed foliation. More typically, and well exemplified in the southern part of the complex, S_1 is defined by parallel, centimetre-wide sheets of granitic/charnockitic material cutting gabbro and leucogabbro. This imparts a compositional layering (Fig. 5) to much of the southern part of the complex, although this layering is essentially intrusive in nature. A shape fabric defined by the elongation of individual mineral constituents is only locally observed to be parallel to S_1 layering (Fig. 5c). The general absence of a shape fabric parallel to S_1 layering (Fig. 5d) is attributed to penetrative realignment of mineral constituents during later deformation. Throughout the complex, a characteristic feature of S_1 is its moderately to shallowly dipping attitude (Fig. 4a). Locally S_1 is subhorizontal. S_1 strikes southeast or southwest, defining folds and monoclinical panels throughout the southern half of the complex (Fig. 4a).

Figure 3. Metavolcanic rocks. *a*) Possible flow-top breccia at interface between pillowed and massive flows. (GSC 1993-260I) *b*) Intermediate pyroclastic breccia, bedding (not shown) is subparallel to fragment elongation direction due to transposed nature of these tightly folded rocks. (GSC 1993-260H)

Figure 4. Planar and linear tectonic elements; a) foliation trajectory map. b) mineral lineation data.

Figure 5. S_1/S_2 fabric elements. **a)** panel of folded S_1 layering enveloped by straight gneisses thoroughly transposed parallel to S_2 . (GSC 1993-260J) **b)** folded S_1 layering with an axial planar foliation (S_2) which is parallel to the shear fabric throughout the right half of the photo. (GSC 1993-260U) **c)** S_1 layering which preserves an S_1 shape fabric and carries an S_2 shape fabric oblique to layering (parallel to pencil tip). (GSC 1993-260L) **d)** S_1 layering with a penetrative S_2 shape fabric oblique to layering (parallel to pencil). (GSC 1993-260M)

Metasedimentary rocks south of the complex generally possess a single foliation defined by both layering and a shape fabric. The attitude of this fabric, which strikes southwesterly and dips moderately northwest (Fig. 4a), is consistent with that of S_1 within the complex. As such, it is tentatively interpreted as S_1 also. The presence of a preferred shape fabric parallel to layering in these amphibolite-grade rocks is interpreted to be a function of their higher structural level, the absence of later penetrative strain, or both. In metasedimentary rocks south of the complex, attitudes of S_1 define a shallow north-dipping homocline (Fig. 4a).

The dominant penetrative planar fabric within the complex is a strong to intensely-developed, east- to east-south-east-striking foliation (Fig. 4a). On the basis of overprinting relationships this foliation is designated S_2 . These relationships include the widespread occurrence of mesoscopic folds defined by S_1 layering to which S_2 is axial planar, and of panels of folded S_1 layering with intervening straight, transposed gneisses whereby the transposition foliation parallels the axial planar S_2 (Fig. 5a, b). Since an S_1 shape fabric is

rarely observed within the complex, a less common relationship is the partial realignment of the S_1 shape fabric by S_2 (Fig. 5c). In contrast to S_1 , S_2 is subvertical to steeply south-dipping, with only local moderately ($38-52^\circ\text{N}$) dipping zones (Fig. 4a). The axial planar relationship of S_2 to folded S_1 layering is well displayed in the field and is reflected in the trajectory map (Fig. 4a) in which S_2 trajectories symmetrically bisect folds defined by the trace of S_1 .

Mineral lineations throughout most of the Kramanitar complex show a remarkably consistent pattern which reflects shallow to moderate ($5-35^\circ$) westward extension (Fig. 4b). This extension direction is observed on all S_2 planar fabrics and as such is interpreted as L_2 . Discrete domains of east-plunging lineations are recognized in the northwest, north-central, and southeast parts of the complex (Fig. 4b). The northwest domain is characterized by shallow to moderate ($10-55^\circ$) east-plunging lineations; the north-central domain by subhorizontal to shallow ($5-20^\circ$) east-plunging lineations; and the extreme southeast part of the complex by a moderate east plunge. A domain of variably plunging mineral lineations

coincides with the eastern boundary of the complex (Fig. 4b). Lineations within this domain show a preferred moderate plunge (25–58°) to the north. This domain is interpreted to be associated with a bend in the northern shear zone (Sanborn-Barrie, 1993) and this is discussed below.

Ductile shear zones

Mappable zones of intensely deformed rock or ultramylonite transect both the northern and southern parts of the complex (Fig. 6a). These deformation zones were initially described by Sanborn-Barrie (1993) and a summary of their structural elements with additional observations from 1993 field work are provided below.

Northern shear zone

Granulite-grade mylonites of gabbroic and gabbroic anorthosite composition delineate the northern shear zone throughout an exposed strike length of approximately 60 km, from James Point in the northwest to the Quoich River in the southeast (Fig. 6a). This anastomosing, 3 to 6 km wide shear zone is complex. It has a curvilinear geometry: from James Point in the northwest, the zone strikes northeast to east across the northern part of the complex, and then bends southeast to south around its eastern margin. Its southeast segment which is east-striking, is separated from the rest of the zone by a right-handed step, or jog (locality j in Fig. 6a). All parts of the northern shear zone, except the gap at the dextral jog, comprise mylonites with granulite-grade assemblages, planar and linear fabric elements of similar intensity, and generally consistent shear-sense indicators (discussed below). The northern shear zone separates the Kramanitu complex from its amphibolite-grade wall rocks.

Throughout the zone, shear fabrics dip steeply. Steep southeast-dipping fabrics are observed in the James Point area in the northwest, whereas steep north- and northwest-dipping fabrics prevail across its central and northeast regions, respectively (Fig. 6a). The southeast segment of the shear zone possesses moderate to steep (40–88°) north-dipping shear fabrics. Mineral lineations on shear planes show patterns that parallel those of L_2 on the S_2 foliation. Shallow to moderate west-plunging mineral lineations are dominant; however, discrete domains of moderate, east-plunging lineations characterize the northwest (and locally north-central) part of the zone. The similar attitude and metamorphic grade of S_2/L_2 fabric elements and those of the northern shear zone suggest that shearing throughout the northern part of the Kramanitu complex is an integral part of D₂ deformation and metamorphism, rather than an unrelated, superimposed event.

Regardless of lithology, rocks within the northern shear zone are characterized by continuous, evenly-spaced, millimetre- to centimetre-scale layering (Fig. 7a) and a marked absence of porphyroclasts, folds, and boudins. The uniform nature of ribbon mylonites of gabbroic, gabbroic anorthosite, and anorthositic protolith (Fig. 7b), as well as of paragneiss and leucogranodiorite attests to the absence of rheological contrasts between these rock types and the remarkably homogeneous strain along the northern margin of the complex

during D₂. Preliminary thin section observations suggest that unannealed textures and equilibrium mineral assemblages (clinopyroxene+garnet±orthopyroxene+plagioclase gabbro and quartz+plagioclase+kyanite+garnet paragneiss) are preserved throughout much of the northern shear zone.

