The Neoarchaean tectonothermal evolution of the SE Nuuk region, Southern West Greenland

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Ph.D.

University of Edinburgh
2009
DECLARATION

Unless otherwise stated, this thesis is entirely my own work and has not been previously submitted for a degree at this or any other institute of learning.

Natasha R. Lee
ABSTRACT

THE NEOARCHAEOAN TECTONOTHERMAL EVOLUTION OF THE SE NUUK REGION, SOUTHERN WEST GREENLAND

The Nuuk region is dominated by complexly deformed Archaean TTG orthogneisses with subordinate supracrustal lithologies that have been variably metamorphosed at upper amphibolite to granulite facies. The region is interpreted as a mid- to lower crustal exposure of a Meso- to Neoarchaean terrane complex, collectively affected by high grade deformation and metamorphism between 2720 and 2650 Ma. New studies of the previously little-understood Kapisillit area, SE Nuuk region, suggest significant exposures of the recently-proposed Kapisillik terrane, a crustal block characterised by a distinct magmatic and metamorphic history between ca. 3100 Ma and 2650 Ma. A fuller understanding of the tectonic units within the Kapisillit area, including the extent and internal age-event relations of the Kapisillik terrane, thus has important implications for our understanding of the tectonothermal evolution of the Nuuk region.

Three field areas were selected in west, east and north Kapisillit (Norsanna, Tummeralik and Aputitooq Mountain respectively). Norsanna is situated at the boundary between the ca. 2825 Ma Tre Brødre and 3800-3300 Ma Færingehavn terranes whereas Tummeralik lies entirely within the proposed Kapisillik terrane. Aputitooq Mountain is situated within 2 km of the purported boundary zone between the Færingehavn and Kapisillik terranes. The tectonothermal evolutions of the three areas were investigated using field mapping, metamorphic petrology of pelitic assemblages and U/Pb zircon geochronology of key orthogneisses and structurally constrained leucosomes. P-T-t paths for each of these areas were constructed and considered in a regional context through integration with complementary geochronological, structural and metamorphic information obtained in parallel studies conducted both in the Kapisillit area and more broadly in the Nuuk region as a whole.

Metamorphic, structural and geochronological data suggest that Norsanna and Tummeralik share a polydeformational history during the Neoarchaean, between ≤2817 Ma and 2659 Ma. Early isoclinal folding (D₃) under kyanite zone (M1)
conditions gave way to peak garnet-sillimanite-biotite-plagioclase-quartz assemblages (M2). Subsequent fold limb attenuation and decompression (D2) at high temperatures (M3) occurred by ca. 2720Ma, constrained by high grade, syn-D2 leucosome. Pelites from Aputitooq Mountain follow a similar P-T path, but dating by other workers has yielded ca. 2650Ma ages from upper amphibolite facies leucosome and metamorphic rims on detrital zircon. Although an isotopic disturbance at 2650 Ma is recorded in a leucosome at Tummeralik, this is interpreted as a partial reset age instigated by a Proterozoic greenschist facies overprint, rather than a high grade metamorphic event.

The study suggests that Norsanna and Tummeralik (the Tre Brødre-Færingehavn and Kapsillik terranes respectively) were juxtaposed by ca. 2720Ma and that the high-grade, ca. 2650Ma event in the Kapsillik terrane becomes less significant moving south towards and across Ameralik Fjord. The ca. 2720 Ma tectonothermal event is recorded in most terranes/crustal units of the Nuuk region and therefore is still considered to represent final amalgamation. Subsequent metamorphism and deformation at ca. 2650 Ma is attributed to post-amalgamation reactivation, shearing and refolding focused along the previously established boundaries between juxtaposed blocks.
ACKNOWLEDGEMENTS

As I understand it, it is customary at the beginning of a thesis to say a big 'cheers' to the multitude of people/institutions that have played a part in the production of the ensuing work of...erm...art. So here we go.

I am indebted to the School of Geosciences at the University of Edinburgh for 1) giving me money (a Geosciences Teaching Scholarship to be exact), 2) giving me office space and access to lots of expensive and shiny instruments and 3) giving me a project in the first place. I must also extend my thanks to the logistical genius of the Geological Survey of Denmark and Greenland, who not only ensured I survived my two field seasons but also saw to it that I had something useful to say for myself at the end of it. They also let me play on their laser.

I am grateful to the small army of people for their help with various instruments – Peter Hill, David Steele and Chris Hayward on the electron probe, Nicola Cayzer and Emma Passmore on the SEM, Nicholas Odling in the XRF lab and Anders Scherstén on the laser at GEUS. In addition, thanks must also go to the Bedlam Outpatients in rooms 400, 401 and 344, whose escapades over the past four years have certainly never lacked in originality.

Greatly appreciated was the help from the Kapisillit mapping team at GEUS (and beyond), whose formal and informal chats during and after field seasons have been invaluable. Of the 20 or so team members I was privileged to work with, I am especially grateful to Venessa Bennett and Yvette Kuiper, who took pity on me when I couldn’t find any pelites and donated theirs instead.

Thanks must also go to my supervisors, Simon Harley, Nigel Kelly and Julie Hollis, for essentially being useful in just about every aspect of the thesis. And of course, a huge thanks goes to various family members (they had to be in here somewhere) for their unfailing support – my parents, who only mentioned the thesis once per conversation, my brother, who never mentioned it at all and my husband Dan, who has been through the whole thing himself and so was allowed to nag. Thanks guys, you’ve all been incredibly patient!
To my brother Ross, who introduced me to the Earth Sciences via the somewhat bizarre medium of woollen blankets...
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INTRODUCTION AND LITERATURE REVIEW

1.1 Introduction and project aims

The North Atlantic Craton in the Nuuk region, southern West Greenland, contains one of the World’s largest exposures of Archaean rocks, covering an area approximately the size of Switzerland (Nutman et al., 1996). The dominantly amphibolite- to granulite facies TTG orthogneisses and associated supracrustal units are thought to reflect the generation of continental crust at active margins throughout the Archaean (Bridgwater et al., 1974; Coe and Robertson, 1984; Garde et al., 2000; McGregor, 1973; Nutman et al., 1984; Nutman et al., 1996; Schiøttte et al., 1988; Talbot and Walton, 1973). These disparate blocks were then juxtaposed by interpreted subduction-accretion processes during the Meso- to Neoarchaean (Crowley, 2002; Friend and Nutman, 1991; Friend and Nutman, 2005b; Friend et al., 1996; Friend et al., 1987; Friend et al., 1988; Garde, 1997; Garde, 2003; Kalsbeek, 1986; McGregor et al., 1991; Nutman, 1991; Nutman et al., 2004a). Since the conception of the terrane theory in the Nuuk region, the identification of terranes has depended largely on the constraining of orthogneiss emplacement ages and metamorphic events through geochronology. Although this is an undeniably vital aspect to the development of the terrane theory, less emphasis has been placed on equally important characterisations of the structural and metamorphic history of individual blocks and the events that juxtaposed them. The large geochronological database currently in existence for the Nuuk region has permitted the identification of regional, high grade metamorphic events at ca. 3600, 3000, 2800, 2720 and 2650Ma and recent studies have begun to focus on their P-T-t evolution (Crowley, 2002; Friend and Nutman, 2001; Friend and Nutman, 2005a; Friend et al., 1988; Garde, 1997; Garde et al., 2000; Garde et al., 2007; Nutman and Collerson, 1991; Nutman and Friend, 2007).

This study focuses on the tectonothermal history preserved in pelitic rocks in the Kapisillit region, the SE quadrant of the Nuuk region that until recently has remained poorly-explored. Pelitic samples from three areas in the central and northern Kapisillit region were collected over two mapping seasons. Metamorphic
assemblage and textural analyses were conducted and integrated with field observations, bulk rock geochemistry, thermobarometry and geochronology of selected zircon and monazite samples to achieve the following aims:

1. Produce a 1:50,000 scale geological map of each selected field area, taking particular note of the area’s structural evolution. Where possible, identify the position and structural evolution of terrane boundaries in each of the selected field areas and integrate these findings with P-T evolution.

2. Characterise the metamorphic assemblage evolution of pelitic units in the three field areas using combined optical and scanning electron microscopy. *Samples from Aputitooq Mountain were donated by V. Bennett and Y. Kuiper.*

3. Attempt to constrain the pressure and temperature of metamorphism in the three areas using thermobarometric analysis of EPMA-quantified and X-ray mapping data.

4. Integrate metamorphic data with structural data from the three areas to produce a P-T-deformation evolution. *Structural and map data from Aputitooq Mountain was provided by V. Bennett and J. Heiss.*

5. Conduct U/Pb zircon geochronology of orthogneisses (in the Norsanna and Tummeralik field areas) to confirm the identity of the orthogneiss (and hence the terrane) hosting the pelitic samples. *Orthogneiss geochronology for Aputitooq Mountain is discussed by Bennett and Heiss (2006).*

6. Constrain the age(s) of metamorphism in the three areas using targeted sampling of leucosome material (U/Pb zircon) and metamorphic monazite populations in pelitic samples (Th-U-total Pb EPMA chemical dating). *For the Aputitooq Mountain field area, zircon ages were donated by V. Bennett.*

7. Integrate (and where possible, correlate) the P-T-t-deformation histories from the three areas into a tectonic model for the Kapisillit region and incorporate these findings into the evolutionary model for the Nuuk region.

The study was undertaken using a combination of field work and laboratory analysis at the University of Edinburgh and the Geological Survey of Denmark and Greenland (GEUS). Details of technique methodologies are given in Appendix A. Field work was undertaken in collaboration with the GEUS-run Kapisillit Mapsheet Project. Techniques using facilities at the University of Edinburgh included
petrological and textural analysis using optical microscopy and SEM imaging, mineral geochemistry and monazite U-Th-Pb chemical dating using EPMA and XRF major element bulk compositional analysis. U/Pb zircon geochronology was undertaken using the LA-SF-ICPMS facility at GEUS and the University of Copenhagen. Zircon imagery using BSE and CL was undertaken using the SEM facility at the University of Edinburgh.

1.2 The application of Phanerozoic plate tectonic processes to Archaean crust

The Archaean represents the part of Earth history from >4600 to 2500 Ma and is divided into the Eoarchaean (>3600Ma), Palaeoarchaean (3600-3200Ma), Mesoarchaean (3200-2800Ma) and Neoarchaean (2800-2500Ma) periods (Gradstein et al., 2004). Well-preserved Archaean rocks are only encountered in stable cratonic interiors. This is because Archaean cratons have typically remained stable since the Archaean or Proterozoic, as the majority of lithologies are high grade crystalline basement and are rheologically strong. This strength ultimately leads to strain focusing around craton margins, leaving the interior relatively unaffected by post-Archaean tectonism. As a result, cratonic areas are often bound by Proterozoic and/or Phanerozoic orogens or crustal collages (Hoffman, 1988; Windley, 1995) (Fig 1.1). When exposed, these cratonic interiors act as a window into the Archaean, permitting the study of early Earth crustal processes without needing to remove multiple phases of post-Archaean deformation and metamorphism.

There are several lines of evidence arguing that Phanerozoic-type plate tectonics did not operate at least during the Palaeoarchaean. Chief among these is the argument that a horizontal tectonic regime was not sustainable during early Earth history. Supporters of this hypothesis favour instead a vertical tectonic regime driven principally by heightened mantle plume activity (Marshak, 1999). This hypothesis has been invoked to explain the formation of km-scale basement domes and their corresponding supracrustal keels that are common in a number of Archaean cratons, for example the Pilbara, Dharwar cratons and Penokean Orogen (Fig 1.1) (Collins et al., 1998; Choukroune et al., 1995; Holm et al., 1998; Marshak et al., 1997).
Arguments advocating a vertical tectonic regime in the Archaean are largely based on the inference that global heat production and heat flux across the Moho were higher at this time (Martin, 1986). This has been attributed to higher (primordial) heat flux from the mantle and greater heat production in the crust as a result of higher concentrations of short-lived radio-isotopes (Collins et al., 1998; Marshak, 1999; Windley, 1995). The effect of increased heat fluxes on plate tectonic process would be seen throughout the plate tectonic cycle, from increased melt volume at mid-ocean ridges and the production of komatiitic magmas, to the inhibition of the blueschist facies stability field in subduction zones and the instigation of partial melting in subducted oceanic crust (Chavagnac, 2004; Drummond and Defant, 1990; Drummond et al., 1996; Kalsbeek, 2001; Ohta et al., 1995; Takahashi, 1990). Studies have also suggested that steeper geothermal gradients in collisional orogens raised the brittle-ductile transition to higher structural levels (Marshak, 1999). This leads to a greater proportion of thermally softened crust, allowing the large-scale doming of granitic plutons and/or sialic basement, with a corresponding subsidence of
surrounding, denser supracrustal material. The resulting structures are termed igneous (I)- or basement (B)-type dome-and-keel complexes (where I- or B- refers to the source of the dome material as new magma or crystalline, meta-igneous basement respectively). These structures have been variably interpreted as the product of partial convective overturn of the crust (Collins et al., 1998), reactivated high-angle thrusts (Choukroune, 1995), large scale crustal diapirism (Bouhallier et al., 1993; Shackleton, 1995) or as steep-sided, rotated decollément zones analogous to Phanerozoic metamorphic core complexes (Holm and Lux, 1996; Marshak, 1999; Marshak et al., 1997a; 1997b). Steeper geothermal gradients also argue that the brittle-ductile transition would occur at shallower crustal levels. This may partially account for the paucity of well-preserved fold-thrust belts in high-grade cratons, as the shallowest crustal levels, which would be most likely to preserve brittle structures, have typically been exhumed and eroded in Archaean orogens (Marshak 1999).

Despite the likely effects of steeper Archaean geothermal gradients on important crustal processes such as magma production and deformation style, it is generally agreed that some form of horizontal tectonics involving subduction did operate during the Archaean (Bridgwater et al., 1974; Moyen et al., 2006; Talbot and Walton, 1973), as this is the principal method of sialic crust production. A major characteristic of Archaean continental crust is the dominance of tonalite-trondjhemite-granodiorite (TTG) orthogneisses (sensu. Jahn et al, 1981), which are considered (with their associated (meta)-volcanic supracrustal sequences) to represent ancient analogues of volcanic arc systems (Condie, 1981; Smithies et al., 2003, 2007). The formation of these magmas has long been contested, due to their unusual chemistry (Kamber et al, 2002; Martin 1999), characterised by a K$_2$O-poor calc-alkaline suite, with Fe+Mg+Mn+Ti oxides and Na$_2$O values of <5%, K$_2$O/Na$_2$O <0.5 and with strong partitioning in REE favouring LREE. Also characteristic are negative Ti and Nb-Ta anomalies, with no anomaly present for Eu or Sr (Martin et al., 1983). Defant and Drummond (1990) suggested a model for tonalite-trondjhemite-dacite formation which involved the melting of the subducted oceanic slab. However this process could not produce the observed trace element profile characteristic of TTG magmas. Kamber et al (2002) recognised the importance of fluids in producing the correct trace element profile and suggested that some
distinction must be drawn between adakites (magmas with a slab-melt component) and TTG magmas. They instead suggested that TTG melts could be produced via the extensive fractional recrystallisation of hydrated mantle wedge. Studies by Smithies et al. (2003), however, found little evidence of mantle wedge influence in TTG terrains that are older than Mesoarchean. This led them to propose a mechanism of ‘flat-subduction’ (Fig. 1.2), which was the dominant crust producing process until at least the Mesoarchean. After this time arc magmas started to show increasing evidence of mantle wedge interaction.

Figure 1.2: Schematic cross sections comparing subduction processes at Phanerozoic and Archaean destructive margins. a) Production of adakite and related melts at a Phanerozoic margin due to low-angle subduction. b and c) Production of TTG melts at Archaean destructive margins during Archaean flat subduction. Adapted from Smithies et al. (2003).
The proposed cause of Archaean flat subduction was again attributed to elevated crustal geotherms, which had the combined effect of increasing both the production of oceanic crust and the spreading rates at ridges, producing large volumes of thick, hot oceanic crust. The resultant buoyancy inhibited subduction at active margins, in a similar way to modern-day low-angle subduction settings, leading to fluid-induced partial melting in the mafic slab rather than the mantle wedge (Smithies et al., 2003).

In recent years, the identification of modified subduction at Archaean active margins, and their sequential amalgamation with other crustal segments, has become a key area of study in the evolution of Archaean cratons. In particular, the process of terrane amalgamation is becoming an increasingly common interpretation for the tectonothermal evolution in a number of cratons; terrane models have been applied to numerous cratons, including the Limpopo Belt (Rollinson, 1993), the Lewisian Gneiss Complex (Kinny et al., 2005), the West Pilbara Block (Krapez and Eisenlohr, 1998) and the North Atlantic Craton in Canada and Greenland (Section 1.3).

The concept that disparate terranes may be tectonically juxtaposed along active margins was originally applied to the complex collage of fault bounded blocks that make up around 70% of the North American Cordillera (Fig. 1.3). Numerous multi-disciplinary studies during the 1970’s introduced the concept of ‘suspect terranes’ (Davis et al., 1978; Helwig, 1974; Jones et al., 1972; Monger et al., 1972; Wilson, 1968), which are defined as tectonically-bound blocks of crust with a different evolutionary history to that of adjacent blocks. Detailed palaeontological, palaeomagnetic, stratigraphic and structural studies identified more than 50 Palaeozoic to Holocene crustal segments in the North American Cordillera, many of which are exotic with respect to North America (Coney et al, 1980; Monger, 1997). These findings led to the suggestion that a major phase of intra-oceanic arc development and subsequent subduction-accretion was initiated during the Mesozoic in response to the development of the massive palaeo-Pacific Ocean (Coney et al., 1980).

In a Phanerozoic system, individual terranes and their sequence of amalgamation to other crustal blocks are identified according to a number of paleontological, stratigraphic, structural, chronological and geomagnetic criteria, which are summarised here:
1. A block may contain fossil species that are exotic to the accreted margin and/or to adjacent blocks.

2. Evidence from pristine palaeomagnetic data may vary between terranes and will usually be different to the magnetic signature of the accreting margin.

3. A terrane will have a discrete stratigraphic succession to adjacent terranes and the accreting margin.

4. A terrane will always be bound by faults (in the North American Cordillera these are typically thrust or transcurrent boundaries).

5. Ages of lithologies may vary between terranes. If two adjacent terranes have different ages, this will be apparent moving across the bounding fault.

6. The pre-accretion metamorphic and/or structural evolution may vary between blocks and early structures will be truncated by the terrane-bounding fault. Pre-accretion tectonothermal histories may also vary between terranes and are also truncated at the terrane-bounding fault.

7. The age of amalgamation between two or more blocks is defined by the youngest lithology that is common to all crustal segments.

8. Adjacent terranes will share a common stratigraphic, structural and metamorphic history after accretion.

The antiquity of the rocks and typically high degree of deformation in many Archaean regions prevents the use of detailed stratigraphy, palaeontology, and palaeomagnetism, and thus the identification of terranes depends largely on the interpretation of deformation and thermal events. Consequently the key criteria used in the identification of Archaean terranes are the protolith ages of TTG gneisses and supracrustal rocks, the timing (both relative and absolute) of high grade metamorphic events and the presence or absence of characteristic rocks, such as mafic dyke suites.

The application of the Phanerozoic terrane hypothesis to an Archaean tectonic system necessitates some form of subduction-accretion process at the accretionary margin during terrane amalgamation. Such a system necessitates the operation of a horizontal tectonic regime, such as those suggested in Bickle et al. (1980) and de Wit et al (1992). In terrains older than Mesoarchaean, the ‘Archaean Flat Subduction’ model of Smithies et al (2003) and Smithies et al (2007) may be cited, after which processes more akin to modern-style subduction would become increasingly significant.
Figure 1.3: The terrane collage of the North American Cordillera, which comprises >50 separate crustal units (denoted by black dots) separated from adjacent blocks by faults. Oblique subduction of the Pacific Plate under the North American plate has resulted in strain partitioning on a crustal scale, leading to the northward migration of accreted terranes along crustal scale strike slip faults. Figure adapted from Coney et al. (1980).

Despite the theoretical applicability of a Phanerozoic-type terrane hypothesis to Archaean active margins, complexities still remain in the identification and definition of individual crustal segments. The complexity of deformation and metamorphism in Archaean cratons may present a problem in distinguishing between terranes and constraining the early history of individual blocks. This is because later tectonothermal events may mask, alter or destroy protolith or early metamorphic relationships, resulting in controversial interpretations. An example of arguments arising from ambiguous field relationships involved the identification of biogenic
carbon signatures in quartz pyroxenite (interpreted as BIF) in Akilia, southern West Greenland (Mojzsis et al., 1996). Later work questioned the interpretation of BIF, preferring instead their formation by repeated metasomatism and deformation of an ultramafic protolith (Fedo and Whitehouse, 2002a). In addition, the ‘biogenic’ graphite grains were also reinterpreted as abiogenic carbon formed from the decomposition of siderite (Fedo and Whitehouse, 2002b; van Zuilen et al., 2002). This debate is still ongoing (Manning et al., 2006; McKeegan et al., 2007; Rosing, 1999; Whitehouse et al., 2009). The example above shows the importance of continual critical assessment and the application of multidisciplinary techniques in the interpretation of complexly deformed terrains.

A further complication involves the difference in exposure level between Archaean and contemporary terrane complexes, as structural levels in the mid- to lower crust are more commonly observed in Archaean cratons. By contrast, modern terrane complexes are characterised by upper crustal processes, such as large scale strike slip motions and the development of fold-thrust belts. Consequently it is often difficult to draw direct comparisons between Phanerozoic and Archaean terrane complexes.

1.3 Greenland and the North Atlantic Craton

The Laurentian shield forms a significant part of the basement of Canada, Scandinavia and Greenland and includes the oldest known exposure of TTG gneisses (Acasta gneisses) (Iizuke et al., 2006; Stern and Bleeker, 1998), as well as the world’s largest known exposure of Archaean rocks (Nuuk region, southern West Greenland) (Windley, 1995). The shield comprises a number of Archaean blocks, including the Superior, Rae, Slave, Hearne, Wyoming, Nain and Karelian cratons, that are thought to have amalgamated along a number of Proterozoic orogenic belts (Fig 1.1) (Hoffman, 1988). The North Atlantic Craton (NAC, also referred to in Canada as the Nain Province) is exposed in parts of northern Labrador, Greenland and northwest Scotland (Bridgwater et al., 1973) (Fig. 1.4). The majority of the craton is exposed in Greenland and consequently its evolution is best constrained here. The NAC is bordered by three orogens that formed between 1990 and 1800Ma (Table 1.1).
Figure 1.4: Detail of the Archaean Nain Province (NAC) with its Proterozoic bounding orogens (Rinkian-Nagssugtoqidian-Ammassalik, Torngat and Ketilidian orogens). A = Ammassalik mobile belt. Red lines denote Phanerozoic orogenic belts. Adapted from Hoffman (1988).

<table>
<thead>
<tr>
<th>Orogen</th>
<th>Age (Ma)</th>
<th>Details</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nagssugtoqidian</td>
<td>~1860-1840</td>
<td>Northern bounding orogens of the NAC.</td>
</tr>
<tr>
<td>Ammassalik</td>
<td></td>
<td>Extensions of the same belt. Also correlated with mobile belts in Scotland and Scandinavia.</td>
</tr>
<tr>
<td>Rinkian</td>
<td>1870-1850</td>
<td>Western bounding orogen of the NAC.</td>
</tr>
<tr>
<td>Torngat</td>
<td>1890-1800</td>
<td>Southern bounding orogen of the NAC. Long-lived Andean-type orogen.</td>
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</table>


The Nagssugtoqidian orogen forms the boundary between NAC and a lesser known Archaean block to the north (Connelly and Mengel, 2000; Kalsbeek, 1986) that has been recently correlated with the Rae Province of northern Canada (Connelly et al., 2006; St-Onge et al., 2007). It is an approximately ENE-trending belt comprising reworked Archaean granulites with smaller amounts of juvenile Proterozoic material (Kalsbeek, 2001) with an island arc affinity (Garde et al., 2007). The Ammassalik mobile belt is an ESE-trending Proterozoic orogen that is exposed in SE Greenland and is interpreted as an extension of the Nagssugtoqidian belt in West Greenland (Henriksen, 2000). Mason et al. (2004) propose that the
Nagssugtoqidian-Ammassalik belts have correlatives in Scotland and Scandinavia. This is based on correlations between the Ammassalik Intrusive Complex, South Harris Complex and Lapland-Kola Granulite Belt (in Greenland, Scottish Outer Hebrides and Baltica respectively). The Rinkian mobile belt is an E-trending Proterozoic fold belt that is exposed to the north of the Nagssugtoqidian orogen. Previous studies had considered the belt to be separate from the Nagssugtoqidian orogen but failed to identify a boundary between the two. Recent work has suggested that the Rinkian belt (which is contemporaneous with the Nagssugtoqidian), formed part of a single intercontinental collisional orogen (Connelly et al., 2006; Mengel et al., 1998). The Torngat orogen defines the western boundary of the NAC and is only exposed in Labrador. Early work by Hoffman (1988) suggested that the Torngat orogen formed during the accretion of the NAC and Rae Provinces between 2300 and 1650Ma. However recent work has identified an Archaean microcontinent in Labrador and has constrained tectonism to 1870-1850Ma (St-Onge et al., 2007). The southern margin of the NAC is bound by the Ketilidian orogen, which comprises high-grade (up to granulite facies) Proterozoic juvenile intrusive gneisses of granitic and metasedimentary affinity (Henriksen, 2000). The belt records a long history of Andean-type accretion along a north-dipping subduction zone (St-Onge et al., 2007).

The NAC itself is a high grade terrain, with the majority of lithologies highly deformed and metamorphosed to amphibolite or granulite facies (Windley, 1995). The craton contains a number of crustal segments, which are dominated by TTG-type orthogneisses with subordinate anorthosite and supracrustal lithologies. Some supracrustal successions, most notably those of Akilia Island and the Isua Greenstone Belt, have Eoarchaean ages. Supracrustal units of this age are also encountered as slivers hosted by Eoarchaean orthogneisses elsewhere in Godthåbsfjord. Orthogneisses in these crustal blocks record a range of magmatic ages from the Eoarchaean Amitsaq gneiss to the Neoarchaean Ikattoq gneiss, and which record numerous tectonothermal events in the Archaean and Proterozoic (section 1.4). The craton is also affected by Proterozoic and Tertiary dyke swarms and preserves Proterozoic and Mesozoic kimberlite, ultramafic lamprophyre and carbonatite provinces (Henriksen, 2000). Despite the post-Archaean thermal and tectonic activity detailed above, the central NAC in West Greenland is relatively unaffected by the events that caused so much deformation at its margins. Consequently the interior of
the craton is an ideal area in which to study crustal processes that operated during the Archaean.

The Nuuk region in the central NAC, southern West Greenland, has been interpreted as a Neoarchaean terrane collage involving a number of discrete crustal segments (Friend and Nutman, 2005b). Good exposure and close proximity to population centres has permitted detailed and repeated studies to be carried out in this area over a long period of time. Consequently the tectonothermal evolution of crustal blocks in the Nuuk region is better constrained than elsewhere in the NAC and represents a window through which we can investigate the tectonic processes involved in an Archaean terrane collage.

1.4 The geological evolution of the Nuuk region, southern West Greenland

The Archaean craton of the Nuuk region in southern West Greenland (Fig. 1.5) covers an area of approximately $3600 \text{ km}^2$ (Nutman et al., 1996) and contains meta-igneous and metasedimentary rocks spanning virtually the entire Archaean eon ($\geq 3850$-$2500\text{ Ma}$). The region is interpreted by some authors to comprise up to 6 terranes, which accreted before the end of the Archaean (Friend and Nutman, 2005b; Rollinson, 2003). All hypothesised terranes have been subject to at least one phase of high-grade metamorphism and commonly exhibit multiple phases of deformation. However the Nuuk region is thought to remain largely unaffected by regional post-Archaean tectonothermal events. A number of localities in the Nuuk region have also been controversially proposed as sites for the preservation of earliest life (McKeegan et al., 2007; Mojzsis et al., 1996; Mojzsis and Harrison, 2002; Monster et al., 1979; Myers and Crowley, 2000; Schidlowski et al., 1979; Rosing, 1999; Rosing and Frei, 2004; van Zuillen et al., 2002). Further studies of the region may consequently have important implications for our understanding of Archaean surface process and the emergence of life.
As a province of Denmark, townships and other geographical locations in Greenland may have a Danish and Greenlandic name, either (or both) of which may be referred to in the published literature. Recent revisions of Greenlandic spellings have also resulted in nomenclature complications in the literature, especially where themes, such as the terrane hypothesis, have been continued over a long period of time. Table 1.2 gives a list of current terms and spellings used in this thesis, and gives the old and/or abandoned equivalent terms that may be encountered in the wider literature.

The identification criteria for Phanerozoic terrane complexes detailed in section 1.2 are largely inapplicable to the highly deformed and metamorphosed lithologies of the Nuuk region. Each terrane is characterised by a specific TTG orthogneiss suite, as it is these units that form the dominant lithology (typically >80%) in the Nuuk region. In the field, a preliminary identification of these units can be made by observing the intensity, character and history of deformation, the presence or absence of mafic intrusives and in some cases, the metamorphic mineral assemblage.
Table 1.2: A list of commonly used place names and lithologies that have been redefined or whose spelling has altered due to changes in orthography. All current terms listed here are encountered during this work but in older literature are used under a different spelling or definition.

<table>
<thead>
<tr>
<th>Current spelling/definition</th>
<th>Old spelling</th>
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<tr>
<td>Nuuk gneiss (Akia terrane only)</td>
<td>Nûk gneiss (All orthogneisses without deformed mafic dyke suite)</td>
</tr>
<tr>
<td>Nuuk (town)</td>
<td>Godhâb/ Nûk</td>
</tr>
<tr>
<td>Amîtsøq gneiss (Færingehavn terrane only)</td>
<td>Amîtsøq gneiss (TTG orthogneiss with deformed mafic dyke suite)</td>
</tr>
<tr>
<td>Kapisillit (town)</td>
<td>Kapisigdlit</td>
</tr>
<tr>
<td>Kapisillik (terrane)</td>
<td>-</td>
</tr>
<tr>
<td>Ikkattoq gneiss</td>
<td>Íkátoq gneiss</td>
</tr>
<tr>
<td>Akulleq terrane</td>
<td>Akugdleq terrane</td>
</tr>
<tr>
<td>Itilleq (fjord)</td>
<td>Itivdleq</td>
</tr>
<tr>
<td>Qarlil nunaat</td>
<td>Qardlit nunât</td>
</tr>
<tr>
<td>Tasiussarsuaq</td>
<td>Tasiusarssuaq</td>
</tr>
<tr>
<td>-</td>
<td>Itsaq gneiss (abandoned term for all Archaean lithologies in the Nuuk region)</td>
</tr>
<tr>
<td>Eoarchaean supracrustal rocks</td>
<td>Akilia Association</td>
</tr>
<tr>
<td>Meso- to Neoarchaean supracrustal rocks</td>
<td>Malene Supracrustals</td>
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For example, the Eoarchaean Amîtsøq gneiss exhibits a well-defined, multiply-folded gneissosity with multiple generations of felsic veining. The unit has also been intruded by the Ameralik dykes (Gill, 1976; Gill and Bridgwater, 1976), a suite of mafic intrusives that have themselves been metamorphosed and strung into parallelism with the host rock gneissosity. This forms a marked contrast with the latest Mesoarchaean Ikkattoq gneiss, which shows an apparently less-complicated
gneissic fabric and has not been intruded by an early mafic suite. The obviously
different state of strain, deformation and intrusive history of these two lithologies
thus allows them to be (relatively) easily distinguished in the field. However it is
stressed that field identifications of high grade lithologies (particularly orthogneisses)
should always be confirmed by geochronology, as factors such as the state of strain
or the exposed area of an intrusive suite may vary significantly within a single
orthogneiss unit.

Each terrane has principally been characterised by distinct U/Pb zircon age
populations that may reflect an age of magmatic crystallisation, a period
of metamorphism or an inherited component in an orthogneiss suite. Whole rock
isotopic dating has been employed to discriminate between the suites, including Rb-
Sr, Pb-Pb and Sm-Nd (Black et al., 1971; Moorbath, 1975; Moorbath et al., 1986;
Nutman et al., 2007). Geochemical studies have also been conducted to ascertain the
influence of crustal contamination in the generation of TTG protoliths in the Nuuk
region (Bridgwater et al., 1974; Coe and Robertson, 1984; Kamber and Moorbath,
1998; Nielsen et al., 2002).

U/Pb geochronology studies of zircon have been of critical importance in the
development of the terrane hypothesis in the Nuuk region. Key works in this area
have been based primarily on SIMS geochronology of the dominant orthogneiss
lithologies (Friend and Nutman, 2005b; Friend et al., 1996; Nelson, 1997; Nutman et
al., 1999; Nutman and Collerson, 1991; Nutman et al., 1993). Ages of metamorphism
specific to certain terranes have also been identified (Crowley, 2002; Nutman et al.,
2004a). Targeted sampling of migmatite, pegmatite and leucosome material has been
employed in a number of studies, in order to identify terrane-specific or -common
metamorphic events. Consequently there is a large database of U/Pb zircon data that
covers much of the Nuuk region, specifically along the west coast in the Tre Brødre-
Færingehavn-Outer Ameralik area (see Fig 1.6), which was one of the first areas to
be interpreted in terms of the terrane hypothesis (Friend et al., 1987; Friend et al.,
1988). The area around the Isua Greenstone Belt has also been extensively studied,
as the supracrustal sequences preserve good examples of Eoarchaean-age
sedimentary and volcanic processes (Appel et al., 1998; Appel et al., 2001; Bolhar et
al., 2004; Crowley, 2003; Furnes et al., 2007; Myers, 2001; Nutman and Collerson,
Rocks with TTG chemistries form the dominant lithologies in the NAC in southern West Greenland. These suites, intruded as juvenile crust (Bridgwater et al., 1974; Coe and Robertson, 1984) between 3850Ma and 2820Ma (Friend and Nutman, 2005b), were then subject to multiple phases of deformation and variably metamorphosed. Early theories of the structural evolution of the craton described multiple periods of horizontal thickening in a single crustal block. This was based on studies by Bridgwater et al., (1974) and Coe and Robertson, (1984) who envisaged a ‘subduction-like’ setting, where ≥3850Ma supracrustal units were progressively intruded by juvenile material in an island arc setting (Garde et al, 2007). The Nuuk region was interpreted to comprise two principal orthogneiss suites; the Amîtsøq and Nûk gneisses. These were initially distinguished in the field by the presence of metamorphosed and deformed mafic dykes, termed the Ameralik dykes (Gill, 1976; McGregor, 1973). Orthogneisses containing the Ameralik dykes were interpreted to belong to the Eoarchaean Amîtsøq gneiss, whereas orthogneisses without a deformed intrusive suite were deemed to belong to the younger, less deformed Nûk gneiss. However, the identification of deformed mafic intrusives in Mesoarchaean orthogneisses (Friend and Nutman, 2001) negated this as a distinctive method for identifying the Amîtsøq and Nûk gneisses in the field. Consequently a re-evaluation of orthogneiss identification criteria was required, after which the term ‘Nûk gneiss’ was restricted to Mesoarchaean rocks north of the Ivinnguit fault, in what is now the Akia terrane (Nutman et al., 1989). Prior to the terrane theory, supracrustal rocks in the Nuuk region were also divided into two broad age groups. These were the Akilia Association, which were intruded by the Amîtsøq gneiss, and the younger Malene Supracrustals, which were argued to preserve an unconformable contact with the basement Nûk gneisses (Bridgwater et al., 1974; McGregor, 1973).

The concept of multiple suspect terranes in the Nuuk region was proposed by Friend et al., (1987). This resulted from the observation that different parts of the craton were composed of rocks with contrasting protolithic ages and which showed markedly different evolutionary histories. Table 1.3 describes the six known terranes and summarises the definitive events in each terrane. The terrane hypothesis initially identified three discrete tectonic blocks that were separated by high grade mylonite zones (section 1.5). These blocks, the Færingehavn, Tre Brødre and Tasiusarsuaq terranes were defined by Friend et al. (1987), with a fourth, the Akia terrane, added
by McGregor et al., (1991). The Akia and Tasiusarsuaq terranes are large crustal bodies, which form the northern and southern bounding blocks to the Nuuk region terrane complex. Although some work has identified discrete crustal blocks within the Akia terrane (Nutman et al., 2004a) and to the south of the Tasiusarsuaq terrane (Friend and Nutman, 2001), the extent and evolution of the terrane complex outside of the Nuuk region remains little-known.


<table>
<thead>
<tr>
<th>Terrane</th>
<th>Definitive Orthogneiss</th>
<th>Emplacement age (Ma)</th>
<th>Tectonothermal events (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Færingehavn</td>
<td>Amîtsoq gneiss</td>
<td>≥3850-3650</td>
<td>3650-3600 (granulite facies)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>3560-3200 (Ameralik dykes)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2720 (amphibolite facies)</td>
</tr>
<tr>
<td>Isukasia</td>
<td>Northern/Southern gneiss</td>
<td>Ca. 3800-3600</td>
<td>&gt;3650 (amphibolite facies)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>3560-3200 (Ameralik dykes)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2950 (zircon record)</td>
</tr>
<tr>
<td>Akia</td>
<td>Nuuk gneiss</td>
<td>3200-3000</td>
<td>2990-2970 (granulite facies)</td>
</tr>
<tr>
<td>Kapisillik</td>
<td>Kapisillik gneiss (unofficial term)</td>
<td>Ca. 3000</td>
<td>Ca. 2950 (zircon record)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2650 (amphibolite-granulite)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2635 (amphibolite facies)</td>
</tr>
<tr>
<td>Tre Brødre</td>
<td>Ikkattoq gneiss</td>
<td>Ca. 2825</td>
<td>2720 (amphibolite facies)</td>
</tr>
<tr>
<td>Tasiusarsuaq</td>
<td>Tasiusarsuaq gneiss (unofficial term)</td>
<td>3100-2840</td>
<td>2795 (granulite facies)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>2720 (amphibolite retrogression)</td>
</tr>
</tbody>
</table>

The Færingehavn and Tre Brødre terranes were interpreted to crop out in the central Nuuk region (Friend et al., 1987; Nutman et al., 1989). The Færingehavn terrane is characterised by the Eo- to Palaeoarchaean Amîtsoq gneiss with ‘Akilia Association’ supracrustal units (McGregor, 1973). These are both intruded by the Ameralik dyke swarms. The ca. 2825Ma Ikkattoq gneiss of the Tre Brødre terrane is the youngest orthogneiss suite recognised in the Nuuk region and underwent peak metamorphism at 2720Ma (McGregor et al., 1991; Nutman, 1991). The boundary between the Amîtsoq and Ikkattoq orthogneisses is highly tectonised throughout the Nuuk region, hence the interpretation that the units represent two tectonically juxtaposed crustal segments. This relationship was questioned by McGregor et al.
(1991), who argued that intrusive contacts are preserved at certain localities in the Kobbefjord, Storø and outer Ameralik areas described by McGregor (1973). This led to the incorporation of the Tre Brødre and Faeringehavn terranes into the composite Akulleq terrane (Crowley, 2002; Friend and Nutman, 1991; Friend et al., 1996; Garde, 2003; McGregor et al., 1991; Nutman, 1991) which amalgamated after 2800Ma (Fig. 1.6).