In spite of the remarkably symmetrical ribbon structure of the ultramylonites, shear-sense indicators although rare are widespread and reveal south-side-up, sinistral movement parallel to west-plunging lineations (Sanborn-Barrie, 1993). If the present attitude of shear planes approximate their position during displacement, then shearing appears to have taken place in an extensional regime across the north-central part of the zone where shear fabrics dip steeply north; and in a contractional regime across the restraining bend in the northeast (locality b in Fig. 6a) where shear fabrics dip steeply west.

Portions of the northern shear zone examined during 1993 include the northwest sector near James Point, and the southeast sector near the Quoich River (Fig. 6a). In the James Point area, spectacular granulite-grade ultramylonites, well-exposed along the shores of Baker Lake, are compositionally and texturally similar to rocks elsewhere in the northern shear zone; however, two aspects of their fabrics are noteworthy. Firstly, these mylonites are characterized by moderately (20–50°) east-plunging mineral lineations. Secondly, they display an S>L fabric, such that the degree of fabric intensity parallel to XZ is virtually identical to that parallel to YZ (Fig. 7c). Generally rocks in the northern shear zone display L=S and less commonly, L>S fabrics. Kinematic indicators are generally ambiguous in terms of shear sense in these highly flattened rocks, but rarely garnet porphyroblasts show anticlockwise rotation which suggests at least local zones of sinistral displacement. Sinistral transcurrent movement coupled with moderate east-plunging mineral lineations may reflect local sinistral, south-side-down movement in the James Point area.

The southeast sector of the shear zone comprises subvertical, east-striking mylonites which display similar "anomalous" attributes: east-plunging mineral lineations and S>L fabrics. Asymmetrical structures reflecting shear sense are rare. Megascopic structural heterogeneities are, however, prevalent in this restricted domain and include tight folds, branching mafic layers and symmetrical boudinage consistent with significant flattening.

Southern shear zone

The southern shear zone is an east-striking ductile deformation zone that transects the southern part of the Kramanitu complex (Fig. 6a). This zone extends roughly 30 km along strike, maintains a width of 0.5 to 2 km, and has been referred to previously as a segment of the Chesterfield fault zone (Schau and Ashton, 1979). Within this zone, granitoid and amphibolite-grade metasedimentary rocks possess strong to intense, north-dipping (55–65°) shear fabrics. Strongly-developed mineral lineations generally plunge 25–50° to the northwest and, locally, are down-dip. In contrast to the northern shear zone, shear-sense indicators are well displayed (Sanborn-Barrie, 1993) and include asymmetric Z-folds,

Figure 6. Deformation zones. *a*) regional-scale shear zones that transect the complex. Restraining bend at *b*, antidiagonal jog at *j*. *b*) distribution and schematic representation of late-stage conjugate structures (S_3).

dextral shear bands, back-rotated boudins, and asymmetrical winged porphyroclasts which consistently reveal north-side-up, dextral displacement.

Two additional aspects of the southern shear zone were recognized during 1993. Toward the eastern end of the exposed zone, apparent curvature of the shear zone is recorded by a progressive change in the strike of mylonitic foliation from 270° to 230° over approximately 7 km of strike length (Fig. 6a). Shear-sense indicators consistent with dextral displacement are commonly recognized in 250° - to 230° -striking segments. Zones of sinistral displacement and flattening are

also recognized by asymmetrical winged porphyroclasts and symmetrical, conjugate extensional shear bands, respectively. The southeast part of the complex thus records flattening strains across both the northern and southern shear zones.

Toward its western extremity the southern shear zone appears to narrow and at several localities could not be mapped as a throughgoing structure (Fig. 6a). These observations suggest either that this part of the shear zone is discontinuous; that the structure steps successively to the right as echelon segments; or that the scale of mapping in this area was not adequate to delineate the structure.

Figure 7. Ultramylonite of the northern shear zone. *a)* millimetre-scale tectonic layering in mylonitized gabbro anorthosite. (GSC 1993-260C) *b)* layer of anorthosite mylonite (white) and adjacent gabbro mylonite. Note ribbon texture of both units and the absence of boudins or folds, reflecting similar rheological behaviour. (GSC 1993-260Q) *c)* S>L gabbro ultramylonite near James Point. (GSC 1993-260B)

Semi-brittle shear zones and kink folds

Spaced, metre-scale structures cut ultramylonitic layering of the northern shear zone. Such structures include ductile shear zones (Fig. 8a), semi-brittle shear zones (Fig. 8b), and kink folds (Fig. 8c, d). These structures, which generally occur at a high angle to mylonitic layering (Fig. 6b), overprint and offset this fabric and are therefore designated S_3 . Granulite-grade mineral assemblages within S_3 structures reveal that P-T conditions during movement on the northern shear zone were maintained during formation of these later structures. An apparent sense of displacement across these structures is consistently provided by sigmoidal curvature of S_2 into these zones as schematically illustrated in Fig. 6b. Curvature consistent with both sinistral and dextral offset is observed (Fig. 8), although typically only one structure of a conjugate set is observed at an outcrop scale. At different localities

either extensional or contractional conjugate structures occur. Extensional structures may be infilled with granitic material (Fig. 8a, b) which commonly carries a weakly to moderately developed foliation. The obliquity of this foliation with the walls of S_3 structures is consistent with injection of granitic material synchronous with, but late, in their development.

Regional-scale brittle-ductile fault

A discrete, northeast-striking brittle-ductile fault zone cuts the southeast part of the complex (Fig. 6a). In general, planar fabrics within this zone strike northeast ($010-035^\circ$), dip steeply east, and possess moderately south-plunging mineral lineations. Deformation style varies along the 10 km exposed strike length of this fault zone. South of the north channel, the zone is characterized by brittle-ductile deformation with associated greenschist-grade alteration. Composite fabrics

Figure 8. S_3 structures. *a)* dextral extensional shear zones. (GSC 1993-260FF) *b)* sinistral, extensional, semi-brittle shear zones infilled with granitoid rock. (GSC 1993-260R) *c)* kink fold of S-asymmetry in tectonized gabbro anorthosite. (GSC 1993-260D) *d)* kink fold of Z-asymmetry, gabbro dyke cutting gabbro anorthosite. (GSC 1993-260F)

reflect sinistral displacement (Fig. 9) which, in conjunction with moderately south-plunging lineations, suggests sinistral, east-side-up movement. Rocks exposed along the north channel and further northeast reveal an increasingly brittle component to the deformation and attenuation of the fault zone (Fig. 6a). Centimetre-scale sinistral shear zones are, however, still locally observed. A sinistral transcurrent component of displacement on this fault zone is consistent with left-lateral offset of metasedimentary rocks (Fig. 2) and the southern

Figure 9. Composite fabrics in greenschist-grade brittle-ductile fault zone, consistent with sinistral movement.

shear zone (Fig. 6a) across this structure. Continuity of gabbroic anorthosite along the northeast margin of the Kramanituar complex (Fig. 2) is consistent with attenuation of this fault toward the northeast.

Timing of this fault zone relative to other deformation zones is clearly established on the basis of overprinting structural and metamorphic relationships. In the south, east-striking fabrics associated with the southern shear zone wrap into the fault zone in a sense consistent with sinistral movement on the latter. In the north, brittle fracturing and cataclasis of mylonitic rocks of the northern shear zone is restricted to this fault zone. Throughout its length, greenschist assemblages reflect retrograde alteration not associated with the other shear zones.

METAMORPHISM

Across the Kramanituar complex, variations in metamorphic assemblages with strain intensity appear to reflect a close relationship between deformation and metamorphism. In the southern part of the complex, gabbroic units that best preserve S_1 are characterized by the assemblage plagioclase+orthopyroxene > clinopyroxene. Toward the north, increasing strain is clearly demonstrated by the increasingly penetrative nature of S_2 which culminates in ultramylonitic rocks of the northern shear zone. A metamorphic reaction reflecting increasing pressure appears to track the increase in strain intensity. The

Figure 10. Preliminary metamorphic isograds and isolated metamorphic mineral occurrences based on field observation only.

reaction: orthopyroxene + plagioclase $\rightarrow$ garnet + clinopyroxene + quartz is reflected by an orthopyroxene-out isograd located north of Chesterfield Inlet, and the garnet-in isograd parallel to, and roughly 3.5 km south of the northern shear zone (Fig. 10). The relationship between strain intensity and metamorphic grade is also mirrored in paragneiss, whereby sillimanite-bearing rocks are prevalent throughout the central part of the complex, while kyanite is restricted to the northern shear zone (Fig. 10).

Field observations pertaining to the transition from granulites to amphibolite-grade wall rocks are as follows. Across the north margin of the complex, the transition appears to occur across a 1 km wide zone. Lower metamorphic grade appears to be indicated initially by the occurrence of sapphirine in clinopyroxene+garnet+plagioclase gabbro and gabbroic dykes. Sapphirine-bearing mafic rocks delineate an isograd which extends from James Point to the northeastern part of the complex (Fig. 10), parallel to, and immediately north of, the northern shear zone. North of this, hornblende porphyroblasts occur in clinopyroxene+garnet gabbro and intermediate tectonite. Hornblende content increases progressively northward with systematic decrease of clinopyroxene and garnet. Plutonic rocks with hornblende $\pm$ biotite $\pm$ epidote are widespread north of the complex.

In the south, the granulite-amphibolite transition appears to occur at, or immediately north of, the southern shear zone. Hornblende-bearing mafic rocks occur immediately north of the southern shear zone (Fig. 10). Within the shear zone, semipelitic rocks contain biotite, lavender garnet and fibrolite sillimanite (Fig. 10). Further south, biotite and garnet persist throughout much of the exposed metasedimentary succession. Coarse-grained sillimanite porphyroblasts are locally observed. Retrograde chlorite after garnet is present along the south channel of Chesterfield Inlet and in gabbroic anorthosite mylonite in the extreme southeast part of the complex (Fig. 10).

DISCUSSION

The field relationships described above are consistent with the Kramanituvar complex as a crustal-scale, granulite-grade tectonic lens, or boudin, bound by asymmetrical shear zones generated upon uplift during southwest-directed shortening. The following structural chronology is preliminary, and based on the observed field relationships. Rare S_1 shape fabrics in gabbroic anorthosite are consistent with southwest-directed shortening during D_1 . Granitic sheets that define S_1 compositional layering may have been injected during D_1 , or may have utilized D_1 anisotropies during magmatic injection prior to D_2 . A second deformation event, D_2 , resulted in folding of S_1 layering; the development of discrete zones of transposition in the southern part of the complex; and the increasingly penetrative development of S_2 toward the northern part of the complex. Highest D_2 strains are recorded by anastomosing ultramylonites of the northern shear zone which delineate the northern margin of the Kramanituvar complex. South-side-up and sinistral shear along much of the

northern shear zone is consistent with uplift of granulite-grade rocks relative to their amphibolite-grade wall rocks during southwest-directed shortening. Extensional shear across steeply dipping shear planes in the north-central part of the zone is contemporaneous with contractional shear across a restraining bend in the northeast part of the zone.

Anomalous east-plunging mineral lineations and marked flattening strains in the northwest and southeast sectors of the northern shear zone may reflect local strains in these quadrants during uplift and sinistral shear. A local dextral regime, reflected by attitudes of conjugate extensional S_3 structures (Fig. 6b) in the northwest sector, may have been imposed on the high-temperature, low yield-strength mylonites by the stiffer, colder wall rocks which envelop them. In the southeast sector of the northern shear zone, contractional S_3 structures generated by layer-parallel shortening similarly may reflect the localized effect of wall rocks near the southeast tip of this crustal-scale boudin.

Along the south boundary of the complex, highest strains are observed in the east-striking southern shear zone. The relative timing of the northern and southern shear zones could not be resolved during field mapping. Granulite-grade assemblages of the northern shear zone, relative to amphibolite-grade assemblages of the southern shear zone demonstrate that these deformation zones were operative at different crustal levels, but does not preclude contemporaneous movement. This problem, which is critical to understanding the uplift history of the Kramanituvar complex may be resolved petrologically, if P-T conditions in the northern shear zone approach those of the southern shear zone in this area. Progressive curvature of the southern shear zone to a trend of 230° is consistent with wrapping of the southern shear zone around the attenuated southeast tip of a crustal-scale boudin.

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Structural investigation of high-grade rocks of the Kramanituar complex, Baker Lake area, Northwest Territories

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Abstract: Granulite-grade metaplutonic and metasedimentary rocks of the Kramanituar complex record strong to intense deformation, although moderate strain is preserved in coarse grained gabbroic anorthosite in the central part of the complex. An outward increase in strain culminates in kilometre-wide zones of ultramylonite which appear to define boundaries to the complex.

Movement on the shear zones is consistent with uplift of granulite-grade material relative to amphibolite-grade rocks exposed north and south of the complex. However, it remains to be demonstrated whether finite strain recorded by the shear zones can account for the juxtaposition of these disparate metamorphic terranes, or whether the shear zones are relatively late structures that transect a previously assembled metamorphic terrane.

Résumé : Les roches métaplutoniques et métasédimentaires du faciès des granulites du complexe de Kramanituar présentent un degré de déformation fort à intense, bien que l'anorthosite gabbroïque à grain grossier de la partie centrale du complexe soit moyennement déformée. Une intensification de la déformation à partir du centre du complexe culmine en des zones de largeur kilométrique d'ultramylonite, qui semblent définir les limites du complexe.