Abbreviations:

A = Akilia
F = Faeringehavn
I = Isua
K = Kobbefjord
OA = Outer Ameralik
S = Storø
T = Tre Brødre

Figure 1.6: The three-terrane model of McGregor et al. (1991) showing the Akia terrane in the north (orange) and Tasiussarsuaq terrane in the south (blue) with the central Akulleq terrane (green) formed of the Tre Brødre and Faeringehavn sub-terrane. Figure adapted from Henriksen et al. (2000).

As more studies are conducted on the rocks of the Nuuk region, the theories of evolution grow more complex. An example of this is the addition of the Kapisillik and Isukasia terranes to the pre-existing Akia, Faeringehavn, Tre Brødre and Tasiussarsuaq terranes by Friend and Nutman (2005) (Fig. 1.7). A notable problem with the advent of the Kapsillik terrane is that many of the proposed outcrop areas, especially in the SE Nuuk region, lie in areas previously interpreted as Tre Brødre terrane. This has the effect of drastically reducing the outcrop area of the Tre Brødre terrane as well as complicating the outcrop pattern of the ca. 2825Ma Ikkattoq gneiss suite, calling into question the affinity of the Tre Brødre terrane. Nutman and Friend (2007) introduce the possibility that the terrane represents the deeply eroded hanging
wall of a low angle detachment zone akin to that of modern day metamorphic core complexes, and suggest that this may account for the complex outcrop patterns between the Tre Brødre and Færingehavn blocks. However until this hypothesis can be substantiated, the Tre Brødre terrane continues to be classed as a suspect terrane in its own right.

Figure 1.7: Terrane map of the Nuuk region after Friend and Nutman (2005) and Nutman and Friend (2007) and showing. Question marks denote uncertainties in the position of key tectonic boundaries or areas of unknown orthogneiss age. Also shown are key place names that are mentioned during this work, in addition to the positions of the three field areas in the southeast quadrant of the Nuuk region (Kapisillit Mapsheet).
A précis of the current understanding of the geological history in the central Nuuk region is illustrated in Table 1.4. The Amîtoq gneisses (and corresponding units in the Isukasia terrane) were intruded into \( \geq 3850\)Ma supracrustal rocks, formerly the Akilia Association, between 3850 and 3600Ma, however the interpreted intrusion history of the Amîtoq gneisses remains controversial. Some propose periodic intrusion of TTG suites over a period of \( \sim 300\)Ma (Friend and Nutman, 2005), based on multiple age populations of magmatic zircon between 3850 and 3600Ma. Others argue for a single intrusion event at ca. 3650Ma (Kamber and Moorbath, 1998; Whitehouse and Kamber, 2005), which subsequently underwent granulite facies metamorphism at ca. 3600Ma (Friend, 2005a). This scenario, the \( >3650\)Ma zircon component in the Amîtoq gneiss is interpreted as inherited magmatic zircon from an older suite that is not exposed at the surface.

The Færingehavn and Isukasia terranes have been correlated with the Uivak Gneiss Complex (Collerson and Bridgwater, 1979) in the Saglek-Hebron block of Labrador (Friend and Nutman, 1994). Potential correlations are also suggested with the Qarliit Tasersuat assemblage and Aisivik terrane south of the Nagssugtoqidian orogen (Nutman et al., 2004a) as all Palaeoarchaean assemblages contain evidence of high grade metamorphism at ca. 3600Ma. The intrusion of the Ameralik dyke swarms at 3500-3450Ma and 3260Ma (Nutman et al., 2004b) in the Færingehavn and Isukasia terranes have been correlated with the Saglek dykes of Labrador (Gill and Bridgwater, 1976) and mafic dykes in the Qarliit Tasersuat assemblage in the Aisivik block (Nutman et al. 2004). This also suggests that the Palaeoarchaean blocks in the NAC share a common history until the Mesoarchaean. The intrusion of the metabasic Ameralik dykes and their correlatives (Gill, 1976; Gill and Bridgwater, 1976) have been interpreted by Friend and Nutman (2005) as a period of continental rifting in an Eoarchaean continent, forming the Isukasia, Færingehavn and Aisivik terranes in Greenland.

The intrusion of the Nuuk gneiss (Akia terrane) in an island arc setting between 3200 and 3000Ma (Garde et al., 2007), with subsequent granulite facies metamorphism at 2990-2970Ma have been interpreted to represent the accretion of the Isukasia terrane and Akia terranes at 2991 \( \pm 2\) Ma by (Hanmer et al., 2002). This counters interpretations by Friend et al. (1996) and is refuted by Friend and Nutman.
<table>
<thead>
<tr>
<th>Paper</th>
<th>Event</th>
<th>Age (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nutman et al., (1997a)</td>
<td>Deposition of Isua and 'Akilia Association' supracrustal rocks in a subaqueous setting. Preserved seawater-altered pillow basalts and BIF.</td>
<td>&gt;3850</td>
</tr>
<tr>
<td>Appel et al., (1998, 2001)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Nutman et al., (1996)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Friend and Nutman, (2005)</td>
<td>Intrusion of Amitsq protoliths.</td>
<td>ca. 3850-3650</td>
</tr>
<tr>
<td>Kamber and Moorbath, (2002)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kamber and Moorbath, (2002)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Gill and Bridgewater (1976)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Nutman et al., (1996)</td>
<td>Intrusion of TTG suites in Akia and Kapisillik terranes</td>
<td>ca.3200</td>
</tr>
<tr>
<td>Friend and Nutman, (2005)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Garde et al., (2007)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>McGregor et al., (1991)</td>
<td>Intrusion of Tasiusarsuaq gneiss Protoliths.</td>
<td>2920-2800</td>
</tr>
<tr>
<td>Friend and Nutman, (2001)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Nutman, (1991)</td>
<td>Intrusion of TTG suite in Tre Brødre terrane.</td>
<td>ca. 2825</td>
</tr>
<tr>
<td>Hanmer et al., (2002)</td>
<td></td>
<td>&gt;2720</td>
</tr>
<tr>
<td>Crowley, (2002)</td>
<td>Granulite facies metamorphism in Tasiusarsuaq terrane.</td>
<td>ca. 2795</td>
</tr>
<tr>
<td>Crowley, (2002)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>McGregor et al., (1991)</td>
<td>Final terrane amalgamation event(s). Emplacement of granitoid sheets (Qarusuk dykes) in all terranes.</td>
<td>2720</td>
</tr>
<tr>
<td>Friend et al., (1996)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Nutman and Friend, (2005)</td>
<td>High pressure granulite facies metamorphism in Kapisillik terrane.</td>
<td>ca. 2650</td>
</tr>
<tr>
<td>Friend and Nutman (2007)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Nutman et al., (2007)</td>
<td>Gold mineralisation on Storø Shear Zone, Kapisillik terrane.</td>
<td>ca. 2635</td>
</tr>
<tr>
<td>Nutman and Friend, (2007)</td>
<td>Earliest Qørqut granite-type intrusions. Qørqut granite (s-type) intruded.</td>
<td>2580, 2500</td>
</tr>
<tr>
<td>McGregor et al., (1991)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
(2005), who propose that the Akia block did not form part of the Nuuk region terrane collage until the Neoarchaean. Instead, Friend and Nutman (2005) argue that the first documented tectono-thermal event between two known terranes in the Nuuk region involved the Kapisillik and Isukasia terranes. This is based on evidence of upper amphibolite facies metamorphism at ca. 2960Ma in both terranes (Friend and Nutman, 2005b; Nutman et al., 2007; Nutman and Friend, 2007). The next significant event involved the juxtaposition of the Isukasia and Kapisillik terranes against the composite Akulleq (Tre Brødre-Færingehavn) terrane. This is inferred to have taken place at ca. 2800Ma, after the intrusion of the ca. 2825Ma Ikkattoq gneiss (Friend and Nutman, 2005b; Polat et al., 2007). However there is little evidence of metamorphism or deformation in either terrane around this time.

The final terrane amalgamation event is argued to have taken place by ca. 2720 Ma, and is correlated with amphibolite facies metamorphism in the Ikkattoq gneiss and contemporaneous retrogression at amphibolite facies in the 2920-2800 Ma Tasiusarsuaq gneisses (McGregor et al., 1991). The metamorphism in this region was attributed to the thrusting of the Tasiusarsuaq granulites over the Ikkattoq gneisses (Fig 1.8) along the Qarliit Nuunat thrust (Crowley, 2002; McGregor et al., 1991). This is followed by the intrusion of the Qârusuk granitic dykes at 2720 Ma, which are common to all terranes (Appel et al., 2003).

Although the final collisional event in the Nuuk region is interpreted at ca. 2720Ma, the Kapisillik terrane exhibits upper amphibolite to granulite facies metamorphism at ca. 2650Ma (Friend and Nutman, 2005) and apparently does not preserve evidence of a 2720Ma event. However the 2650Ma event does not appear to be significant in the neighbouring Isukasia or Tre Brødre terranes. This non-record of 2650Ma overprints in the Ivisaartoq Supracrustal Belt (at the Kapisillik-Isukasia boundary) has led to the suggestion that the Kapisillik terrane underlay both the Isukasia and Tre Brødre terranes at this time (Nutman et al., 2007; Nutman and Friend, 2007). Upper amphibolite facies metamorphism at 2650Ma is widespread in the Kapisillik terrane, and is associated with gold mineralisation along the Storø Shear Zone, a high-strain zone on the island of Storø to the northeast of Nuuk (Fig 1.7) (Nutman et al., 2007). In addition to high grade metamorphism at 2650Ma, Nutman and Friend (2007) suggest that tectonism associated with accretion
continued until the intrusion of the largely post-tectonic Qorqût granite between 2550 and 2500Ma (Friend et al., 1985; Friend et al., 1987; McGregor et al., 1991).

Figure 1.8: (Top) Block diagram (adapted from McGregor et al., 1991) of the tectonic environment in the Nuuk region at ca. 2720Ma and its modern-day outcrop pattern with field areas shown as red stars (bottom). At ca. 2720Ma, amphibolite facies metamorphism in the Færingehavn/Tre Brødre blocks (shown here as the Akulleq terrane diagram of Henriksen (2000) for simplicity), is instigated by the northward thrusting of the granulite facies Tasiusarsuaq terrane along the Qarliit Nunaat Thrust. The Tasiusarsuaq block experiences simultaneous amphibolite facies retrogression. By contrast, the Akia terrane amalgamation is driven predominantly by strike-slip motion (McGregor et al., 1991).

The Nuuk region has been affected by a number of post-Archaean tectonothermal events, including the intrusion of large picritic to doleritic dykes. These are correlated with the 2.5-1.9 Ga North Atlantic Dyke Swarm that affects much of West Greenland, Canada and Scotland (Bridgwater et al., 1995; Escher et al., 1995). Multiple phases of faulting at low metamorphic grade are also recorded, the most notable of which are the ca. 1800Ma Ivinguit and Kobbefjord faults (Hollis et al., 2004). These rework the high grade southern boundary of the Akia terrane on Bjørneøen and Sermitsiaq Islands (Kelly and Frei, 2004), forming steep, greenschist facies mylonitic and brittle fabrics in the immediate vicinity of the faults.
However no regional post-Archaean metamorphic events have yet been recognised at outcrop scale.

The précis above gives one interpretation of the Nuuk region evolution, however full characterisation of the tectono-thermal history of the region has, until recently, been hampered by the almost sole use of geochronology as a terrane indicator and complicated by contrary interpretations of the data. This is illustrated by contrasting models for the timing of accretion in the Isukasia terrane described above (Friend and Nutman, 2005b; Hanmer et al., 2002). Recent studies have started using alternative geochronological methods, such as monazite and rutile dating (Nutman and Friend 2007, Nutman et al. 2004) in addition to U/Pb zircon dating by LA-ICPMS. Such techniques are useful not only in supporting pre-existing interpretations from earlier geochronological studies, but may also begin to provide more accurate constraints on timing of metamorphism. Monazite chemical dating has particularly significant implications as careful analysis can place constraints on a particular point on the P-T curve. Although some works have used thermobarometry to gain an approximate understanding of P-T conditions of metamorphism in the different terranes (Nutman et al. 1996, Nutman and Friend 2007), little detailed petrographic work has been undertaken to quantify and characterise the P-T history for the individual blocks. This work aims to provide an integrated study of the metamorphic evolution of key areas in the Kapisillit region using combined structural and metamorphic mapping, detailed petrology and petrography, mineral geochemistry and U/Pb zircon dating using LA-ICPMS.

1.5 Terrane boundaries in the Nuuk region

The bounding faults that are a pre-requisite for identifying terranes in Phanerozoic terrane complexes are much more difficult to identify in high grade Archaean terranes, of which the Nuuk region is a prime example. In this area, the terrane bounding faults first were characterised by McGregor et al (1991) as mylonitic zones with a structural thickness between 10 and 50m, which were often refolded by later events and have been recrystallised in places (Friend, Pers. Comm). Detailed mapping has identified the positions of the terrane boundary regions over large parts of the Nuuk region, most notably in the extensively studied areas of south
of Nuuk between Tre Brødre and Færingehavn (Friend et al., 1987) (Fig 1.7). Other important boundaries are the Qarliit Nunaat Thrust, identified and characterised by Rollinson (2002) as the northern boundary of the Tasiusarsuaq terrane (Fig 1.8) and the Ivinnguit Fault, a brittle reworking of an older structure that forms the southern boundary of the Akia terrane (Kelly and Frei, 2004).

Although fairly well defined along coastal areas, these terrane-bounding mylonite zones become difficult to trace inland, where the extent of exposure decreases and access becomes more problematic. This is exemplified by the Qarlitt Nunaat Thrust in the southeast Nuuk region (Kapisillit Mapsheet), where the trace of the boundary is assumed to follow the Kangeriuerarsunngup Tasersua hydrolake (Appendix D; Friend and Nutman, 2005b). However recent mapping of this region suggests that the boundary does in fact crop out north of this lake and may even extend as far north as Ameralla Fjord (Kapisillit Mapsheet, unpublished GEUS report). A key issue regarding our understanding of Nuuk region evolution is thus dependent on being able to identify and follow these boundaries across the whole of the region. A further complication to be considered is the function of the terrane bounding mylonites, whose origins are rarely, if ever, considered in the majority of literature on the subject. A 10-50m wide mylonite zone would suggest a mid- to upper crustal origin. However the high metamorphic grade (upper amphibolite to granulite facies) associated with the suggested terrane amalgamation events, would more likely be associated with large-scale zones of ductile deformation and crustal interleaving. The possibility must therefore be considered that the 10-50m wide mylonite zones of McGregor et al (1991) do not represent the original collisional boundaries, but instead formed at a later stage of the terrane collage evolution. It is thus possible that two types of equally valid terrane boundary exist in the Nuuk region, both of which are characterised below:

1. High grade terrane boundaries characterised by regionally significant ductile high strain zones with structural thicknesses on the scale of >100m. Such zones may display any/all of isoclinal or rootless folds, shear folds and boudinage and will exhibit a regional gneissosity that transposes older structures. The boundary may also preserve evidence of thrust faulting, large scale tectonic interleaving or regionally significant strain partitioning,
such as that recorded by Kelly and Frei (2004) on the island of Bjørneøen in Godthåbsfjord.

2. Lower grade mylonite zones with a structural thickness \( \leq 50 \text{m} \). Zones are mica-rich, with recrystallisation responsible for an increase in grain size from typical mylonitic textures. In some cases, the only evidence of former mylonitic textures is relict plagioclase stringers, signifying blastomylonite development. Such textures are recorded on Kapisillit Fjord by Friend (Pers. Comm).

The above descriptions give an indication of the types of terrane boundaries that may be present in the Nuuk region, however it is emphasised that the same structures may also be formed in the absence of terrane amalgamation. It is also compatible with the terrane theory of Coney et al (1980) to refer to any tectonised boundary between rocks of two terranes as a terrane boundary, even if the boundary in question refers to a small scale imbricate or interleaved slice. Consequently no assumption is made about the nature of any tectonic boundary encountered during this project, without alluding to accompanying geochronological evidence and considering the published literature.

1.6 The SE Nuuk region: The Kapisillit Mapsheet Project

Currently the most detailed published geological map of the Kapisillit region is the 1:500,000 Frederikshåb Isblink – Søndre Strømfjord (Allart, 1982) (Fig. 1.9). This map not only predates the terrane theory and much of the current geochronological knowledge of the area, but by necessity is grossly over-simplified, identifying only the Amitsoq and Nûk orthogneisses as the dominant orthogneiss lithologies. In addition, the 1:500,000 map sheet shows a significantly overestimated amphibolite outcrop. This is especially the case in the Kapisillit region, which predicts large bodies of amphibolitic lithologies south of Ameralik Fjord (Fig. 1.9 inset). The contrast in outcrop patterns can be seen by comparing the Allart (1982) map and the new 1:100,000 scale Kapisillit map in Appendix D.
Key (for SE Nuuk region):

- Khaki = Palaeoarchaean TTG orthogneiss (Amitsoq gneiss)
- Beige = Meso-Neoarchaean TTG orthogneiss (Nūk gneiss)
- Orange = Granulite facies orthogneiss
- Dark green = Amphibolite
- Dark grey = Anorthosite
- Brown = Metasediment
- Pink = Neoarchaean Qôrqut Granite
- Pale grey = Quaternary deposits

Coloured stars denote the field areas investigated during this study; Norsanna (red), Tummeralik (blue) and Aputitooq Mountain (yellow).

Figure 1.9: 1:500,000 geological map of the Nuuk region, southern West Greenland (Allart, 1982). Inset: Geological sketch map of the Kapisillit area (SE Nuuk region) enlarged from the 1:500,000 scale Frederikshåb Isblink – Søndre Strømfjord map of Allart (1982).
Geological maps of the Nuuk region exist at 1:100,000 scale in three of the four quadrants - Nordlandet, Qôrqut and Ivisaartoq. The fourth area, Kapisillit, covers the southeast quarter of the Nuuk region, and is currently being mapped at this scale by GEUS. The area is key to current investigations concerning the evolutionary history of the Kapsillik terrane, as the map area covers much of what is currently inferred to be the southern Kapsillik terrane (Fig 1.7). The lack of data from this area, due in part to logistical constraints on previous field projects may also be a factor in the apparent absence of 2720Ma metamorphic ages in the Kapsillik terrane (see section 1.4). A greater understanding of the geology of the Kapisillit region thus has significant implications for our understanding of the tectonothermal evolution of the Nuuk region as a whole.

1.7 The selection of field areas in the Kapisillit region

In order to address the above issues, three study areas were selected (Fig 1.9 inset); two were visited by the author and mapped at 1:50,000 scale (Norsanna and Tummeralik) and the third (Aputitooq Mountain) was investigated using field observations and samples donated by V Bennett and J Heiss. Due to limited pre-existing knowledge of the Kapisillit region as a whole, field areas were initially chosen based on studies of the 1:500,000 geological map of Ahart (1982) and will focus on petrology and geochemistry of low variance metapelitic assemblages. However regional mapping has shown outcrops of pelitic material to be small and typically below the resolution of the 1:500,000 map. Consequently, areas were selected that showed large potential outcrops of amphibolitic material, under the working hypothesis that metapelitic rocks may be spatially associated with supracrustal amphibolite. The occurrence of metapelitic assemblages in the Nuuk region has also been attributed (in some cases) to the metasomatic alteration of a metavolcanic protolith. This was first suggested by Garde et al (2007), who proposed that metapelitic assemblages on the Qussuk Peninsula (of the Akia terrane) were the result of pre-metamorphic alteration of andesitic tuffs in a Mesoarchaean arc system. The study described 'interfingered' outcrops of alternating 'pelitic' and 'tuffaceous' material, both of which preserve trace element profiles compatible with an andesitic to mafic protolith. Similar 'interfingered' outcrops of amphibolite and pelite have
also been observed in the recently mapped Kapisillit region (Bennett and Hollis, Pers. Comm.). Consequently the term ‘metasediment’ is not used to describe the garnet-sillimanite-bearing gneisses in this project unless further evidence of a sedimentary origin is present. Similarly, the term ‘pelite’ refers to the mineral composition of a rock and makes no assumption of the protolith material. For the purposes of this project, the protolith of the pelitic assemblages in the Kapisillit region is not considered in detail, as the end-product lithologies preserve classic high grade pelitic mineral assemblages, which are the main focus of the project.

The position of field areas relative to suspected terrane boundaries was also taken into consideration during the selection of study areas. This has a twofold advantage; firstly, it presents an opportunity to fully characterise the tectonic evolution of a Nuuk region terrane boundary. Secondly, previous studies have shown that such areas may contain significant amounts of interleaved supracrustal material (Windley, 1995). Of the half a dozen areas mapped by the author in the central and southern Kapisillit region over two summer field seasons, only Norsanna and Tummeralik contained appreciable supracrustal material with metapelitic sequences and so became the main focuses of the study.

Norsanna, at the western edge of the central Kapisillit mapsheet, is one of the few parts of the Kapisillit region to have been incorporated into the terrane theory. The site is arguably situated on an isoclinally folded and overturned portion of the Færingehavn-Tre Brødre boundary (Nutman and Friend, 2007). The location of this area allows for a structural and metamorphic characterisation of a Nuuk region ‘terrane boundary’, with a view to compare observations to that of other areas. Tummeralik lies approximately 40km east of Norsanna and has yet to be incorporated into the terrane theory of Friend and Nutman, (2005b), Friend et al., (1987) and Nutman and Friend, (2007). The area lies east of the purported boundary between the Færingehavn-Tre Brødre and Kapsillik terranes and so is currently interpreted as part of the latter. The area consequently provides a good opportunity to test the predictions of the Nuuk region terrane hypothesis, through correlating the dominant TTG orthogneiss with orthogneisses of known age and terrane affiliations and, where possible, by tracing the terrane bounding structures across the map sheet. A third area, Aputitooq Mountain, is situated in the central northern Kapisillit Mapsheet and lies near to the purported Færingehavn-Tre Brødre and Kapsillik
boundary. The area should thus preserve good metamorphic evidence of a 2650Ma event, with which to compare assemblages from the other two field areas.

In the following chapters, detailed structural and lithological observations are presented to characterise the deformational history of the three field areas. Selected geochronological analyses are presented to address the initial terrane interpretation of the each area and to identify key thermal and tectono-thermal events. Initial interpretations of field and geochronological data from Norsanna and Tummeralik are presented in Chapter 5 prior to a detailed petrological and mineralogical study of pelitic samples from all three areas in Chapter 6. In the final chapters, field and geochronological data is combined with metamorphic data to present P-T-t-deformation paths for all three field areas, which are then discussed in context with Nuuk region evolution as a whole.


Barker (Editor), Trondjhemites, dacties and related rocks. Elsevier, Amsterdam.


Iizuke, T. et al., 2006. 4.2Ga zircon xenocrysts in an Acasta gneiss from northwestern Canada: evidence for early continental crust. Geology, 34: 245-248.


Atlantic Craton, southern West Greenland: Complexities of Neoarchaean collisional orogeny. Precambrian Research, 155(3-4): 159-203.


CHAPTER 2
THE GEOLOGY OF NORSANNA

2.1 Overview

The western field area in this study, hereafter termed ‘Norsanna’, has been previously identified as a boundary zone between the Palaeoarchaean Færingehavn and Neoarchaean Tre Brødre terranes (Crowley, 2002; Friend and Nutman, 2005; Friend et al., 1987). Recent geochronological studies (Nutman, 2005, Personal Communication) from the shores of Ameralik and Itilleq Fjords have identified orthogneisses with ca. 2840 Ma emplacement ages at a number of localities, suggesting that the Norsanna area may also contain outliers of the Tasiusarsuaq terrane (Friend, 2006, Personal Communication) in addition to the purported Færingehavn-Tre Brødre boundary. Norsanna is consequently a key area in which to study the tectonic and metamorphic evolution of at least two distinct orthogneiss suites and their associated lithologies.

The data presented here relates to the structural and metamorphic field observations obtained from Norsanna during the summer seasons of 2005 and 2006. The study area is situated in the westernmost part of the central Kapisillit Mapsheet (Fig. 2.1) and covers an area of approximately 57km$^2$. It extends east to west from UTM 22W 519301E, 7123336N to 22W 513250E, 71023543N and north to south from UTM 22W 514831E, 7126431N to 22W 515863E, 7111028N. Norsanna is bound to the south by Ameralik Fjord and to the east by Itilleq Fjord and extends inland ~10km north of Ameralik Fjord. The area is dominated by orthogneisses with subordinate supracrustal and late intrusive material.
Figure 2.1: 1:50,000 geological map showing the major lithologies and structure in Norsanna. Inset: Star symbol denotes the position of Norsanna in the Kapisillit region.
2.2 Lithologies

*Polyphase Orthogneiss*

A polyphase, migmatitic, medium to coarse grained (up to 1mm) orthogneiss forms the dominant lithology, which crops out in two belts north of Ameralik Fjord and west of Itilleq Fjord (Fig. 2.1). The unit is generally tonalitic in composition and contains the assemblage plagioclase, quartz, biotite ± hornblendic amphibole. It exhibits a characteristic mm-cm scale gneissosity, which is principally defined by the segregation of the mafic and felsic minerals, so that biotite and amphibole (where present) form monominerallic selvedges lining the felsic component of the tonalitic gneiss (Fig. 2.2a).

Figure 2.2: The Polyphase orthogneiss of Norsanna showing a) mm-scale gneissic banding defined by leucocratic and melanocratic minerals (UTM 22W 514938E 7124597N), b) early generations of intensely deformed invasive felsic material, cut by late, undeformed granitic pegmatite (UTM 22W 523483E, 7128844N) and c) felsic material ponding in the axial planes of early folds (white arrow) and accentuating shear (UTM 22W 22W 523483E, 7128844N). A pen and compass are shown for scale in a) and b) respectively. Scale bar in c) measures approx 10cm.
In addition to the mineralogical variation on scales of 0.5-2cm that defines the gneissic banding, larger-scale heterogeneity is produced by variable amounts of leucosome. This leucosome is generally granodioritic in composition and forms 2-10cm wide, discontinuous layers and lenses parallel to the gneissosity. These leucosomes have sharp boundaries with the host tonalitic orthogneiss and are thus considered part of an early invasive melt (Fig. 2.2b) that pre- or syn-dates the development of the regional gneissosity rather than an in situ partial melt phase. The host gneiss and invasive material have been intensely deformed and completely recrystallised by subsequent tectonothermal events, resulting in relatively uniform grain size throughout the unit.

The gneissic fabric itself has been complexly deformed by at least three deformation events at high grade. These are suggested by the presence of early felsic pods that are discordant to the regional gneissosity, but are themselves deformed by the post-Dn events described in section 2.3. In areas of relatively low late-stage strain, the unit forms tight to isoclinal folds on cm-dm scale with granodioritic to granitic leucosome pooling parallel to the axial planes of these folds (Fig. 2.2b and c). In areas of relatively high late-stage strain, these early folds are completely overprinted by a new planar gneissic fabric.

A characteristic of the polyphase orthogneiss is the presence of layers, pods and lenses of homogeneous amphibolite, interpreted as an early mafic intrusive suite. These intrusives occur as deformed, linear bodies, usually less than 2m wide, which have been subject to at least two phases of deformation and metamorphosed to amphibolite facies. Three mineralogically distinct ‘suites’ are recognised in the mafic intrusives in the polyphase orthogneiss (Fig. 2.3a-c);

1. ‘spotted’ amphibolites (Fig 2.3a), which contain 1cm diameter spots of granular plagioclase, interpreted here as relict plagioclase phenocrysts,
2. ‘streaky’ amphibolites (Fig 2.3b) containing 0.5-1cm wide discontinuous stringers of felsic material, and
3. ‘rusty’ amphibolite (Fig 2.3c), which has a higher plagioclase content than the other variants, but with a rusty colouration as a result of sulphide and oxide weathering.
Figure 2.3: Variants of the early intrusive mafic suite in the polyphase orthogneiss. a) ‘spotted’ amphibolite (UTM 22W 514196E, 7123433N) where spots are flattened and aligned parallel to the regional gneissosity. b) ‘streaky’ amphibolite (UTM 22W 513409E, 7123714N). Felsic partial melt streaks are aligned parallel to the regional gneissosity. c) ‘rusty amphibolite’ (UTM 22W 513409E, 7123714N) has an orange weathering colour due to sulphide weathering.

The intrusive amphibolite suites are highly deformed, forming boudinaged, rotated lenses, which are typically strung out parallel to the gneissosity in the host orthogneiss.

Rare layered amphibolites, interpreted to be of supracrustal origin, were identified in the polyphase orthogneiss near its boundary with the homogeneous orthogneiss (UTM 22W 519310E, 7123207N). These are distinct from the deformed, homogeneous intrusives described above as they contain 20-50cm lenses of diopside-plagioclase calc-silicate or garnet-quartzite and are interpreted as metamorphosed marl and chemical sediments respectively. The amphibolite itself also tends to be heterogeneous with cm-scale compositional banding defined by differences in the modal abundance of plagioclase and amphibole. The largest amphibolite body in the polyphase orthogneiss occurs at UTM 22W 523483E, 7128844N, where it forms a 100-200m wide linear body. The unit here is predominantly mafic to ultramafic in composition with lesser amounts of sulphide-bearing garnetites, garnet-hornblende-
clinopyroxene-plagioclase gneiss and garnet-magnetite quartzite. Like the polyphase orthogneiss, the amphibolite at this locality displays m-scale, refolded isoclinal folds (Fig. 2.4).

Figure 2.4: Refolded isoclines in supracrustal units entrained within the polyphase orthogneiss UTM 22W 523481E, 7128915N). F1 folds are defined by quartz veins and the gneissic fabric in the amphibolite. The m-scale F1 isoclinal fold is refolded by open F3 folds. No D2 structures were observed as the locality is in an area of low D2 strain.

Homogeneous Orthogneiss

The central part of the Norsanna study area is dominated by a homogeneous orthogneiss (Fig. 2.1) with an apparently simpler deformational history than that of the polyphase orthogneiss. The unit forms a tectonised sliver and is bound above and below by the polyphase orthogneiss. The homogeneous orthogneiss contains the assemblage plagioclase, quartz, biotite ± K-feldspar ± hornblende and is homogeneous on cm- to m-scales, although the overall composition may vary between outcrops from tonalitic to granodioritic. In contrast to the polyphase orthogneiss, the granodioritic component is not confined to leucosome veins, suggesting that the observed compositional variation is a primary feature, rather than the result of later fluid invasion. The gneiss is medium to coarse grained, (0.5-1mm) and in areas of relatively low strain, may exhibit a sugary texture but develops a flaggy cleavage in areas of higher strain. The lithology does not exhibit the intense segregation of mafic and felsic minerals and so does not form the strong gneissic banding that is characteristic of the polyphase orthogneiss. A planar gneissic fabric in
Figure 2.5: Variation in the homogeneous orthogneiss a) mm-cm scale leucocratic veining (white arrow) defining the regional gneissosity in the homogeneous orthogneiss (UTM 22W 516073E, 7121505N). The tonalitic host rock is homogeneous on a cm-scale. Structurally above the orthogneiss in the photograph is an enclave of leucogabbro. b) 30-50 cm wide granodioritic pegmatitic sheets (white arrow) define the gneissosity in the homogeneous orthogneiss and are spaced on m-scales (UTM 22W 513146E, 7142704N). c) Highly sheared and recrystallised leucogabbro-anorthosite enclaves hosted by homogeneous orthogneiss near the boundary of the homogeneous orthogneiss with the polyphase orthogneiss (UTM 22W 516741E, 87121442N).

The homogeneous orthogneiss is principally defined by invasive felsic veins, which constitute 10 to 30% of the exposures of the gneiss. These mm- to cm-scale leucocratic layers occur on a spacing of 5 to 20 cm, are typically concordant or subcordant with the biotite fabric in the host orthogneiss and form discrete veins in an otherwise homogeneous host rock (Fig. 2.5a). In addition to these, dm-scale granodioritic veins also define the gneissosity and are spaced on a m- to 10m-scale (Fig. 2.5b). These are usually concordant to subcordant to the main fabric, but cross-cutting relationships are seen in areas of relatively low strain, suggesting intrusion during the later stages of the gneissosity-forming event.

The only significant heterogeneity in this unit is encountered within ~100 m of its upper and lower boundary regions. At these structural levels the unit becomes increasingly tectonised and locally contains dm- to m-scale rafted enclaves of
gabbro, leucogabbro and anorthosite (Fig. 2.5a and c). Competency contrasts between the mafic enclaves and quartzofeldspathic host rock have resulted in the boudinage and block rotation of the enclaves and the preferential partitioning of strain into the host gneiss. This strain partitioning between different lithological units has also permitted the preservation of primary igneous textures in some larger enclaves. Smaller leucogabbro rafts have been dismembered by strain partitioning on an individual grain scale, where shearing has caused ductile flow around igneous plagioclase phenocrysts, resulting in hybrid orthogneiss-leucogabbro patches in places. The boundary regions of the homogeneous orthogneiss are also host to the majority of supracrustal amphibolite in the Norsanna field area. No intrusive relationships between the polyphase and homogeneous orthogneisses are preserved in Norsanna as the boundary between the two units is invariably tectonised.

Anorthosites

Anorthosites and related rocks form a minor component of the lithologies in Norsanna. Rafts and enclaves have already been described in the homogeneous orthogneiss near its boundary with the polyphase orthgneiss on the north shore of Amealik Fjord (UTM 22W 518145E, 7122672N). Tectonised slivers and enclaves of dominantly anorthositic material form 1 to 10m scale bodies (also within the homogeneous orthgneiss) south of Amealik Fjord (UTM 22W 516920E, 7113516N). The anorthosite is leucocratic, with variable but minor amounts of mafic minerals (amphibole, chloritised amphibole) and local leucogabbro (Fig. 2.6). Plagioclase augen define a fabric that is conformable with the regional gneissosity in the homogeneous orthgneiss where it occurs both as larger tectonic slices and as individual enclaves. No large scale (>100m length) anorthositic bodies were encountered in the Norsanna field area. The closest large anorthosite body lies east of Norsanna and trends approximately N-S, subparallel to the eastern shore of Itilleq Fjord (see Appendix D).
Supracrustal lithologies

Supracrustal lithologies form a minor (<5% total outcrop) albeit significant component of the lithologies in Norsanna (Fig. 2.1). They are almost invariably contained within the homogeneous orthogneiss and are generally concentrated near the boundary between the homogeneous and polyphase orthogneisses. Three distinctive lithological units (amphibolite\(^1\), ultramafic\(^2\) and pelitic gneiss\(^3\)) occur in close association with each other. A supracrustal protolith is interpreted for the amphibolitic and ultramafic lithologies because of their intimate spatial association with pelitic lithologies. Although no sedimentary structures are preserved in the pelitic units, these have been interpreted as possible metasedimentary rocks because they form distinct boundaries with amphibolitic rocks, rather than exhibiting the alteration fronts described in section 1.7, which may suggest a metasomatic origin. This interpretation may change in the face of detailed geochemical analysis but is not considered further in this work. In addition, the supracrustal interpretation for amphibolitic and ultramafic lithologies may encompass shallow level mafic and ultramafic intrusives as well as extrusive material. However no primary structural evidence remains to discriminate between the two.

The three lithologies typically form the structural order 1-2-3 (see above) moving from SW to NE. An exception to this is seen adjacent to the inferred boundary between the homogeneous and polyphase orthogneisses to the SW of
Norsanna summit, where the sequence is reversed (3-2-1 moving SW to NE). The sequence was first recognised in a series of boudins at UTM 22W 514167E, 7122704N and the same geometry was encountered ~5km along strike to the south on the northern shore of Ameralik Fjord (UTM 22W 515202E, 7121271N). This systematic relationship, where ultramafic units are bound by amphibolite and pelite, is considered a primary feature of the supracrustal rocks in Norsanna, rather the result of later deformation. However the change in structural stacking order from 1-2-3 to 3-2-1 is considered the result of regional folding, where the order 1-2-3 lies to the west and 3-2-1 to the east of the Norsanna Isocline fold axis (Fig 2.1 and section 2.3). Two instances of distinctive supracrustal rocks occurring in the polyphase orthogneiss have been described in a previous section and are not further considered.

Amphibolites

Heterogeneous amphibolite forms the dominant supracrustal lithology in Norsanna. The largest bodies occur in linear belts that can be traced along strike for up to a kilometre and may vary from 10 to 100m wide. These belts are often highly attenuated, forming strings of boudinaged lenses, which typically occur within a few hundred metres of the homogeneous-polyphase orthogneiss boundary. The majority of supracrustal amphibolites occur in or adjacent to the boundary high strain zones and no primary relationships (neither intrusive nor sedimentary) between the amphibolite and orthogneiss are preserved. The characteristic linear outcrop pattern commonly means that the amphibolites define regional scale structures, and are therefore considered important structural marker horizons. Supracrustal amphibolite in Norsanna can be distinguished from the mafic intrusives in the polyphase orthogneiss in two ways:

1. Supracrustal outcrops are typically much thicker (individual boudins may be >10m wide, compared with the intrusives, which are typically <2m wide).
2. The amphibolite itself is characteristically heterogeneous, with compositional banding on cm- to m-scales.
3. Supracrustal amphibolite is spatially associated with pelitic and ultramafic lithologies.
Figure 2.7: Variation in supracrustal amphibolites in Norsanna. a) F₁ isoclinal folding defined by compositional banding in garnet-absent amphibolite. Th banding is defined by modal differences in hornblende and plagioclase (white arrow). b) Coarse garnet porphyroblasts in heterogeneous amphibolite. c) Coarse diopside-plagioclase layers in amphibolite showing evidence of block rotation towards parallelism with the gneissosity in the amphibolite d) Dip-surface view of felsic sheets intruding the heterogeneous amphibolite parallel to the gneissosity in the amphibolite. All photographs were taken at UTM 22W 516073E, 7121505N.