Le déplacement des zones de cisaillement concorde avec le soulèvement des matériaux du faciès des granulites par rapport aux roches du faciès des amphibolites qui affleurent au nord et au sud du complexe. Il reste toutefois à démontrer si la déformation finie des zones de cisaillement peut expliquer la juxtaposition de ces terranes métamorphiques disparates, ou si ces zones de cisaillement sont des structures tardives qui recoupent un terrane métamorphique s'étant constitué antérieurement.

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INTRODUCTION

Prominent linear anomalies in horizontal gravity gradient and magnetic anomaly maps of Canada (Goodacre et al., 1987; GSC, 1987) extending from central Alberta to the northwest shore of Hudson Bay correspond to discontinuous occurrences of granulite-grade rocks (highs), and metasedimentary rocks and/or paragneiss (lows) (Gibb, 1968; Wallis, 1970). Speculation on the geological nature of these anomalies across the central part of the (former) Churchill province has led to two interpretations: 1) that the Churchill Province comprises two Archean microcontinents, the northern Rae and the southern Hearne, joined along a northeast- to east-striking suture, designated the Snowbird tectonic zone (STZ) (Hoffman, 1988; Hanmer et al., 1992), which resulted from their collision in Early Proterozoic time (Walcott and Boyd, 1971; Gibb and Thomas, 1976; Thomas et al., 1988; Hoffman 1989); or, 2) that they represent an intra-plate boundary which corresponds to the northwestern limit of crustal reactivation associated with the Trans-Hudsonian orogeny (Lewry and Sibbald, 1980; Thomas and Gibb, 1985).

An early Proterozoic age for the development of the suture was proposed by Hoffman (1989), based on the apparent truncation of the circa 2.02 to 1.9 Ga Taltson magmatic arc of northeastern Alberta by the Snowbird Tectonic Zone in the subsurface; and the age of sedimentary rocks, lamprophyre dykes and related volcanic rocks of the Dubawnt Supergroup (circa 1.875 Ga), interpreted to form an overlap assemblage at the Rae-Hearne interface near Baker Lake. However, recent work along a segment of the Snowbird Tectonic Zone by Hanmer et al. (1992) has demonstrated that deformation under granulite facies conditions occurred at ca. 3.2 and ca. 2.6 Ga.

Previous studies on segments of the Snowbird Tectonic Zone

Previous work has focused on five segments of the 2800 km length of the proposed Snowbird Tectonic Zone. These include the East Athabasca mylonite zone and the Tulemalu fault zone southwest of Baker Lake; the Uvauk and Daly Bay complexes east of Baker Lake; and the Kramanituak complex near Baker Lake (Fig. 1). These studies have emphasized different aspects of the geology of each particular segment, and the structural history of the Snowbird Tectonic Zone as a whole remains enigmatic. Results of these studies are summarized briefly below.

The East Athabasca mylonite zone in northern Saskatchewan (Fig. 1) is an anastomosing system of granulite- and amphibolite-grade mylonites that developed in response to bulk pure shear during northwest-directed shortening (Hanmer et al., 1991, 1992). On a local scale, these rocks record a complex kinematic history which is interpreted to reflect partitioning of the bulk pure shear flow into coeval zones of dextral, sinistral and extensional (southwest-side-down) ductile flow. About 300 km southwest of Baker Lake, the Tulemalu fault zone (Fig. 1) contains high-pressure mafic granulite xenoliths hosted in ductilely deformed granitoid gneiss (Tella and Eade, 1986). Movement on the fault is not supported by direct field relationships but is postulated to

involve reverse, west-side-up ductile displacement owing to the occurrence of granulite gneiss west of the fault; followed by normal, west-side-down faulting (Tella and Eade, 1986).

About 200 km east of Baker Lake on Chesterfield Inlet, anorthositic rocks and granulite-grade orthogneiss were reported during reconnaissance mapping by Tella et al. (1992). These rocks, designated the Uvauk complex (Fig. 1), were mapped in more detail during the 1992 field season and are described in this volume (Tella et al., 1993). The Daly Bay complex on Hudson Bay (Fig. 1) comprises felsic to mafic granulite gneiss, gabbroic anorthosite and migmatite and is interpreted as an allochthonous complex thrust northwestward from lower crustal levels (0.55 GPa, 750°C) onto amphibolite-grade gneiss during the Early Proterozoic (Gordon, 1988).

Figure 1. Geology of the western Canadian Shield after Hoffman (1989). Segments of the proposed Snowbird tectonic zone, at the interface between the Rae and Hearne provinces, include the East Athabasca mylonite zone (EAmz), the Tulemalu fault zone, the Kramanituak complex (Kc), the Uvauk complex (Uc) and the Daly Bay complex (DBc).

Layered granulites and anorthositic rocks of the Kramanitu complex near Baker Lake (Wright, 1967; Schau and Hulbert, 1977; Schau and Ashton, 1980; Schau et al. 1982; Schau, 1983) are situated close to a significant bend in the regional-scale geophysical lineament from northeast- to east-trending (Fig. 1). These high-grade rocks were described as prominently layered, strongly foliated and conspicuously lineated (Schau and Hulbert, 1977; Schau and Ashton, 1980), features that suggested these rocks had undergone significant tectonic modification.

This study

A detailed structural and petrological study of the Kramanitu complex was commenced during the 1992 field season in order to clarify the tectono-metamorphic history of this segment of the Snowbird Tectonic Zone. Previous mapping of well exposed and accessible rocks of this complex (Schau and Hulbert, 1977; Reinhardt et al., 1980; Schau and Ashton, 1980; Schau, 1983) provides an excellent geological framework from which detailed studies can proceed. The location of this complex close to a bend in an interpreted regional-scale crustal break lends added importance to this segment of the Snowbird Tectonic Zone. An understanding of the structural and metamorphic history of this segment is critical if comparisons are to be made between the Kramanitu complex and high-grade rocks of the East Athabasca mylonite zone to the southwest, or the Uvauk and Daly Bay complexes to the east (Fig. 1).

LITHOLOGICAL UNITS WITHIN THE COMPLEX

The Kramanitu complex, exposed over roughly 60 km by 30 km, was proposed by Schau and Ashton (1980) to include rocks exposed east of Baker Lake with mineral assemblages reflecting granulite facies metamorphism. The complex consists of a central suite of anorthositic rocks enveloped to the north and south by layered granulite-grade rocks. The anorthositic suite comprises mainly gabbroic anorthosite, gabbro and leucogabbro, with minor anorthosite, anorthositic gabbro and ultramafic rock. Adjacent granulite-grade rocks include paragneiss, diatexite and various metaplutonic rocks. The geology of the complex and adjacent rock units is shown in Figure 2.

Anorthositic suite

The anorthositic suite is dominated by light buff to white weathering gabbroic anorthosite (Fig. 2a) composed of sodic labradorite (Schau and Hulbert, 1977) with 12% to 20% orthopyroxene+clinopyroxene+garnet. Magnetite+ilmenite are accessory phases. Where these rocks are not penetratively strained and recrystallized, they are generally coarse grained and contain up to 60% phenocrysts of plagioclase (average 2 cm by 1 cm). Orthopyroxene porphyritic phases occur at several localities along the northeast margin of the complex and in its centre (Schau, pers. comm., 1992). Rarely are both plagioclase and orthopyroxene phenocrysts observed (Fig. 3a).

Gabbro and leucogabbro occur as 1 to 100 cm-wide layers in gabbroic anorthosite and as homogeneous units up to 2 km wide (Fig. 2a). Centimetre-scale layers are rusty brown weathering, equigranular, fine- to medium-grained (0.5-2 mm), with 30-40% pyroxene, 0-15% garnet and locally up to 2% magnetite. These rocks generally comprise 2-15% of exposures of predominantly gabbroic anorthosite. The gabbroic layers are generally tectonized, straight, uniform in composition across their breadth, with margins that typically are in discrete contact with adjacent gabbroic anorthosite. Locally, gabbroic layers are observed to thin and pinch out parallel to strike. Although these layers are tectonized, they are believed to reflect primary magmatic layering. Rarely, modal layering is present in less strained rocks (Fig. 3b).

Gabbro and leucogabbro, 1 to 2 km wide, occur as discrete units in the northeast part of the complex (Fig. 2a). These dark green to rusty brown weathering rocks contain plagioclase, 15-25% pyroxene, 2-20% garnet, and may contain <2% magnetite and <5% quartz. Patches with up to 50% garnet and 30% interstitial quartz occur locally.

Rocks of true anorthosite composition (colour index <10) form a minor component of the anorthositic suite. They occur mainly as subangular, metre-scale xenoliths of fine grained rock hosted in gabbroic anorthosite (Fig. 2a; locality A). Anorthosite ($\pm$ gabbroic anorthosite) also occurs as tabular units, 10 to 30 m wide and several kilometres long, within paragneiss. These units are generally layered and locally such layering is thought to be primary. By virtue of their tabular geometry, layered nature and the presence of modal layering locally, these distinctive marker horizons are interpreted as sills and considered as part of the anorthositic suite.

Granulite suite

Granulite-grade rocks exposed north and south of the anorthositic suite are well-layered rocks composed mainly of various proportions of plagioclase+orthopyroxene+clinopyroxene+garnet+quartz+K-feldspar. These rocks are subdivided according to their protolith as paragneiss and metaplutonic rocks. Further subdivision of the metaplutonic rocks is based on textural and compositional variations.

Paragneiss

Layered paragneiss is exposed as two elongate units north and south of the anorthositic suite, and in the extreme northeast part of the complex at locality B (Fig. 2a). These rocks consist of fine grained, rusty weathering, centimetre- to metre-wide layers interpreted as feldspathic wacke which alternate with white-weathering, quartzo-feldspathic leucosome, 1-30 cm wide (Fig. 4). The feldspathic wacke (paleosome) is composed mainly of plagioclase, 15-20% biotite, 5-15% quartz and up to 8% garnets. These rocks may contain up to 10% graphite, 1% to 5% sillimanite, 2% to 10% kyanite, K-feldspar, pyrite and rutile. Locally hypersthene is present in addition to biotite, or 5% to 15% hypersthene occurs in the absence of biotite. Graphite and sillimanite appear to be fairly widespread, while kyanite-bearing rocks are restricted to two

Figure 2. General geology of the Kramanituur complex and adjacent units. a) Lithological units modified from Schau and Ashton (1980) and Schau (1983). Localities A, B, C and D referred to in the text. b) Major structural elements.

narrow panels, several hundred metres wide and several kilometres long, in the northeast part of the complex (locality B in Fig. 2a). Leucosome is dominantly composed of plagioclase, 20-30% ribbon quartz, up to 10% K-feldspar and 1-10% garnet. Minor phases are graphite and sillimanite. Kyanite occurs in leucosome at locality B. Leucosome generally comprises between 2% and 20% of paragneiss (Fig. 4a), but may comprise up to 60%.

Centimetre-scale layers of pink weathering, garnet-bearing leuco-monzogranite (described below) form a minor component of the paragneiss locally. Anorthosite sills (described above) and gabbroic dykes and/or sills crosscut paragneiss.

Metaplutonic rocks

Diatexite

Pale buff weathering, quartzo-feldspathic diatexite is a distinctive unit which is widespread throughout the complex (Fig. 2a), and occurs northwest of Jigging Point where it is interpreted as an outlier of the complex (Schau et al., 1982). The diatexite is a texturally distinct unit owing to its medium

to coarse grain size and pock-marked appearance created by the presence of 1-5 mm red anhedral garnets draped with ribbon quartz. These rocks contain plagioclase, 10-15% quartz (locally up to 30%), 2-15% garnet, 10-15% very fine pale grey-brown pyroxene, and up to 10% K-feldspar. Finely disseminated biotite may be present in addition to pyroxene. At several localities, rare porphyroclasts of resinous brown pyroxene <2 mm in length were observed. These rocks generally possess a mylonitic foliation. In contrast to other units, the diatexite displays a relatively homogeneous texture with indistinct centimetre- to metre-scale banding reflecting subtle variations in colour index and strain intensity. These rocks commonly occur as 5-10 m wide sheets or panels that alternate with paragneiss or garnetiferous gabbro.

Lithological relationships observed in less strained and metamorphosed rocks in the southeast part of the complex suggest that the diatexite is a composite unit of two plutonic rocks: a dioritic host cut by pervasive centimetre-wide dykes of leuco-monzogranite (described below). Metamorphism with attendant anatexis and deformation of these two units are believed to have resulted in the homogenization of these two units to produce this distinctive garnet-bearing quartzo-feldspathic diatexite. A transition from relatively unstrained,

Figure 3. Anorthositic suite. **a)** Plagioclase and orthopyroxene porphyritic gabbroic anorthosite (GSC 1992-272W). **b)** Modal layering in deformed gabbroic anorthosite showing systematic decrease in mafic minerals toward right side of photo. (GSC 1992-272J)

Figure 4. Tectonized, granulite-grade paragneiss. **a)** Centimetre- to metre-scale layers of feldspathic wacke alternating with white weathering leucosome (GSC 1992-272S). **b)** detail of a. (GSC 1992-272I)

unmetamorphosed plutonic protoliths to diatexite was observed northwest of Cross Bay (Fig. 2) along several across-strike traverses.

Quartz diorite and diorite

Magnetite-bearing leucocratic quartz diorite ($\pm$ diorite) is widespread across the southern part of the complex (Fig. 2a). These light grey to buff-weathering, medium-grained rocks are composed of plagioclase, 5-15% pyroxene and amphibole, 5-10% quartz, and notable (2-8%) magnetite which occurs as disseminated grains and rarely as layers <1 cm wide. These rocks are moderately to strongly foliated to gneissic.