The host amphibolite itself is generally garnet absent, consisting mainly of hornblende-plagioclase ± epidote and some minor occurrences of clinopyroxene, where the principal heterogeneity is defined by differences in modal abundance of hornblende and plagioclase on a scale of 2-10cm (Fig. 2.7a). Garnet-bearing amphibolite occurs occasionally as pods and boudins and as cm- to m-thick layers within garnet-absent amphibolite. The lithology may also exhibit a more marked mafic-felsic compositional banding on scales of 0.5 to 2cm and may contain quartz. These variations are considered to reflect primary differences in amphibolite composition rather than metamorphic segregation, and support a volcano-
sedimentary protolith interpretation. Where present, garnet forms deep red porphyroblasts up to 3cm in diameter (Fig. 2.7b). Semi-continuous horizons and cm- to m-scale layers and lenses of diopside-plagioclase ± epidote ± carbonate ± sulphides add to this heterogeneity. The diopside-plagioclase lenses are wrapped by the regional gneissic foliation, and are rotated and boudinaged (Fig. 2.7c), suggesting that these calc-silicate horizons formed prior to deformation, and may represent early carbonate or calc-silicate alteration.

Both garnet-bearing and garnet-absent amphibolite are cut by net-veined and anastomosing felsic leucosome (Fig. 2.7d). Much of the leucosome lies parallel to the regional gneissosity, only cutting high grade fabrics in local domains where it forms dm-scale pools or patches. At UTM 22W 515251E, 7121289N, the felsic leucosome intruding supracrustal amphibolite contains clots of orthoamphibole, suggesting a degree of host rock-melt interaction. However inclusions of this nature are uncommon in the leucosome as a whole. The structural constraints placed on the leucosome suggest invasion during the late regional gneissosity-forming event in the supracrustal amphibolites and homogeneous orthogneiss (section 2.3). Hence, the syn- to late-D₂ leucosomes are important targets for geochronological sampling.

**Ultramafic gneiss**

Rocks of ultramafic affinity often form the core lithology in the 1-2-3 supracrustal sequence in the Norsanna field area. They occur as boudinaged lenses, usually in the structurally highest levels of the homogeneous orthogneiss package (near its boundary with the polyphase orthogneiss), and vary in size from 10 to >500m in length. The largest ultramafic body, which exceeds 550m in length, occurs around UTM 22W 513880E, 7124443N (Fig. 2.8a). Ultramafic material is typically highly altered to various combinations of actinolite, tremolite, talc, anthophyllite, chlorite and biotite with minor carbonates and rare diopside. The majority of ultramafics are dominated by coarse (1-2mm), homogeneous talc-tremolite/actinolite ± serpentine ± anthophyllite ± carbonate. Extremely coarse (individual grains exceeding 10cm in length) areas of pale-weathering, monomineralic anthophyllite and/or tremolite (Fig. 2.8b) are well-developed near narrow shear zones. At UTM 22W 514007E, 7124910N, orange-coloured, tabular to elliptical grains 1-5cm in diameter are preserved within an actinolite, talc and chlorite matrix. These are
Figure 2.8: Variations in ultramafic lithologies in Norsanna. a) Large (>500m in length), highly altered ultramafic boudin structurally overlain by pelitic units. The interface between the two units comprises a zone of coarse grained, altered ultramafic material (UTM 22W 513844E, 7124420N). b) Very coarse, randomly oriented, monominerallic amphibolite (UTM 22W 5138442E, 7124553N). Image shown is probably orthoamphibole but monominerallic tremolite and actinolite were also encountered. c) Biotite-orthoamphibole schist at highly strained contact between ultramafic boudin and semipelitic gneiss (UTM 22W 513811E, 7124447N). d) Tourmaline-rich rind in late granitic pegmatite where it intrudes ultramafic boudin (UTM 22W 513837E, 7124544N).
interpreted as orthoamphibole-bastite after igneous orthopyroxene. Strain partitioning at boudin rims often forms a highly schistose rind, which is most often biotite or chlorite-rich. At UTM 22W 513811E, 7124447N a 2-4cm thick layer consisting entirely of coarse, schistose pale amphibole-biotite-plagioclase (Fig. 2.8c) forms a mantle between less-deformed, phyllosilicate-poor ultramafic gneiss and underlying semi-pelitic gneiss. Significant strain partitioning between the ultramafic and surrounding lithologies has produced less pronounced gneissic fabrics in boudin centres, with granofelsic textures forming in the centres of the largest bodies.

Ultramafic units in Norsanna are cut by the same felsic leucosome, granitic pegmatite and narrow shear zones that cut other lithologies in the area. Interactions between the wall-rock and this evolved material have resulted in the intense alteration of the ultramafics, forming the blackwall assemblages of amphibole and tourmaline at the margins of invasive felsic veins (Fig. 2.8d). The Mg-rich mineralogy surrounding these veins and shear zones (tremolite-anthophyllite) strongly suggests the leaching of Fe from the ultramafic host into the fluids associated with felsic veining and shearing.

**Metapelitic and related rocks**

Rocks with a pelitic and semipelitic composition, interpreted to be of sedimentary origin, form the smallest component of the supracrustal units in Norsanna. They are intimately associated with the amphibolite and ultramafic gneisses described in previous sections, and are commonly bound to one side (usually at their upper boundary) by the amphibolite or ultramafic gneiss and to the other by the homogeneous orthogneiss (Fig. 2.1). The unit is dominated by rusty-weathering metapelites containing the assemblage quartz-plagioclase-biotite-garnet ± sillimanite ± muscovite. Quartzofeldspathic and garnet-biotite gneiss are subordinate lithologies. The peak gneissic fabric in the metasedimentary units is represented by an S>L tectonite defined by coarse (1-2mm) intergrown biotite and sillimanite (Fig. 2.9a and b). The foliation wraps 1-5mm pink-mauve garnet porphyroblasts, which occur in virtually all outcrops of metasedimentary material. Cordierite is present locally. Pelitic material is generally homogeneous at outcrop scale, although weak compositional banding defined by modal differences in quartz or plagioclase and ferromagnesian minerals is present on cm-scales. This is considered to reflect a
primary compositional variation. No unequivocal sedimentary sequences were observed in any of the lithologies described here. A single outcrop of garnet-orthoamphibole gneiss at UTM 22W 515521E, 7121289N suggests a more Fe- and Mg-rich bulk composition than the other pelitic units in Norsanna, suggesting a mafic component in the sediment source in this instance.

![Figure 2.9: Rusty weathering in pelitic units in Norsanna. a) A planar S2 gneissosity in a high-strain zone defined principally by variations in the modal abundance of ferromagnesian and quartzofeldspathic minerals (UTM 22W 514938E, 7124597N). b) F3 folds of the S2 gneissosity in pelitic orthogneiss at UTM 22W 515202E, 7121271N. F3 fold axes are denoted in yellow. At both localities, rust weathering is more intense in sillimanite-biotite-rich foliae than the quartzofeldspathic foliae.](image)

**Pegmatites**

Granitic pegmatites are abundant throughout Norsanna and cut all lithologies except the mafic dykes described below. A group of sub-parallel pegmatites trend NNE, forming steeply WNW-dipping sheets in the northern part of the study area (Fig. 2.1), which includes the flank of Norsanna mountain, cutting the large ultramafic boudin where they are associated with the formation of tourmaline (Fig 2.8d). The pegmatites are undeformed by high grade events, but are locally cut by greenschist facies mylonite zones south of Ameralik Fjord (Fig. 2.10).
Figure 2.10: Granitic pegmatite intruding the homogeneous orthogneiss with amphibolite enclave at a high angle to the gneissic fabric. The gneissic fabric is defined by the orientation of thin felsic veins within the amphibolite enclave and by the enclave itself (UTM 22W 516675E, 7111551N). The granitic pegmatite shows extensive recrystallisation, especially at the edges of the sheet. Deformation in the granitic pegmatites is associated with $D_{4N}$. Scale bar measures 50cm.

**Mafic intrusions**

Two generations of mafic dykes were identified in Norsanna. These generally trend NW-WNW and cut all lithologies identified in Norsanna. The early generation preserves structurally discordant relationships to the gneissosity in the orthogneisses but is deformed by late stage greenschist facies mylonitic shear zones. They exhibit a schistose fabric at their margins (Fig. 2.11a) and are often invaded by epidote veins (Fig. 2.11b). The later generation of chemically similar mafic dykes were intruded in entirely post-deformational settings and so preserve well-defined, unsheared chilled margins. The largest intrusion is a ~50m wide doleritic dyke, seen south of Ameralik Fjord at UTM 22W 517262E, 7112036N, which belongs to this later generation. The body was identified near its north-westerly termination but can likely be traced southeast for 40km to the western shore of the lake Isortuarsuk (Appendix D).
2.3 Structural evolution of Norsanna

The Norsanna field area has been subject to multiple high grade deformation events, resulting in various levels of overprinting across the whole area. In addition to the complex events that formed the gneissic fabrics in the polyphase orthogneiss, a further two deformation phases in the amphibolite facies are recorded, as well as at least one phase of deformation in the greenschist facies. The following section gives a detailed description of the latest phases of deformation in the Norsanna region. A summary of the deformation history in the area is given in Table 2.1.

Gneissosity-forming events ($D_n$)

The polyphase orthogneiss is the most complexly deformed unit in Norsanna, where the earliest deformation event(s) led to the development of an early $S_n$ gneissosity. The $S_n$ gneissosity is defined by the segregation of mafic and felsic minerals in the orthogneiss and also by the layer-parallel alignment of coarse felsic material that is interpreted as an early invasive suite (Fig. 2.2). The $S_n$ gneissosity has been subject to multiple phases of folding, however little evidence remains of the number of deformation events involved in the initial formation of the fabric. The definition of the gneissosity by amphibole suggests that the fabric formed under at least amphibolite facies conditions.
### Table 2.1: A descriptive structural evolution of the Norsanna area

<table>
<thead>
<tr>
<th>Event</th>
<th>Grade</th>
<th>Description</th>
<th>Affected lithologies</th>
</tr>
</thead>
<tbody>
<tr>
<td>$D_n$</td>
<td>?</td>
<td>Initial formation of gneissic fabrics and compositional banding. Multiple folding events in older orthogneisses.</td>
<td>Polyphase orthogneiss Deformed mafic dykes Supracrustals in polyphase orthgneiss</td>
</tr>
<tr>
<td>$D_1$</td>
<td>Upper amphibolite facies</td>
<td>Km-scale isoclinal folding of $D_n$ gneissosity in narrow high-strain zones. Development of sillimanite-biotite peak fabric.</td>
<td>All units except: Granitic pegmatite Late mafic intrusions</td>
</tr>
<tr>
<td>$D_2$</td>
<td>Amphibolite facies</td>
<td>Tectonic interleaving of polyphase and homogeneous orthogneisses and supracrustal units. Development of pervasive platy fabric in high-strain zone and dominant gneissic fabric in homogeneous orthogneiss. Boudinage of supracrustal units and attenuation of isoclinal fold limbs.</td>
<td>All units except: Granitic pegmatite Late mafic intrusions</td>
</tr>
<tr>
<td>$D_3$</td>
<td>Amphibolite facies</td>
<td>Refolding of $D_1$ structures into km-scale open structure.</td>
<td>All units except: Granitic pegmatite Late mafic intrusions</td>
</tr>
<tr>
<td></td>
<td>Granite pegmatite intrusion late syn- to post $D_{n+2}$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$D_4$</td>
<td>Greenschist facies</td>
<td>Localised cataclasis and shearing affecting late granite pegmatites. Single NE-trending chlorite grade shear zone partitioned into at least three mylonitic zones less than 50m width.</td>
<td>All units except: Latest generation of mafic dykes</td>
</tr>
</tbody>
</table>

**Isoclinal folding ($D_1$): The Norsanna Isocline**

The $D_1$ event is the first recorded deformation event to affect the homogeneous orthogneiss. The event is manifest by the formation of early isoclinal folds parallel to the boundary zone between the polyphase and homogeneous orthogneisses, and good evidence for the $D_1$ deformation event tends to be limited to within this boundary zone. The isoclinal fold structure (hereafter termed the Norsanna Isocline) is contained wholly within the homogeneous orthogneiss but within 50m of the boundary with the polyphase orthgneiss. The structure is defined by the amphibolite-ultramafic-pelite assemblage (1-2-3) described in section 2.2, where it is responsible for the reversal in structural geometry of the supracrustal sequence from 1-2-3 to 3-2-1 across an inferred fold hinge (Fig. 2.1). Later deformation has resulted in the smearing of $F_1$ hinge regions into parallelism with the regional fabric, and therefore no unambiguous $F_1$ fold closures were encountered in
Norsanna. A likely locality for the hinge region is, however, seen at UTM 22W 513954 E, 7124731 N where a very large ultramafic boudin and associated amphibolite and pelitic lithologies forms a single tectonised package around UTM 22W 513880E, 7124443N, but appears to branch into two ‘limbs’ moving east and south of this locality. The intense post-D_1 deformation in this region precludes a more detailed interpretation. The initial amplitude of the Norsanna Isocline is also difficult to gauge, but the fold limbs can be traced along strike some 5km to the shore of Ameralik Fjord, suggesting km-scale F_1 fold amplitudes.

**Progressive shearing (D_2): The Norsanna High Strain Zone**

The onset of shearing during D_2 resulted in the development of a flaggy, planar gneissosity, which pervasively overprints the earlier, complexly folded D_n gneissosity in the polyphase orthogneiss (Fig. 2.12a). The relatively simple gneissic banding in the homogeneous orthogneiss is synonymous with the S_2 fabric in the polyphase orthogneiss, and is inferred to have formed during the same event. The S_2 foliation is the earliest deformation fabric observed in the homogeneous orthogneiss and incorporates cm-scale felsic segregations and dm- to m-wide felsic veins, suggesting the attainment of high grade metamorphic conditions prior to D_2. The foliation is defined by hornblende and biotite in the orthogneisses and amphibolites, and by sillimanite-biotite in the pelitic units. The orthogneisses and supracrustal units are intruded by syn-D_2 felsic leucosomes that form anastomosing networks that are oriented overall parallel to the S_2 gneissosity (Fig. 2.7).

Although locally pervasive, the intensity of deformation associated with D_2 is highly variable across Norsanna, and is concentrated at major lithological boundaries. These zones may vary in thickness from <10m to >100m and are characterised by tectonic interleaving on 1m to 100m scales. The largest of these is the Norsanna High Strain Zone (NHSZ), which is centred at the boundary between the polyphase and homogeneous orthogneisses and has an estimated structural thickness between 120 and 150m. The NHSZ also includes the majority of supracrustal units and anorthosite enclaves in the homogeneous orthogneiss, including the supracrustal sequence that defines the Norsanna Isocline. The limbs of the Norsanna Isocline have been intensely attenuated, forming a string of boudinaged...
lenses between 1 and 10m in length, and the hinge zone has been largely sheared into parallelism with the \( S_2 \) gneissososity. The fold runs parallel to the NHSZ along its entire exposed length (ca. 7km), and displays no evidence of discordance with \( S_2 \) or transposition by the NHSZ on a larger scale. Shear indicators in the NHSZ are rare, however a top-to-SW sense of motion can be discerned from rare m-scale \( F_2 \) tight to isoclinal folds, an example of which is seen in the north of the field area at UTM 22W 513389E, 7124108N (Fig. 2.12b).

A second, smaller high-strain zone is encountered on the lower boundary of the homogeneous orthogneiss where it is once again juxtaposed against the polyphase orthogneiss. In this area (UTM 22W 519310E, 7123207N), the polyphase orthogneiss displays the platy \( S_2 \) fabric over a 50m wide outcrop moving NE towards its boundary with the homogeneous orthogneiss. The deformation is not as marked here as in the NHSZ, suggesting that strain was concentrated into the NHSZ. Shear indicators are not preserved along this zone.

**Regional folding (\( D_3 \))**

The whole of Norsanna is refolded into a km-scale open synformal \( F_3 \) fold, whose axial trace can be followed north for >50km. Quartz elongation and crenulation lineations (Fig. 2.13) predict a shallowly SSE-plunging fold (\(-20/170\)) in the Norsanna area, but this swings to trend NE, north of Norsanna. The fold is
inclined, with an axial plane dipping 35/120 and an interlimb angle of 90-100° (Fig. 2.14). Mesoscale F$_3$ structures occur sporadically in the homogeneous orthogneiss (UTM 22W 513146E, 7142704N). These refold early isoclinal folds in the polyphase orthogneiss at UTM 22W 523481E, 7128915N). L$_3$ quartz lineations are also developed in crenulated granodioritic pegmatite that defines the S$_2$ gneissosity in the homogeneous orthogneiss.

Figure 2.13: D$_3$ crenulations and quartz lineations (yellow dashed lines) in a granodioritic pegmatite intruding the homogeneous orthogneiss (UTM 22W 513146E, 7142704).

Figure 2.14: The F$_3$ structure in Norsanna is defined by folding of the S$_2$ gneissosity and controls the outcrop pattern of the whole area.
Greenschist facies shearing ($D_4$)

Northeast-trending greenschist facies shear zones cut the $S_2$ gneissosity in the homogeneous orthogneiss. All $D_4$ shear zones deform the late granitic pegmatites and deform the early generation of mafic dykes, but are cut by the later generation (Fig. 2.11). The largest $D_4$ shear zones occur south of Ameralik Fjord where a series of three parallel zones trend NE (Fig. 2.1) and form subvertical mylonite zones between 50 and 100m wide. Isoclinal folds of the regional $S_2$ gneissosity are rotated in these zones, becoming subvertical in the centre of the shear zones. Rotated fold hinges and intense L<<S tectonites in mylonitised pegmatites have an orientation of 128/80 (Fig. 2.15). A schistose $S_4$ foliation pervades the homogeneous orthogneiss in the $D_4$ mylonite zones (Fig. 2.15), which is defined by chlorite, showing shearing occurred under greenschist facies conditions. This is supported by epidote patches and veins in $D_4$-deformed mafic dykes at UTM 22W 516500E, 7111220N.

![Diagram showing orientation of shear zones and gneissosity](attachment:image)

Figure 2.15: Rotation of $D_2$ fold hinges (squares) into near vertical attitudes by $D_4$Norsanna.

A number of <1m wide blastomylonite shear zones crop out on the north shore of Ameralik Fjord (UTM 22W 518143E, 7122667N). These exhibit a biotite-defined foliation with stringers of quartz and plagioclase augen that may reach 15mm diameter (Fig. 2.16a and b). S-C fabrics in these zones trend towards ~140° with a West towards the SE sense of motion. No cross-cutting relationships are seen between the large $D_4$ mylonite zones south of Ameralik Fjord and the smaller blastomylonitic shear zones on the northern shore of Ameralik.
2.4 A summary of the geological evolution of Norsanna

From field evidence alone it appears that Norsanna displays a protracted tectonothermal evolution, involving several episodes of intrusion, deformation and metamorphism. The principal stages in the evolution of Norsanna are summarised below:

1. Intrusion of the precursor to the polyphase orthogneiss into an early suite of supracrustal lithologies, followed by intrusion of mafic dykes (evidence of deformation and metamorphism prior to dyking event is not preserved in Norsanna). High grade metamorphism and deformation of Polyphase orthogneiss forming the complex S_n gneissosity, with associated boudinage and shearing of the mafic dykes suite into parallelism with S_n.
2. Intrusion of homogeneous orthogneiss.
3. D_1 isoclinal folding in the polyphase-homogeneous orthogneiss boundary zone, forming the Norsanna Isocline. Upper amphibolite facies conditions reached during D_1.
4. D_2 shearing along lithology boundaries forming the NHSZ and other smaller scale high-strain zones. Upper amphibolite facies conditions maintained. Intrusion of felsic leucosome generally parallel to S_2 fabric.
5. Regional D_3 open folding of NHSZ. No record of retrogression from upper amphibolite facies.
6. Post-tectonic intrusion of coarse, granitic pegmatite.
2.5 A summary of the geological evolution of Aputitooq Mountain

The Aputitooq Mountain field area was mapped by Bennett and Heiss (2006) and lies east of Itilleq Fjord and north of Norsanna and Tummeralik (between latitude -50°21'W and -50°10'W and longitude 64°17'N and 64°25'N) (Fig. 2.17). The following section gives a brief overview of the field area. Field relations and deformation event nomenclature are sourced from an unpublished field report by Bennett and Heiss (Pers. Comm).

Figure 2.17: A simplified geological map of Aputitooq Mountain from Bennett and Heiss (2006). The key is ordered according to structural level, with the Bt-Hbl TTG orthogneiss and pink granite forming the bottom and top of the structural pile respectively. All lithology boundaries are highly tectonised.
Eight principal lithological units were identified; three variably deformed orthogneiss suites, amphibolite, pelitic and semi-pelitic units, a weakly deformed granite and an anorthosite package. The unit forming the bottom of the structural pile is a tonalitic to granodioritic orthogneiss containing variable amounts of biotite and hornblende. U/Pb SIMS dating of the unit yields an emplacement age of 2925Ma (ages from Bennett and Heiss (unpublished field report). The biotite-bearing granodioritic orthogneiss in the middle of the structural pile contains recumbent folds and shows evidence of migmatisation. No age constraints exist for this unit. The third orthogneiss, which forms the uppermost unit displaying intense deformation, is a polyphase TTG orthogneiss and is correlated with the polyphase orthogneiss of Norsanna (Bennett and Heiss, unpublished field report). A weakly deformed pink granite occupies the highest structural level in Aputitooq Mountain.

Pelitic material is sourced from two units in the area; 1) as discontinuous 1-15m thick lenses in a heterogeneous amphibolite unit and 2) as a heterogeneous linear belt between 5 and 50m thick. The heterogeneous amphibolite is a locally garnet-bearing hornblende-biotite-plagioclase ± quartz rock with a gneissic fabric defined by amphibolite-diorite compositional banding on cm- to dm-scales. The unit is intruded by felsic to granitic leucosome, which is gneissosity-parallel and locally garnetiferous. The heterogeneous pelitic unit is placed structurally higher than the heterogeneous amphibolite and is characterised by abundant gossanous and locally mineralised horizons. The unit displays a compositional zoning from felsic-intermediate at the base, which grades into dominantly pelitic material in the core and becomes mafic at the highest levels. The pelitic core contains a range of mineral assemblages including garnetites (Grt-Bt-Q), semi-pelites (Grt-Bt-Pl-Q) and aluminosilicate and/or cordierite-bearing pelites (Grt-Bt-Pl-Q ± Sil ± Crd). All units with the possible exception of the weakly deformed granite have been subject to amphibolite facies conditions. The supracrustal units show localised evidence of retrogression from granulate facies suggested by cordierite decompression rims around garnet that are evident at outcrop scale.

All units with the exception of the weakly deformed granite are affected by multiple phases of deformation, including (an) early phase(s) of gneissosity
formation under amphibolite facies conditions (Dn). This initial phase was followed by a period of thrusting and mesoscopic isoclinal fold formation (Dn+1), which took place at upper amphibolite (and locally granulite) facies, indicated by pegmatite intrusion parallel to the peak gneissosity. The tectonic intercalation of all the complexly deformed lithologies took place during Dn+1. The area was then folded by regional, SE-plunging open folds (Dn+2) and subject to dilatational brittle-ductile shearing with associated syn-shear melt intrusion (D3) at unknown metamorphic grade. A final, brittle faulting event is associated with greenschist facies NW-striking fracture systems (D4).
CHAPTER 3
THE GEOLOGY OF TUMMERALIK

3.1 Overview

The data presented here relates to the structural and metamorphic field observations at Tummeralik, in the eastern central Kapisillit region (Fig. 3.1). Data was collected during the 2006 summer season. The eastern field area lies ~40km east of Norsanna, adjacent to the Inland Ice and covers an area of approximately 61km². Tummeralik structurally underlies lithologies in Norsanna. The area extends east to west from UTM 22W 562883E, 7124087N to 22W 550812E, 7124940N and north south from UTM 22W 557312E, 7127768N to 22W 560155E, 7124217N. Orthogneisses of tonalitic chemistry comprise the majority of the study area, with subordinate dioritic orthogneiss and supracrustal sequences (Fig. 3.1). Prior to this study, detailed lithological and structural studies had not been undertaken in Tummeralik. Previous work on the 1:500,000 Frederikshåb Isblink – Søndre Strømfjord mapsheet (Allart, 1982) has suggested the presence of large linear bodies of amphibolitic and potentially metasedimentary material. The area thus holds considerable potential for constraining the pressure-temperature evolution of this part of the NAC. This study will present the first detailed structural and metamorphic field study of the Tummeralik area, allowing the tectonothermal evolution of Tummeralik to be placed in a regional context.
Figure 3.1: 1:50,000 geological map of Tummeralik showing key lithologies and structures. UTM Grid references North = 22W 557312E, 7127768N; South = 22W 560155E, 7124217N; East = 22W 562883E, 7124087N; West = 22W 550812E, 7124940N. 
Inset: The location of Tummeralik in the Kapisillit Mapsheet. R = Position of the reconnaissance area southeast of Tummeralik.
3.2 Lithologies

Tonalitic orthogneiss

Tummeralik is dominated by a tonalitic orthogneiss suite. The unit is heterogeneous on scales of 10m, with heterogeneity defined by variations in the modal abundances of plagioclase versus quartz, and also by variation in the proportion of mafic and felsic minerals. The orthogneiss is medium to coarse grained (0.5-1mm) and contains the general assemblage plagioclase-quartz-biotite-hornblende. The cm- to dm-scale gneissic foliation is defined by differences in modal abundance of mafic and felsic minerals and cm-thick selvedges of near-monomineralic biotite and hornblende accentuate this fabric. The tonalitic orthogneiss is characterised by the common occurrence of pods, layers and lenses of amphibolite, some of which may be 5-10m in length (Fig. 3.2a). The amphibolite itself consists entirely of hornblendic amphibole and plagioclase with little variation in the modal abundances of these minerals although the rafts may contain stringers of felsic material parallel to the gneissosity that are interpreted as intrusive to the amphibolite. These homogeneous amphibolites lie parallel to the regional gneissosity and are intensely boudinated and veined, forming horizons up to 20m wide in which rafted and boudined trails of hornblende-plagioclase amphibolite form up to 70% of the outcrop in the tonalitic orthogneiss host rock. Strain partitioning preferentially into the tonalite has resulted in the intensification of the gneissic fabric surrounding mafic boudins. Discordant relationships between the tonalitic host and the mafic component were not widely observed due to the high state of strain in both lithologies. However at one locality the amphibolite preserves isoclinal fold hinges (Fig. 3.2b) that are truncated against the tonalite, which may suggest parts of the mafic units to be older than the host. The horizons of boudinaged, homogeneous amphibolite are more common within the tonalitic orthogneiss in the north of the mapping area, but become less significant moving south. At UTM 22W 563043E, 7125212N, homogeneous amphibolite pods and boudins are totally absent, and the tonalitic orthogneiss forms a homogeneous horizon some 200m thick. This occurs adjacent to a boundary between the tonalitic orthogneiss and a supracrustal amphibolite body. This zone of amphibolite-absent orthogneiss can be followed along strike (parallel to the boundary) over a distance of several kilometres.
Rare dm-scale boudins of garnet-biotite or garnet-biotite-quartz occur sporadically throughout the tonalitic orthogneiss. These are considered to be remnants of supracrustal lithologies entrained during the emplacement of the tonalitic orthogneiss and do not constitute a major lithological unit.

The tonalitic orthogneiss is invaded by several generations of felsic to granodioritic veins and sheets. Discontinuous, 1-2cm thick layers and stringers of hornblende-plagioclase leucosome commonly occur on dm-scales throughout the unit. In most cases, this leucosome is broadly concordant with the regional gneissosity (Fig. 3.3a) however where they occur in strain shadows created by amphibolite pods and rafts, the leucosomes mantle and cut the amphibolite (Fig. 3.3b). This may suggest that the leucosome initially formed in association with the homogeneous amphibolite, but was later strung into parallelism with the regional gneissosity. This may have occurred during two separate events or as part of a single, progressive tectonothermal episode. Alternatively, the intrusion of a syn-deformational leucosome may have pooled in the low strain domains associated with the more competent amphibolite enclaves.

An early granodioritic phase intrudes the tonalitic orthogneiss in the central and southern parts of the area. In the central area (UTM 22W 563457E, 7125097N),
these veins form cm- to dm-scale interconnected systems that are parallel or subparallel to the regional gneissic fabric, suggesting intrusion during the formation of the main gneissosity. These become increasingly voluminous moving south and forms a zone of >80% of the outcrop at UTM 22W 560413E, 7123839N, which can be traced along strike for ~1km in either direction (Fig. 3.3c). The pegmatite exhibits a weak foliation defined by biotite that is parallel to the regional gneissosity in the host orthogneiss, and has also been largely recrystallised, so that the grain size is similar or slightly coarser than that of the tonalitic orthogneiss. The granodioritic pegmatite and homogeneous amphibolite are not observed in the same outcrop and so appear the relationship between the amphibolite and pegmatite could not be ascertained. The tonalitic orthogneiss, mafic horizons and granodioritic leucosome are cut by cm-scale, epidote-bearing shear zones, which are mylonitic in places.
**Dioritic orthogneiss**

A dioritic orthogneiss forms the core of a regional fold in central and eastern Tummeralik (Fig 3.1). The unit forms boundaries with the tonalitic orthogneiss, supracrustal amphibolite and a small granodioritic body. The dioritic orthogneiss is a medium grained (~0.5mm) garnet-absent hornblende-plagioclase gneiss unit with a matrix that is mineralogically homogeneous but which contains cm-scale bands of tonalitic and felsic pegmatite spaced at dm- to m scale intervals, which are concordant with the regional gneissosity (Fig. 3.4a and c). Decimetre-scale boudins and enclaves of amphibolitic material are common throughout the unit, some of which exhibit potential relict igneous textures in the form of coarse hornblende spots. The diorite gneiss is invaded by >10cm-thick felsic pegmatite that develops a medium grained (0.5mm) reaction rim of hornblende, plagioclase and biotite where in contact with the gabbroic enclaves entrained within the dioritic host rock (Fig. 3.4b).

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Figure 3.4: Variation in the dioritic orthogneiss a) Dip-surface view of the dioritic orthogneiss with interleaved tonalitic orthogneiss slivers and intruded by deformed granodioritic sheets that running parallel to the regional gneissosity (UTM 22W 560952E, 7124616N). b) Raft of dioritic orthogneiss with gneissic fabric defined by aligned hornblende and intrusive felsic veins (UTM 22W 560952E, 7124616N). The pale, fine grained rim to the left of the raft comprises micaceous alteration products of amphibole (delineated by red dashed lines). c) Platy gneissic fabrics in the dioritic orthogneiss (UTM 22W 560179E, 7124638N).
Although the gneissic banding in the dioritic orthogneiss is principally defined by invasive, pegmatitic material, an increase in strain at UTM 22W 560179E, 7124638N causes the orthogneiss to develop a platy fabric over a structural thickness of ~45m (Fig. 3.4c). High strain zones of this scale and smaller are encountered within and between lithologies throughout Tummeralik and are correlated with the gneissosity-forming event.

Granodioritic orthogneiss

A small, mineralogically homogeneous granodioritic orthogneiss was first encountered at UTM 22W 561004E, 7124894N, forming a sliver ~200m wide that is traceable for ~5km along strike and is bound above by the dioritic orthogneiss and below by the supracrustal amphibolite. The unit pinches out westwards and is not encountered west of UTM 22W 558990E, 7125056N (Fig. 3.1). The unit contains the assemblage biotite-plagioclase-K-feldspar-quartz with little variation in the modal abundance of these minerals, where aligned biotite-bearing granodioritic pegmatite veins define a weak foliation (Fig. 3.5a). The mineralogy of this pegmatite is the same as the host, and is distinguished from the host solely by a coarser grainsize. The unit is not as intensely deformed as other lithologies in Tummeralik, displaying a much weaker gneissic fabric than any other orthogneiss suite in Tummeralik. This may be the result of a predominantly quartzofeldspathic mineralogy, which would hinder the development of a gneissic fabric, however other characteristics of the lithology may indicate a lower state of finite strain. Near its upper boundary with the dioritic orthogneiss, the granodioritic orthogneiss contains rafted enclaves of dioritic material (Fig. 3.5b). This is interpreted as an intrusive contact and suggests that the granodioritic orthogneiss is younger than the dioritic orthogneiss. This may also account for the apparent low state of strain observed in the granodioritic orthogneiss relative to adjacent rocks.
Supracrustal lithologies

Numerous enclaves, pods and laterally continuous layers of amphibolite, metapelite and ultramafics crop out throughout Tummeralik, forming 10-15% of the total outcrop. Rocks with a classic upper amphibolite facies pelitic mineralogy are hypothesised to have metasedimentary protoliths because they form well-defined boundaries with amphibolitic material, with no evidence of metasomatism along an alteration front (see Chapter 1.7). Protoliths to amphibolitic and ultramafic units are interpreted to be supracrustal because of their close spatial association with interpreted metasedimentary and chemical sediment lithologies, however they may represent either shallow-level intrusive rocks or volcanic sequences.

The supracrustal lithologies are almost exclusively contained within the tonalitic orthogneiss, where they occur as boudins or linear bodies a few metres wide. A few m-scale outcrops were encountered as tectonised slivers in the granodioritic orthogneiss, however it remains uncertain as to whether these were rafted during granodiorite emplacement or whether they were tectonically intercalated during peak deformation. The largest occurrence of supracrustal lithologies forms a semi-continuous belt up to 100m wide on fold limbs, and up to
200m wide in fold noses. This forms near the boundary between the tonalitic and dioritic orthogneiss suites and defines the km-scale Tummeralik Isocline. The body can be traced across the entire Tummeralik study area from ESE to WNW and continues outside of Tummeralik for at least another 5km to the NW.

Supracrustal amphibolite

Amphibolite constitutes >75% of the supracrustal lithologies in Tummeralik and typically occurs as linear belts a few metres to tens of metres wide, which can sometimes be followed along strike for several kilometres. Alternatively, amphibolite forms boudinaged layers, up to a few tens of metres in length within the tonalitic orthogneiss. These amphibolites have a heterogeneous composition that is described below and are discrete from the homogeneous amphibolite that forms horizons in the tonalitic orthogneiss. The amphibolite unit forms the host rock for most occurrences of pelitic, semipelitic and ultramafic units, which occur as tectonised slivers and boudinaged lenses respectively.

The unit comprises almost entirely garnet-amphibolite, although some garnet-absent layers and bands occur on cm- to m-scales (Fig. 3.6a). Garnet-rich amphibolite occurs as layers within the amphibolite, and as garnet-hornblende pods within garnet-absent amphibolite and occasionally within meta-pelitic and semipelitic units. Porphyroblasts are deep red in colour and subhedral to anhedral at outcrop scale (Fig. 3.6b). They typically range from 2-10mm in diameter, however very coarse garnet (up to 50mm diameter) is encountered at UTM 22W 557848E, 7127709N and 22W 562147E, 7125097N (Fig. 3.6c).
Figure 3.6: Variations in supracrustal amphibolite compositions. a) Highly weathered largely garnet-absent amphibolite with bands of diopside-plagioclase and a fissile gneissic fabric (UTM 22W 563261E, 7125257N). b) Coarse garnet in massive, hornblende-rich amphibolite (UTM 22W561713E, 7125092N) d) Very coarse garnet in a more plagioclase-rich (dioritic composition) unit within the supracrustal amphibolite (UTM 22W 557848E, 7127709).

The amphibolite exhibits cm- to dm-scale compositional banding defined by modal differences in mafic and felsic minerals but the character of this variation does not vary greatly over scales of 10m or more. At UTM 22W 554002E, 7124965N, well developed cm-scale mafic-intermediate-felsic (with respective amphibolite-diorite-granodiorite compositions) banding defines the regional gneissosity (Fig. 3.7a). The degree of heterogeneity is largely dependent on the presence of garnet in the amphibolite, as garnet-amphibolites in Tummeralik are generally massive (Fig. 3.7b), whereas marked compositional banding is more apparent in garnet-poor or -absent amphibolite. Boudinaged layers of calc-silicate (diopside-plagioclase ± garnet), typically <10cm thick, occur in the compositionally banded garnet-absent amphibolite (Fig. 3.7c). Calc-silicate pods were not encountered within massive garnet amphibolite.
Figure 3.7: Variations in compositional banding in supracrustal amphibolite. a) Mafic-intermediate-felsic banding in possible volcano-sedimentary units constitutes a rare variant of the supracrustal package in Tummeralik. Banding is defined by differences in the modal abundance of hornblende and plagioclase ± quartz (UTM 22W 554002E, 7129665N). b) Massive amphibolite with the gneissic banding defined by deformed invasive granodioritic leucosome. (UTM 22W 562167E, 7125097N). c) Diopside-plagioclase pods and lenses parallel to the regional gneissosity (yellow boxed area in Fig. 3.6 - 22W 563261E, 7125257N). d) Felsic leucosome invading the supracrustal amphibolite and reacting with the host rock to form very coarse garnet (garnet diameters measure approximately 3-5cm).

The gneissosity in the amphibolite is defined by compositional banding and also by the orientation of plagioclase-quartz leucosomes, which form cm-thick anastomosing veins that are intruded broadly parallel to the regional fabric. The felsic leucosome often contains coarse, red garnet where it intrudes the amphibolite (Fig. 3.7d), suggesting intrusion during amphibolite facies metamorphism. This is the only garnet-bearing intrusive phase encountered in Tummeralik and the presence of garnet is attributed to interactions between the invasive veins and local host rocks.

At its boundary with the tonalitic orthogneiss, one of the major supracrustal amphibolite belts grades into a massive, homogeneous, leucocratic granodioritic lithology with virtually no mafic mineral content (Fig. 3.8a). Immediately adjacent to the supracrustal amphibolite, this lithology develops a characteristic hornblende
banding, which can be traced at least 1km along strike (Fig. 3.8b and c). The affinity of this unit is uncertain, however the boundary between the tonalitic orthogneiss and supracrustal amphibolite is highly tectonised, leading to the suggestion that the unit represents a hybridised boundary lithology. Alternatively, it may be a more felsic part of the supracrustal succession, perhaps of volcanic or volcanosedimentary origin.