Narrow panels of fine grained diorite are exposed north of North inlet (Fig. 2a). These are dark grey-green weathering, weakly to moderately magnetic rocks with a colour index of 30 to 40. These rocks are believed to represent the dioritic component of the diatexite which occurs adjacent to, and is transitional with, these rocks.

Granitic rocks

Pink-weathering granitic rocks are exposed at two localities along the eastern shore of Baker Lake (locality C and D) and in the southern part of the complex (Fig. 2a). At locality C, quartz-monzodiorite and monzodiorite contains 5-20% ribbon quartz, 5-10% very fine grained pyroxene $\pm$ biotite, plagioclase, K-feldspar $\pm$ magnetite $\pm$ ilmenite. K-feldspar generally makes up 10-50% of the total feldspar. These rocks are generally very highly strained. They are intruded by a swarm of moderately foliated, 5-20 m wide gabbro dykes, which limit exposures of the granitic rocks to narrow panels less than 3 m wide. At locality B, moderately strained, pink-to orange-weathering monzogranite and leuco-monzogranite (charnockite) is composed of plagioclase, K-feldspar, 20-25% quartz and 5% very fine grained mafic minerals which generally appear to be pyroxene. K-feldspar typically comprises 30-40% of the total feldspar. These rocks are commonly interlayered with dioritic to quartz dioritic rocks, however, in contrast to locality C, the granites here cut their dioritic host. The granitic rocks are interpreted to be the granitic component of the diatexite observed to the north.

Granitoid rocks in the southern part of the complex are similar in composition to those at locality B and C. These are strongly deformed rocks that appear to form a discrete plutonic body (Fig. 2a).

UNITS EXTERNAL TO THE COMPLEX

Amphibolite-grade metaplutonic and metasedimentary rocks occur external to the Kramanituar complex (Fig. 2a). North and east of the complex, foliated to gneissic metaplutonic rocks are dominant. Metasedimentary rocks are exposed south of the complex.

Grey-weathering, foliated to gneissic rocks of dioritic to granodioritic composition (Fig. 5a), exposed north and east of the complex, are designated the Ingilik Point gneiss complex (Schau et al., 1982) and interpreted to be ca. 2675 Ma old (U-Pb; Schau, 1980). These are typically hornblende-bearing rocks composed of plagioclase, 15-20% hornblende $\pm$ biotite, 5-20% quartz, and K-feldspar, which comprises up to 20% of the total feldspar. These rocks are moderately foliated to gneissic. In contrast to the straight layering typified by units within the complex, these rocks display fabrics of moderate intensity and variable attitude.

Moderately foliated, amphibolitic gabbro with a colour index of 40 to 50 occurs as discrete, elongate bodies north and east of the complex (Fig. 2a). The geometry and distribution of these bodies suggest they may represent a boudinaged mafic unit(s).

Pink weathering, K-feldspar porphyritic syenogranite, east of Jigging Point (Fig. 2a), contains up to 30% blocky subhedral K-feldspar phenocrysts (up to 2 cm by 5 cm) in a medium-grained groundmass of plagioclase, 15-20% quartz

Figure 5. Lithological units external to the complex. a) Gneissic granodiorite north of the complex (GSC 1992-272M). b) Partially melted metasedimentary rocks south of the complex. (GSC 1992-272Q)

and 10-15% elliptical aggregates of biotite. Locally these rocks are cut by quartz veins with associated pyrite±chalcopyrite±fluorite.

Throughout the southern part of the map area amphibolite-grade metasedimentary rocks are widespread (Fig. 2a, 5b). These rocks form a relatively homogeneous sequence of moderately to strongly foliated, biotite- and/or garnet-bearing feldspathic wacke. These are light grey weathering rocks mainly composed of plagioclase, 10-20% black to reddish-brown biotite, 5-12% quartz and <10% pale pink garnet. Locally these rocks contain minor tremolite, talc, sillimanite and possible pyrophyllite. Magnetite ironstone and magnetite-bearing wacke occur as 1-100 cm wide units at several localities.

STRUCTURE

Planar and linear elements within the complex

Gabbroic anorthosite in the central part of the complex is relatively unstrained and displays some of the earliest fabrics recognized. These include primary magmatic and modal layering (S_0) and a weak to moderately developed tectonic foliation (S_1). S_0 strikes east-southeast and dips steeply south (Fig. 2b). Rare modal layering, showing a systematic decrease in colour index across layers toward the south, suggests that this part of the intrusion may young to the south. In these relatively unstrained rocks, S_1 strikes east-southeast, dips steeply (73-85°) south, and is subparallel to S_0 or forms a small (2-15°) counterclockwise angle. S_1 rarely displays a weak, shallow, west-plunging mineral lineation defined by aligned pyroxene.

South of the low-strain domain, gneissic layering (S_g) is pervasive throughout quartz diorite and diorite. Most commonly S_g strikes southeast and dips moderately (40-60°) to the southwest. Several fold closures (ie., domal pattern) are suggested by variations in the strike and dip of S_g (Fig. 2b), alternatively, these variations may reflect an anastomosing character of the gneissic fabric. Irrespective of the attitude of S_g , mineral lineations in these rocks plunge moderately (15-50°) to the west. The relative timing of development of S_g and S_1 is not yet established, however both of these fabrics predate, and are overprinted by S_2 .

A strong to intense (mylonitic) foliation is regionally developed throughout the Kramanituur complex, with the exception of the low-strain domain in its centre. North and east of the low-strain domain, a progressive increase in strain intensity culminates in kilometre-wide zones of ultramylonite that parallel, and appear to define, the northern and eastern boundaries to the complex (Fig. 2b). This fabric is designated S_2 , since an overprinting relationship with S_1 can be established at several localities.

The nature of S_2 is variable, and appears to be largely controlled by the mineralogy and relative competency of the lithological units in which it is developed. In gabbroic anorthosite, S_2 is manifest as a streaky, discontinuous foliation formed by recrystallized aggregates of feldspar, pyroxene and garnet. In paragneiss, S_2 is defined by straight,

continuous, alternating layers of feldspathic wacke and leucosome (Fig. 4), and by ribbons of quartz and aligned mafic minerals within these units. In gabbro and leucogabbro, S_2 is typically manifest as a mylonitic foliation characterized by millimetre-scale laminations.

A progressive change in the attitude of S_2 can be mapped across the north and east parts of the complex (Fig. 2b). Across the northern part of the complex, S_2 strikes east-northeast to east and dips steeply to the north and south. Mineral lineations are strongly developed, and plunge shallowly to the west and east. Toward the eastern part of the complex, S_2 progressively swings in strike from east to south-southeast and dips steeply (70-85°) to the southwest. Strongly to intensely developed mineral lineations plunge west and northwest, and progressively steepen to a maximum plunge of 55° north-northwest along the eastern boundary of the complex.

South of the low-strain domain, an abrupt increase in strain intensity is revealed by a mylonitic foliation (S_2) displayed by granitic rocks at locality C (Fig. 2). These fabrics strike east, dip south (60-85°) and possess shallow (2-30°) east- and southwest-plunging mineral lineations. Farther south, a moderate to strong, east-striking, steeply south-dipping S_2 foliation overprints S_g in plutonic rocks (Fig. 2b).

In the southeast part of the complex, north of Cross Bay (Fig. 2), a strong to intense east-striking tectonic foliation is observed. In contrast to the steep dip of other high-strain parts of the complex, foliations here dip moderately (40-70°) north. Strongly to intensely developed mineral lineations plunge 15-45° to the northwest.

Ultramylonite zones

The degree of strain recorded by most units throughout the Kramanituur complex is generally very strong. Rocks characterized by intense ductile deformation appear to form discrete zones across the northern and southern margins of the complex (Fig. 2b). These zones are developed in a variety of rock types, including gabbro, gabbroic anorthosite, paragneiss and granite; all of which display ultramylonitic fabrics reflecting extreme reduction in grain size. Ultramylonitic paragneiss, gabbro, and gabbroic anorthosite occur along the northern boundary of the complex (Fig. 6), and mainly mylonitized granitic rocks occur along its southern boundary (Fig. 7). Weathered exposures of rocks within these zones are generally characterized by remarkably straight, continuous millimetre-scale laminations. Fresh surfaces are typically aphanitic and, at a hand sample scale, are virtually textureless.

The northern zone of mylonitic rocks is about 2 km wide and is curvilinear in its geometry. This zone of rocks strikes east and dips vertically in the northwest, and progressively changes to southeast- and south-striking with 70-80°SW dips in the eastern part of the complex (Fig. 2b). Strongly developed mineral lineations generally plunge shallowly to moderately to the west and northwest but may plunge shallowly east. At one exposure, east- to west-plunging lineations displayed across a single foliation surface demonstrate

the curvilinear nature of the displacement vectors along this part of the zone. Kinematic indicators within the northern mylonite zone were mainly recognized in the northeast part of the complex. Asymmetric wings on garnet and pyroxene (Fig. 6d) in gabbro, and shallow plunging S-folds in paragneiss, in conjunction with generally moderate west-plunging mineral lineations indicate sinistral, south-side-up displacement.

The southern ultramylonite zone is about 500 m wide and strikes uniformly east and dips 30-65°N (Fig. 7a). Mineral lineations vary in plunge from down-dip in the west, to moderate (30°NW) in the eastern part of the zone. Kinematic indicators, including dextral offset along shear bands

(Fig. 7b), asymmetric wings of K-feldspar phenocrysts (Fig. 7c), and folds of Z-asymmetry (Fig. 7d), record oblique, north-side-up, dextral shear.

Planar and linear elements external to the complex

Foliated and gneissic plutonic rocks north of the complex display southeast-striking foliations that dip moderately to shallowly (60-25°) to the southwest and northeast (Fig. 2b). Mineral lineations plunge moderately (10-35°) to the west. East of the complex, dioritic orthogneiss strikes north-northeast and dips steeply to the east and west (Fig. 2b).

Figure 6. Mylonitic rocks along the northern margin of the complex. **a)** Tectonic layering in mylonitized gabbroic rocks (GSC 1992-272K). **b)** Gabbro ultramylonite with two pyroxene bands alternating with plagioclase bands (GSC 1992-272V). **c)** Moderately west-plunging linear fabric displayed by diatexite (GSC 1992-272Y). **d)** Sinistrally rotated orthopyroxene porphyroclast with fibrous fringe of amphibole (GSC 1992-272U). **e)** Shear bands showing sinistral shear parallel to pencil (GSC 1992-272D).

Mineral lineations plunge to the southwest and southeast at 50-80°, representing the steepest linear fabrics observed in the map area.

Metasedimentary rocks south of the complex display a moderately developed, west- to southwest-striking foliation that dips 40-60°NW. Moderately developed mineral lineations consistently plunge 15-35° to the west and northwest.

DISCUSSION

One of the most striking features of rocks of the Kramanitar complex, apart from their high metamorphic grade, is the pervasive development of strong to intense, penetrative tectonic fabrics. This study will investigate the local and regional significance of these fabrics and their kinematics, and will attempt to unravel as much of the early history of the area as possible. Several preliminary interpretations relating to the magmatic, structural and metamorphic history of the area are summarized below.

The anorthositic suite, as noted by Schau and Hulbert (1977), displays evidence of a cumulate history and is interpreted to have originated as a sill-like intrusive complex

owing to the presence of magmatic layering, and particularly modal layering, in relatively unstrained parts of the intrusion. A change to more gabbroic composition toward the northeast part of the intrusion could be considered to be consistent with an interpretation of southward younging.

Two lines of evidence suggest that anorthositic magma was emplaced at some depth, and then crystallized prior to the deformation events recorded by these rocks. The first concerns the presence of leucosome, not only as a significant component of paragneiss, but also interlayered with gabbroic rocks in the northeast part of the complex. Both of these units lack textural evidence (such as restite development) of advanced partial melting required to generate such volumes of garnet+sillimanite±kyanite quartzo-feldspathic leucosome. This suggests that the leucosome was not generated in situ, but was injected (intruded) into these units. The volume of injected leucosome suggests significant melting occurred at depth. The mylonitized nature and straight gneissic layering of the paragneiss (Fig. 4) reveals that melting took place prior to significant deformation. The second line of evidence concerns the anorthositic units in paragneiss, interpreted to be sills. These sills are not fault bounded, though undoubtedly they are transposed in parallelism with the tectonic fabric

Figure 7. Mylonitic rocks along the southern margin of the complex. a) shallow-dipping mylonitic fabric in monzogranite, note strongly developed lineations parallel to hammer (GSC 1992-272F). b) back-rotated leucocratic pods resulting from dextral shear parallel to pencil (GSC 1992-272E). c) asymmetric wings of K-feldspar porphyroclast indicating dextral shear (1992-272A). d) Z-folds in granitic mylonite (GSC 1992-272D).

which penetrates the complex. Their presence strongly supports an intrusive relationship between the anorthositic suite and a metasedimentary host.

A number of lithological and structural similarities within and external to the complex were noted which should be considered, particularly in light of a tectonic model involving collision of two terranes (Hoffman, 1988). Granulite-grade rocks of metasedimentary origin are exposed within the complex, both north and south of the anorthositic suite (Fig. 2a). Amphibolite-grade feldspathic wacke, exposed south of the complex is compositionally similar (ie., quartz content) and may represent a lower grade equivalent. Some structural fabrics are similar in attitude inside and outside of the complex (Fig. 2b). For instance, moderate to shallow dipping gneissic fabrics (S_g) are similar in attitude to gneissic fabrics displayed by metaplutonic rocks north of the complex, and to foliations in metasedimentary rocks south of the complex.