Figure 3.8: Hybrid unit at the boundary between the supracrustal amphibolite and tonalitic orthogneiss. a) A homogeneous, massive, leucocratic granodiorite forms a minor lithology in the hybrid unit (UTM 22W 563456E, 7125096N), which grades into b) the hornblende-plagioclase-quartz-banded gneiss, showing a prominent gneissic banding defined by varying modal abundances of these three minerals (UTM 22W 561713e, 7125092N). c) Epidote-bearing brittle fractures in the hornblende-banded hybrid associated with $D_{JT}$ (UTM 22W 563456E, 7125096N).
Ultramafics

Ultramafic rocks were encountered as individual boudins (UTM 22W 560850E, 7124266N) or as strung out lenses or linear bodies in western Tummeralik. Boudins are typically 5-10m in length and are contained within or at the boundary of the supracrustal amphibolite. Boudins have a characteristic mottled green, honey and dark brown weathering colour, denoting the presence of chlorite, phlogopite, orthoamphibole and actinolite as metamorphic minerals replacing primary olivine and pyroxene. At UTM 22W 560850E, 7124266N a 4m long boudin in contact with the upper boundary of a garnet amphibolite layer has been altered to orthoamphibole, phlogopitic biotite and actinolitic amphibole (Fig. 3.9). Some boudins also contain considerable hornblende and at UTM 22W 563377E, 7124880N, a hornblendite boudin entrained in tonalite exhibits a marginal hornblende-corundum assemblage, suggesting local metasomatism between the boudin and adjacent orthogneiss.

Figure 3.9: Rare ultramafic boudin altered to orthoamphibole, biotite and talc at the boundary between the supracrustal amphibolite and tonalitic orthogneiss (UTM 22W 560850E, 7124266N). The brown weathering colouration is due intense weathering of orthoamphibole.

A single linear belt of dominantly ultramafic gneiss occurs at UTM 22W 550812E, 7124940N, where a ~40m wide belt strikes SE-NW for ~1km. As with other ultramafic occurrences in Tummeralik the unit is intensely altered to varying amounts of phlogopite, actinolite, and orthoamphibole but with some additional tremolite and diopside. The unit is intensely deformed and displays a schistose fabric as a result of the alteration described above. This schistose fabric is parallel to the regional gneissosity. The altered material is prone to physical disintegration, and
most ultramafic outcrops are surrounded by mica-rich gravel. No primary textures or structures are preserved in the ultramafic units in Tummeralik.

The only occurrence (in Tummeralik) of ironstone is encountered on the lower (northern) boundary of the large ultramafic body at UTM 22W 552232E, 7124551N, where a single 10m horizon of magnetite-quartz is documented. In places, the magnetite forms mm-scale layers interbanded with quartz, but otherwise forms massive magnetite. The lithology is similar to other documented ironstone deposits in the Nuuk region, which are themselves interpreted as chemical precipitates. The occurrence of magnetite-quartzite is thus considered to represent a metamorphosed chemical sediments and is considered (along with the associated linear body of ultramafic gneiss) to have a supracrustal origin.

Metapelitic and related rocks

Lenses, layers and boudins of pelitic and semipelitic material commonly occur throughout Tummeralik. These are usually hosted by or adjacent to garnet amphibolite, although a ~10m wide psammitic band does occur in isolation at UTM 22W 553321E, 7123971N. The largest and most continuous units are found within the major amphibolite belts that define the limbs of the Tummeralik Isocline. They occur mainly as strung-out, well-defined, boudinaged lenses a few tens of metres in length, and may be up to 20m wide and occasionally form laterally continuous layers, which can be traced up to 1km along strike. All pelitic and semi-pelitic rocks display a characteristic yellow- to rust-brown weathering colour and most outcrops (especially sillimanite-bearing units) are intensely weathered. Metasedimentary packages in Tummeralik have the general assemblage garnet-biotite-plagioclase-quartz ± sillimanite ± muscovite. The lithology is medium to coarse grained (0.5 to >1mm) and compositionally banded on very fine scales. Garnet occurs almost ubiquitously in the pelitic and semipelitic packages, where it usually forms red, euhedral porphyroblasts 2-20mm in diameter (Fig. 3.10a-e). Sillimanite is present mainly as acicular needles concentrated into 1-2mm thick sillimanite-biotite-rich foliae (Fig. 3.10b and e), which alternate with garnet-bearing quartzofeldspathic bands 1-3mm thick (Fig. 3.10d). This small-scale compositional variation is repeated on m-scales, producing an alternating pelite-semipelite sequence. No primary
sedimentary structures are preserved. Regular 1-3mm compositional banding is best developed in pelitic lithologies. Sillimanite-poor or absent compositions tend to be more massive, with little variation in the modal abundances of constituent minerals.

Figure 3.10: Variation in pelitic units in Tummeralik. All photographs except 'b' are taken from UTM 22W 557239E, 7127886N. a) Coarse garnets in pelite surrounded by coarse quartz-plagioclase in low strain zones associated with the garnet porphyroblasts (looking onto foliation surface). The characteristic rusty weathering colour is most prominent in sillimanite-bearing areas. b) Prominent sillimanite (white arrow) defining the regional lineation (orientation delineated by the yellow arrow) in garnet-sillimanite gneiss (UTM 22W 562757E, 7125171N). c) Paler rust brown weathering pattern typical of sillimanite-absent semi-pelite. The hammer handle measures approximately 6cm length. d) The $S_{1-2}$ gneissic fabric (running parallel to the pencil) defined in a garnet-sillimanite gneiss by aligned biotite, sillimanite and plagioclase-quartz pressure shadows associated with garnet. e) Coarse garnet porphyroblasts wrapped by the biotite-sillimanite fabric in a biotite-rich pelitic gneiss.
Pelitic compositions also tend to contain coarse quartz-plagioclase lenses, which are parallel to the regional gneissosity on mm- to cm-scales throughout the study area (Fig. 3.10c). Biotite is the principal foliation-forming mineral in almost all pelitic and semipelitic horizons in Tummeralik, with muscovite occurring in addition to biotite in a few localities. Where sillimanite is present, it always forms a prominent lineation, which in all cases wraps garnet porphyroblasts, suggesting that sillimanite growth largely progressed subsequent to garnet growth.

**Pegmatites**

All lithologies in Tummeralik are cut by three generations of granodioritic to granitic pegmatite (Fig. 3.11a and b). The earliest generation, which forms the voluminous pegmatite intruding the tonalitic orthogneiss, is granodioritic and forms cm- thick sheets and vein networks parallel to the regional foliation, although cross-cutting relationships are seen in areas of relatively low strain. The pegmatite is deformed by the peak deformation events and exhibits a biotite foliation, parallel to the regional gneissosity. As a result of this deformation, the pegmatites have been recrystallised to some degree, although average grain size (2-3mm) is still coarser than that of the host tonalitic orthogneiss. The first generation pegmatite, hornblende-bearin g leucosome in the tonalitic orthogneiss and garnet-bearing leucosome in the supracrustal lithologies are all intruded parallel or sub-parallel to the regional gneissosity (except in low-strain zones surrounding enclaves of more competent material, such as the homogenous amphibolite horizons in the tonalitic orthogneiss) and so the three types of felsic leucosome are tentatively considered to have been intruded prior to or during the development of the main gneissosity (Fig. 3.3 and 3.7). The relationship between the first generation pegmatite, hornblende- and garnet-bearing leucosomes cannot be constrained, as the different leucosomes were never encountered in the same outcrop. Consequently no cross-cutting relationships could be established. However it remains possible that the development of three distinct leucosome groups may be the result of reaction between a single felsic leucosome generation and host rocks of varying bulk composition. This is exemplified by the development of hornblende-bearing leucosome only in the tonalitic orthogneiss and garnet-bearing leucosome only in the supracrustal units.
Figure 3.11: a) Three generations of granodioritic pegmatite intruding the supracrustal amphibolite at UTM 22W 563456E, 7125096N). b) Field sketch of ‘a’, showing generation 1 pegmatite parallel to the gneissosity in the host rock. This is cut at a high angle by recrystallised generation 2 pegmatite. Coarse grained, undeformed generation 3 pegmatite cuts both earlier pegmatites but runs approximately parallel to generation 1.

The second pegmatite generation also has granodioritic to granitic compositions, but tends to form vein systems ∼30cm in width. These are less voluminous than the earliest pegmatite generation. They also show some evidence of partial recrystallisation but do not exhibit a gneissosity-parallel biotite foliation; they may form at a high angle to the regional fabric. Late granitic intrusions form sheets up to 1m thick. These cut all other intrusions and fabrics in Tummeralik and show no evidence of recrystallisation.

Mafic intrusions

A number of undeformed mafic (doleritic to picritic) dykes are encountered in west Tummeralik, which vary in width from 10-30m, with a subvertical dip and approximately north-south trend. The picritic dykes contain orthopyroxene and altered olivine with very little plagioclase and tend to be very coarse grained (>1mm). They are intensely weathered and disintegrate to form characteristic sculpted outcrops (Fig. 3.12a), often surrounded by orthopyroxene gravel. The
doleritic dykes tend to form the largest intrusions and comprise plagioclase and pyroxene, which is coarse grained in the central intrusion but forms a pronounced chilled margin (Fig. 3.12b). The dykes are segmented, overstep each other at their terminations and are undeformed by ductile deformation, although mapping of adjacent areas shows them to be offset by brittle faults (see Appendix D).

Figure 3.12: a) Characteristic sculpted outcrops of late, undeformed mafic dykes surrounded by weathered orthopyroxene gravel (UTM 22W 552577E, 7106979N). b) chilled margin with slightly coarse plagioclase (cream spots) phenocrysts (UTM 22W 551798E, 7106964N). These photographs are the best exposed examples of features associated with the late mafic dyke suite. They were taken to the south of the Tummeralik mapping area but similar structures are encountered in most mafic dykes in Tummeralik.

3.3 Reconnaissance area southeast of Tummeralik

An area to the south and east of the main Tummeralik mapping area was visited during helicopter reconnaissance (Fig. 3.1 inset). The area was chosen for reconnaissance sampling because previous work had suggested the presence of large outcrops of supracrustal amphibolite and metapelitic lithologies. The reconnaissance area appears to be structurally continuous with lithologies to the west (Appendix D), but the relationship between Tummeralik and the reconnaissance area is not apparent from previous maps. Based on field observations the NW-SE trending fault, which runs across Tummeralik, was hypothesised to continue into the reconnaissance area. The area consequently holds considerable potential for constraining the tectono-thermal events that led to both peak and post peak metamorphism and deformation, in addition to investigating the relationship between this area and neighbouring Tummeralik.

The reconnaissance area is dominated by a tonalitic orthogneiss very similar to the tonalitic orthogneiss in Tummeralik. Peak metamorphic assemblages of
biotite-plagioclase-quartz form a mm-scale gneissic banding, which is interspersed on cm-dm scales by felsic pegmatite material that is intruded parallel to the regional gneissosity (Fig. 3.13a). By contrast to the tonalitic orthogneiss of Tummeralik, much of the biotite in the reconnaissance area tonalitic orthogneiss has been replaced by chlorite, suggesting that the greenschist facies retrogressive event(s) are more widespread in the reconnaissance area than in Tummeralik.

Mapping in the northeast of the reconnaissance area identified a large linear body of amphibolite, with significant amounts of metapelitic material incorporated within the amphibolite. These metapelitic rocks were discrete from those encountered in Tummeralik and Norsanna in that the general assemblage was garnet-sillimanite-(kyanite)-muscovite-biotite-plagioclase-quartz, whereas pelitic rocks in Norsanna and Tummeralik tend to be muscovite-absent. Garnet is present in varying amounts and occurs as small (1-2mm) purple porphyroblasts. The rocks were also extremely fresh, despite showing very intense pyrite, bornite and sulphur surface weathering (Fig. 3.13b). The outcrop patterns of pelitic units in the reconnaissance area are generally oriented layer-parallel, but form irregularly-shaped lenses and patches with an apparent gradation from garnet-amphibolite to pelite. The bodies pinch and swell and become discontinuous in all orientations, not just in the direction of maximum attenuation. By contrast, pelitic units in Tummeralik and Norsanna occur as sharply-defined, boudinaged lenses. The irregular geometry of pelitic bodies in the reconnaissance area bear closer similarities to those on the island of Storø to the northwest and those of Nunataarsuaq, a nunatak ~12km to the NNE still within
the Kapisillit Mapsheet (Hollis and Bennett Pers. Comm.). These pelitic units have been interpreted as metasomatised amphibolitic rocks due to the presence of alteration fronts at the boundary between amphibolitic and pelitic rocks (Chapter 1.7). As a consequence of these observations no correlation is drawn between pelitic units in the reconnaissance area and Tummeralik, however the pelitic assemblages are still of importance in constraining the pressure-temperature history of the area, and so are still considered important samples for petrological study.

3.4 Structural evolution of Tummeralik

The Tummeralik field area has been subject to multiple phases of deformation at amphibolite facies and further deformation under greenschist facies conditions. The area is cut by a number of small scale high strain zones, however there is no evidence to suggest the presence of regionally significant tectonised boundaries. A summary of the structural evolution of Tummeralik is given in Table 3.1.

Gneissosity forming events ($D_n$)

An approximately E-W striking gneissosity is exhibited by all lithologies in Tummeralik except pegmatite generations 2 and 3 and the late mafic dyke suite. The gneissosity is principally displayed by the tonalitic and dioritic orthogneisses and supracrustal units, where it is defined by the orientation of biotite and hornblende and by mafic-quartzofeldspathic banding. The lower state of strain apparent in the granodioritic orthogneiss relative to the other orthogneisses and supracrustal units may indicate a later, post-$D_n$ emplacement age. This $S_n$ gneissosity is typically planar, although some rare isoclinal folds are preserved that are dismembered by later events. The $S_n$ gneissosity is folded by folds described here as $D_1$, however the exact number of events leading to the development of $S_n$ is uncertain.
Table 3.1: Table showing the structural evolution of Tummeralik

<table>
<thead>
<tr>
<th>Event</th>
<th>Grade</th>
<th>Description</th>
<th>Affected lithologies</th>
</tr>
</thead>
<tbody>
<tr>
<td>$D_n$</td>
<td>?</td>
<td>Initial formation of gneissic fabrics and compositional banding. Unknown number of events.</td>
<td>Tonalitic orthogneiss and entrained homogeneous amphibolite. Dioritic orthogneiss Supracrystal units?</td>
</tr>
<tr>
<td></td>
<td></td>
<td>No open folds observed</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Granite pegmatite intrusion late syn- to post $D_2$</td>
<td></td>
</tr>
<tr>
<td>$D_3$</td>
<td>Greenschist facies</td>
<td>Tummeralik - Brittle faulting with small-scale, epidote/chlorite-bearing mylonitic textures. Reconnaissance area – Epidote-chlorite (blasto-) mylonitic shear zones with moderately to steeply-plunging open folds of gneissosity and well-pronounced associated lineation. Greenschist facies faults zones truncate the late mafic dykes to the north of Tummeralik</td>
<td>All units except: Mafic dykes?</td>
</tr>
</tbody>
</table>

Isoclinal folding ($D_1$): The Tummeralik Isocline

The $D_1$ event resulted in the formation of the Tummeralik Isocline (Fig 3.1); a major fold in the $S_0$ gneissosity that dominates the outcrop geometry of all units in Tummeralik, indicating that intrusion of granodioritic orthogneiss predates the onset of $D_1$. Poles to gneissosity data define a profile plane of 209/52NW, from which the plunge of the large-scale fold has been calculated at 119/38 (Fig. 3.14). This moderately inclined and plunging structure closes westwards, indicating a synformal
geometry. No corresponding antiformal structure was encountered in the field area. Parasitic $F_1$ folds are rare, so that the dominant fabric dips $\sim 60^\circ$ towards the south, however structural data from minor fold vergences encountered at UTM 22W 557629E, 7126643N are consistent with the values predicted in Fig. 3.14. The limbs of the Tummeralik Isocline are defined by outcrops of supracrustal amphibolite, (semi-) pelitic, and minor ultramafic material, which can be traced from the hinge zone in central Tummeralik (Fig. 3.1) due east for $\sim 7$km. The supracrustal units that define the isocline suggest a fold wavelength between 1 and 2km. Isoclinal folding during $D_1$ has resulted in the development of the axial planar $S_1$ foliation parallel to the dominantly planar $S_n$ fabric, thus forming a composite $S_n$-$S_1$ gneissic fabric. $L_1$ sillimanite lineations in pelitic units and anastomosing garnet-plagioclase leucosomes in supracrustal rocks, which run broadly parallel to $S_1$ suggests upper amphibolite facies conditions were reached during $D_1$.

![Figure 3.14: Poles to Sn gneissosity defining the profile plane of the Tummeralik Isocline.](image)
Boudinage and fold limb attenuation ($D_2$)

$D_2$ involved a period of stretching, which resulted in the intensification of the $S_n$-$S_1$ gneissosity and the attenuation of fold limbs to the Tummeralik Isocline, causing large scale boudinage on the limbs of the $F_1$ fold and the rotation of the fold hinge into parallelism with the regional $D_2$ fabric (UTM 22W 557327E, 7125629N). Hence, by the end of $D_2$, the regional fabric is a composite gneissosity of $D_n$, $D_1$ and $D_2$. There is no evidence of $F_2$ refolding of $F_1$ structures. The $D_2$ event is associated with the strain locking and shearing of rare parasitic $F_1$ fold hinges at UTM 22W 571204E, 7119195N (Fig. 3.15) and the concentration of strain at lithology boundaries.

![Figure 3.15: Strain locked isoclinal fold hinge in the tonalitic orthogneiss in the Ice Lake area (UTM 22W 571204E 7119195N). Hornblende-plagioclase leucosomes are strung parallel to the gneissosity and truncated by the sheared limb.](image)

The attenuation and boudinage of pelitic units during $D_2$ has in many cases, resulted in a slight block rotation, demonstrated by a spread in $L_1$ sillimanite lineation angles form the calculated hinge orientation for the Tummeralik Isocline (Fig. 3.16). This is also supported by field evidence, in which foliation and lineation orientations in pelitic boudins are discordant (by 10-20°) to the fabrics in the host material. Strain concentration at lithology boundaries has resulted in the formation of 1 to 10m wide high strain zones with m-scale tectonic interleaving. No high strain zones wider than this are encountered within the Tummeralik area. However the composite $S_{n-1-2}$ fabric intensifies and all units become parallel towards the south of the field area.
Greenschist retrogression ($D_3$)

A phase of faulting and shearing took place post-$D_2$, and is associated with localised chloritisation and epidote staining adjacent to fractures. Small-scale, branching fractures with cm- to m-scale displacement are seen throughout much of Tummeralik, and are particularly marked in the leucocratic, 'hybrid' lithology at the boundary between the supracrustal amphibolite and tonalitic orthogneiss at UTM 22W 563456E, 7125096N (Fig. 3.8b). However the extent of greenschist facies strain increases considerably moving eastwards. The largest $D_3$ structure in Tummeralik is a NW-trending fault with a normal and sinistral sense of motion, which dips moderately (~50°) to the SW. Offset on this structure exceeds 50 m at UTM 22W 561713E, 7125092N, and preferential weathering of the fault plane creates a lineament that can be traced for several kilometres NW of the mapping area. The character of $D_3$ deformation changes from brittle to ductile moving towards the SE and becomes mylonitic at UTM 22W 561498E, 7125138N. This change in deformation character is also associated with increasing frequency of epidote veining and chloritisation in the country rocks. The reconnaissance area SE of Tummeralik
lies directly along strike of the large D₃ fault and although the continuation of the fault itself cannot be seen, the area as been subject to intense greenschist facies deformation and thus exposes an arguably deeper crustal section at the time of D₃. The tonalitic orthogneiss in the reconnaissance area has developed a chlorite schistosity and the planar composite S₁-2 gneissose has been further intensified and transposed into this (Fig. 3.17a). The ENE-trending regional gneissosity is rotated into an approximately S- to SE-trending attitude with associated S-plunging L₃ stretching lineations (180/38) defined principally by quartz in the tonalitic orthogneiss and in quartz veins. SE-trending protomylonitic fabrics are locally developed at UTM 22W 567133E, 7115566N (Fig. 3.17b). D₃ deformation or alteration in the mafic dyke suite was not observed in Tummeralik, however the large D₃ fault zone that runs across Tummeralik displaces at least one mafic dyke to the northwest of the field area, indicating that parts of the mafic dyke suite had been intruded prior to D₃ in Tummeralik.

Figure 3.17: D₃ greenschist facies deformation in the reconnaissance area a) Intense gneissic foliation in tonalitic orthogneisses of the reconnaissance area that has been transposed from an initial strike of approximately E-W to a post-D₃ strike of approximately NW-SE. Matrix biotite at this locality has also been partially replaced by chlorite, indicated by the grey-green colour of the rocks (UTM 22W 567923E, 7115579N). b) Protomylonitic fabrics in the tonalitic orthogneiss with plagioclase augen (UTM 22W 567133E, 7115566N).
3.5 Summary

The evidence discussed in the previous section indicates that Tummeralik has evolved as a single crustal block throughout most, if not all of its history. The following section presents a summary of the key events in the evolution of Tummeralik.

1. Intrusion of the tonalitic and dioritic orthogneisses prior to Dn. The character of the country rock into which the tonalitic orthogneiss was intruded is uncertain. However rare truncated folds in the homogeneous amphibolite horizons in the tonalitic orthogneiss suggest that at least some of these bodies may predate the main orthogneiss suite. These folds may also indicate that at least one stage of Dn occurred prior to the intrusion of the tonalitic orthogneiss.

2. Deformation and tectonic intercalation of tonalitic and dioritic orthogneisses and supracrustal units leading to the development of the Sn gneissosity.

3. Potential intrusion of the granodioritic orthogneiss unit into tonalitic, dioritic and supracrustal gneisses post-Dn.

4. N-S directed shortening during D1 resulted in the further intercalation of the orthogneiss and supracrustal units and km-scale isoclinal folding, forming the Tummeralik Isocline. Upper amphibolite facies reached by end-D1. Intrusion of early pegmatite.

5. Near coaxial D2 shearing resulting in F1 fold limb attenuation, strain focussing at lithology boundaries and block rotation of metapelitic boudins on fold limbs. Upper amphibolite facies in the sillimanite field maintained during D2. Intrusion of garnet-bearing felsic leucosome and post-tectonic granitic pegmatite.

6. D3 greenschist facies NW-SE trending faulting and mylonitisation. Faults have a sinistral sense of motion. D3 displaces mafic dykes to the northwest of Tummeralik, but the intrusion of the mafic dyke suite continues post-D3.


Hollis, J. and Bennett, V. Personal Communication 2006. Informal field discussion.
CHAPTER 4

THE TIMING OF ORTHOGNEISS EMPLACEMENT AND METAMORPHISM IN THE KAPISILLIT REGION

4.1 Overview

U/Pb geochronology studies of zircon have been of critical importance in the development of the Nuuk region terrane hypothesis, with key works in this area based on the SIMS geochronology of magmatic zircon populations in key orthogneiss suites. Ages of metamorphism for the different terranes have also been identified, often as secondary age populations within orthogneiss samples (Friend and Nutman, 1994; Nutman et al., 2000; Nutman et al., 2004; Nutman et al., 1996; Whitehouse et al., 1999). Targeted sampling of migmatite, pegmatite and leucosome material has also been employed in a number of studies (Nutman et al., 2007; Nutman and Friend, 2007; Whitehouse and Fedo, 2003; Whitehouse and Kamber, 2005) in order to establish timings of metamorphism for parts of the Nuuk region.

The large variation in magmatic and metamorphic zircon populations within the Kapisillit region suggests a structurally and isotopically complex area, which is better constrained in some parts than others. For example, orthogneiss emplacement ages and timings of metamorphism are well constrained in (and to) the north and west of the Kapisillit Mapsheet and are the result of a number of studies in the Norsanna area. However, few studies in Norsanna have focussed on the timing of motion and metamorphism on the Norsanna High Strain Zone (hereafter NHSZ), which has long been proposed as a part of the purported Færingehavn-Tre Brødre terrane boundary. Samples from Norsanna are thus directed at confirming the emplacement age of orthogneisses on either side of the NHSZ and constraining the timing of motion and associated metamorphism on the NHSZ itself.

Geological mapping has identified a pronounced structural break running approximately parallel to the east shore of Itilleq Fjord that separates approximately N-S and E-W striking structures (see Appendix D). The presence of $>3600\text{Ma}$ and ca. $2825\text{Ma}$ orthogneisses to the west and ca. $3000\text{Ma}$ orthogneisses to the east of this break have identified the area as a potential terrane boundary, with the Eo-
Neoarchaean Færingehavn-Tre Brødre to the west and Mesoarchaean Kapsillik terrane to the east (Nutman and Friend, 2007). However the geochronological record in the area east of this boundary remains poorly studied away from Kapisillit Fjord and the ages of a large proportion of orthogneisses in the central, south and east of the Kapisillit region remain currently unconstrained. An understanding of the age of the dominant orthogneiss suite(s) in this area is consequently vital in furthering our understanding of the region’s evolution. Similarly, there are few constraints placed on the metamorphic history of the central Kapisillit region. The samples from Tummeralik and one sample from the northern shore of Ameralla Fjord (in between Norsanna and Tummeralik) address the question of orthogneiss age and metamorphic history in parts of the central and east Kapisillit Mapsheet using targeted and structurally constrained orthogneiss and leucosome samples.

4.2 Methodology

Bulk samples of 1-4kg were crushed in a tungsten-carbide jaw crusher. Zircons were separated from <500μm coarse crush using sequential Wilfley (gravity), heavy liquid (TBE) and Frantz magnetic separation at the mineral separation laboratory at St Andrews University and GEOTRACK, Victoria, Australia.

Once zircon fractions had been obtained, grains were hand-picked and their dimensions measured prior to mounting. Using a microdrill, one trench per sample was drilled into one inch epoxy mounts. The depth of a trench was determined by the typical depth of the zircon fraction of that sample. Where any single sample contained a range of zircon depths, a stepped topography was drilled in the trench bottom so that the mid-point of each zircon was approximately level, regardless of the absolute grain size (Fig 4.1). Each mount held between one and three samples. During mounting, zircons were held in place in their trenches by a thin film of epoxy resin before being backfilled and placed under vacuum to remove air bubbles from the resin. The advantage of mounting grains using this method allows all zircons to be polished to the same level and reduces the risk of the largest grains being polished away or plucked before the smallest zircons are exposed. Once mounted and polished, the grains were imaged under BSE and CL using the Philips XL30CP SEM.
facility at the School of Geosciences, University of Edinburgh. The importance of SEM imaging in the interpretation of zircon growth patterns and the identification of poor quality analyses is discussed in the following section.

Figure 4.1: a) Trenches of differing depths are drilled in 1 inch epoxy rounds that reflect the size of zircon morphologies in sample. Trench depths are calculated so that the centres of all grains lie at approximately the same depth (shown by the 'mid-point'). During mounting, zircons are held in place by a thin film of epoxy resin. b) After mounting, trenches are back-filled with epoxy resin and polished to the ‘mid-point’, exposing central sections of all zircons in the mount.

Zircon dating using Laser Ablation Inductively Coupled Plasma Mass Spectrometry (LA-ICPMS) was carried out on seven selected samples from Norsanna and Tummeralik. The dating of zircons in the Nuuk region has previously been largely obtained using SIMS geochronology. This study carried out some of the first analyses on the new LA-ICPMS facility at GEUS and the University of Copenhagen in early 2005. The technique has a key advantage over SIMS geochronology in that a much smaller analysis time is required (around two minutes per analysis) and entire sessions can be automated provided periodic checks for beam and peak drift are carried out. This allows a large amount of data to be collected in a short amount of time and the technique is consequently much quicker and cheaper than SIMS. However the beam size used in laser dating is highly destructive compared to standard ion beam techniques, often creating an ablation pit several tens of microns deep as opposed to ~6μm for SIMS (Fig 4.2). This presents a particular problem for complex zircons, as the technique carries a high risk of hitting domain boundaries, which may result in hybrid and geologically meaningless isotopic age. A
large beam size also increases the risk of introducing common Pb ($^{204}$Pb) into the analysis because of the greater likelihood of the beam hitting metamict zircon or a fracture. The combination of a short analysis time and a (potentially) high common Pb contamination usually results in higher errors on individual analyses for samples dated using LA-ICPMS when compared to samples analysed using SIMS. However, careful analysis of SEM imagery and consideration of the strength of the beam during analysis (a poor, unsteady response suggests damaged zircon) greatly increases the accuracy of results. This means that errors on population ages may be as low as $\pm 5$Ma ($2\sigma$), although on average, population errors tend to be around $\pm 15$Ma ($2\sigma$). The high probability of hitting a fracture, domain or grain boundary during an automated session (where the beam can drift significantly over the course of an analysis session) also highlights the importance of SEM imaging after analysis, to identify any potentially hybrid ages in a dataset.

Figure 4.2: Ablation pits in zircon analysed using a) SIMS (Secondary Ionizing Mass Spectrometry) and b) LA-ICPMS, showing the latter to be a more destructive technique.

U/Pb isotopic analysis was carried out using the LA-SF-ICP-MS facility at GEUS, Copenhagen, using a NewWave Research/Merchantek UP213 frequency quintupled Nd-YAG laser emitting at a wavelength of 213 nm coupled to an Element2 (ThermoFinnigan) single-collector double focusing magnetic sector field ICP-MS. Automated analysis sessions used a beam diameter of 20-30$\mu$m, starting and finishing with six analyses of the GJ-1 zircon standard, with three standard analyses for every ten unknowns during the course of the session.

Data were reduced using the Zirchron age-calculation spreadsheets developed in-house by GEUS (Frei and Scherstén, 2006) and final analysis using IsoplotEx 3.0 (Ludwig 2003). The following decay constants were used with a $2\sigma$ (%) uncertainty:
$^{238}U = 1.55125 \times 10^{-10} \pm 0.107$, $^{235}U = 9.8485 \times 10^{-10} \pm 0.136$ and $^{232}Th = 4.9475 \times 10^{-11} \pm 0.3$ (from $^{230}Th$) with a natural $^{238}U/^{235}U = 137.88$. Uncertainties on ages quoted in the text and in tables from individual age analyses are at the 1σ level. All uncertainties in calculated group ages are reported at 2σ confidence levels. Concordia diagrams are plotted according to the textural and zoning populations described per sample. Further details of instrument set-up are given in Appendix A and isotopic data and calculated ages are presented in Appendix B.

SEM imagery and zoning patterns in zircon

Scanning Electron Microscopy was used to establish the number and type of zircon morphologies present in each sample prior to isotopic analysis. The identification of discrete zoning patterns in zircon using backscattered electron (BSE) and cathodoluminescence (CL) imaging is vital throughout the age dating process. It provides not only an initial interpretation of zircon origin, but is also critical in identifying target areas and ensuring the accurate analysis and interpretation of isotopic data. The following section examines the criteria used to distinguish different zircon morphologies and outlines how SEM imagery may be used in identifying poor analyses that should not be used in age calculations.

BSE displays contrasts in the average atomic number across a sample, where areas of higher atomic number appear brighter than areas of relatively low atomic number. In zircon, variations in BSE response are chiefly influenced by Hf and U concentrations (Hanchar and Rudnick, 1995). CL images displays the luminescence given off by the excitation of certain ions in the crystal lattice. Used in tandem, these two methods can reveal the growth history of a mineral. Zircon growth occurs under a number of conditions, but zircon formed by magmatic and metamorphic crystallisation is only considered here as these account for the vast majority of the mineral in the Kapisillit region samples (discussed below).

Zircons in an igneous rock can undergo recrystallisation during metamorphism by dissolution-reprecipitation, or by point defect migration during sub-solidus recrystallisation (Cherniak and Watson, 2003). Thus, zircons in polymetamorphic lithologies may be expected to preserve complex zoning patterns.
These can be distinguished using established models of zoning in zircon (Corfù et al., 2003), which are detailed below.

![Figure 4.3: Typical attributes of magmatic and metamorphic zircon in CL and BSE. Scale bars measure 100µm. a) Magmatic zircons (CL) with elongate (left) to acicular (right) morphologies, exhibiting faint oscillatory zoning and poor CL responses. b) The same zircons as ‘a’ shown in BSE. Oscillatory zoning is picked out by zones of damage (fracture and radiation damage) in certain bands. c) Equant-elongate magmatic zircon showing well-pronounced oscillatory zoning and small pyramidal terminations (white arrow). Strong CL response. d) High grade metamorphic zircon with an ovoid morphology and moderate CL response. Sector zoning is shown by the white arrow. e) Typical sub-ovoid to ovoid metamorphic zircons from a felsic leucosome. White arrow shows patchy zoning in addition to sector zoning. All zircon images obtained from samples dated by the author.](image)

Primary magmatic zircons typically display tight, sharply-defined bands that are concentric about the grain centre. These are best-defined in CL, as they are characterised by sharply alternating bands of strong (bright) and weak (dark) response. The bands represent progressive growth zoning in a melt and the concentric banding is termed ‘oscillatory zoning’ (Fig 4.3 a-c). Primary magmatic grains may also be characterised by euahedral, prismatic and elongate crystals with pyramidal end terminations (Corfù et al., 2003; Pidgeon, 1992). However a
magnetic interpretation should not be made on external morphology alone as subsequent recrystallisation and Pb loss in an initially magmatic grain may result in an isotopic age unrelated to magmatic emplacement. Where this occurs, recrystallisation can usually be identified by the blurring, truncation or disappearance of oscillatory zoning.

Metamorphic zircons are far more variable in terms of morphology and zoning pattern. They may form as new grains, as mantles on pre-existing cores or from the sub-solidus recrystallisation of a pre-existing zircon. In the Kapisillit sample suite, most interpreted new metamorphic grains form subovoid to ovoid grains (Fig 4.3d and e) and are the dominant morphology in felsic leucosome samples. These morphologies are typical of high grade zircon crystallised in the presence of felsic leucosome (e.g. Corfu et al., 2003). In some cases however, zircon grown in a metamorphic leucosome may exhibit various morphologies, and may sometimes even resemble magmatic zircon (although oscillatory zoning is not common. See sample 481438).

A nomenclature problem exists in the classification of metamorphic zircon as zircon crystallised from a melt is technically magmatic, regardless of whether the melt is igneous or metamorphic in origin. It would thus be correct to refer to such zircon as igneous zircon crystallised from metamorphic leucosome. However for ease of discussion, zircon recording the age of crystallisation of a metamorphic leucosome is referred to in this work as ‘metamorphic’.

In terms of BSE and CL response, zoning in metamorphic zircons is typically more subtle than oscillatory zoning, often forming regular ‘sector zoning’ or more irregular ‘patchy zoning’ (Fig 4.3d and e respectively). Due to the subtlety of zoning, these features are often only visible in CL. It is also possible to identify ‘ghost’ or ‘relict’ zoning, where recrystallisation has partly destroyed a pre-existing zoning pattern (Fig 4.4a). This is most common in recrystallised magmatic zircon, where relict zoning may be manifest as faint, occasionally wavy bands with indistinct edges that run approximately parallel to grain boundaries. Where this type of zoning is regularly spaced throughout the grain or zircon domain, it is termed ‘planar banding’ (Fig 4.4b).
Figure 4.4: Scale bars measure 100μm. a) CL image of a recrystallised rim mantling an oscillatory zoned magmatic zircon. Relict zoning occurs in the rim as faint concentric lines following the line of zoning in the core. b) Thick ‘planar bands’ in zircon run parallel to the grain boundary. These may represent either metamorphic growth zoning or relict zoning in a recrystallised magmatic zircon.

Figure 4.5: a) CL image of a complex zircon showing a rounded, inherited core with a significantly older $^{207}\text{Pb} / {^{206}}\text{Pb}$ age than that of the metamorphic rim. b) Complex zircon with an oscillatory zoned core mantled by a recrystallised rim. The bright spot above the scale bar is a laser analysis pit that inadvertently sampled both the core and rim domains. c) BSE image of a stubby, euhedral magmatic grain with well-pronounced pyramidal end terminations and metamict core. The centre of the grain exhibits chaotic zoning, characterised by swirling zoning patterns.
In polymetamorphic zircon from meta-igneous rocks, grains may often be cored by magmatic zircon and mantled by metamorphic material. In an undisturbed grain, the core will preserve the magmatic age and the rim the age of metamorphism (Fig 4.5a). Care should be taken to analyse complex grains as far from domain boundaries as possible, as an analysis that hits zones of different ages will yield a hybrid and geologically meaningless age. Some analyses will inevitably sample material from a different textural domain, be it the result of beam drift or an unseen domain boundary running just below the target area (Fig 4.5b). This highlights the importance of post-analysis imaging, so that data from these points is not erroneously interpreted.

In some cases, all or part of a zircon may be metamict. This is caused by accumulative radiation damage to the zircon lattice over time (Meldrum et al, 1998) and is characterised in BSE and CL by 'chaotic' zoning (Fig 4.5c). In undamaged zircon, bright BSE images will typically be accompanied by weak CL responses and vice versa. However where the structure of a zircon lattice is seriously compromised, for example through metamictisation or break-down reactions, grains are characterised by poor responses in both BSE and CL. Such areas are to be avoided during analysis because of the high risk of contaminants such as common Pb in the damaged lattice, which may result in anomalously old ages.

The interpretation of isotopic data

Age data are graphically presented in this thesis using U/Pb concordia diagrams. As noted above, all analyses are grouped according to the textural group that has been ascertained prior to isotopic analysis using SEM imagery. Each textural group in a sample is assigned a colour, but all textural groups in a single sample are plotted on the same diagram in order to put the entire dataset in context. Where two textural groups can be argued to record the same event, a combined age is calculated, otherwise ages are calculated using individual textural groups. Rejected analyses are also presented on the concordia diagram and are always plotted as unfilled black circles (filled black circles are used occasionally and are labelled accordingly). The reasons for their rejection are defined in the analysis description for that sample and the data are not used in age calculations. However they are presented graphically as
their presence may, in some cases, play an important role in the interpretation of the sample. This is especially the case in rocks where zircons have undergone multiple phases of Pb loss. The following section examines some of the isotopic features that can be produced in complex zircon populations, using a hypothetical sample with magmatic and metamorphic zircon populations.

As the zircon lattice does not easily accommodate Pb, the element will be expelled from the lattice during recrystallisation along recrystallisation fronts or through fluid-enhanced diffusion along fractures. This type of Pb loss is commonly associated with renewed thermal activity in the host rock or events that instigate physical fracture, such as exhumation (Chen et al., 2002). In an unaltered rock containing one magmatic and one metamorphic population, the U/Pb concordia diagram displays two discrete populations; one recording the age of emplacement and one recording the age of metamorphism (Fig 4.7a). If the U/Pb system is reset by recent Pb loss (time = ~0 years), the affected analyses tend to track towards zero (i.e. the age of disturbance) on the U/Pb concordia and will become discordant. In this situation a line can be projected from zero, through the discordant data. The point where the projected line intercepts the concordia curve is taken as the age of that event (Fig 4.7b).

The interpretation of samples that involve a Pb loss event at some point in the past become more complex, as discordant data track towards an age of disturbance that is not the present day (Fig 4.7c). In this case, the upper intercept of the cord should still represent a minimum age of emplacement, whereas the lower intercept ideally represents the age of disturbance. Where the concordia has a shallow curve, it remains possible for zircons experiencing ancient Pb loss to maintain an apparent concordance, leading to large scatters of data (high MSWD) along the concordia. It should also be noted that thermal events that cause new zircon growth may also be responsible for ancient Pb loss in an older population.

The most complex geometries on concordia occur where a sample has experienced both ancient and recent Pb loss (Fig 4.7d). Where this occurs, data from pre-existing zircons (i.e. the magmatic population) are pulled back along an ancient Pb loss cord towards the new metamorphic zircon population (Fig 4.7d, ‘1’). Following this, recent Pb loss causes analyses from both populations to track towards
zero (Fig 4.7d, ‘2’). This creates a fan-shaped geometry and makes it impossible to confidently use a concordant population or upper intercept age as the age of emplacement. Instead, a minimum age of emplacement can be estimated from the oldest concordant magmatic grain. All analyses that are demonstrably affected by both episodes of Pb loss are removed from age calculations, as they will likely preserve a hybrid age. Because of the problem of recent Pb loss superimposed on ancient Pb loss in the Kapisillit sample suite, caution is exercised when estimating ages of ancient Pb loss events. Lower intercepts are not used to interpret the age of ancient Pb loss the position of the lower intercept may itself have been disturbed by the recent Pb loss event.

Figure 4.6: Schematic U/Pb concordia demonstrating different Pb loss scenarios. a) In a closed an isotopic system, different zircon populations will lie on the concordia line at the point where the U/Pb system on crystallisation. b) Recent Pb loss wholly or partially resets the U/Pb values and datapoints track towards the graph origin. c) Ancient Pb loss causes the datapoints to track towards the age of resetting, resulting in a smear of apparently concordant points along the concordia line. d) Combined ancient and recent Pb loss, where datapoints initially track towards the ancient resetting event (1). All datapoints will track towards zero during recent Pb loss (2) resulting in a ‘fan shaped’ geometry.
Some samples in the Kapisillit suite exhibit varying degrees of reverse discordance, which occurs where data points plot above the concordia curve and yield a concordance >100%. This amounts to a relative loss of U or gain in Pb and is difficult to reconcile in zircon. This is because U diffusion coefficients in zircon are very sluggish and the zircon lattice will not easily accommodate Pb, effectively hindering the diffusive loss of U and preventing the lattice from taking in excess Pb. Occurrences of reverse discordance may be the result of inadvertent sampling of Pb-bearing material in microfractures or small inclusions. Another potential source may be the sampling of Pb-bearing reaction products if a metamict grain is analysed (Chen et al., 2002; Williams et al., 1984). This last scenario is the preferred interpretation for most analyses showing >105% concordance (and is mostly visible in SEM imaging) and these are not used in age calculations. However reversely discordant samples are used in age calculations if they show no visible evidence of hitting a microfracture, inclusion or metamict area and improve the precision of age calculations.
4.3 Age of orthogneisses in Norsanna

Figure 4.7: 1:50,000 geological map of Norsanna showing key lithologies, F1 and F3 fold hinges and the localities of geochronology samples 482401, 482406, 482426, 482424, 482444 and 481438.
Sample 482426 was collected from a hornblende-biotite-bearing tonalite belonging to the polyphase orthogneiss (Fig. 4.7). At this locality within the NHSZ deformation zone, the Polyphase orthogneiss contains the planar, pervasive S₂ gneissosity (Chapter 2), which is defined by the alignment of mafic minerals. The sample was collected from a relatively homogeneous part of the orthogneiss, in order to minimise contamination from later, invasive leucosomes and thus avoid mixing zircon populations.

The sample yielded abundant pale brown, medium to coarse (100 to 400μm) zircons with ovoid to stubby and elongate morphologies (Fig. 4.8). Three distinct textural populations were identified based on grain morphology and internal zoning patterns. Population 1 zircon comprises strongly luminescent grains characterised by distinct oscillatory zoning. Rare whole grains are preserved, which exhibit pyramidal end terminations, however the majority of Population 1 zircon comprises equant or rounded cores, which in most cases are embayed and zoning is truncated along sinuous fronts. U concentrations are low to moderate (46-414 ppm) and Th/U values.
range from 0.23-0.87. Population 2 zircon comprises individual, sub-ovoid grains or wide rims overgrowing oscillatory zoned cores. Zircon in this group is moderately luminescent and preserves widely spaced, diffuse and discontinuous planar banding or sector zoning. The group has low to moderate U concentrations (103-466 ppm) and Th/U ratios range from 0.17-0.97. Population 3 zircon is rare and is characterised by small, stubby to acicular grains, with moderate luminescence and patchy zoning. Only two grains of Population 3 were analysed and these have moderate U concentrations (263 and 349 ppm) with Th/U ratios of 0.35 and 0.31 respectively.

Key: Red = Population 1 (oscillatory zoned); Green = Population 2 (sector and patchy zoned); Blue = Population 3 (acicular, patchy zoned)

Figure 4.9: U/Pb concordia diagrams for samples 482426. Black unfilled ellipses represent rejected grains that were not used in age calculations. Population ages are quoted to 2σ. The oldest U/Pb ages are recorded by Population 1. Highly discordant rejected analyses and Populations 1, 2 and define a combined ancient and recent Pb loss trend, denoted by the fan-shaped geometry. Three analyses of Population 2 define a younger metamorphic population at 2701 ± 31Ma (lower intercept = -260 ± 1300Ma; MSWD = 0.0027).

Analysis of 20 points on 18 grains revealed a complex data distribution with at least two age populations apparent (Fig. 4.9). 8 analyses were rejected, 4 of which
overlapped domain or grain boundaries and a further 3 intersected fractures. 3 analyses of Population 1 zircon yield \(^{207}\text{Pb}/^{206}\text{Pb}\) ages between 3623 and 3707Ma, of which the oldest concordant grain gives a \(^{207}\text{Pb}/^{206}\text{Pb}\) age of 3707 ± 7Ma. The remaining 4 analyses were rejected. Population 2 zircon yields two separate age populations, with the majority of analyses defining a complex Pb loss trend. 8 of 11 Population 2 analyses form a loose cluster at ca. 3560Ma but this age cluster is interpreted as a combined ancient and recent Pb loss trend. This is evidenced by 5 rejected analyses, which define an ancient Pb loss line that tracks along the concordia curve, which has then been affected by recent Pb loss, resulting in the fan-shaped spread described in section 4.2. Consequently no U/Pb population age was calculated for this group. The remaining 3 analyses of Population 2 zircon define a distinct, younger age population. The grains are texturally indistinguishable from the ca. 3560Ma grains described above (sector-zoned rim overgrowths on partially resorbed, oscillatory-zoned cores) but yield a U/Pb upper intercept age of 2701 ± 31 Ma (MSWD = 0.0027). One grain of Population 3 was rejected, with the remaining grain yielding a \(^{207}\text{Pb}/^{206}\text{Pb}\) age of 3531 ± 23Ma, within error of the Population 2 age spread at ca. 3560 Ma.

The oscillatory zoning in Population 1 zircon is compatible with a primary magmatic origin. However the blurring of oscillatory zoning (Hoskin and Black, 2000) and dispersion of Population 1 along the concordia curve indicates significant post-crystallisation modification and associated ancient Pb loss. This precludes the calculation of a population age for the magmatic component in this sample. Consequently the \(^{207}\text{Pb}/^{206}\text{Pb}\) age of 3707 ± 7Ma, taken from the oldest concordant grain, is interpreted as a minimum age of emplacement for the protolith of the polyphase orthogneiss. The characteristic morphologies of Population 2 zircon are comparable to those associated with new zircon growth under high grade metamorphic conditions (Corfu et al., 2003). The occurrence of multiple ‘metamorphic’ populations (ca. 3560 and 2700Ma) indicates that the polyphase orthogneiss has undergone a complex tectonothermal history. However, the large amount of scatter in the ca. 3560Ma cluster suggests that the older populations of zircon have been variably reset. Consequently, this age is not considered to record a true age for early metamorphism and is interpreted to represent a minimum age of metamorphism. The acicular morphology of Population 3 zircon is more compatible
with a magmatic than a metamorphic origin, suggesting that these are magmatic grains that underwent recrystallisation during $\leq 3560$ Ma metamorphism. The ca. 2700 Ma sub-population in Population 2 is interpreted as the timing of a second phase of metamorphism.

482424 – Homogeneous orthogneiss

Sample 482424 was collected from a biotite-bearing part of the homogeneous orthogneiss (Fig. 4.7), which displays a weak gneissosity defined by the orientation of deformed granodioritic pegmatite dykes. The sampling locality was situated outside of the main NHSZ deformation zone, but the gneissic fabric displayed in the tonalitic orthogneiss is interpreted to be equivalent to the $S_2$ platy fabric associated with the NHSZ.

Figure 4.10: Highly fractured, stubby to acicular grains with oscillatory zoning visible in BSE. The dark grains in zircon centres are quartz inclusions (a, b, d). Oscillatory zoning is embayed by recrystallisation fronts at grain rims (a, white arrow) and metamict zones have developed parallel to oscillatory zones (b, white arrow).

The sample yielded abundant, medium to coarse (100-400$\mu$m), pale brown zircons with stubby to acicular morphologies (Fig. 4.10). Grains are commonly embayed and heavily fractured, with rounded end terminations. Most grains preserve
well pronounced oscillatory zoning, which is locally blurred and chaotic in grain centres, especially where quartz inclusions are present. Modification may also be confined to individual oscillatory zones and is interpreted to reflect partial recrystallisation of metamict zircon. Thin rims are present and bound by sinuous fronts which appear to move from the grain edge towards the centre, however these were too small to target using LA-ICPMS. Uranium concentrations are variable (57-1217 ppm) and Th/U ratios range from 0.12-0.99.

![Image of U/Pb concordia diagram for sample 482424](image)

**Figure 4.11:** U/Pb concordia diagrams for sample 482424. Population ages are quoted to 2σ. A single stubby to acicular, oscillatory zoned population defines a recent Pb loss trend with an upper intercept at 2806 ± 15 Ma (lower intercept = 141 ± 181 Ma; MSWD = 11.7).

Analysis of 20 points on 20 grains yields a single age population, which lies on a relatively tight Pb-loss trend (78-104% concordant) that intersects the concordia curve at 2806 ± 15 Ma (MSWD 10.7, n=20) with a lower intercept at 141 ± 180 Ma (Fig. 4.11). Grains with acicular morphologies and well-preserved oscillatory zoning tend to yield marginally older ages when compared with stubby grains with embayed margins and those that contained metamict domains or sinuous fronts.

The oscillatory zoning and stubby to acicular morphologies predominant in 482424 are compatible with a primary magmatic origin. The development of blurred
oscillatory zoning and metamict cores in magmatic grains is considered an artefact of post-crystallisation diffusion and radiation damage, but is not necessarily instigated by subsequent metamorphism. The sinuous fronts which propagate from grain rims towards grain centres are considered to reflect recrystallisation of the original magmatic grain, however the absolute time gap between magmatic crystallisation and modification are not constrained by the isotope data. The suggestion that zircon has undergone post-crystallisation modification does, however suggest that the age of 2806 ± 15 Ma should be considered minimum estimate for the intrusion of the precursor to the Homogeneous orthogneiss.

482444 – Felsic leucosome

Sample 482444 was collected from a hornblende-garnet-plagioclase-quartz leucosome that invades the homogeneous orthogneiss within the NHSZ (Fig. 4.7). At this locality, the homogeneous orthogneiss is tonalitic in composition and has a well developed $S_2$ gneissosity, which intensifies moving west into a supracrustal amphibolite unit. Both homogenous orthogneiss and amphibolite are intruded by the felsic leucosome, which contains coarse garnet and hornblende porphyroblasts and lies parallel or subparallel to the $S_2$ gneissosity. Host rock material was removed from sample 482444 prior to zircon separation in an attempt to target only zircon that was formed within the leucosome itself.

The sample yielded abundant, medium to coarse (100-300 μm), pale to dark brown zircon grains, with subovoid to stubby and acicular morphologies (Fig. 4.12). Four textural groups were defined according to zircon morphology and internal zoning patterns. Population 1 zircon comprises strongly luminescent cores with blurred and convoluted oscillatory zoning. Patchy zoning is locally present. Cores are
Figure 4.12: SEM (CL) images of 482444. Scale bars measure 100µm. $^{207}$Pb/$^{206}$Pb ages are quoted to 1σ. a) Strongly luminescent, partially recrystallised inherited core with oscillatory zoning (Population 1). b and c) Subovoid metamorphic grains (Population 2) with pronounced patchy zoning. d) Stubby grains with faint planar banding (Population 3), potentially reflecting remnant oscillatory zoning. e) Recrystallised rim (Population 4) with concentric planar banding mantling a partially resorbed core of Population 2.

ellongate to rounded and in some cases oscillatory zoning is truncated by mantles of Population 3 zircon. Uranium concentrations are low to moderate (109-199ppm) and Th/U ratios are variable (0.11-0.67). Population 2 zircon forms stubby to sub-ovoid grains and constitutes the dominant textural group. The group is characterised by weak to moderately luminescent zircon that comprises both individual grains and large cores (> 150µm) with well-defined sector zoning. Planar banding is commonly developed near grain boundaries. The population yields variable but moderate to high U concentrations (248-2167ppm) and low Th/U ratios (0.01 – 0.11). Population 3 zircon comprises stubby to acicular grains with rounded end terminations, which may also contain cores of Population 1 zircon. Population 3 is weakly luminescent with faint planar banding, which becomes discontinuous towards grain end terminations Uranium concentrations are moderate (296-676ppm) and Th/U ratios are low (0.01-0.05). A rare, fourth population forms thin rims on embayed Population 2 cores. The population is weakly to moderately luminescent (although generally more strongly luminescent than Population 2) and exhibits weak planar banding. Two analyses of Population 4 zircon were made, which yielded moderate Uranium concentrations (450 and 425ppm) and low Th/U ratios (0.04 and 0.08 respectively).
Analysis of 30 points on 28 grains yielded a single age population with some scatter to older ages (Fig. 4.13). 11 analyses were rejected; 2 for extreme discordance that is attributed to high U concentrations (>1100ppm). A further 9 analyses were centred over domain or grain boundaries. The one remaining analysis from Population 1 yielded a concordant $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2846 ± 9 Ma (1σ) – the oldest analysis of the whole sample – and lies outside the main age population.

12 of 18 analyses of Population 2 zircon form a near concordant age population (95-102% concordant) with a U/Pb upper intercept age of 2716 ± 15 Ma (MSWD = 6.2). 4 of 6 analyses of the stubby to prismatic Population 3 zircon yield a poorly defined U/Pb population age of 2722 ± 71 Ma (MSWD = 25) and 2 of 3 analyses from Population 4 rims also plot within error of Populations 2 and 3. The ages yielded by Populations 2, 3 and 4 are similar within error. In addition, the morphologies of all three textural groups are compatible with metamorphic zircon (both new growth and recrystallised). Consequently all three groups are considered to represent a single event and define a Pb-loss trend with a combined age of 2720 ± 14 Ma (MSWD = 10.8) and a lower intercept of 339 ± 630 Ma. Analyses from all three populations are variably affected by recent Pb loss, however the majority of points are >95% concordant. There is some scatter along the concordia, with ovoid, sector zoned zircon (Population 2) giving marginally younger concordant ages than stubby, planar banded Population 3 zircon, suggesting either a lengthy period of crystallisation or some minor ancient Pb loss.

The occurrence of Population 1 zircon as small, embayed cores with patchy and blurred oscillatory zoning suggests a magmatic origin for this textual group that has been subject to considerable alteration, resorption, re-precipitation and recrystallisation similar to that described by Corfu et al., (2003), Hoskin and Black, (2000), Pidgeon et al., (1998) and Schaltegger et al., (1999). The single $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2846 ± 9 Ma is notably older than other zircon in the sample, and the textural constraints on this population suggest that it represents an inherited component that became entrained in the felsic leucosome either at the source of melting or during emplacement. This is also evidenced by apparent ancient Pb-loss trends between the inherited cores and the main metamorphic population (see below). U/Pb zircon ages
Figure 4.13: U/Pb concordia diagrams for sample 482444. Black unfilled ellipses represent rejected grains that were not used in age calculations. Population ages are quoted to 2σ. a) Whole sample age spread for sample 482444 including rejected grains. b) A single age population defining a recent Pb loss trend (with some minor ancient Pb loss), with an upper intercept age of 2720 ± 14Ma (lower intercept = 339 ± 630Ma; MSWD = 10.8). The red Population 1 analysis is not included in the age calculation due to its interpretation as an inherited grain.
of ca. 2840 Ma have been obtained from an orthogneiss in Inner Ameralik (Nutman, 2006), suggesting that the inherited component in the leucosome could be locally sourced.

As discussed above, the zircon morphologies in Population 2, 3 and 4 are compatible with new zircon growth and/or recrystallisation during high grade metamorphism in the presence of partial melt (as described by Corfu et al., 2003). The inclusion of inherited magmatic zircon (Population 1) that is resorbed and mantled by metamorphic rims suggests that at least some of the metamorphic component in the sample is derived from inherited magmatic zircon. The presence of sub-ovoid grain morphologies does suggest that some new zircon growth occurred at 2720 ± 14Ma, which is interpreted to represent the age of crystallisation of the leucosome. The occurrence of hornblende and garnet in the leucosome at the sample locality confirms that leucosome injection took place under amphibolite facies conditions and as such is interpreted to record the age of amphibolite facies metamorphism.

481438 – Felsic leucosome

Sample 481438 was collected from a felsic melt segregation intruding supracrustal amphibolite within the NHSZ (Fig. 4.7). The felsic material forms ~5 cm-wide vein systems, which lie parallel or subparallel to the S₂ gneissosity. The veins are undeformed by D₂, indicating their intrusion during late-D₂ and have poorly defined, diffuse walls and conduit margins, suggesting an in situ partial melt source for the leucosome (Fig. 4.14a). The leucosome is dominated by plagioclase-quartz but contains coarse clots of a pale amphibole, which is rimmed by a thin mantle of hornblende (Fig. 4.14b). Host rock material was removed from the sample prior to zircon separation, in order to reduce contamination from the host amphibolite.

The sample yielded abundant, very coarse (250-1000μm), dark brown zircons, with equant, stubby and elongate morphologies. Thin section examination showed the coarsest zircon to be associated with anthophyllite-cummingtonite clots. Imaging identified 3 distinct textural populations (Fig. 4.15). Population 1 zircon forms weakly luminescent, stubby to elongate cores with faint planar banding or
patchy zoning. Core margins are rounded and embayed and internal zoning is truncated and mantled by zircon from Population 2. Uranium concentrations are variable (132-646ppm) but the majority of grains average 200ppm. Th/U ratios are low (0.04-0.27). Population 2 zircon forms moderately luminescent mantles around Population 1 cores that display broad planar banding or sector zoning at pyramidal grain terminations.

Figure 4.14: a) Dip slope exposure of leucosome sample 481438 at UTM 22W 516073E, 7121505N b) Coarse clots of anthophyllite-cummingtonite with thin rims of hornblende in leucosome sample 481438 at the sample locality. Leucosome is intruded into host garnet amphibolite parallel to S2 fabric, defined by hornblende.

Figure 4.15: Stubby grains typical of 481438 exhibiting the dominant textural populations. Scale bar measures 100μm. a-d) Dark, partially resorbed cores (Population 1) and moderately luminescent, planar banded and sector zoned mantles (Population 2). e) Sinuous recrystallisation fronts (Population 3) migrating across a Population 1 core.
Uranium concentrations are low (64-180 ppm) and Th/U ratios are moderate to high (0.22-0.6). Population 3 zircon forms weakly luminescent rims around Population 2. The weak planar banding in Population 3 is typically concordant to zoning in Population 2 and the group also exhibits pronounced pyramidal end terminations. Uranium concentrations in the population are variable (89-551 ppm) and Th/U ratios range from 0.003 to 0.39.

Analysis of 40 points on 28 grains yielded a single age population (Fig. 4.16). Six analyses were rejected; 2 analyses exhibited errors too large to be of use in statistical analysis and 4 analyses were centred over domain boundaries. All analyses yielded near concordant data (>96% concordant). 12 analyses of Populations 1 form a tight cluster with an upper intercept age of 2721 ± 5 Ma (MSWD = 1.7). 16 analyses of Population 2 zircon define a Pb loss trend with an upper intercept age of 2718 ± 5 Ma (MSWD = 0.81) and 6 analyses of Population 3 yield a poorly-defined upper intercept age of 2709 ± 54 (MSWD = 6.7). All three textural populations yield ages within error of each other. Zoning patterns in all populations suggest that zircon in the leucosome crystallised during a single event and as such yield a combined U/Pb age with an upper intercept at 2716 ± 6 Ma (MSWD = 3.7) and lower intercept at 312 ± 280 Ma. The large error on Population 3 zircon is considered a result of higher U contents combined with a small sample size (n = 5 of 6 analyses) and a relatively small target area for individual analyses (where errors – specifically common Pb contamination – may increase, due to the inability to avoid damaged areas or fractures on grain tips).

Textural analysis identified several phases of zircon growth in sample 481438. The most apparent break in growth is encountered between Populations 1 and 2, evidenced by irregular, embayed Population 1 cores combined with the truncation of zoning. In many cases, the textural discordance between core and rim material is continuous around the entire core boundary, indicating partial resorption and subsequent re-precipitation of zircon. The formation of pyramidal end terminations and a tendency to form concordant contacts between earlier zircon suggests that Population 3 represents a minor phase of new zircon growth. The Pb-loss projection towards zero in all textural groups indicates that all zircon in the leucosome crystallised during a single thermal event, which has not been significantly affected by ancient Pb loss. The identification of multiple zircon growth
phases within a single event may be the result of chemical interactions between a felsic melt and a mafic host rock, resulting in rapid fluctuations in Zr saturation in the melt. The occurrence of a tightly defined, concordant age suggests that all phases of zircon growth, resorption and re-precipitation occurred in rapid succession during felsic leucosome crystallisation under upper amphibolite facies conditions at $2717 \pm 6\text{Ma}$.

Figure 4.16: Concordia diagram of zircon in leucosome sample 481438. Black unfilled ellipses represent rejected grains that were not used in age calculations. Population ages are quoted to $2\sigma$. A tightly defined single age population gives an age of $2716 \pm 6\text{Ma}$ (lower intercept $= 312 \pm 280\text{Ma}$; MSWD $= 3.7$).
Sample 481418 was collected from a biotite-bearing granodioritic orthogneiss on the northern shore of Ameralla Fjord (Fig. 4.17) that contains minor invasive pegmatitic material and leucogabbroic inclusions. The sample was taken from a homogeneous part of the granodioritic gneiss, avoiding contamination from felsic or mafic enclaves.

The sample yielded, medium to coarse (150-400µm), pale brown zircons, with equant to acicular morphologies (Fig. 4.18). Two textural populations were defined on the basis of SEM images. Population 1 zircon exhibits well-defined oscillatory zoning with some sector zoning at grain terminations and the group forms equant to acicular whole grains and cores that are mantled by Population 2. Grain terminations are typically rounded, but pyramidal end terminations are preserved locally, suggesting minor resorption of Population 1 zircon. CL responses vary from weak to strong as a result of the prominent zoning but strongly luminescent material predominates.

Figure 4.18: SEM (CL) images of 481418. Scale bars measure 100µm. $^{207}$Pb/$^{206}$Pb ages are quoted to 1σ. a) Acicular, oscillatory zoned magmatic zircon (Population 1). b-d) Grains cored by oscillatory zoned Population 1 and mantled by weakly luminescent, planar banded Population 2 zircon. e) Whole grain of metamorphic Population 2 exhibiting faint planar banding.

A number of Population 1 grains mantle embayed, weakly luminescent and chaotically-zoned cores, which are interpreted here as metamict and partially resorbed inherited grains. The inherited component was not analysed due to the significant damage to the zircon lattices, evident from SEM images. A subset of
Population 1 is very strongly luminescent, and oscillatory zoning is locally accompanied by patchy zoning. These textures tend to form in Population 1 grain centres and are concordant with normal Population 1 zircon, which tends to form immediately outside of these cores. In addition, some moderately luminescent Population 1 grain centres are homogeneous or exhibit patchy zoning but show no evidence of resorption between the grain centre and the surrounding oscillatory zoned Population 1 zircon. These areas and the strongly luminescent cores described above are considered to represent magmatic zircon that homogenised during magmatic (or later re-) crystallisation and are thus grouped with Population 1. U concentrations are variable (33-660ppm) and the lowest U concentrations are associated with the strongly luminescent subset of the population. Th/U is also variable and ranges from 0.16-0.84. Population 2 zircon is defined by stubby whole grains and rims on Population 1. The group is characterised by weak to moderate CL responses and may display patchy zoning or diffuse planar banding. The latter is more common where Population 2 forms mantles around Population 1 cores. The zoning in these mantles is usually concordant with the oscillatory zoning exhibited by Population 1 cores, however truncation of oscillatory zoning does occur locally. U concentrations are generally higher than Population 1 zircon (320-739ppm) and Th/U is less variable (0.02-0.09).

Analysis of 40 points on 32 grains yielded a mix of concordant to significantly discordant data (49-101% concordant). Six analyses were rejected prior to age calculation due to intersections of fractures, domain or grain boundaries (Fig. 4.19a). 20 analyses of Population 1 zircon define a Pb loss trend with an upper intercept age of 3027 ± 15Ma (MSWD = 11.5) and a lower intercept age of 103 ±120Ma. 13 analyses of Population 2 zircon also define a Pb-loss trend but with an upper intercept age of 2722 ± 8Ma (MSWD = 2.4). However the lower intercept of this population yields a negative age of -462 ± 140Ma, which is incompatible with a recent Pb loss trend. Further inspection of the Pb loss trends in Population 2 suggests that the four most discordant grains may lie on a different Pb loss trend to the rest of Population 2, so that they project to an upper intercept intermediate between Populations 1 and 2. The 9 remaining Population 2 analyses are all >99% concordant and intercept the Concordia curve at 2718 ± 7Ma (MSWD = 0.48) and 721 ±1200Ma (Fig. 4.19b).
Figure 4.19: U/Pb concordia diagrams for samples 481418. Black unfilled ellipses represent rejected grains that were not used in age calculations. Population ages are quoted to 2σ. a) Oscillatory zoned magmatic zircon (Population 1) defines a recent Pb loss trend with an upper intercept at 3027 ± 15 Ma (lower intercept = 103 ± 120 Ma; MSWD = 11.5). Patchy and sector zoned zircon (Population 2) zircon forms a younger metamorphic population. b) >99% concordant analyses of Population 2 (boxed area, Fig 4.19a) yield an age of 2718 ± 7 Ma (lower intercept = 721 ± 1200 Ma; MSWD = 0.48) after rejection of the four significantly discordant analyses given in (a).

Key: Red = Population 1 (oscillatory zoned); Green = Population 2 (patchy and sector zoned).
The stubby to acicular grain morphology, coupled with the prevalence of oscillatory zoning in Population 1 is compatible with a magmatic origin. However the presence of a younger zircon population and a potential ancient Pb loss trend in Population 2 suggests that modification of the magmatic population is likely. This is supported by 3 of the rejected grains interpreted to define the ancient Pb loss curve described above. These grains form mantles on Population 1 cores that exhibit planar banding parallel to the oscillatory zoning in the magmatic cores, and as such are interpreted as relict oscillatory zones resulting from recrystallisation of magmatic zircon. As such, the age of 3027 ± 15Ma is interpreted as a minimum age of emplacement of the precursor to the orthogneiss. The morphology and zoning patterns of Population 2 zircons are compatible with a metamorphic origin. Whereas some grains are considered to represent recrystallised magmatic zircon (as above), evidence also supports some new metamorphic zircon growth. This latter interpretation applies chiefly to single phase grains in Population 2, the majority of which are concordant and show no evidence of ancient Pb loss. The age of 2718 ± 7Ma is consequently interpreted to reflect metamorphism in the orthogneiss, which was associated with variable Pb loss in the magmatic population, in addition to new zircon growth and recrystallisation.

496831 – Tonalitic orthogneiss

The sample was collected from a homogeneous part of the tonalitic orthogneiss of Tummeralik at (Fig. 4.17). The gneiss at the sample locality contains the assemblage hornblende-biotite-plagioclase and is intruded by deformed granodioritic veins, which are subcordant to the S1,2 gneissosity, and also by later granitic pegmatite. Boudinaged trails and rafts of the early mafic suite (Chapter 3) are common near the sample locality, which are occasionally agmatitic. Sampled material was selected so as to eliminate as much non-tonalitic material as possible.

The sample yielded abundant, medium to coarse (100-600µm), pale to dark brown zircon. Three distinct textural populations were identified (Fig. 4.20). Population 1 zircon comprises strongly luminescent, equant to stubby cores and individual grains with pronounced oscillatory zoning. Zircon in this population is commonly embayed and mantled by Population 2 zircon, where the boundary
between the two groups is defined by a sinuous front. Uranium concentrations are moderate (107-268ppm) and Th/U ratios vary from 0.55-0.99. Population 2 zircon forms the dominant textural group and is characterised by very weakly luminescent, equant to stubby grains. The morphology of whole grains are similar to Population 1,

but the group also forms rims on Population 1 cores. The group typically exhibits planar banding or patchy zoning. Chaotic zoning is preserved locally, especially in the central regions of grains. Where Population 2 zircon forms a mantle around Population 1, the planar banding often lies parallel to oscillatory zoning in the Population 1 cores. Many Population 2 grains surround a small, rounded or stubby-elongate 'seed-like' core of highly luminescent zircon; these cores were too small for analysis using LA-ICPMS but may be highly resorbed remnants of Population 1. Uranium concentrations are moderate to high (267-1742ppm) and Th/U ratios range from 0.03-0.42. A third, very minor textural population occurs as small rims on Population 2. The boundary between Populations 2 and 3 is defined by sinuous fronts progressing from grain rims towards their centres, which truncate the planar banding of Population 2. Population 3 zircons tend to have stronger CL responses than Population 2, but are still weakly to moderately luminescent and exhibit faint planar banding that is usually approximately parallel to the front. Two analyses of Population 3 were obtained, which gave moderate Uranium concentrations of 246

Figure 4.20: SEM images (CL) of zircon in sample 496831. Scale bars measure 100μm. a) Prominent oscillatory zoning in magmatic Population 1. Population 1 forms a partially resorbed core in (i), with oscillatory zoning truncated by mantling Population 2. g-i) Weakly luminescent Population 2 zircon with planar banding forming whole grains or mantles on Population 1 cores. j) Planar banded metamorphic grain (Population 2) with a strongly luminescent 'seed-like' core of Population 1. Zoning in the Population 2 grain is truncated by a recrystallised rim of Population 3 exhibiting planar banding parallel to the recrystallisation front.
and 336ppm, and Th/U values of 0.32. Zircons from all three populations exhibit serrated and embayed grain boundaries.

Analysis of 50 points on 30 grains yielded two distinct age populations (Fig. 4.21). Nine analyses were rejected prior to age calculation. Four analyses were centred on grain or domain boundaries and a further 4 analyses showed significant discordance that correlated directly with high U concentrations (1357-2954ppm). One analysis on a highly damaged grain exhibited significant reverse discordance (concordance = 113%). This is considered to be the result of inadvertent analysis of Pb-bearing microparticles in a metamict zircon lattice. The strongly luminescent, oscillatory zoned zircons of Population 1 yielded the oldest age population, which has an upper intercept with the concordia curve of $3046 \pm 10$ Ma (MSWD = 2; n=9 of 10). Analyses of Populations 2 and 3 form a tight concordant to near-concordant cluster, with a tendency towards slight reverse discordance (103-105% to a maximum of 109%). The two textural groups yield age populations within uncertainty of each other and give a combined upper intercept age of $2680 \pm 11$ Ma (MSWD = 7.2; n = 32 of 40).

The equant to stubby, oscillatory zoning that characterises Population 1 is compatible with a magmatic origin, and as such, $3046 \pm 10$ Ma is taken to represent the age of emplacement of the precursor to the tonalitic orthogneiss. Common embayment of Population 1 cores and the truncation of oscillatory zoning suggest significant resorption-reprecipitation and/or recrystallisation of Population 1, however there is little scatter along the concordia curve, suggesting a minimal amount of ancient Pb loss. However, this age is considered a minimum estimate considering the evidence for resorption in the magmatic population and the presence of a younger ‘metamorphic’ population.

The zoning patterns which typify Populations 2 and 3 are compatible with a metamorphic origin. Similarities in grain morphology between Populations 1 and 2 and the occurrence of planar banding (Population 2) parallel to primary oscillatory zoning (Population 1) suggest that the former represents recrystallised magmatic zircon that has been reset during metamorphism. In this case, the planar banding typical of Population 2 is interpreted to reflect relict oscillatory zoning. A problem
with this interpretation is that the widespread metamorphic recrystallisation of Population 1 zircon should be associated with significant Pb loss within that group.

Key: Red = Population 1 (oscillatory zoned); Green = Population 2 (patchy and sector zoned); Blue = Population 3 (recrystallised rims)

Figure 4.21: U/Pb concordia for sample 496831. Black unfilled ellipses represent rejected grains that were not used in age calculations. Population ages are quoted to 2σ. a) U/Pb concordia plot of the entire sample including rejected analyses that were not used in age calculations. b) Detail of the main age populations showing rejected analyses that were not used in age calculations. c) The main age populations showing oscillatory zoned (Population 1, red) zircons defining a recent Pb loss trend with an upper intercept at 3046 ± 10Ma (lower intercept = 355 ± 740Ma; MSWD = 2.0). Stubby elongate grains with patchy zoning (Population 2, green) and recrystallised rims (Population 3, blue) yield a combined age of 2680 ± 11Ma (lower intercept = 393 ± 300Ma; MSWD = 7.2).

However the lack of scatter along the concordia curve in between Populations 1 and 2 refutes recrystallisation of magmatic zircon as a mechanism for producing a metamorphic population unless affected magmatic grains underwent 100% Pb loss at 2680Ma. An alternative mechanism would be the growth of new metamorphic zircon
around the partially resorbed Population 1 nuclei, where the Population 2 planar banding is the product of new metamorphic grains, potentially crystallised in the presence of partial melt. Zircon in the sample is interpreted to have formed by both mechanisms. New zircon growth on partially resorbed magmatic cores is considered the dominant process forming Population 2. However recrystallisation of pre-existing material is evident in Population 3, which is characterised by migrating, sinuous boundaries interpreted as recrystallisation fronts (e.g. Cherniak and Watson, 2003; Corfu et al., 2003). The age of 2680 ± 11Ma is consequently considered to represent new growth and recrystallisation of zircon in the tonalitic orthogneiss under high grade metamorphic conditions.

478334 – Garnet-bearing felsic leucosome

Sample 478334 was collected from a felsic leucosome intruding the supracrustal amphibolite on the south limb of the Tummeralik Isocline (Fig. 4.17). The leucosome contains the assemblage garnet-plagioclase-quartz, with garnet occurring as very large (>30mm) porphyroblasts within the leucosome (Fig. 4.22). Leucosome in the outcrop forms an anastomosing vein network that is parallel or subparallel to the local S2 gneissosity. In order to avoid contamination, host rock material was removed from the sample prior to zircon separation.

Figure 4.22: Partial melting of host garnet amphibolite at the sample locality for 478334 (UTM 22W 560155E, 7124217N). Leucosome stringers lie parallel to the S1,2 fabric (white arrow) and coarse garnet form where the leucosome ponds. Scale: hammer handle ~20cm.
The sample yielded abundant, medium to coarse (100-400μm), pale to dark brown zircons with ovoid to stubby morphologies. A single textural group was identified, which is characterised by ovoid to stubby grain morphologies. The grains are weakly luminescent (in most cases, luminescing less than that of the epoxy grain mount). In BSE images zircon grains show weak planar banding or sector zoning, with the boundaries between zones often defined by fractures (Fig. 4.23). Grains also display radial fracture in addition to those defining zone boundaries. In all cases, fractures are infilled by highly luminescent material, which was too small to analyse. Uranium concentrations are moderate to very high (147-5301ppm) and Th/U ratios range from 0.05-0.61. A single grain was encountered that exhibited a stubby-elongate morphology, with very weak luminescence and faint planar banding.

Analysis of 40 points on 34 grains revealed a highly damaged sample with a single apparent age population defined by a small proportion of points (Fig. 4.24). Of the 40 grains analysed, 10 grains disintegrated during the analysis and so yielded internally inconsistent results. A further 12 were rejected, of which 6 grains exhibited extreme discordance and $^{207}\text{Pb}/^{206}\text{Pb}$ ages in excess of 4000 Ma, 3 grains produced $^{207}\text{Pb}/^{206}\text{Pb}$ ages in excess of 3000 Ma. Three grains were centred over fractures or metamict areas and consequently yielded large errors. All rejected analyses had associated errors that were too large to provide geologically meaningful interpretations and were associated with the highest Uranium concentrations (>1000ppm). The remaining 18 analyses define a Pb-loss trend (83-108% concordance), which intersects the concordia curve at 2659 ± 17 Ma (MSWD = 7.3).
An element of ancient Pb loss in the sample is suggested by the $^{207}\text{Pb}/^{206}\text{Pb}$ ages in some analyses, which yield older ages than their corresponding U/Pb ages.

Figure 4.24: U/Pb concordia diagrams for sample 478334. Black unfilled ellipses represent rejected grains that were not used in age calculations. Population ages are quoted to 2σ. a) A high number of rejected analyses exhibiting extreme discordance and unacceptably large errors. b) The remaining near concordant analyses define a Pb loss trend with upper intercept at 2659 ± 17Ma (lower intercept -706 ± 850Ma; MSWD 7.3).
The ovoid morphology and presence of planar banding and sector is compatible with a metamorphic origin. It is suggested that high initial U concentrations resulted in metamictisation in the majority of grains, leaving many grains vulnerable to post-crystallisation alteration (e.g. Pidgeon, 1992; Pidgeon and Kalsbeek, 1978; Pidgeon et al., 1998). Common Pb contamination is considered to be a significant contributor to the high errors associated with many of the analyses, of which the metamictisation of grains is likely to be a key factor (as described in Meldrum et al., 1998; Mezger and Krogstad, 1997). Additional sources of common Pb contamination may also be associated with the radial and zone-parallel fracturing evident in most grains.