Ultramylonitic rocks occur in zones that are parallel to, and appear to define the boundaries of the Kramanituar complex. Movement on the north and south mylonite zones is consistent with uplift of granulite-grade material relative to the amphibolite-grade rocks that envelop the complex. However, it remains to be demonstrated whether the finite strain recorded by the shear zones reflect movement responsible for juxtaposing these disparate metamorphic terranes, or whether the shear zones are relatively late features that dissect a previously assembled metamorphic terrane in which earlier events are cryptic in nature.

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Appendix III. Crystallization ages of plutonic and volcanic rocks of the western Churchill Province.

PLUTONIC ROCKS

---ARCHEAN---

Rae granitoid rocks	3.3-2.8 Ga	Wanless (1979)
gneiss, Kasba Lake	ca. 3270	Loveridge et al. 1988
high grade gneiss of the hearme	3.48-2.67	Eade (1986)
granodiorite, Edehon-Hyde lakes	ca. 3190 Ma	Loveridge et al., 1988
Anorthosite-mafic complex, Sask.	3149 ±100 Ma	Hanmer, 1997
Striding River granite, Athabasca, NWT	ca. 3300-3100	
Chipman dykes and granites	3126 +6/-5 Ma	Hanmer et al., 1994
granodiorite, Hyde Lake	2737 +13/-9 Ma	Loveridge et al., 1988
mylonitic orthogneiss, Tavani	2713 ±13 Ma	Park and Ralser (1992)
mylonitic orthogneiss, Tavani	2704 ±4 Ma	Park and Ralser (1992)
Kaminak batholith	2700 ±11 Ma	Cavell, 1992
Last Lake pluton	2686 ±4 Ma	Davis, pers. comm. 1998
porphyritic tonalite dykes, Last Lake	2685 ±3 Ma	
granodiorite, Kramanituvar	2675 +33/-11 Ma	Schau, 1980
Gill Lake granite	ca. 2670 Ma	
syenite, Kaminak Lake	2659 ±5 Ma	Cavell et al., 1992 ?
granodiorite, Yathkyed Lake	2653 +8/-7 Ma	Loveridge et al., 1988
McGillivray granite, Athabasca	2621 ±3 Ma	Hanmer et al., 1994
granite, Amer Lake area	2621 ±2 Ma	Ashton, 1988
foliated granite, Chesterfield Inlet	2615 ±17 Ma	
Whitehills late- to post-tectonic granite	2612 ±4 Ma	
pegmatite, Eyehill High	2612	
Clut and Melby-Turnbull granites, Sask.	2610 ±2 Ma	Hanmer, 1997
Deep Rose Lake leucogranite	2610 +13/-12 Ma	LeCheminant and Roddick, 1991
quartz-feldspar porphyry, Pukia Lake	2610 +11/-13 Ma	" "
megacrystic monzogranite, Dubawnt Lake	2605 +5/-4 Ma	" "
Hawkes granite, Athabasca, Sask.	2604 ±1 Ma	Hanmer, 1997
granodiorite gneiss, Eyehill High	2601 Ma	
Godfrey granite	ca. 2601 Ma	Hanmer et al., 1994
Rea granite	ca. 2600 Ma	
Axis mafic complex	ca. 2600 Ma	
Bohica mafic complex	ca. 2600 Ma	
Fehr granite	2598 ±3 Ma	
granite, Amer Lake	2599 ±5 Ma	Ashton, 1988
megacrystic granite, Chantrey Inlet	2596 +13/-10 Ma	
granodiorite, Barclay volcanic belt	ca. 2595 Ma	
megacrystic augen gneiss, Wharton Lake	2595 +14/-13 Ma	LeCheminant and Roddick, 1991
Dacite porphyry, Akiliniq Hills	2581 +10/-9 Ma	" "
Selwyn Lake granite, NWT	2585 ±2 Ma	Hanmer, 1997
Kramanituvar charnockite gneiss	2573 +28/-5 Ma	Schau, 1980

granodiorite, Edehon Lake	2544 ±11/-8 Ma	Loveridge et al., 1988
alkaline complex, Kaminak Lake	2540 ±76 (Sm-Nd) 2540 Ma ²⁰⁷ Pb/ ²⁰⁶ Pb	Cavell et al., 1987 Wanless, unpub. data
—PROTEROZOIC—		
garnet-pyroxene granite, Wholdai Lake Robillard-Patterson gneissic plutons	1907 ±8 Ma ca. 1900 Ma	Hanmer 1997 Hanmer et al., 1994
quartz syenite Amer Lake	1850 +30/-10 Ma	Tella et al., 1985a
isotropic granite, Athabasca	ca. 1840 Ma	Hanmer et al., 1994
fluorite granite, Daly Bay fluorite granite, Chesterfield Inlet fluorite granite, Chesterfield Inlet	ca. 1832 Ma 1826 ±3 Ma 1803 +16/-13	Tella and Eade, 1986
rapakivi granite granite, Christopher Island rapakivi granite	1760 1753 +3/-2 Ma 1748	LeCheminant et al., 1987a Loveridge et al., 1987

VOLCANIC ROCKS

—ARCHEAN—

<i>Prince Albert Group</i> felsic volcanic, Committee Bay	2879 Ma	Frisch, 1982
<i>Woodburn Lake Group</i> dacite porphyry, Amer Lake dacite, Whitehills Lake	2798 +24/-21 2786 Ma	Tella et al., 1985b Henderson et al., 1991
<i>MacQuoid-Gibson Supracrustal Belt</i> dacite pyroclastic breccia, Chesterfield Inlet	>2655 ±12 Ma <2691 ±10 Ma	Davis and Sanborn-Barrie, unpub. data
<i>Rankin Inlet Group</i> felsic (formerly 2644 Ma)	2665 ±3 Ma	Tella unpub. data?
<i>Ennadai Greenstone Belt</i> felsic volcanic, N. Saskatchewan rhyolite, N. Saskatchewan	2708 +3/-2 Ma 2682 ±6 Ma	Devaney et al., 1990 Chiarenzelli and Macdonald, 1986
<i>Kaminak Greenstone Belt</i> Whiterock Lake tuff rhyolite, middle strat. section tuff rhyolite, Kaminak Lake, top strat. section rhyolite, Quartzite Lake felsic tuff, Yathkyed Lake Gill Lake rhyolite porphyry	2700 ±1 Ma 2697 ±1.4 Ma 2695 ±3 Ma 2692 ±1 Ma 2681 ±3 Ma 2680 +28/-24 Ma 2666 ±9 Ma	Davis, pers comm., 1997 Mortenson and Thorpe, 1987 Davis pers. comm., 1997 Mortenson and Thorpe, 1987 Patterson and Heaman, 1990 Loveridge et al., 1988

---PROTEROZOIC---

quartz-feldspar phryic flows, Pitz Formation	1760 Ma	LeCheminant et al., 1987b
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