High common Pb contamination from metamictisation and fracturing, combined evidence from $^{207}\text{Pb}/^{206}\text{Pb}$ ages and Pb loss trends indicates isotopic alteration and (partial-) resetting in the majority of grains in the sample. This has serious implications for the geological significance of the age given by the 17 accepted analyses, as it is unlikely that such a damaged sample set would yield a geologically meaningful age. The age 2659 ± 17Ma is thus interpreted as a minimum age of leucosome crystallisation and provides a minimum age constraint for amphibolite facies metamorphism in Tummeralik.

4.5 Preliminary ages from monazite chemical dating in Norsanna

Two semi-pelite samples from Norsanna yielded appreciable accessory monazite, and were dated using the U-Th-Pb chemical dating method (Montel et al., 1996; Suzuki and Adachi, 1991). Samples were imaged in BSE using a Philips XL30CP SEM, prior to whole grain X-ray mapping and analysis using a Cameca SX100 EPMA. Both SEM and electron microprobe facilities are housed at the School of Geosciences, University of Edinburgh. X-ray mapping of monazite was carried out prior to dating to identify zoning patterns in Ce, Pb, Th, U and Y. Details of the analytical technique are presented in Appendix A. The data presented here has not been examined in detail and should be considered as preliminary data only.

Samples 482401 (UTM 22W 514938E, 7124597N) and 482406 (UTM 22W 513811E, 7124447N) were collected from boudinaged semi-pelite lenses in the NHSZ (Fig. 4.7). Both samples contain the assemblage Grt-Bt-Pl-Qtz with accessory
zircon and monazite, with the $S_2$ foliation defined by biotite. Monazite in the samples ranges from 20-200μm and occurs as inclusions in garnet and as a matrix mineral, often intergrown with biotite or spatially associated with garnet (Fig. 4.25).

![BSE images of monazite in the Norsanna semi-pelites. a) Irregular monazite grains included in garnet in sample 482401. b) Monazite in contact with garnet in sample 482406. A fine grained monazite is included in garnet rims, whereas the coarse grain is adjacent to garnet but contained within matrix biotite.]

Both samples exhibited a brabantite composition and typically exhibited complex zoning (Fig. 4.26), although some of the smaller grains tended to be homogeneous (Fig. 4.27). This may be a function of either late stage growth or the orientation of the grain in the thin section. Complex zoning was either patchy or concentric about a central domain (although the central domain was not necessarily the oldest domain). Individual domains tended to be elongate, and rarely measured >10μm width, which presented a problem for targeting points for analysis. The small size of the domains also limited the number of analyses that could be made per domain.
Figure 4.26: Complex zoning in monazite. a) Relatively homogeneous Th zoning in sample 482401. b) Concentric zoning in Y in the same grain in sample 482401. c and d) Concentric zoning in Th (c) and Y (d) in sample 482406.

Figure 4.27: Homogeneous zoning profiles in Th (a) and Y (b) in sample 482406.
Between 6 and 15 points per grain were analysed and processed using a modified data reduction programme by Berry (unpublished) and Pyle et al., (2005) with modifications by Steele and Kelly (Pers. Comm). $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average ages of both samples yielded a spread in ages between 2720 and 2500Ma, with broad individual peaks occurring at ca. 2700, 2650 and 2500 Ma (Fig. 4.28).

![Weighted probability density curves showing the spread in $^{207}\text{Pb}/^{206}\text{Pb}$ age for all analyses in sample 482401 (a) and 482406 (b).](image)

A comparison of textures suggests that the majority of grains yielding $>$2700Ma domains are included in garnet, whereas younger monazites tend to occur in the matrix. A number of monazite inclusions in garnet also yield younger ($<$2650Ma) ages, but these are demonstrably connected to the matrix via fractures in the host garnet. Monazite domains with the older age of ca. 2720Ma may reflect monazite growth during metamorphism at this time, in accordance with the metamorphic
zircon populations observed in leucosome and orthogneisses in Norsanna. However the spread in ages after 2650Ma suggests that Norsanna may have been affected by later thermal overprints.

It should be noted the spread in age data may also be a function of errors caused by low counts in key elements such as Th. The majority of monazite analysed contains low Th, which has resulted in 1σ errors in excess of 50Ma for most individual points. Caution should also be exercised when interpreting the significance of the observed spread in ages, as hybrid ages may be produced if an analysis hits a domain boundary. This, as for complex zircons, may be a problem given the small width of the majority of domains combined with the possibility of beam drift over the period of an analysis session. In addition, the polygenetic nature of the monazites themselves suggests a real risk of partial Pb loss, which may also account for some of the spread in ages (Cocherie et al., 1998). This is supported by the presence of dominantly young ages from grains included in garnet but connected to the matrix. Time constraints prevented a detailed analysis of these datasets and consequently no specific interpretations are made regarding the construction of P-T-t paths in Norsanna. However the ages suggested by the data in this section is compared with other geochronological data from the Kapisillit region and discussed in a regional context in Chapter 7.
4.6 Summary

The following points give a summary of the findings from this chapter. A full discussion, in which the geochronology is incorporated with structural data and field relations, is given in Chapter 5.

1. The polyphase and homogeneous orthogneisses in Norsanna form discrete suites, separated in time by ca. 900 Ma and were emplaced $\geq 3707 \pm 7$ Ma and $\geq 2806 \pm 15$ Ma respectively.

2. The polyphase orthogneiss underwent high grade metamorphism at or prior to ca. 3560 Ma.

3. Partial melting under upper amphibolite facies conditions affected Norsanna between 2720 and 2700 Ma, as indicated by leucosome invasion and the ca. 2700 Ma population in the polyphase orthogneiss (482426).

4. Tonalitic to granodioritic orthogneiss with (a) Mesoarchaean emplacement age(s) are recognised east of Itilleq Fjord, yielding ages of $3046 \pm 10$ and $3027 \pm 15$ Ma.

5. The Mesoarchaean orthogneisses east of Itilleq Fjord show a disparity in metamorphic age population, where the Ameralla Fjord sample suggests metamorphism at $2718 \pm 7$ Ma, whereas the tonalitic orthogneiss in Tummeralik sample preserves evidence of metamorphism at $2680 \pm 11$ Ma.

6. The crystallisation of amphibolite facies leucosome in Tummeralik is recorded at or prior $2659 \pm 17$ Ma.

7. Most samples are affected to some degree by ancient Pb loss, although physical damage to the lattice, such as metamictisation, severe partial resorption and physical fracture are more apparent in Tummeralik than Norsanna or Ameralla Fjord.

8. Preliminary data suggests a polymetamorphic origin for monazite in semipelitic units in Norsanna starting at ca. 2720-2700 Ma, which is suggested to reflect metamorphism at this time.

9. The large spread in $^{207}\text{Pb}^{206}\text{Pb}$ monazite ages are treated with caution, however several potential peaks are observed between 2700 and 2500 Ma, which may reflect later thermal events. However these may be influenced partial Pb loss in polygenetic monazite and hybrid analyses.


Hanchar, J/M/ and Rudnick, R.L. 1995. Revealing hidden structures: The application of cathodoluminescence and backscattered electron imaging to dating zircons from lower crustal xenoliths. Lithos, v. 36; 189-203


Nutman, A.P., 2006, Reconnaissance dating in the Kapisillit Mapsheet, in GEUS, ed.: Copenhagen, workshop discussion.


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CHAPTER 5
A COMPARATIVE STUDY OF NORSANNA AND TUMMERALIK

5.1 Overview

The most recent evolutionary hypothesis for the Nuuk region proposes that it is formed of up to 6 discrete tectonostratigraphic terranes that were amalgamated during the Neoarchaean. The terrane boundaries are preserved as narrow, deformed and mylonitised zones up to 50 m thick (Friend and Nutman, 2005a; Friend and Nutman, 2005b; Friend et al., 1987; McGregor et al., 1991; Nutman and Friend, 2007; Nutman et al., 1989). Chapter 1 discussed the hypothesis that a boundary region between the Tre Brødre-Færingehavn and Kapisillik terranes runs along the eastern coast of Itilleq Fjord on an approximate north-south trend. The Norsanna field area is thought to represent part of the boundary between the Tre Brødre and Færingehavn terranes, whereas Tummeralik lies well within the area interpreted as belonging to the Kapisillik terrane (using the terrane boundaries of Nutman and Friend, 2007). According to the criteria for identifying suspect terranes outlined in Chapter 1, the two field areas should therefore preserve distinct lithological and structural histories. The relationship between Norsanna and Tummeralik may therefore be addressed to some degree by a comparison of field observations and geochronology from the two areas (described in Chapters 2, 3 and 4). This chapter integrates lithological, structural and geochronological data in Norsanna and Tummeralik and critically examines any relationships that may point to individual or shared evolutionary histories.

5.2 Constituent lithologies in Norsanna and Tummeralik

The lithologies that make up Norsanna and Tummeralik – dominant tonalitic to granodioritic orthogneisses with subordinate supracrustal assemblages described in Chapters 2 and 3 – are typical of high grade Archaean cratonic areas (Windley, 1995). It is stressed that the identification of independently evolved tectonic blocks using lithological comparisons should always be supported by additional isotopic and metamorphic evidence. This is particularly crucial in high grade terrains composed
dominantly of metaigneous material, where there are no preserved stratigraphic sequences or palaeontological records. However a lithological comparison between Norsanna and Tummeralik is presented here, prior to more detailed discussions that integrate isotopic data.

The modal proportions of lithologies may be used as a criterion to distinguish geologically discrete areas in the field. For instance one terrane may be characterised by lithologies or modal proportions of lithologies that are markedly different to that of an adjacent block. Comparison of the southern Kapisillit Mapsheet (Kangeriuarsungup Tasersua and Qarliit Nunaat) with the central and northern Kapisillit Mapsheet (areas encompassing Norsanna and Tummeralik) is such an example (Fig. 5.1 and Appendix D). All areas in the Kapisillit region contain amphibolite of interpreted supracrustal origin. However Kangeriuarsungup Tasersua and Qarliit Nunaat in southern Kapisillit contain no observed occurrences of pelitic or semi-pelitic material. In addition, a greater proportion of amphibolite from these areas tends to exhibit relict cumulate textures (Lee, 2005, unpublished field report), suggesting an intrusive origin. This is notably different from the pelite-bearing supracrustal sequences in Norsanna and Tummeralik. More specifically, pelite-bearing supracrustal sequences have not been recorded south of the purported boundary of the Tasiusarsuaq terrane (Fig. 5.1). This observation, combined with geochronological evidence may suggest a fundamental difference in evolutionary history between the southern and northern Kapisillit areas.

A comparison of the supracrustal lithologies in Norsanna and Tummeralik show that both areas contain rocks of pelitic, mafic and ultramafic affinity, albeit in differing proportions. Pelitic units in Norsanna constitute a far smaller proportion of the supracrustal package than in Tummeralik. Conversely, ultramafic lithologies are more significant in the Norsanna supracrustal package. The consistent mafic-ultramafic-pelite structural stacking of supracrustal units is also an important characteristic of the Norsanna supracrustals that is not encountered in Tummeralik, where pelites occur principally as strung out boudins hosted by amphibolite.
Field studies have shown that Norsanna and Tummeralik contain mineralogically similar lithologies in terms of TTG orthogneisses and supracrustal sequences. The observed differences in the supracrustal lithologies may result from a number of factors, including exposure level, deformation or local differences in sediment source. However none of these need be directly related to the evolution and amalgamation of disparate terranes and consequently these criteria should not be used as an independent means of identifying different such terranes. Thus in the absence of geochronology there is no persuasive reason to propose the presence of distinct tectonic blocks, where this distinction is based solely of comparative lithology.

5.3 Orthogneiss suites in Norsanna and Tummeralik

Norsanna and Tummeralik are dominated by orthogneiss units, which often preserve good evidence of the deformational history of a region, as their extensive outcrop area generally gives a clear (and well-exposed) record of the area’s structural history. It is therefore worthwhile to consider the deformational history and boundary relationships of each of the three main orthogneiss units (polyphase and
homogeneous orthogneisses of Norsanna and tonalitic orthogneiss of Tummeralik). A comparison of the emplacement and metamorphic ages yielded by the geochronological study of the orthogneisses in Chapter 4 is given in Table 5.1.

Table 5.1: A comparative table summarising the ages of emplacement and metamorphism in the polyphase, homogeneous and tonalitic orthogneisses as discussed in Chapter 4.

<table>
<thead>
<tr>
<th></th>
<th>Norsanna</th>
<th>Tummeralik/Ameralla Fjord</th>
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<tbody>
<tr>
<td>Polyphase orthogneiss</td>
<td>Homogeneous orthogneiss</td>
<td>Tonalitic orthogneiss</td>
</tr>
<tr>
<td>Metamorphism ca. 2700Ma?</td>
<td>Emplacement 2806 ± 15 Ma</td>
<td>Metamorphism 2680 ± 11Ma</td>
</tr>
<tr>
<td></td>
<td>Emplacement 2806 ± 15 Ma</td>
<td>Metamorphism 2722 ± 8Ma</td>
</tr>
<tr>
<td>Metamorphism ≥3560Ma</td>
<td>Emplacement 3046 ± 10Ma</td>
<td>Emplacement 3027 ± 15Ma</td>
</tr>
<tr>
<td></td>
<td>Emplacement 3707 ± 7Ma</td>
<td></td>
</tr>
</tbody>
</table>

The polyphase orthogneiss appears to have the most complex structural history of the three orthogneiss suites, manifest as a complexly folded gneissic fabric, which is in turn locally overprinted by the planar S2 fabric in the Norsanna High Strain Zone (NHSZ). The unit shows evidence of D₆ gneissosity formation prior to the intrusion of a mafic suite, all of which were subsequently metamorphosed and deformed. Polyphase orthogneiss sample 482426 yielded magmatic zircon with an emplacement age of 3707 ± 7Ma, with significant high grade metamorphic zircon growth and recrystallisation ≥3560Ma. Comparable units correlated with the polyphase orthogneiss structurally underlie the homogeneous orthogneiss to the NE of Norsanna and have also yielded Eoarchaean magmatic zircon with ages up to ~3850 Ma (Heiss, 2006). This supports the interpretation that the polyphase orthogneiss lies structurally above and below the homogeneous orthogneiss in Norsanna (Allart, 1982; Friend et al., 1987).
The complexly deformed, Eoarchaean orthogneisses in Inner Ameralik have previously been correlated with the Amitsoq gneisses that crop out over much of the southern and western Nuuk region (Friend and Nutman, 2005a; Friend et al., 1987; Nutman and Friend, 2007; Nutman et al., 2004a). As discussed in Chapter 1, the Amitsoq gneiss is characterised by initial TTG emplacement between 3850 and 3660Ma, with granite sheet emplacement and high grade metamorphism between 3650 and 3600Ma. The suite is intruded by the 3500-3260Ma Ameralik dyke swarms (Gill, 1976; Gill and Bridgwater, 1976; McGregor, 1973; Nutman et al., 2004b) that have themselves been deformed and metamorphosed to amphibolite or granulite facies. The polyphase orthogneiss in Norsanna supports this correlation, as the unit displays the complexly deformed gneissosity and deformed mafic intrusives that characterise the Amitsoq gneiss. This interpretation is further supported by the Eoarchaean emplacement age (3707 ± 7Ma) yielded by sample 482426. In addition, the significant metamorphic zircon growth and recrystallisation at ≥3560Ma is correlated with the high grade thermal event observed in the Amitsoq gneiss at the base Palaeoarchaean. However the ca. 3560Ma population is interpreted as an ancient Pb loss trend in Norsanna and so represents an absolute minimum age of metamorphism. This Pb loss trend is considered to represent partial Pb loss in sample 482426 principally during D₂, as the sample locality lies within the NHSZ. The presence of older metamorphic ages (3600-3650Ma) in the polyphase orthogneiss (Heiss, 2006) in areas of low D₂ strain further supports this ancient Pb loss argument.

The tonalitic orthogneiss of Tummeralik shares some similarities with the polyphase orthogneiss in Norsanna. The unit exhibits an amphibolite facies D₅₉ gneissosity that deforms an early mafic suite. Prior to the conception of the terrane theory in the Nuuk region (Friend and Nutman, 1991; Friend et al., 1987; Nutman, 1991; Nutman and Collerson, 1991; Nutman et al., 1989; Nutman et al., 1993), the presence or absence of these early mafic suites was used as a key feature to distinguish between the Palaeoarchaean Amitsoq gneiss and the Meso- to Neoarchaean Nuk gneisses in the field (Bridgwater et al., 1974). These criteria were used to correlate the tonalitic orthogneiss in Tummeralik and the Amitsoq gneiss (Allart, 1982). However more recent geochronology has recognised these to be distinct magmatic suites (Friend and Nutman, 2005b; Nutman, 1991; Nutman and Friend, 2007; Nutman et al., 1989) with emplacement ages separated by ca. 700Ma
This is supported by geochronology from Tummeralik sample 496831, which yields an emplacement age of 3046 ± 10 Ma.

Sample 481418 from a biotite-tonalite/granodiorite from the northern shore of Ameralla Fjord yields an emplacement age of 3027 ± 15 Ma, statistically equivalent to the emplacement age of the tonalitic orthogneiss in Tummeralik. Regional mapping (Appendix D) suggests structural continuity between the geochronology sample localities on Ameralla Fjord and Tummeralik. Thus a correlation is suggested between the orthogneisses in Ameralla Fjord and Tummeralik, where orthogneisses from these two areas form part of a Mesoarchaean TTG orthogneiss suite. This interpretation is based on similarities in composition, structural setting and emplacement age and is in agreement Nutman and Friend (2007), who suggest the Mesoarchaean Kapsillik terrane covers much of the Kapisillit Mapsheet east of Itilleq Fjord. However this interpretation requires further work to properly establish the relationship between orthogneisses in Ameralla Fjord and Tummeralik. Dating work to address this issue is currently being undertaken as part of the Kapisillit Mapsheet Project. For the purposes of this study, an interpretation of a single Mesoarchaean orthogneiss suite is preferred as a working hypothesis, as this is the simplest interpretation at present to be supported by the existing geochronology. If, after further work, the two lithologies are found to be unrelated, then some distinction must be made between ‘Kapisillik’ and ‘non-Kapisillik’ Mesoarchaean orthogneisses in the southeast Nuuk region.

Unlike the polyphase and tonalitic orthogneisses, initial deformation in the homogeneous orthogneiss took place during D1. The homogeneous orthogneiss does not contain a deformed mafic suite, suggesting either that the lithology evolved independently of the polyphase orthogneiss, or that its emplacement postdates that of the mafic suite.

Homogeneous orthogneiss sample 482424 yields an emplacement age of 2806 ± 15 Ma, indicating that the unit would not host the Palaeo- to early-Mesoarchaean Ameralik dyke swarms in the event that homogeneous orthogneiss intruded the polyphase orthogneiss. Previous studies (McGregor et al., 1991; Nutman, 1991) have interpreted the orthogneiss in central Norsanna as the latest-Mesoarchaean Ikkattoq gneiss of the Tre Brødre terrane, which has a well-
constrained emplacement age of ca. 2825Ma (Crowley, 2002; Friend and Nutman, 2005b; Friend et al., 1987; Nutman and Friend, 2007; Nutman et al., 2004a). However recent mapping by Nutman and Hollis (Pers. Comm.) has identified a third orthogneiss hosted within the homogeneous orthogneiss. The unit has an emplacement age of 2858 ± 17Ma, and is indistinguishable in the field from the homogeneous orthogneiss. This emplacement age is compatible with the Tasiursarsuaq terrane orthogneisses to the south and has been interpreted as a tectonically interleaved slice of Tasiursarsuaq terrane (Nutman and Friend, 2006, Pers. Comm). The 2806 ± 15Ma age of the homogeneous orthogneiss is considered a minimum emplacement age due to evidence of recrystallisation in the sample that indicates post-crystallisation disturbance. A degree of ancient Pb loss has been identified in the polyphase orthogneiss, which also lies within the NHSZ, and therefore the possibility remains that the U/Pb systematics of the homogenous orthogneiss has also been affected by movement on the NHSZ. The emplacement age of the homogeneous orthogneiss is more compatible with the Ikkattoq gneiss than the Tasiursarsuaq orthogneisses, however the similarity in age of the Ikkattoq and Tasiursarsuaq suites suggests that further geochronology of the homogeneous gneiss in Norsanna is required, in order to establish the number of orthogneiss components within the unit. The homogeneous orthogneiss dated in inland Norsanna (sample 482424) is interpreted as Ikkattoq gneiss as ages of ca. 2858Ma orthogneiss have only been encountered in a small area near the northern shore of Ameralik Fjord. It is suggested that future geochronology should focus on any metamorphic zircon populations occurring in the 2858Ma orthogneiss, as a suite of Tasiursarsuaq affinity may preserve evidence of metamorphism at ca. 2795Ma, which would not be recorded in the Ikkattoq gneiss.

5.4 Ages of metamorphism in Norsanna and Tummeralik

The two leucosome samples from the northern shore of Ameralik Fjord constrain the age of leucosome crystallisation to ca. 2720 Ma. In both cases, the leucosome contains porphyroblasts of garnet and amphibole and demonstrates leucosome crystallisation under amphibolite facies conditions, while the S2-parallel vein networks also suggests that leucobolite emplacement was contemporaneous with movement on the NHSZ. In addition, the leucosomes themselves are not deformed.
by D₂, which may suggest their emplacement during the later stages of D₂ deformation. U/Pb geochronology of these two samples thus places a minimum constraint on the timing of D₂ and its associated sillimanite-grade metamorphism at 2720 Ma. A metamorphic zircon population at ca. 2720Ma is also encountered in the tonalitic–granodioritic orthogneiss on Ameralla Fjord (sample 481418). This finding is significant because the orthogneisses in this area lie to the east of the purported boundary between the Færingehavn-Tre Brødre and Kapisillik terranes (Nutman and Friend 2007; Friend and Nutman 2005) in a unit interpreted as belonging to the Kapisillik terrane. The presence of a 2720Ma event in ca. 3000Ma Kapisillik orthogneisses (Table 5.1) consequently places a minimum constraint of 2720Ma for the juxtaposition of the Færingehavn-Tre Brødre and Kapisillik blocks.

Despite the occurrence of 2720Ma metamorphism in Mesoarchaean orthogneisses on Ameralla Fjord, metamorphic zircon populations of this age are not encountered in Tummeralik, with orthogneiss and amphibolite facies leucosome samples yielding 2680 ± 11 and 2659 ± 17Ma respectively. This result may be produced by:

1. Two separate metamorphic events at ca. 2680 and 2659Ma.
2. A single metamorphic event at ≥2680Ma with subsequent Pb loss in sample 478334 to produce an apparent age of ca. 2659Ma. As demonstrated in Chapter 4, zircon in sample 478334 is metamict and altered. The age is thus considered to represent a minimum constraint on metamorphism in Tummeralik.
3. One event at ≥2680Ma, followed by a second event at ca. 2659Ma, where the ca. 2659Ma event re-opens the U/Pb system of the ≥2680Ma event, leading to partial resetting of the earlier metamorphic population. In this scenario, the 2680Ma is a hybrid age and does not represent the true age of metamorphism.

Some ca. 2680Ma ages are encountered for example near Kapisillik village (Appendix D) and on the west coast near Akilia Island. Zircon in the latter example also exhibits inclusions of sillimanite (Nutman and Friend, 2007), however it is considered that not enough evidence exists to correlate a relatively small number of geochronological samples dispersed over such as large area. Therefore, a metamorphic event at 2680Ma in Tummeralik is not preferred in this instance,
especially given the state of alteration and metamictisation in all zircons from this field area.

A major problem with the correlation of high grade metamorphic events in Norsanna and Tummeralik is the apparent absence of ca. 2720 metamorphic zircon in Tummeralik. Scenarios 2 and 3 allow for a significant degree of ancient Pb loss to produce an apparent event at 2680Ma. It should therefore be considered that a hybrid age of 2680 may plausibly be sourced from a ca. 2720Ma metamorphic population. This possibility is suggested based on the presence of 2720Ma metamorphic zircon the Ameralla Fjord sample, however this hypothesis should be investigated further before a confident interpretation can be made. In the event that Tummeralik and Norsanna formed a contiguous piece of crust by 2720Ma, the non-occurrence of 2720Ma metamorphic zircon populations may be result from:

1. Ameralla Fjord occupying a deeper structural level than Tummeralik at 2720Ma.
2. Tummeralik undergoing metamorphism at 2720Ma but with all evidence of the event being destroyed by subsequent metamorphic events.

As discussed in Chapter 1, the Kapisillik terrane is characterised by high grade metamorphism at ca. 2650Ma. Whilst the metamorphic age of 2659Ma from sample 478334 falls within error of this, the quality of the analysis is called into question by the high analytical error and significant discordance in a large proportion of analysed points. This is considered to reflect lattice damage through metamictisation (Pidgeon, 1992; Pidgeon and Kalsbeek, 1978; Pidgeon et al., 1998), combined with Pb loss (Chemiak and Watson, 2003), common Pb contamination through physical fractures (Meldrum et al., 1998; Mezger and Krogstad, 1997) and inadvertent analysis of zircon alteration products. Although possible that Tummeralik did undergo metamorphism ca. 2659Ma, further geochronological samples from less damaged grains would be required to confirm this. Thus the initial interpretation for sample 478334 remains, which suggests that 2659Ma represents an absolute minimum age for the timing of metamorphism in Tummeralik.

It is clear that the number of metamorphic events in Tummeralik cannot be conclusively identified based on the current knowledge of geochronology and metamorphic assemblages in the area. Given the apparent damage in the zircon
populations in Tummeralik, the study of the pressure-temperature paths of units in both Norsanna and Tummeralik becomes vital, in order to ascertain any link between the tectonothermal evolution of both areas. In addition, such studies of the metamorphic assemblages may also assist in identifying the causes of the severe Pb loss in the Tummeralik geochronology samples.

5.5 The development of high strain zones in Norsanna and Tummeralik

The position of Norsanna at the boundary between the Færingehavn-Tre Brødre terranes is a long-standing interpretation of the Nuuk region terrane geometry (Friend et al., 1987; Friend et al., 1988) and is discussed in Chapter 1. Also detailed is the suggestion that Tummeralik lies within the interior of a crustal block. The differing positions of the two field areas with respect to purported terrane boundaries is considered here as a mechanism to explain the variation in the distribution of strain in Norsanna and Tummeralik.

The principal difference in deformation between Norsanna and Tummeralik is the scale of the high strain zones in the areas. These zones are formed at lithology boundaries in both areas but are developed on 50-100m scales in Norsanna versus 10-50m scales in Tummeralik. A typical Nuuk region terrane is separated from adjacent blocks by a marked tectonic break, usually a 10- to 50m wide mylonite zone, which may itself be deformed (McGregor et al., 1991). Mylonites associated with the NHSZ were encountered on the north shore of Ameralik Fjord in Norsanna and interpreted as ‘recrystallised mylonite zones’ (Nutman and Friend, 2007). However it is suggested here that the mylonites of Nutman and Friend (2007) may be the cm- to m-scale D4N mylonite and blastomylonite zones, which at localities on the north shore of Ameralik Fjord run at a low angle (<20°) to the regional S2 gneissosity. The absence of such zones in Tummeralik may be explained by its interpreted position in a terrane interior. However no mylonitic shear zones are encountered in either area that could be interpreted as ‘terrane boundaries’.

The NHSZ stands out as a candidate for a terrane boundary because the zone separates two orthogneiss suites of markedly different age (ca. 900Ma separates the emplacement ages of the polyphase and homogeneous orthogneisses) and state of
deformation. However the large scale of the NHSZ and the apparent absence of associated mylonites sets it apart from other proposed terrane boundaries in the Nuuk region, as a structural thickness of >120m is considerably greater than the 10-50m wide zones indicated elsewhere (McGregor et al., 1991). It should be noted however, that a high strain zone of the scale of the NHSZ need not form as a result of terrane juxtaposition. The following section discusses the nature of the boundary between the homogeneous and polyphase orthogneisses and considers other mechanisms of formation for the NHSZ.

Previous studies (Allart, 1982; unpublished GEUS field maps) suggested that the polyphase and homogeneous orthogneiss suites and an anorthosite body were folded about a large isocline, which was cored by the homogeneous orthogneiss and anorthosite. However neither the anorthosite body nor the isoclinal fold closure (Fig. 5.2a) were observed in Norsanna, suggesting that the homogeneous orthogneiss lithology may instead be a tectonically emplaced sliver (Fig. 5.2b).

![Figure 5.2: Possible structures represented by the NHSZ. a) NHSZ is a high strain zone developed on the limb of a km-scale isoclinal fold cored by the homogeneous orthogneiss. The NHSZ forms the upper and lower boundary to the homogeneous orthogneiss. b) Homogeneous orthogneiss forms a sliver that has been tectonically interleaved with the polyphase orthogneiss along the NHSZ. A subordinate high strain zone forms on the lower boundary of the homogeneous orthogneiss.](image-url)
Alternatively, the boundary between the two orthogneiss suites may be a highly tectonised igneous contact (Fig. 5.3), where the NHSZ developed due to the competency contrast between the strain-hardened polyphase orthogneiss and the previously undeformed homogeneous orthogneiss. Strain associated with the NHSZ is preferentially partitioned into the homogenous orthogneiss, suggested by a larger D2 deformation zone in the homogeneous orthogneiss than the strain hardened polyphase orthogneiss. It is suggested that competency contrasts are an important factor in the development of high strain zones in both Norsanna and Tummeralik, as the competency of a lithology governs its response to stress and hence may influence the distribution of strain over a large area. An example of this is Tummeralik, where lithologies are dominated by the tonalitic orthogneiss. The unit shares a similar deformational history with most of the other major lithological units in the study area (with the exception of the granodioritic orthogneiss). The greater lithological and structural homogeneity, caused by the dominance of a single lithology in Tummeralik, may limit the amount of strain partitioning across the area by reducing regional competency contrasts between units of different composition and state of strain. In the absence of regional scale high strain zones, the distribution of strain in Tummeralik is more homogeneous, forming m- to 10m-scale high strain zones at lithology boundaries where competency contrasts are most marked as a result of compositional variation.

This forms a marked contrast with Norsanna, where the NHSZ straddles a boundary with a steep strain gradient. The mechanisms that produce high strain zones through strain focussing do not require the juxtaposition of disparate terranes in order to operate. It is argued here that the formation of high strain zones as a result of competency contrast is a key process occurring during D2, which may occur in conjunction with terrane amalgamation as well as independently.
5.6 Structural evolution of Norsanna and Tummeralik

Lithologies in both Norsanna and Tummeralik exhibit significant $D_n$ deformation. As discussed above, no correlation is drawn between $D_n$ events in the polyphase and tonalitic orthogneisses (Fig. 5.4a). This is because the polyphase orthogneiss records a history of emplacement, deformation and intrusion of a mafic dyke suite starting 700Ma before the emplacement of the tonalitic orthogneiss. The following section considers the post-$D_n$ structural evolution of Norsanna and Tummeralik, as these are the first events that are quantifiable in both areas, in addition to sharing a number of common features. A comparison of the structural evolution in Norsanna and Tummeralik is given in Table 5.2.

In Norsanna and Tummeralik, $D_1$ is characterised by a phase of isoclinal folding (Fig. 5.4b), with a single km-scale fold recognised in each area (Norsanna
and Tummeralik Isoclines respectively) and few preserved parasitic folds. The F₁ Tummeralik Isocline follows an E-W trend, however the Norsanna Isocline has been refolded by later events and thus its initial orientation must be estimated. Reconstructions of the Norsanna Isocline, in which the effects of the F₁₃N structure were removed, suggest that the F₁ fold may have followed an E-W trend prior to D₃N. In both areas, D₁ would thus indicate a period of N-S compression.

The D₂ events in Norsanna (D₂₃N) and Tummeralik (D₂T) are characterised by intense F₁ fold limb attenuation and boudinage (Fig. 5.4c), however the state of D₂ strain in both areas varies in style (section 5.5). The S₂ gneissosity in Tummeralik and the orientation of the NHSZ have developed parallel to the F₁ axial plane. In addition, the attenuation of F₁ fold limbs is approximately coaxial with F₁ axial planes, as boudinaged fold limbs only undergo minor D₂ rotation and are not refolded. These structures all suggest that the orientation of the principal stress field did not alter greatly between D₁ and D₂ and the two events are consequently interpreted as two phases of a single progressive event. Evidence from the extensive boudinage in both areas, and shear indicators from the NHSZ may suggest a degree of extension associated with D₂. This hypothesis would require further evidence of decompression and is addressed in Chapters 6 and 7. The dating of late-D₂ at 2720Ma supports previous interpretations that deformation in Norsanna is related to the juxtaposition of the Tasiusarsuaq at this time. The identification of 2720Ma metamorphism in Ameralla Fjord, combined with evidence of N-S compression in Tummeralik also indicates that the southern part of the purported Kapisillik terrane was also affected by the Tasiusarsuaq collision at this time. This suggests that Norsanna and Tummeralik share a common history prior to 2720Ma or if not, that the two blocks were in close enough proximity to develop similar responses to stresses imposed by the Tasiusarsuaq block.

After D₂₃N attenuation, Norsanna was subject to a phase of folding that is not encountered in Tummeralik. This resulted in the formation of a km-scale synformal structure that controls the outcrop pattern in Norsanna and caused the folding of the Norsanna Isocline and NHSZ (Fig. 5.4d) to its present N-S strike. Despite the correlation of the D₁ and D₂ events in Norsanna and Tummeralik, the restriction of a regional fold-forming event, such as D₁₃N to Norsanna could argue against a common post-D₄ tectonothermal history for the two areas.
Figure 5.4: Comparative schematic structural evolution of Norsanna (left) and Tummeralik (right). a) Speculative pre-$D_n$ intrusive relationships. The primary relationship between the polyphase and homogeneous orthogneiss is not considered here. b) $D_1$ deformation forming the km-scale Norsanna and Tummeralik Isoclines. c) $D_2$ deformation coaxial to $D_1$ causes attenuation of fold limbs and extensive boudinage. The NHSZ develops in Norsanna whereas smaller scale high strain zones form in Tummeralik. d) Open folding of the NHSZ during $D_{3N}$ is not encountered in Tummeralik although both areas are intruded by coarse granite pegmatites. Block diagrams in Norsanna look S after folding by $D_{3N}$. e) Norsanna and Tummeralik are intruded by the doleritic to picritic dykes that are approximately contemporaneous with $D_{4N}$ and $D_{2T}$. 
### Table 5.2: Comparative structural evolution of the Norsanna and Tummeralik study areas

<table>
<thead>
<tr>
<th>Event</th>
<th>Grade</th>
<th>Description</th>
<th>Event</th>
<th>Grade</th>
<th>Description</th>
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</thead>
<tbody>
<tr>
<td>$D_n$</td>
<td>?</td>
<td><strong>Initial formation of gneissic fabrics and compositional banding.</strong> Multiple folding events in older orthogneisses</td>
<td>$D_n$</td>
<td>?</td>
<td><strong>Initial formation of gneissic fabrics and compositional banding. Unknown number of events</strong></td>
</tr>
<tr>
<td>$D_1$</td>
<td>Upper amphibolite facies</td>
<td>• Km-scale isoclinal folding of $D_n$ gneissosity in narrow high strain zones • Development of sillimanite-biotite peak fabric</td>
<td>$D_1$</td>
<td>Upper amphibolite facies</td>
<td><strong>Km-scale isoclinal folding of $D_1$ fabrics</strong> • Development of sillimanite-biotite peak fabric</td>
</tr>
<tr>
<td>$D_2$</td>
<td>Amphibolite facies</td>
<td>• Interleaving of polyphase and homogeneous orthogneisses and supracrustal units. • Development of pervasive platy fabric in high strain zone and dominant gneissic fabric in homogeneous orthogneiss. • Boudinage of supracrustal units and attenuation of isoclinal fold limbs</td>
<td>$D_2$</td>
<td>Amphibolite facies</td>
<td><strong>Intensification of gneissic fabrics</strong> • Boudinage of supracrustal lithologies • Attenuation of $F_1$ isoclinal fold limbs and locking/shearing of fold hinges • Rotation of boudinaged supracrustal lenses.</td>
</tr>
<tr>
<td>$D_{3N}$</td>
<td>Amphibolite facies</td>
<td>• Refolding of $D_1$ structures into km-scale open structure</td>
<td></td>
<td></td>
<td><strong>No open folds observed</strong></td>
</tr>
</tbody>
</table>
| $D_{4N}$ | Greenschist facies | • Localised cataclasis and shearing affecting late granite pegmatites • Single NE-trending chlorite grade shear zone partitioned into at least three mylonitic zones less than 50m width. | $D_{3T}$ | Greenschist facies | **Tummeralik - Brittle faulting with small-scale mylonitic textures bearing epidote-chlorite in** • **Reconnaissance area – Epidote-chlorite (blasto-) mylonitic shear zones with moderately to steeply-plunging open folds of gneissosity and well-pronounced associated lineation**

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Grain pegmatite intrusion late syn- to post $D_{n+2}$
A corresponding F3N antiform is encountered east of Itilleq Fjord (Appendix D), which folds Mesoarchaean rocks correlated with the tonalitic orthogneiss in Tummeralik. However, the D3N event does not appear to extend as far east as Tummeralik itself. It is therefore suggested that D3N is a relatively localised folding event whose influence decreases moving east, potentially as a result of distance or structural level at the time of folding.

The phase of deformation associated with greenschist facies retrogression (D3T and D4N for Tummeralik and Norsanna respectively) is apparent in both areas as brittle faulting and localised mylonitisation (Fig. 5.4e). This event best preserved in Norsanna, where chlorite grade mylonitisation of almost all units show D4N shear zones to be some of the youngest structures in the area. The D4N mylonite zones are demonstrably Proterozoic in age, indicated where they cut, and are cut by, different generations of a Proterozoic mafic dyke swarm (sensu Cadman et al., 1993). The age of greenschist facies shearing is more difficult to determine in Tummeralik, but is thought to be Proterozoic or younger, based on fault truncations of mafic dykes to the north of Tummeralik (Appendix D). The number of low grade shearing events that occurred in Norsanna and Tummeralik remains uncertain. Both areas display NW-trending chlorite ± biotite-grade shear zones on a variety of scales. In Tummeralik, faults of this orientation are dominant and form the largest D3T structures, whereas in Norsanna, NW-trending faults are not recorded on scales larger than 10cm thickness and are subordinate to the NNE-trending mylonites that are exposed south of Ameralik Fjord. There were no observed cross cutting relationships between the two orientations of fault, and therefore it is difficult to constrain the relative timing of formation of the two sets of fractures. In terms of regional significance, the D4N shear zones run approximately parallel to the Proterozoic Kobbefjord and Ivingnguit faults (Hollis et al., 2004), which are major lineaments in the Nuuk region. This suggests that D4N is regionally significant west of the Kapisillit Mapsheet, but the intensity and regional significance of this event decreases moving east, becoming apparently subordinate to D3T in Tummeralik.
5.7 Summary

Emplacement ages of orthogneisses in Norsanna and Tummeralik correlate the units in the two areas with the Amitsoq-Ikkattoq and ‘Kapisillik’ orthogneiss suites. These relate to the Færingehavn-Tre Brødre and Kapisillik terranes, respectively (Crowley, 2002; Friend and Nutman, 2005b; Friend et al., 1987; Nutman and Friend, 2007). The evidence presented above argues for a common post-D₄ structural evolution in Norsanna and Tummeralik based on the following conclusions:

1. N-S compression during D₁ and D₂ resulted in initially E-W-trending regional isoclinal folds, followed by F₁-coaxial high strain zone development and intense attenuation (possibly extensional) of fold limbs.
2. Attenuation and shearing had commenced by 2720Ma in Norsanna and extended at least as far east as Ameralla Fjord.

A single metamorphic age of 2659Ma in Tummeralik may correlate with upper amphibolite facies metamorphism in the central Kapisillik terrane at this time. However this age remains contentious due to the questionable quality of the analysis. Norsanna was subsequently affected by a semi-localised phase of open folding that did not extend as far east as Tummeralik but is encountered in Mesoarchaean gneisses to north and west of Tummeralik. Both areas were then variably affected by at least two phase of faulting at greenschist facies conditions, which was coincident with the intrusion of the mafic dykes in Norsanna and possibly Tummeralik.

A preliminary correlation between post-D₄ Norsanna and Tummeralik is presented in this chapter based on structural and geochronological correlations. This hypothesis is tested in the following chapters using metamorphic assemblage analysis to identify the P-T paths followed in both areas. In chapter 6, detailed petrology of pelitic gneisses from Norsanna, Tummeralik and Aputitooq Mountain (at the purported Færingehavn/Tre Brødre-Kapisillik boundary) will be discussed, in order to better constrain the pressure-temperature path of the central Kapisillit mapsheet. The study will also attempt to identify any relict metamorphic assemblages, which may indicate independent tectonothermal evolution prior to juxtaposition. The integration of these studies will allow for a critical evaluation of the P-T-t-deformation evolution of the central Kapisillit mapsheet its implication for the tectonic evolution of the Nuuk region in general.


CHAPTER 6

METAMORPHIC PETROLOGY AND P-T RECORD IN METAPELITIC UNITS OF THE KAPISILLIT REGION

6.1 Overview

Pelitic and semi-pelitic rocks constitute a small but significant proportion of the lithological units in the Kapisillit region. Although comprising <1% of the total outcrop, the chemistry of these lithologies allows the development of low variance mineral assemblages, which may be used to accurately constrain the P-T path of the immediate area. Samples were collected from Norsanna, Tummeralik and Aputitooq Mountain for detailed petrographic analysis using both optical and scanning electron microscopy. In situ major element analyses of the major minerals and x-ray maps of zoning and inclusion patterns in garnet (section 6.2) were obtained from the same thin sections using the SX100 EPMA facility at the University of Edinburgh. Details of instrument set up and analytical technique are given in Appendix A. Mineral abbreviations are in accordance with Kretz, (1983).

6.2 Interpretation of zoning patterns in garnet

X-ray mapping of Fe, Mg, Ca, Mn and Y was carried out on garnets from all pelitic and semi-pelitic samples in the Kapisillit sample suite, in order to ascertain major and trace element zoning patterns. The garnet lattice substitutes Fe, Mg, Ca and Mn into three M-sites, so producing the almandine (alm), pyrope (pyp), grossular (grs) and spessartine (sps) end members respectively. The differing size, mobility and compatibility of these ions require that particular garnet compositions are favoured under certain pressure-temperature (P-T) conditions. Thus the composition of garnet is sensitive to changes in pressure and temperature. These changes may be recorded as zoning patterns in garnet, thus providing information on the evolution of the P-T conditions and the character of metamorphism.

Mineral zoning may represent changes in P-T conditions throughout the prograde and retrograde paths. In garnet, zoning patterns produced during garnet growth are termed ‘growth zoning’ and are usually manifest on the prograde path.
Zoning patterns produced under increasing pressure and/or temperature are referred to as prograde zoning and exhibit a number of defining characteristics that are briefly outlined here. Garnet is a sink for Mn and thus prograde zoning typically involves high initial spessartine. This decreases from core to rim as the effective bulk composition of the rock becomes depleted in Mn. Increasing pressure is usually reflected in an increasing pyrope (Mg) component moving from core to rim. This is because the Mg$^{2+}$ ion is the smallest of the four M-site ions and so is most stable at higher pressures. Greater entropy associated with increasing temperature results in increasing grossular (Ca) contents, as Ca$^{2+}$ is the largest ion to substitute into the M-sites. Retrograde zoning – produced by garnet growth during decreasing pressure and/or temperature – will display the opposite zoning patterns i.e. increasing spessartine moving towards grain rims, with decreasing grossular and pyrope (Bucher and Frey, 1992; Pyle and Spear, 1999; Spear 1993; Yang and Rivers, 2002).

Garnet may also develop zoning as a result of elemental diffusion. This is common at grain edges and along fractures. It is marked principally by changes in Mg and Mn, which remain highly mobile in the garnet lattice. In particular, the diffusion of Mg out of garnet is greatest where a garnet is in contact with a phase that preferentially partitions Mg, such as biotite or cordierite. Diffusive zoning is rarely seen in Ca and Y, as these have a low mobility in the garnet lattice. Consequently, growth zoning in Ca and Y may be preserved after prograde zoning in Fe, Mn and Mg is destroyed by retrograde or diffusive zoning. It is for this reason that Ca and Y are included in routine garnet maps for the samples in the Kapisillit sample suite.

6.3 Common mineral assemblages and textures in the Kapisillit region

The majority of samples analysed in Norsanna, Tummeralik and Aputitooq Mountain are characterised by the assemblage garnet-sillimanite-biotite-plagioclase-quartz, with variable development of staurolite and cordierite. Accessory minerals include ilmenite, rutile, pyrite, apatite, tourmaline, zircon and monazite. Appendix C contains all mineral assemblage and mineral geochemistry tables referred to in this chapter. Comparative tables for major and accessory mineral assemblages from all samples are presented in Appendix C; Tables 1 and 2, whilst representative EPMA analyses of the major minerals are presented in Appendix C; Table 3.
Modal abundances of each mineral vary between samples but this is particularly marked by sillimanite, which may form solely as inclusions in garnet, (e.g. sample 492234), but more often forms a key framework mineral. Biotite is another mineral that shows significant variation in modal abundance, exemplified by samples 481439 and 478343, which contain approximately 10 and 30 modal% respectively.

Biotite and sillimanite are commonly intergrown and define the dominant fabric, which in most cases is a prominent gneissose or schistose SL tectonite. Gneissic banding in pelitic units is defined at outcrop and thin section scales by alternating layers of biotite-sillimanite and plagioclase-quartz (Fig. 6.1) that invariably wrap garnet porphyroblasts. These quartzofeldspathic layers may be continuous, with a similar grainsize to the rest of the matrix or may form discontinuous, annealed pods and lenses that are coarser than other framework minerals. The former texture is interpreted to represent compositional banding, whereas the coarse grained pods and lenses are considered to reflect small volumes of partial melt and are not present in all samples. The gneissic fabrics described above correspond to the regional $S_{1,2}$ fabric in Norsanna and Tummeralik and the dominant fabric in Aputitooq Mountain.

![Figure 6.1: Typical gneissic fabrics in pelitic material in the Kapisillit region. Lithology shown is a garnet-sillimanite gneiss from Tummeralik at UTM 22W 557239E, 7127886N.](image-url)
Four samples in Norsanna, Tummeralik and the Ice Lake area exhibit an additional sillimanite texture in addition to the dominant acicular grains that are intergrown with biotite. Samples 482402, 482435, 478343 and 478325 contain rare, blocky to lozenge-shaped grains of sillimanite, which lie parallel to the gneissic fabric (Fig. 6.2). These are characterised by the development of small ($\leq 50\mu m$) subgrains, which are best defined in grain centres. Rim regions are typically formed of individual sillimanite grains or fibrolite with a similar size to the subgrains, suggesting their formation by subgrain rotation. Where subgrained sillimanite domains have retained a coherent grain shape, the squat, lozenge-shaped morphology is reminiscent of kyanite. It is suggested that these domains represent sillimanite pseudomorphs after early kyanite, where the development of subgrains resulted from crystal lattice distortion associated with the kyanite-sillimanite transition (Goergen et al., 2008).

Figure 6.2: Examples of sillimanite showing subgrained textures in samples 478325 (a), 482402 (b) and 482435 (c). Scale bars measure 500$\mu m$. d) STEM image showing lattice distortion and subgrain development during the kyanite-sillimanite transition in experimentally deformed kyanite (Goergen et al., 2008).
6.4 Pelitic and semi-pelitic units of Norsanna

Figure 6.3: Simplified geological map of Norsanna showing key lithologies, F1 and F3 fold hinges and the localities of pelite samples 482402, 482435, 481439 and 481440.
*Garnet-aluminosilicate ± staurolite gneiss*

Sample 482402 was collected from a ~5m thick boudinaged lens hosted by the homogeneous orthogneiss within the NHSZ and sample 481439 from a pelitic body hosted by supracrustal amphibolite ~7km along strike of 482402 (Fig. 6.3). In hand specimen, both samples display mm- to cm-scale gneissic banding defined by changes in the modal percentage of ferromagnesian and quartzofeldspathic minerals and are medium to coarse (0.5-1mm), although sillimanite may measure up to 2mm. In thin section, the matrix is dominated by plagioclase, quartz, biotite and sillimanite in textural equilibrium, with accessory ilmenite growing in association with matrix biotite. Sillimanite in both samples is intergrown with coarse matrix biotite, where it forms coarse, acicular laths that define the peak L-S gneissosity. Blocky, subgrained patches of former kyanite (section 6.3) constitute a second sillimanite texture in 482402 (Fig. 6.4a). A single grain of staurolite is observed in sample 481439 and is the only occurrence of matrix staurolite in the entire Kapisillit sample suite. The staurolite here lies in between garnet and sillimanite, and is partially pseudomorphed by sillimanite (Fig. 6.4b).

Numerous, pale pink garnet porphyroblasts (1-3mm) contain inclusion-rich core domains containing rutile, biotite and quartz. Inclusion trails in cores are often aligned, but the orientation varies between garnets. Rims tend to be largely inclusion-free, however isolated inclusions of fine grained zircon and monazite occur as do rare clusters of sillimanite (Fig. 6.5).
Figure 6.4: a) (XPL) Sil pseudomorphs after Ky (section 6.3) lying parallel to the peak Sil-Bt-Pl-Q fabric in sample 482402. b) (PPL) Photomicrograph of sample 481439, showing the development of retrograde St from the reaction between Grt and Sil that resides in a dominantly Bt-Pl-Q matrix. Scale bar measures 500μm.
Figure 6.5: Inclusion assemblage in garnet in sample 482402. Grt cores contain Rut and Q, while rims contain clusters of Sil. The Grt porphyroblast is hosted by Q and Bt in the matrix, and fractures are infilled by Ms-Chl.

Where in proximity to a coarse, annealed plagioclase-quartz horizon in sample 482402, biotite is partially or completely replaced by chlorite (Fig. 6.6a), and sillimanite is mantled by fine aggregates of white mica. In addition the biotite-sillimanite fabric is cut at a high angle by white mica veins (Fig. 6.6b). This alteration becomes less pronounced with distance from the plagioclase-quartz horizon; however biotite invariably displays a degree of chloritisation, demonstrated in Fig. 6.7. The marked trend of biotite towards chlorite in Fig 6.7 is due to electron probe analyses sampling mixed layers of biotite and chlorite, producing an overall hybrid chlorite-biotite composition. These textures are interpreted to reflect alteration by fluids that have utilised the plagioclase-quartz horizon as a conduit. In addition, biotite may contain small inclusions of ilmenite that are concentrated along cleavage planes but may also form larger (≤600μm) irregular grains overgrowing sillimanite. The peak mineral assemblage in 481439 remains largely unaltered.
Figure 6.6: a) SEM image (BSE) showing the effects of the greenschist facies partial overprint in sample 482402. Matrix Ht yields variable HSE responses due to variable levels of chloritisation and is dusted with coarse- and fine-grained Ilm. Sil is intergrown with matrix Bt but is partially pseudomorphed by fine-grained aggregates of white mica. b) Micrograph (XPL) showing muscovite veins (red) cutting the peak fabric defined in yellow. Scale bar measures 500μm.

Garnets in both samples have dominantly almandine-pyrope compositions and variable grossular and spessartine. Garnets in 482402 exhibit homogeneous core regions with respect to the major elements and Y. However some zoning is seen in Fe, Mn and Mg within 100μm of grain rims, which is especially pronounced along fractures infilled with biotite and/or chlorite (Fig. 6.8a and b; Fig 6.9a and b). Fe and Mn increase moving towards grain rims, which is matched by a corresponding decrease in Mg. These features are consistent with compositional change induced by retrograde reaction (increasing Mn), coupled with diffusional re-equilibration (Fe-Mg exchange), as observed commonly in medium to high grade garnets worldwide.
Garnet zoning in 481439 is variable, with a number of garnets exhibiting homogeneous Ca and/or Mn profiles. However the compositions of core material in homogeneous garnets are the same as rim compositions in zoned grains, suggesting the variability in zoning may be due to sectioning effects. Zoning in Fe is more prominent in 481439. X-ray maps and major element zoning profiles in the most zoned garnet (Fig. 6.8c and d; Fig 6.9c and d) show marked zoning in Ca which decreases across an irregular boundary separating core regions from rims. This is interpreted as an overgrown resorption surface. Although this surface is also depicted by a steep count gradient on x-ray maps in Fe, Mn and Mg, the irregular surface of the boundary is only preserved in Ca, suggesting that pre-existing growth zoning in Fe, Mn and Mg has been affected by later diffusion. The preservation of zoning in Ca and not Mg, Fe and Mn is attributed to the significantly lower immobility of the Ca\(^{2+}\) ion in the garnet lattice relative to Fe, Mg and Mn.

Figure 6.7: AFM projection from the retrograde Ms-bearing pelite 482402 showing biotite compositions tracking towards chlorite as a result of low grade alteration.

Plagioclase varies in composition between 482402 and 481439, but is unzoned and constant within the individual samples. Matrix plagioclase in 482402 has an oligoclase composition (X\(_{ab}\) = 0.77-0.81) and plagioclase in 481439 has an andesine composition (X\(_{ab}\) 0.55-0.59). Plagioclase in both samples is unzoned with little compositional variation between grains. Analyses of the least altered biotite grains (near to or more than 9 wt% K\(_2\)O and >95wt% oxides) suggest dominantly phlogopitic compositions, with appreciable eastonite and annite components (Table
6.3). $X_{\text{Mg}}$ ranges from 0.56-0.6 and ferric iron varies from 8.7 to 12.6% of $\text{Fe}_{\text{tot}}$. Biotite in 481439 is less altered, with marginally higher $X_{\text{Mg}}$ (0.63-0.68) and $\text{Fe}^{3+}$, which represents 12.5-20.46% of $\text{Fe}_{\text{tot}}$. Staurolite in 481439 is homogeneous, with $X_{\text{Mg}}$ 0.23-0.24 and Zn 0.26-0.27 per formula unit (p.f.u).

Figure 6.8: Comparative X-ray maps of major elements in garnet in samples 482402 (a, b) and 481439 (c, d). a) Mg in garnet dropping within 10-20µm of grain edges and fractures, which are most pronounced where garnet is in contact with biotite. b) Corresponding increase in Mn near grain boundaries and fracture. c) Ca growth zoning in 481439 separated from lower-Ca rims by a sinuous resorption front. d) Fe zoning opposite to Ca zoning in the same garnet.
Figure 6.9: Major element traverses across garnet in samples 482402 (a, b) and 481439 (c, d). X_{gpn} and X_{sps} traverses rescaled in b and d to show small-scale zoning patterns. A sharp increase between cores and rims in 481439 is coincident with the resorption front in Fig 6.9c.

**Garnet-sillimanite-cordierite gneiss**

Sample 482435 was collected from the northern shore of Ameralik Fjord, east of the main deformation zone of the NHSZ but still showing pervasive D_{2} deformation (Fig. 6.3). The outcrop forms a ~3m wide band, bound to the west by the heterogeneous amphibolite and to the east by the polyphase orthogneiss. The sample is medium grained, with an average matrix grain size of 0.2-0.5mm and a gneissic fabric dipping moderately to the east. Acicular sillimanite forms the dominant aluminosilicate texture, but the sample also contains rare subgrained sillimanite pseudomorphs after kyanite (section 6.3) that are rimmed by fibrolitic sillimanite (Fig. 6.10a). These grains are typically adjacent or near to garnet porphyroblasts and lie parallel to the main fabric. Cordierite occurs in the matrix, where it pseudomorphs sillimanite (Fig. 6.10b).
Figure 6.10: a) (XPL) Matrix assemblage in sample 482435 showing garnet porphyroblasts and Sil pseudomorphs after Ky in a Bt-Sil-Pl-Q matrix  b) (XPL) Two Sil textures: blocky, subgrained Sil denoting pseudomorphs after Ky and acicular, Sil intergrown with Bt to form the peak Sil-Bt-Pl-Q fabric. Peak Sil is mantled by Crd, which also overgrows the Ky pseudomorphs. Image ‘b’ has been to optimise birefringence in both Sil textures. Scale bars measure 500μm.

Cordierite is also locally replaced by chlorite in the presence of biotite (Fig. 6.11a), although textures elsewhere in the sample show cordierite and biotite in stable contact. Garnet forms abundant, pale pink, sub- to anhedral porphyroblasts (500-1000μm) that exhibit three textural domains (Fig. 6.11b). Domain 1, which is not present in all garnets, forms inclusion-poor cores containing rare inclusions of fine (<10μm) plagioclase and zircon. Domain 2 is present in the majority of garnets and forms ~200μm wide inclusion-rich bands around domain 1. Where domain 1 is
absent, domain 2 forms cores. EDX of domain 2 identified fine grained inclusions of chlorite, in addition to coarser grained quartz and rutile, with minor aluminosilicate, biotite and small (<5μm) monazites. Domain 3 is present in all garnets, forming >100μm thick rims and is inclusion poor. The domain may truncate domains 1 and 2, suggesting a period of resorption between the growth of 2 and 3 but is in textural equilibrium with the biotite-sillimanite fabric.

Figure 6.11: a) SEM (BSE) image of sample 482435, showing Crd pseudomorphs of acicular Sil that defined the peak Sil-Bt-Pl-Q fabric. The Crd itself is subsequently partially replaced by Chl during greenschist retrogression. b) SEM image (BSE) showing textural domains in garnet in sample 482435, with textural domains 1 and 2 truncated by domain 3 rims.

Garnet compositions in this sample are dominated by almandine-pyrope mixes with low grossular and spessartine (Fig. 6.12). Despite the observed textural complexity, x-ray mapping revealed only minimal zoning in major elements and Y. Domain 1 is slightly depleted in Mn and elevated in Y (Fig. 6.12a and b). Domain 2 shows some minor fluctuations in Mn and Ti adjacent to rutile inclusions but otherwise exhibits a similar chemistry to domain 3 (Fig. 6.12a and c). A discontinuous zone of elevated Mn is apparent at the outermost rim of domain 3, which is most pronounced where garnet is in contact with biotite. Zoning in Mn (and Ti) is interpreted as retrograde diffusive exchange between garnet and biotite (and rutile).
Figure 6.12: a-c) EPMA X-ray maps of garnet in sample 482435. a) Typically homogeneous Mn profiles increase marginally around inclusions in domain 2 and sharply at grain edges, especially where garnet is in contact with biotite. b) Elevated Y in domain 1, which drops off moving into domain 2, after which Y-counts remain steady. c) Ti X-ray map showing rutile inclusions in domain 2. Garnet domains are denoted by white dashed lines and the garnet outline is denoted by yellow dashes. d) Major element garnet traverse for the mapped grain along the traverse presented in c (solid line).

Plagioclase is unzoned with an andesine composition ($X_{ab} = 0.66-0.69$) which varies little with across the sample. Biotite, which may be partially replaced by chlorite, is also unzoned and displays little compositional variation ($X_{Mg} 0.62-0.65$) with predominantly phlogopitic compositions. Ferric iron in biotite varies between 16 and 19% Fe$_{tot}$. Cordierite compositions are also homogeneous with respect to Fe and Mg ($X_{Mg} 0.73-0.74$).
Orthoamphibole-bearing gneiss

Sample 481440 was collected from the supracrustal assemblage in the NHSZ at the same locality as geochronology sample 481439 (Chapter 4). The unit is medium grained, with matrix grain size varying from 0.1 to 0.3mm. In contrast to the other samples in Norsanna, 481440 is sillimanite-absent and the gneissic fabric is instead defined by cm scale biotite-amphibole and plagioclase-quartz banding (Appendix C, Table 1; Fig. 6.13). Plagioclase forms two textures; an early generation that is in textural equilibrium with the biotite-amphibole fabric, and a later generation that mantles garnet and is intergrown with a second generation of biotite (Fig. 6.13a). Ilmenite is present in the matrix as coarse grains that are often associated with biotite in the matrix. The gneissic fabric wraps coarse (≤10mm diameter), pink, anhedral, garnets that exhibit complex growth and inclusion textures. Garnets may be poikiloblastic (Grt₁) or porphyroblastic (Grt₂). Grt₁ grains contain ‘centres’ that are inclusion-poor and interconnected by ‘mantles’ of poikiloblastic garnet containing large, closely-spaced inclusions of quartz. Randomly oriented ilmenite and sphene are present in garnet ‘centres’ and ‘mantles’ but may develop a preferential orientation moving towards grain rims (Fig. 6.14a). Grt₁ textures are rimmed by porphyroblastic garnet that may contain Fe-Ti oxides but lack the large quartz inclusions. The rims are commonly embayed by plagioclase and biotite at their margins. Grt₂ grains are porphyroblastic and similar in appearance to the
porphyroblastic rims on Grt₁, containing oriented inclusions of ilmenite, sphene and quartz. In addition to these dominant garnet textures, a few small (<0.5mm), inclusion-free garnets are present, which are highly embayed and replaced by biotite-plagioclase (Fig. 6.14b).

Figure 6.14: a) (PPL) Micrograph showing the inclusion-rich Grt₁ morphology in sample 481440. Large quartz inclusions give the grain a poikiloblastic texture in places, which mantle relatively inclusion-poor centres. Black ilmenite inclusions are randomly oriented in the grain centre but develop a preferential orientation moving towards the rim. b) Fine grained garnet in 481440 that is partially pseudomorphed by plagioclase and biotite. Scale bars measure 500µm.

Garnets are dominated by the almandine-pyrope compositional mixes with moderate grossular and minor spessartine components. Major element zoning is observed in poikiloblastic grains and is focussed around the inclusion-poor grain centres. Ca and Mg increase from grain centres moving into poikiloblastic mantles, coupled with a corresponding decrease in Mn and Fe (Fig. 6.15). This is interpreted to reflect sequential growth zoning of the grain centres, followed by the development
of poikiloblastic mantles. The textures and zoning patterns associated with the poikiloblastic grain suggests that the inclusion-poor grain centres may have initially been clusters of individual garnets or multiple nucleation points for garnet growth that have become connected during the subsequent growth of the poikiloblastic garnet rims. The fine grained, embayed garnets are homogeneous with respect to major elements and Y.

Figure 6.15: a and b) Irregular zoning patterns in Grt, from sample 481440. Both Ca and Fe preserve prograde growth zoning from a number of centres within any one grain.

Plagioclase in this sample is much more calcic than other samples in the area, with a bytownite composition (Xab 0.12-0.17). The mineral is unzoned with no systematic compositional variation across the sample and no difference in composition between plagioclase formed in the matrix and that formed from the breakdown of garnet. Matrix biotite is unzoned and relatively homogeneous (X_Mg 0.67-0.68) with a dominantly phlogopite-eastonite composition. Fe^{3+} represents 12-15% Fe_{tot}. Amphibole in 481440 has an anthophyllite-cummingtonite composition, computed using the AMPH-CLASS amphibole classification of Esawi (2004) but is otherwise unzoned and unvaried across the sample (X_Mg 0.59-0.6).
Figure 6.16: a) Major element traverse from core to rim in a Grt porphyroblast (see Fig 6.14a for traverse line) shows largely homogeneous zoning profiles that fluctuate slightly moving across a poikiloblastic zone near the garnet rims. b) X_grs and X_sps traverses rescaled to show small-scale zoning patterns.
Figure 6.17: Simplified geological map of Tumemark formation showing the location of late intrusions and the localities of pelite samples. Also: Map of the Kapshak region showing the location of Orthogneisses (red star) and the reconnaissance area (R).

- Orthogneisses
  - Tonalitic orthogneiss
  - Deformed granodiorite
  - Granodiorite orthogneiss
  - Late intrusions

- Supracrustal units
  - Supracrustal amphibolite
  - Metasediment

- Structure
  - Tumerial isocline
  - Dip/Dip direction of S2 gneissosity
  - L2 mineral (sillimanite) lineations

- Ultramafic (in BIF)
Garnet-alumino-silicate gneisses

Garnet-alumino-silicate-bearing lithologies represent the dominant pelitic rock type in Tummeralik. Eight samples were collected from various localities outlined in Fig. 6.17. The following samples are grouped according Y zoning in garnet.

Figure 6.18:  a) (PPL) Coarse biotite-sillimanite fabric wrapping garnet porphyroblasts in 478330. b) (PPL) Quartzfeldspathic matrix with coarse biotite and garnet porphyroblasts in 496830. Plagioclase is dotted with fine grained white mica at grain boundaries and contains clusters of fine grained sillimanite. c) (PPL) Garnet porphyroblasts in 496833 wrapped by the gneissic biotite-sillimanite fabric. Plagioclase is commonly altered to fine aggregates of white mica. Scale bars measure 500µm.

Samples 478330, 496830 and 496833 were collected from 10m-scale pelitic lenses housed by the supracrustal amphibolite unit (Chapter 3). The matrix in all samples is medium grained (0.2-0.5mm) (Fig. 6.18), with gneissic banding on mm- and cm-scales produced by alternating biotite-sillimanite foliae with biotite-plagioclase-quartz bands. Garnets form small (1-2mm), pale pink porphyroblasts, which occur throughout the samples but tend to be more common and larger in the biotite-sillimanite foliae. Garnet cores are small, with inclusions of oriented biotite, quartz and rutile, whereas rims are largely inclusion-free. An exception to this is
sample 478330, which contains needles of fibrolitic sillimanite in garnet rims. Biotite is partially replaced by chlorite (Fig. 6.19), and sillimanite and plagioclase are replaced by fine aggregates of white mica, suggesting a partial overprint at greenschist facies. Sample 496833 contains the only occurrence of coarse grained K-feldspar encountered in the Kapisillit sample suite.

![Figure 6.19: AFM projection from muscovite illustrating the spread in A values for partially chloritised biotite (blue dots). Red = Garnet, Yellow = aluminosilicate, Green = approximate range of the chlorite compositional field in the lower amphibolite facies (Bucher and Frey, 2002).](image)

Garnets in samples 478330, 496830 and 496833 are largely homogeneous with respect to major elements and Y, however small increases in Fe and Mn, with a corresponding decrease in Mg, are apparent within 20-50μm of grain edges and fractures (Fig. 6.20). These zoning patterns are especially pronounced where garnet is in contact with biotite and are interpreted as late stage diffusive zoning. Garnet compositions are almandine-rich with subordinate pyrope and minor grossular and spessartine components (Fig. 6.21). There is little variation in biotite chemistry ($X_{Mg} 0.6-0.62$), where biotite in all three samples is dominated by phlogopite-annite with $Fe^{3+}$ ranges from 12-25% $Fe_{tot}$. However biotite inclusions in garnet in 496833 have dominantly phlogopite-eastonite compositions ($X_{Mg} 0.74-0.76$) with a much smaller annite component. Plagioclase has an andesine composition in samples 478330 and 496833 ($X_{ab} 0.66-0.7$) but is slightly more sodic in 496830 ($X_{ab} 0.74-0.76$). K-feldspar in 496833 has $X_{ab} 0.06-0.09$ and $X_{or} 0.91-0.94$. 181
Figure 6.20: X-ray maps of garnet in samples 478330, 496830 and 496833. a, b) Fe and Mn in 478330, c, d) Mg and Mn in 496830 and e, f) Mg and Mn in 496833. Garnets in all samples show homogeneous major element zoning moving from inclusion-rich cores to inclusion-free rims. Retrograde zoning occurs at grain rims, especially where in contact with biotite and along fractures. Scale bars measure 1 mm.
Figure 6.21: a) Core to rim (left to right) traverse showing homogeneous major element zoning profiles across a typical garnet in 478330. Zoning apparent at the rim is due to exchange between garnet and adjacent biotite. Length of traverse is approximately 2mm. b, c) Rim to rim traverses showing major element zoning profiles in 496830 b) and 496833 c). Lengths of traverses are approximately 3mm. Inset a, b and c) Rescaled plots of X_grs (triangles) and X_sps (crosses).
Samples 478323, 478336, 478339 and 478340 were collected from a pinching and swelling linear pelitic body hosted by garnet amphibolite in northern Tummeralik. Sillimanite forms coarse, acicular grains intergrown with biotite (but modal abundances vary between samples and in 478339 the mineral is rare, only occurring as inclusions in garnet. Garnets are wrapped by the biotite-sillimanite (Fig. 6.22) fabric and form blocky, pale pink porphyroblasts, which vary in size from 1-5mm diameter.

Figure 6.22: a) (PPL) Garnet porphyroblast with inclusion-rich core is wrapped by the peak biotite-sillimanite fabric in 478323 b) (PPL) Partially resorbed garnet in 478336 is replaced by intergrown biotite-plagioclase. Sillimanite occurs intergrown in coarse matrix biotite and as fibrolitic inclusions in the garnet rim. c) (PPL) Garnet in 478339 showing embayment and partial replacement by biotite-plagioclase (yellow arrow). The sample is the most quartzofeldspathic of the aluminosilicate-bearing samples. d) (PPL) Garnet in 478340 in association with the peak Bt-Sil-Pl-Qtz assemblage. Scale bars measure 500μm.
Garnets exhibit an inclusion-rich core that is mantled by an inclusion-poor rim. Cores typically contain quartz, rutile (Fig. 6.23a and b), monazite and zircon, where rutile forms laths, which typically cluster around the core edge. Minor ilmenite occurs in garnet in 478339. Inclusion-poor mantles contain fine-grained zircon (<20μm) often with patches or annuli of fibrolitic sillimanite near grain edges. Blue-green pleochroic tourmaline occurs as an accessory mineral in the matrix and in garnet rims in sample 478336. Sample 478336 also contains symplectic intergrowths of plagioclase and quartz in stable contact with biotite and sillimanite in the matrix (Fig. 6.24).
Figure 6.24: a) (XPL) Micrographs showing relationships between matrix minerals in sample 478336. Note the segregation of biotite-sillimanite-rich foliae and intervening layers of dominantly quartzofeldspathic material. b) Plagioclase-quartz symplectites showing stable contacts with the biotite-sillimanite fabric in sample 478336 (yellow boxed area in figure a). Scale bars measure 500µm.

Figure 6.25: AFM projections from muscovite illustrating the spread in A values for partially chloritised biotite in samples a) 478323, b) 478336, c) 478339 and d) 478340. All projections include quartz, muscovite and H₂O in excess.
Biotite is commonly replaced by chlorite (Fig. 6.25) and sillimanite is embayed and pseudomorphed by fine aggregates of white mica. This is especially pronounced where sillimanite is in contact with biotite or plagioclase. Garnet rims in all samples are fractured, embayed and replaced to some extent by plagioclase and biotite or white mica. The replacement of garnet by biotite and plagioclase is most apparent in 478336 and 478340, where plagioclase and biotite replace garnet to the point where the wrapping biotite-sillimanite fabric is no longer in contact with garnet (Fig. 6.26).

Figure 6.26: Replacement of garnet in sample 478336 by coronae of plagioclase and biotite. Resorption is such that the garnet is no longer in contact with the biotite-sillimanite fabric. Scale bar measures 500µm.

In sample 478323, small (<100µm wide), anastomosing vein networks infilled by fine grained chlorite and white mica cut the gneissic fabric at a high angle. In 478339, chlorite alteration of biotite is focussed immediately adjacent to coarse quartz layers and lenses, which rapidly decreases moving into the matrix. This low grade alteration is most apparent in 478323 and 478339 and is interpreted to be intimately related to the invasion of hydrous fluids associated with quartz and muscovite veining.
Garnets in 478323, 478336, 478339 and 478340 are homogeneous with respect to the major elements and are almandine-rich with subordinate pyrope and minor grossular. An exception to low grossular contents is seen in 478339, where $X_{\text{grs}}$ decreases from core to rims ($X_{\text{grs}}$ 0.11-0.05 respectively) (Fig. 6.27). The defining characteristic of garnets in these four samples is the presence of Y-rich cores (Fig. 6.28), whose boundaries coincide with the rutile-quartz-bearing garnet cores.
Figure 6.28: X-ray maps showing Y growth zoning in garnets from samples a) 478323, b) 478336, c) 478339 and d) 478340. In all cases, the highest Y counts are coincident with inclusion-rich cores and decreases in inclusion-free rims.

Figure 6.29: X-ray maps of representative zoning in garnet in samples 478323 and 478340. a and b) Mg and Mn in 478323. c and d) Mg and Mn in 478340. Zoning in both samples is homogeneous throughout most of the grain except for retrograde diffusive zoning near grain edges and along fractures.
Weak zoning in garnet occurs within 10µm of grain edges and fractures (Fig. 6.29), manifest as increasing Mn and Fe and decreasing Mg. Whereas zoning in Y is considered to reflect initial growth zoning from core to rim, the small scale zoning along grain edges and fractures is interpreted as retrograde diffusive exchange in Fe, Mn and Mg. Biotite in all samples is unzoned (X_{Mg} 0.5-0.62) with no systematic variation in composition between matrix biotite and biotite replacing garnet. Analysis of the least altered grains suggests dominantly phlogopite-annite compositions with Fe^{3+} constituting 21-25% Fe_{tot}. Matrix plagioclase is unzoned with andesine compositions (X_{ab} 0.61-0.71) in samples 478336, 478339 and 478340. Plagioclase in sample 478323 is more sodic, with oligoclase compositions. Weak zoning in plagioclase is also preserved in this sample, which becomes more anorthite-rich moving from core to rim (X_{ab} 0.82-0.74).

**Garnet-staurolite-aluminosilicate gneisses**

Sample 478335 was collected from northern Tummeralik from a 10m thick, pelitic layer hosted by the supracrustal amphibolite unit (Chapter 3). Sample 492237 was collected from a large, linear pelitic unit collected west of and structurally higher than units in Tummeralik (Fig. 6.17, inset). Both display a gneissic fabric defined by mm to cm-scale bands and lenses of biotite-sillimanite separated by larger (5cm-scale) plagioclase-quartz ± biotite. The average matrix grain size in both samples is 0.5-1mm. Sillimanite forms coarse, acicular grains (up to 3mm) (Fig. 6.30). In both samples, garnet forms very large (5-20mm), pale pink porphyroblasts that are wrapped by the peak biotite-sillimanite fabric and are associated with the development of symmetrical pressure shadows that contain very coarse biotite, plagioclase and quartz (up to 3mm).
Figure 6.30:  
a) (PPL) Very coarse, peak biotite-sillimanite fabric adjacent to garnet porphyroblasts in sample 478335.  
b) (PPL) Inclusion-rich core in garnet in 492237 with the minerals staurolite, biotite, quartz, rutile and ilmenite. Scale bars measure 500µm.

Figure 6.31:  
Thin section scan of sample 478335, showing garnet morphologies Grt and Grt₂. Box ‘a’ refers to the garnet inclusion maps given in Figure 6.32. Thin section measures 2x3cm.

Garnet in 478335 displays complex morphologies and inclusion textures (Fig. 6.31). The first morphology (Grt₁) is subhedral with one apparent growth phase and contains sequential bands of inclusions containing 1) Rut + Ilm + St + Zrc + Mnz 2) quartz and 3) Sil + Bt + Pl + Rut + Ilm + Zrc moving from core towards the rim. The second morphology (Grt₂) has an irregular shape and is interpreted to have been pulled apart at high grade, as the spaces in between garnet sections are infilled by biotite and sillimanite. However some parts of Grt₂ are elongate and appear to exhibit
core and rim inclusion assemblages, suggesting preferential growth parallel to certain horizons prior to attenuation. The inclusion zone 1 assemblage is present in Grt₂ (Fig. 6.32). Inclusion zone 3 is also apparent in grain rims, suggesting that the two garnet morphologies grew contemporaneously.

Figure 6.32: Mg (a) and Ti (b) x-ray maps for the Grt₂ morphology in sample 478335. Maps show the distribution of rutile (small red, elongate inclusions) and staurolite throughout the irregularly-shaped porphyroblast. Retrograde Mg diffusion between garnet and biotite are also shown in ‘a’.

Garnet textures in 492237 contain cores with the assemblage St + Bt + Qtz + Rut + Ilm and may be circular or irregularly shaped. Inclusions are randomly oriented in the centre of the cores, but develop a preferential orientation approximately parallel to the matrix gneissosity at the core edge. Cores are surrounded by a mantle of inclusion-poor material, which in turn is terminated by an inclusion-bearing rim. This rim is often defined by a sharp, euhedral annulus of fibrolitic sillimanite, quartz and minor rutile rimmed with ilmenite. Garnet rims in both samples are embayed and replaced by intergrown plagioclase and biotite and in some cases this results in the re-exposure of inclusion domains to the matrix (Fig. 6.33).
Garnets in 478335 are almandine-rich with subordinate pyrope and minor grossular and spessartine. Zoning patterns are variable due to the complex garnet morphologies described above, such that Grt<sub>1</sub> exhibits distinct zoning patterns from Grt<sub>2</sub>. Grt<sub>1</sub> preserves marked zoning in Mn, which decreases towards the rim from an off-centre core. Fe, Ca and Mg zoning are more complex, with Fe levels falling marginally in and around inclusion zones 1 and 2, but increasing again near the grain rim (Fig. 6.34a). The opposite pattern is seen in Mg and Ca. Zoning patterns in Grt<sub>2</sub> are more difficult to characterise, due to irregular grain shapes that often truncate zoning patterns. However in a number of Grt<sub>2</sub> segments, Ca appears to be elevated in cores and decreases moving towards the rims (Fig. 6.34b). Fe, Mg and Mn levels are homogeneous throughout much of the grains, but Fe and Mn tend to increase at grain rims, especially where the garnet is in contact with biotite. This is accompanied by a corresponding decrease in Mg. Ca profiles in Grt<sub>1</sub> and Grt<sub>2</sub> are compatible with growth zoning, and Mg, Fe and Mn zoning profiles are considered to reflect retrograde diffusive zoning at grain boundaries and along fractures.
Figure 6.34: Major element core to rim (left to right) traverses for representative garnets in a) Grt₁, sample 478335. Traverse line shown in Fig 6.31. b) Grt₂, sample 478335. Traverse line shown in Fig 6.32b. c) Sample 492237. Traverse not shown. The analysed garnet is the same as shown in Fig 6.33 and the traverse crosses the Sill annulus in a different part of the grain. Traverses measure approximately 1mm.

Garnets in 492237 are almandine-rich with subordinate pyrope, grossular and spessartine (Fig. 6.34c) and garnet cores retain elevated Y (Fig. 6.35a). Xgrs increases marginally moving across the sillimanite-bearing rim but major element
compositions are otherwise homogeneous. Mn and Fe increase within 100μm grain rims along with a corresponding decrease in Mg (Fig. 6.35b). This zoning is discontinuous around the grain rim and most prominent where garnet is in contact with biotite and is thus interpreted to reflect late diffusive zoning.

Figure 6.35: X-ray maps of garnet in 492237 showing a) growth zoning in Y. b) Retrograde diffusive zoning in Mn, which is most pronounced adjacent to biotite.

Matrix plagioclase is unzoned with an andesine composition (X\textsubscript{ab} 0.61-0.67) and is homogeneous across the samples with no systematic variation between plagioclase in the matrix and that formed from the breakdown of garnet. Biotite in both samples is unzoned and homogeneous, and is dominated by the phlogopite-annite exchange with Fe\textsuperscript{3+} constituting 20-25% Fe\textsubscript{tot}. Biotite inclusions in garnet cores in 492237 are more Fe-rich than biotite in the matrix (X\textsubscript{Mg} 0.72 and 0.75) with Fe\textsuperscript{3+} contents of ~32% Fe\textsubscript{tot}.

**Garnet-sillimanite-cordierite gneiss**

Sample 478343 was collected from a ~5m wide linear pelitic body hosted by garnet amphibolite in western Tummeralik (Fig. 6.17). In hand specimen the sample displays a mm- to cm-scale gneissosity defined by alternating biotite-sillimanite and quartzofeldspathic horizons. The typical matrix grain size ranges from 0.5-1mm, however quartz and plagioclase-rich layers are on average coarser than biotite and sillimanite. Sillimanite forms two distinct textures in the matrix. Rare subgrained
sillimanite pseudomorphs after kyanite are encountered in garnet pressure shadows (section 6.3) (Fig. 6.36) but the dominant sillimanite texture forms coarse, acicular grains (up to 1mm length).

![Figure 6.36: Sillimanite pseudomorph after kyanite in the matrix of 478343. The central part of the grain shows lattice distortion due to subgrain development, whilst the outer grain is formed of sillimanite aggregates developed during subgrain rotation recrystallisation.](image)

Cordierite is common in the matrix and also forms two textures. The principal cordierite texture grows at the expense of acicular sillimanite, but the reaction is incomplete, leaving sillimanite grains embayed and mantled by cordierite (Fig. 6.37a). Cordierite also forms randomly oriented grains, which form mantles round embayed garnet (Fig. 6.37b). These textures suggest the involvement of both sillimanite and garnet in the formation of cordierite and show the development of cordierite to post-date the Grt-Sil-Bt-Pl-Qtz assemblage.
Figure 6.37: Cordierite textures in sample 478343. a) (XPL) Cordierite mantles formed from the breakdown of sillimanite. Scale bar measures 500µm. b) Sillimanite and garnet breaking down to form cordierite. White arrow denotes a cordierite partial pseudomorph of sillimanite.

Figure 6.38: SEM (BSE) images of rutile and ilmenite textures in 478343. a) Coarse rutile rimmed by ilmenite, b) coarse rutile grain with associated ilmenite in stable contact with garnet and c) coarse ilmenite intergrown with biotite.
The sample also contains coarse (200-500\(\mu\)m) rutile, which grows in contact with garnet and biotite and is mantled by \(~100\mu\)m thick rims of ilmenite (Fig. 6.38a). Some of these form asymmetric pressure shadows on garnet (Fig. 6.38b) and are in textural equilibrium with garnet and matrix biotite, plagioclase, sillimanite and quartz, suggesting their development during deformation at high metamorphic grade. Coarse ilmenite is also intergrown with matrix biotite (Fig. 6.38c). Garnet forms small (<1mm), pale pink porphyroblasts, which exhibit up to three textural domains that are based on inclusion assemblages (Fig. 6.39).

![Figure 6.39: X-ray maps showing rutile inclusions (small, red points) in garnet domains 1 and 2 in sample 478343.](image)

Domain 1 garnet forms inclusion-poor cores, with rare, very small rutile and quartz inclusions. Domain 2 forms a ring of coarser quartz with small, rare monazites. These are mantled by domain 3, which again is largely inclusion-free. All garnets are embayed to some degree, in some cases resulting in the exposure of inclusion domains 1 and 2 to the matrix. Cordierite breaks down to fine aggregates of intergrown plagioclase, chlorite and other clay minerals, suggesting low grade partial overprinting. The breakdown products of cordierite are intergrown with the coarse ilmenite mantles on rutile, suggesting that the development of the ilmenite rims to be contemporaneous with retrogression in cordierite (Fig. 6.40).
Figure 6.40: SEM (BSE) imaging of cordierite resorption textures in sample 478343. a) Cordierite that initially grew at the expense of sillimanite in the matrix is broken down to fine grained chlorite where in contact with biotite and sillimanite. b) Cordierite breaking down to fine grained intergrowths of plagioclase chlorite and ilmenite rims on coarse rutile (see Figure 6.38 for larger image).

Garnet in 478343 is generally homogeneous with respect to major element and Y, but exhibits weak zoning within 50μm of grain boundaries and fractures, where Fe and Mn increase and Mg decreases (Fig. 6.41). This is considered to represent late stage diffusive zoning and is most marked where garnet is adjacent to biotite or cordierite. Matrix plagioclase exhibits a range of compositions from oligoclase to andesine (Xab 0.68-0.78), however individual grains are unzoned and there is no systematic variation in composition according to texture or a grain’s location in the section. Matrix biotite is unzoned with phlogopite-eastonite compositions and is relatively Fe-rich (XMg 0.7-0.72). Biotites are unzoned with Fe\textsuperscript{3+} representing 21-26% Fe\textsubscript{tot}. Cordierite is unzoned and relatively homogeneous across the entire sample (XMg 0.78-0.81).

Figure 6.41: X-ray Mg (a) and Mn (b) maps of representative garnet in sample 478343. Samples are homogeneous throughout much of the grain with retrograde zoning at grain edges and along fractures. Inclusion trails of textural domain 2 are also visible on both maps.
Garnet-chlorite-muscovite schist

Sample 496826 was collected from a pelitic lens hosted by garnet amphibolite. In hand specimen the sample displays a biotite-defined S>L schistosity with an average matrix grainsize of 0.2-0.5mm. The peak upper amphibolite facies foliation is defined by former biotite and sillimanite; however the sample is characterised by the extensive breakdown of biotite to chlorite and of sillimanite to white mica. This is most pronounced where the two minerals are in contact, and the peak biotite-sillimanite fabric has been almost pervasively overprinted by chlorite and white mica (Fig. 6.42). A second biotite texture is associated with the breakdown of garnet, where it forms coarse plates that embay garnet rims.

Figure 6.42: (PPL) Matrix-porphyroblast relationship in sample 496826. Garnet is intensely fractured, with fractures infilled by fine grained muscovite. The peak biotite-sillimanite fabric is replaced by chlorite and muscovite, which pseudomorph biotite and sillimanite respectively. Scale bar measures 500μm.
Key: Red = Garnet; Purple = Staurolite; Green = Chlorite; Blue = Biotite (showing varying states of chloritisation).

Figure 6.43: AFM projection from muscovite showing variations in biotite chemistry in sample 496826. Biotite compositions track towards chlorite with increasing alteration and reflect a loss of K in biotite.

This second biotite texture is also chloritised to some degree, although not as intensely as typical matrix biotite (Fig. 6.43). The sample contains very coarse (up to 15mm), pink garnet porphyroblasts which are wrapped by the biotite-sillimanite fabric and contain three textural domains. The first domain is characterised by roughly circular, inclusion-rich cores, containing randomly oriented staurolite, quartz and rutile (Fig. 6.44a). Staurolite also contains rare inclusions of rutile (Fig. 6.44b). The second domain is denoted by inclusion-free rims, which make up the majority of garnet in the sample and contain a diffuse annulus of sillimanite (Fig. 6.44c). The annulus measures ~300μm in thickness with the sillimanite inclusions largely oriented parallel to the garnet edge. A third, inclusion-free domain continues rimwards of the sillimanite annulus, which marks the boundary between domains 2 and 3. Garnet is replaced by intergrown biotite and plagioclase (Fig. 6.44d), which has removed large swathes of domain 3, exposing domain 2 to the matrix in some places.
Figure 6.44: a and b) X-ray maps of a large staurolite-bearing garnet in 496826. a) Mg map of the boundary area between the inclusion-rich core and inclusion-free rim shows a flat zoning profile in garnet. Staurolite inclusions are shown in blue and the black inclusions are dominantly quartz. b) Ti map of the same area as 'a' with rutile picked out in red and staurolite in blue. c) SEM (BSE) image of a garnet rim containing an annulus of fibrolitic sillimanite oriented approximately parallel to the perimeter of the garnet and d) garnet in 496826 with embayed rims replaced by intergrown biotite and plagioclase. Biotite replacing garnet has also been partially chloritised.

Figure 6.45: Core to rim (left to right) major element zoning in garnet in sample 496826. Length of traverse 4mm. Inset: Rescaled traverse to show detailed variation in X_{grs} and X_{spe}.
Garnet in 496826 is largely homogenous with respect to major elements and Y, with no compositional variation between textural domains 1 and 2 (Fig. 6.45). Garnet is dominated by almandine-pyrope with minor grossular and spessartine. A gradual increase in Ca is observed in domain 2, however the most prominent zoning occurs within 50μm of grain edges or fractures and is marked by increasing Fe and Mn and decreasing Mg. Plagioclase has an oligoclase composition (X_{ab} 0.74-0.75) and is unzoned with no systematic variation in composition across the sample. Biotite (where preserved) has X_{Mg} 0.61-0.63(AFM) and Fe^{3+} varies from 19-24% Fe_{tot}. Biotite formed from the breakdown of garnet is more compositionally variable (X_{Mg} 0.57-0.62), with a lower Fe^{3+} (9-14% Fe_{tot}).

6.6 Pelitic material of the Ice Lake reconnaissance area

Pelitic material in the Ice Lake reconnaissance area (Fig. 6.17) has a similar mineralogy to the pelites of Tummeralik despite showing fundamental differences in outcrop pattern (Chapter 3). Pelitic units form large, discontinuous belts housed within heterogeneous garnet amphibolite, which are characterised both at outcrop scale and in thin section by pervasive alteration of high grade mineral assemblages to white mica and chlorite. Sample 588528 contains garnet, white mica, biotite, plagioclase and quartz. The sample is invaded by white-mica veins and the mineral completely pseudomorphs former sillimanite and largely replaces garnet and plagioclase (Fig. 6.46). Biotite is visibly chloritised. No geochemical analyses were made of sample 588528 due to the high level of alteration.

Figure 6.46: Development of retrograde Ms in sample 588528. a) (XPL) Ms pseudomorphing Sil with extensively chloritised Bt. b) (XPL) Grt is highly fractured and replaced by Ms.
Sample 478325 was collected from a ~15m wide pelitic lens, whose mottled rust and purple weathering colour is attributed to intense sulphide surface weathering. The sample is a fine to medium grained schist with an average matrix grain size of 0.1-0.5mm. Sillimanite forms acicular grains up to 1mm long, which define the gneissic fabric with coarse matrix biotite (Fig. 6.47a). Rare, subgrained patches of former kyanite constitute a minor second sillimanite texture (Fig. 6.47; section 6.3). Both aluminosilicate textures are replaced along fractures and at their rims by fine aggregates of white mica (Fig. 6.47b). One characteristic of sample 478325 is that it does not contain ilmenite as an accessory Fe-Ti oxide. Instead, rutile is often associated with matrix biotite and may form grains up to 100μm length. In addition, pyrite forms coarse overgrowths on garnet and in the matrix where it is intergrown with white mica (Fig. 6.47c). Chlorite- and white mica-bearing veins cut the biotite-sillimanite schistosity at a high angle and are associated with the alteration of biotite and sillimanite, indicating invasion after the development of the gneissic fabric (Fig. 6.48).
Garnet forms small, pale pink porphyroblasts that are typically finer grained than the former kyanite in the matrix (<1mm diameter). The biotite-sillimanite fabric wraps garnet and matrix biotite forms distinct, equant pressure shadows. In places, the porphyroblasts are pulled apart, with the intervening space filled by biotite and sillimanite. Most garnets preserve a single textural domain, although some small cores with plagioclase or chlorite inclusions are present. A number of garnets contain patches of oriented sillimanite and biotite towards the grain rims (Fig 6.49).
Garnets are almandine-rich with low grossular and spessartine contents and are unzoned with respect to major elements and Y, with the exception of the extreme rims and areas adjacent to fractures, which show an increasing Mn and Fe and decreasing in Mg within 100μm of grain edges and fractures (Fig. 6.50). Plagioclase is unzoned, with an oligoclase composition (X_{ab} 0.71-0.75) and is homogeneous throughout the section, where it is extensively altered to fine aggregates of white mica. Biotite is dominantly phlogopite (X_{Mg} 0.64) with subordinate annite and eastonite components. Fe^{3+} accounts for 21-26% Fe_{tot}.

Figure 6.50: a and b) Mg and major element zoning in 478325 showing flat growth zoning patterns but a significant element of late Fe-Mg exchange.
6.7 Pelitic and semi-pelitic units of Aputitooq Mountain

Garnet-biotite-orthoamphibole gneiss

Figure 6.51: Simplified geological map of Aputitooq Mountain (Bennett and Heiss, personal communication, 2006) showing key lithologies and sample localities. Scale bar measures 1km. Red star denotes the sample locality of sample 492710.

Sample 492206 was collected from the heterogeneous pelitic unit that structurally underlies the heterogeneous TTG orthogneiss (Fig. 6.51) (Chapter 2). The sample is sillimanite-absent, with the dominant S>L schistosity defined instead by intergrown biotite-amphibole and has a bimodal composition, where garnet and amphibole are mutually exclusive. Pale green to colourless pleochroic amphibole with a sodic anthophyllite-cummingtonite composition (using the classification spreadsheet AMPH-CLASS of Esawi, (2004) and amphibole classification criteria of Leake et al., (1997)) forms coarse (up to 1mm) grains intergrown with green-brown pleochroic biotite, (Fig. 6.52a). The biotite-orthoamphibole assemblage is in textural equilibrium with matrix plagioclase and quartz but has been altered to chlorite and fine aggregates of white mica respectively. Biotite is stable where it is not in contact
with orthoamphibole. Small amounts of ilmenite are present in the matrix and are associated with the breakdown of biotite. Garnet is present as anhedral, pale pink porphyroblasts, forming the stable assemblage garnet-biotite-plagioclase-quartz (Fig. 6.52b). Grains are inclusion-free and typically pulled apart, with the space filled by biotite, plagioclase and quartz.

Figure 6.52: Matrix ± porphyroblast textures in the two compositional domains in sample 492206. a) (PPL) intergrown Bt-Amph-Pl-Qtz fabric with some chloritisation of Biotite. b) (PPL) Garnet-bearing domain in 492206, with the assemblage Grt-Bt-Pl-Qtz. Biotite is stable in this assemblage but reacts with anthophyllite to form Chl. c) Major element profile of Grt from 492206. Scale bars measure 500µm. Traverse measures ~1mm.

Garnets are homogeneous and dominated by almandine-pyrope with subordinate spessartine and grossular. Minor fluctuations in Fe, Mg and Ca occur moving towards rim regions of garnets (Fig. 6.52c). Matrix plagioclase has an andesine composition ($X_{ab}$ 0.57-0.63) and is unzoned and homogeneous throughout. Biotite is the principal schistosity-forming mineral and is dominated by phlogopite ($X_{Mg}$ 0.64-0.65) with $Fe^{3+}$ accounting for 23-26% $Fe_{tot}$.
Sample 492234 was collected from within a semi-pelitic horizon entrained in the biotite-hornblende TTG gneiss that forms the bottom of the structural pile in Aputitooq Mountain (Chapter 2). The unit exhibits a mm-cm scale gneissosity defined by relative abundances in quartzofeldspathic minerals and biotite. Garnet forms pale pink porphyroblasts up to 3mm diameter (Fig. 6.53). These are generally inclusion-poor, but locally contain patches of fibrolitic sillimanite and acicular ilmenite near their rims. Garnets are almandine-rich and homogeneous throughout much of the grain, although increasing Mn and Fe is observed within 500μm of the grain edge (Fig. 6.54) and is interpreted as retrograde zoning. More marked zoning in Fe and Mg is present within 50μm of grain edges and fractures, especially where rims are in contact with biotite. This second zoning pattern is interpreted as late stage diffusive zoning. Matrix plagioclase has an andesine composition and is unzoned ($X_{ab} \approx 0.67-0.69$) as is matrix biotite, which has a dominantly phlogopitic composition ($X_{Mg} \approx 0.57-0.59$) with subordinate annite-eastonite and $Fe^{3+}$ constituting 14-18% $Fe_{tot}$. Biotite that forms rims around garnet tends to have a higher phlogopite component ($X_{Mg} \approx 0.6-0.62$) and a corresponding decrease in the annite component, with $Fe^{3+}$ constituting 15-21% $Fe_{tot}$. 

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**Garnet-biotite schist**

![Figure 6.53: Stable peak assemblage Grt-B-Pl-Qtz in sample 492234. Scale bar measures 500μm.](image)
Garnet-sillimanite gneiss

Sample 492209 was also collected from the heterogeneous pelitic gneiss package that underlies the heterogeneous TTG suite. The lithology is schistose and homogeneous on m-scales, but heterogeneity emerges on mm- to cm-scales, where coarse (up to 5mm) plagioclase-quartz leucosome ponds in pressure shadows around large (10-15mm diameter) garnets parallel to the peak schistose fabric. Coarse (0.5-1mm), acicular sillimanite defines the fabric, although fine grained needles (~0.1-0.3mm) of sillimanite also occur as inclusions in plagioclase, where it is often mantled by aggregates of white mica (Fig. 6.55). Ilmenite is the dominant Fe-Ti oxide in the sample, where it forms coarse (up to 1mm) grains that are often associated with garnet but also overgrow minerals in the matrix. Pale pink garnets form anhedral porphyroblasts that are wrapped by the biotite-sillimanite fabric or
enveloped by coarse plagioclase-quartz leucosome. Garnets are highly fractured and extensively altered to chlorite, white mica and coarse ilmenite (Fig. 6.55) to the extent that inclusion assemblages could not be identified. A single grain of staurolite is preserved in the core of one garnet in addition to a single grain of tourmaline.

Figure 6.55: (PPL) Matrix-porphyroblast relationships in 492209. Intense greenschist alteration has resulted in the retrogression of the peak Grt-Sil-Bt-Pl-Qtz assemblage to chlorite-muscovite. Garnet is particularly affected by the greenschist overprint. Scale bar measures 500μm.

Grains are dominated by almandine-pyrope with minor grossular and spessartine. Compositions are variable but no zoning is preserved due to intense retrogression of garnet in the sample. Plagioclase has a relatively uniform oligoclase composition ($X_{ab} 0.75-0.78$) and is unzoned. Plagioclase in felsic leucosome has the same composition as plagioclase in the matrix and that replacing garnet. Biotite is invariably replaced by chlorite, however the least altered grains suggest dominantly phlogopite-eastonite compositions ($X_{Mg} 0.63-0.68$) with Fe$^{3+}$ representing 18-21% Fe$_{tot}$. 
Garnet-staurolite-sillimanite gneiss

Samples 492202 and 492219 were collected from pelitic horizons in the heterogeneous amphibolite and heterogeneous pelite units respectively. The matrix in both samples is medium to coarse grained (0.5-1 mm) and interspersed with garnet porphyroblasts up to 5 mm diameter. Both samples are gneissic, with prominent SL tectonites defined by alternating quartz-plagioclase and biotite-sillimanite foliae. Aluminosilicate in 492202 is largely confined to gneissosity-parallel knots and foliae, which may be formed almost entirely of acicular sillimanite (Fig. 6.56a).

![Image](image_url)

**Figure 6.56**: a) (PPL) Partially resorbed garnet porphyroblast on the fringe of a concentrated sillimanite layer in 492202 and b) (PPL) Staurolite-bearing garnet porphyroblasts wrapped by the coarse, intergrown biotite-sillimanite fabric in 492219. Scale bars measure 500 μm.

The peak sillimanite-bearing assemblage is preserved virtually unretrogressed in both samples, however a second generation of biotite forms at the expense of garnet. Pale pink, subhedral garnets almost always form marginal to- or within biotite-sillimanite bands in 492202 and are wrapped by the peak biotite-sillimanite fabric in both samples. Garnets typically display large, inclusion-bearing cores, which are mantled by smaller, inclusion-free rims. The inclusion assemblage in the cores is typically quartz, with lesser amounts of staurolite, biotite and ilmenite (Fig. 6.56b, Fig. 6.65a-d). Rutile, zircon and monazite are also present in small abundances. Inclusion assemblages in both samples are preferentially oriented, but orientations vary between grains, suggesting porphyroblast rotation subsequent to core formation. Rim zones in 492202 garnets contain trails of sillimanite oriented parallel to the garnet grain boundary. Core domains and inclusion trails in a number of grains have been exposed to the matrix and are accompanied by embayment at grain edges, suggesting
garnet resorption. Where embayed, the grains are replaced by intergrown biotite and sillimanite.

Figure 6.57: X-ray maps of garnet from samples 492219 (a-c) and 492202 (d). a) Homogeneous Fe profile with weak retrograde zoning where garnet is adjacent to biotite. b) Ti map showing rutile and ilmenite inclusions in garnet but only ilmenite inclusions in the matrix (compare with Fe map). Blue staurolite inclusions are visible in garnet cores in both maps. c) Mg map shows a slight increase from core to rim, with the greatest increase occurring at grain boundaries or along fractures. d) Homogeneous Mn profile in 492202 except for very weak retrograde rim zoning where garnet is in contact with biotite. e, f) Core to rim (left to right) major element traverse of garnet in 492219 (e) and 492202 (f). Inset e,f: Rescaled garnet traverse to show detailed zoning patterns in $X_{grs}$ and $X_{sps}$. Traverses measure ~1mm.
Garnets are almandine-rich with subordinate grossular and spessartine and have homogeneous Ca profiles but Mn increases within 100μm of garnet edges (Fig. 6.57e and f). Increasing Fe and a corresponding decrease in Mg is observed on a smaller scale, within ~20μm of grain rims. These zoning patterns are considered to reflect retrograde zoning in Mn and late stage diffusive zoning in Fe and Mg. Plagioclase is unzoned and has andesine to oligoclase compositions (Xab 0.68-0.76) with plagioclase in 492219 marginally more sodic than 492202 (Xab 0.75-0.77). Biotite in these samples is largely unaltered and unzoned. Compositions are dominantly phlogopite-annite with variable eastonite (X_Mg 0.57-0.6) and Fe^{3+} accounting for 21-26% Fe_{tot} in 492202 and 14-21% in 492219. Biotite embaying garnet has a marginally higher phlogopite component than matrix biotite (X_Mg 0.6-0.63) with Fe^{3+} constituting 18-27% Fe_{tot}.

**Garnet-sillimanite-cordierite gneiss**

Sample 492710 was collected southeast of Aputitooq Mountain (Fig. 6.51 inset). The sample is unique in the Kapisillit sample set in that it is plagioclase-absent but very silica-rich (>70wt% SiO₂) (Appendix C; Table 4). Acicular sillimanite is intergrown with biotite and in textural equilibrium with quartz, forming a medium to coarse (0.5-1mm) schistosity that represents the peak metamorphic fabric. Mg-cordierite is a common matrix mineral, which is in stable contact with biotite and contains appreciable inclusions of zircon. Cordierite also forms coronitic textures around garnet and sillimanite, which is especially marked where the two minerals are in contact (Fig. 6.58a). Symplectic intergrowths of cordierite-quartz are often associated with coronae surrounding resorbed garnet (Fig. 6.58b). Ilmenite is present in the matrix and is stable with biotite, where it grows parallel to cleavage planes. Rutile forms rare grains in association with cordierite and biotite. The central part of the sample contains a cm-wide vein that is parallel to the peak biotite-sillimanite foliation. The vein mostly comprises coarse (>1mm), annealed quartz, but also contains garnet and sillimanite, with the associated cordierite textures described above (Fig. 6.59). The majority of garnet in the sample is contained within and adjacent to this vein, although a single garnet is present in the matrix that exhibits the same textures as those within the vein.
Figure 6.58: a) (XPL) Corona of cordierite mantling and replacing sillimanite in the Grt-Sil-Crd-bearing quartz vein in sample 492710. Cordierite appears to overprint garnet, but does not exhibit the same texture as the garnet in ‘b’. b) (XPL) Garnet mantled by a corona of symplectic cordierite-quartz. Scale bars measure 500μm.

Figure 6.59: Large X-ray map showing textural domains in 492710. Central to the section is a Grt-Sil-Crd-bearing quartz vein (dotted white lines denote vein boundaries), which intrudes parallel to the peak biotite-sillimanite fabric of the host pelitic rock.
The close proximity of the garnet-cordierite textures to the quartz-rich vein suggests that the development of these textures may be associated with vein fluid-host rock interactions during the invasion of the quartz vein. Garnet forms anhedral, pale pink porphyroblasts up to 5mm in diameter, which exhibit a single growth phase and exhibit irregular morphologies that are interpreted as pull apart textures (Fig. 6.60).

Figure 6.60: X-ray map of garnet in sample 492710. The garnet is partially resorbed and replaced by symplectic intergrowths of cordierite-quartz. Retrograde zoning occurs where garnet is in contact with cordierite or biotite. The grain also exhibits pull-apart textures denoted by the dashed white lines, which postdate the development of cordierite.

Discontinuous bands of sillimanite and quartz are present throughout small garnet grains but are restricted to the rim regions of large garnets. Garnet compositions are dominantly almandine-pyrope, with small grossular and very low spessartine contents. An X-ray map of a coarse garnet with sillimanite-bearing rims shows a marginal increase in Ca moving from core to rim, suggesting relict growth zoning (Fig. 6.61a). Appreciable zoning is encountered in Fe and Mg (Fe increases and Mg decreases) within 100μm of grain edges and fractures and is most prominent where garnet is in contact with cordierite and/or biotite (Fig. 6.61b). This zoning profile is interpreted to reflect late stage Fe-Mg exchange between garnet, cordierite and biotite. Biotite in the sample is very Mg-rich and dominantly phlogopite-eastonite.

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(X_{Mg} 0.69-0.71), with the highest X_{Mg} occurring in biotite adjacent to garnet and cordierite. Biotite is stable with cordierite, which exhibits zoning in Mg-Fe adjacent to garnet (X_{Mg} 0.78-0.84) attributed to late diffusive exchange. There is no compositional difference between the two cordierite textures described above, although compositions do vary along traverses where cordierite is in contact with garnet (X_{Mg} 0.78-0.86). This is also interpreted to reflect Fe-Mg exchange between garnet and cordierite.

Figure 6.61: X-ray maps of garnet from the Grt-Sil-Crd-bearing quartz vein in sample 492710. a) Ca map suggests a slight increase in X_{grs}, moving from core to sillimanite-bearing rim. b) Mg map displays retrograde zoning in Mg at grain boundaries or along fractures. Zoning is most prominent where garnet is in contact with cordierite or biotite.
6.8 Summary of petrological observations

The detailed petrological study has identified a number of common features in pelitic and semi-pelitic lithologies in the Kapisillit region. The key features are summarised below:

1. The early assemblage with Grt-Bt-Pl-Qtz ± Ky ± St ± Rut ± Ilm is preserved in garnet cores and as rare relict matrix kyanite. Kyanite, where present, is blocky and has been replaced by subgrained sillimanite.

2. Garnet inclusion patterns are similar in virtually all samples. Staurolite forms inclusions in garnet cores and is often associated with rutile. Cordierite-bearing samples exhibit distinctive 3-domain inclusion patterns in garnet (inclusion-free core – Chl-Bt-Rut-Qtz-Mnz rings – inclusion-free rim), where a significant phase of resorption separates domains 2 and 3. All garnets in sillimanite-bearing lithologies contain clusters or annuli of fibrolitic sillimanite in their rims.

3. Growth zoning in garnet is largely absent from most samples. However retrograde zoning in Fe, Mg and Mn is very common at grain rims and often encroaches into sillimanite-bearing annuli. Garnet in most samples contains $X_{gfr}$ between 0.03 and 0.05.

4. The dominant gneissic fabric in all aluminosilicate-bearing samples is defined by coarse, intergrown biotite-sillimanite, which wraps garnet in all cases. The peak mineral assemblage is Grt-Sil-Bt-Pl-Qtz ± Ilm ± Rut ± Zrc ± Mnz.

5. Development of cordierite at the expense of garnet and sillimanite and is not associated with the development of a new fabric.

6. Biotite in all samples has undergone some degree of chloritisation, which is most pronounced in samples within or adjacent to $D_{4N}$ or $D_{3T}$ fault zones.

7. Muscovite is only developed as a retrograde phase and is associated with $D_{4N}$ or $D_{3T}$. 
6.9 **The effect of bulk rock composition on muscovite and K-feldspar in the pelitic units**

One notable characteristic of all pelitic and semi-pelitic lithologies sampled in Norsanna, Tummeralik, the Ice Lake reconnaissance area and Aputittooq Mountain is that none contain appreciable amounts of prograde to peak muscovite or K-feldspar. Where present, muscovite/white mica forms as a breakdown product of aluminosilicate or plagioclase, or as infill material in post-peak veins, that is commonly intergrown with chlorite. X-ray mapping over a \(~4\text{cm}^2\) area per thin section, combined with EDX mineral identification showed no evidence of K-feldspar above the resolution of the maps (100μm) in 20 of 24 samples. A further 3 samples contained rare (1-3 grains per section) K-feldspar grains \(~100\mu m\) wide. A fourth (496833) contained 4-6 grains of coarse K-feldspar (100-500μm) in a coarse grained felsic leucosome layer in textural equilibrium with sodic plagioclase and quartz (Fig. 6.62). Thompson AFM projections are consequently unsuitable for representing the muscovite- and K-feldspar-absent assemblage which typifies the Kapisillit pelites. However the Thompson projection from muscovite can be employed to describe muscovite-producing retrograde assemblages, where such reactions can be texturally and chemically constrained.

**Figure 6.62:** The single occurrence of coarse grained Kfs in a felsic leucosome in sample 496833.
In the upper amphibolite facies, K-feldspar is principally formed during muscovite dehydration melting reactions 1 and 2, or by muscovite dehydration reaction 3, which operates at >650°C given high $a$H$_2$O. An absence or paucity of muscovite, which in typical low-Al pelitic compositions first appears in the lower greenschist facies, thus limits the development of K-feldspar at higher metamorphic grades. Muscovite (and K-feldspar at higher grades) are K-bearing phases often present in excess in pelitic bulk compositions. Thus the development of these two minerals may be inhibited by a low-K bulk composition. AKF diagrams of the Kapisillit samples suggest a low-K bulk composition when compared to average high-Al and low-Al pelites (Spear 1993), with variable Al and Fe content (Appendix C, Table 4; Fig. 6.63). It is emphasised that the K-values presented here have not been corrected for retrograde muscovite. As such, the effective composition of the Kapisillit pelites prior to the muscovite veining event(s) may be even more depleted in K$_2$O than the bulk compositions given in Appendix C, Table 4.

<table>
<thead>
<tr>
<th>No.</th>
<th>Reaction</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Ms + Qtz = Kfs + Al$_2$ + L</td>
</tr>
<tr>
<td>2</td>
<td>Ms + Ab + Qtz = Kfs + Al$_2$ + L</td>
</tr>
<tr>
<td>3</td>
<td>Ms + Qtz = Kfs + Al$_2$ + V</td>
</tr>
</tbody>
</table>

Figure 6.63: AKF and AFM (from muscovite) bulk compositional plots of the Kapisillit sample suite compared to average values for high-Al and low-Al pelites (adapted from Spear, 1993).
However comparisons with published material (Berman et al., 2007; Garcia-Casco et al., 2001; Hollis et al., 2006; Hollis et al., 2004; Tenczer et al., 2006; White et al., 2004; Zeh, 2001) show the Kapisillit pelites to have a normal K2O content but are relatively enriched in MgO and Fe2O3. This enrichment is particularly noted in MgO and is also suggested by AFM projections from muscovite, where the majority of samples plot to the high-Mg side of the high-Al pelite field (Spear, 1993) (Fig. 6.63). These observations suggest that the paucity of muscovite and/or K-feldspar may be a combined result of low-K and increased Mg and Fe in the bulk composition.

The muscovite and K-feldspar-deficient compositions observed in the Kapisillit pelites may be explained by two contrasting scenarios. The first relates to an initial bulk composition that is low in K2O relative to FeO and MgO. This would result in the preferential formation of biotite relative to muscovite, due to a reduced K/FeM ratio and may be produced by the input of a mafic source into the sedimentary protolith. The effect of a mafic source on the bulk composition of a pelitic or semi-pelitic unit is compatible with the increased ferromagnesian components encountered in the Kapisillit pelites. An increase in Mg, in particular, seems to exert a significant control on the mineral assemblage. This is exemplified by the presence of anthophyllite-cummingtonite in two samples, a mineral that is commonly associated with metavolcanic assemblages (Spear, 1993) and which is only developed in the two most Mg-rich samples. A metavolcanic or sedimentary lithology with a mafic source is a plausible depositional environment for the protoliths to the Kapisillit pelites, given their close spatial association with mafic rocks of an interpreted supracrustal origin. This is especially applicable where supracrustal units have been correlated with episodes of calc-alkaline magmatism in the area. The second scenario involves the evolution of high-K pelitic compositions, and their subsequent depletion in K2O. In this scenario, equations 1, 2 and/or 3 produce K-feldspar from the breakdown of muscovite. The subsequent removal of the melt or vapour phase – which includes appreciable amounts of dissolved H2O and K leaves the restitic/dehydrated bulk composition depleted in these components.

The formation of K-feldspar by the muscovite melting reactions would produce the coarse, annealed grains, such as those observed in the leucosome pod in
In most cases, the coarse K-feldspar in the leucosome would then be removed from the system. This leaves only fine grained (<100μm) K-feldspar that had been anchored by surrounding matrix minerals, such as biotite. Regardless of the type of reaction (dehydration or melting), K must be removed from the system either as a liquid or vapour phase to produce the observed bulk compositions.

The K-depletion scenario may apply locally, where reaction 3 is preferred for the majority of samples, as there is little evidence of migmatisation at larger scales that would indicate large-scale partial melting of the host rock. The apparent non-development of prograde or peak muscovite is attributed principally to low-K and high Fe-Mg bulk compositions, where the development of <100μm K-feldspar results from local variations in bulk composition, which permit the limited operation of reaction 3. These in turn, are attributed to a sedimentary protolith with a mafic source.

6.10 P-T paths and thermobarometry

The role of thermobarometry and pseudosection modelling in defining P-T paths

Geothermobarometry provides a quantitative method with which to constrain a particular mineral assemblage or texture to a point in P-T space. Numerous techniques use mineral chemistry from equilibrated mineral pairs to ascertain the pressure or temperature that equilibrium was attained. This is based on the knowledge that the chemical composition of phases in equilibrium will change with changing P-T conditions, where the recorded pressure or temperature relates to the point at which the system becomes closed. Care must therefore be taken to establish the extent of alteration in any sample, as subsequent metamorphic episodes or a slow cooling and/or decompression path may (re-)open key chemical systems. This may lead to partial or total resetting, which would ultimately result in anomalous pressure and/or temperature estimates.

The following section describes the geothermobarometry techniques applied to the pelitic and semi-pelitic samples in the Kapisillit sample suite. Two types of reaction are used in geothermobarometry; exchange and net transfer reactions. The former involves the equilibrium diffusion of atoms between two mineral phases.
whereas latter involves product-reactant equilibria. These reactions may be continuous or discontinuous but are always multivariant in P-T space. One or more of the involved phases must also exhibit variable composition in the form of solid solution (Bucher and Frey, 2002). The dominant mineral assemblage throughout the Kapisillit sample suite is the upper amphibolite facies assemblage Grt-Pl-Bt-Qtz ± Als ± Fe-Ti oxide. However the absence of muscovite or K-feldspar from the samples reduces the total number of thermobarometry systems that could be applied. Thus the most suitable systems for thermobarometry in the Kapsillit samples are the garnet-biotite thermometer and the GASP and GRAIL barometers. The reasons for their selection are outlined below.

The garnet-biotite geothermometer is the most widely used system for estimating temperatures in medium grade pelitic rocks (Holdaway et al., 1997). Versions of the garnet-biotite thermometer have been reported since the first temperature estimates based on Thompson (1976). The calibration of Holdaway (2000) is used here. The thermometer calculates equilibrium temperature through Mg-Fe exchange between co-existing garnet and biotite. As both minerals exhibit substitutions involving multiple end-members, the thermometry calculation must also take into account the effects of intra-phase substitutions in addition to inter-phase Mg-Fe exchange (Bucher and Frey, 2002; Holdaway, 2000; Holdaway et al., 1997). Errors in the garnet-biotite systems are notoriously difficult to estimate by error propagation, because errors for a number of garnet models and Margules parameters (thermodynamic datasets for each significant component in garnet) are not available and errors for Margules parameters in biotite tend to be intimately linked to each other and to the garnet models. Errors on temperature estimates are therefore estimated using the reproducibility of the experimental dataset, which amounts to a 2σ absolute error of ±25°C (Holdaway, 2000). The geothermometer has been largely calibrated for high temperature-moderate pressure systems (greenschist and amphibolite facies), and so provides the most accurate data for samples within this range. The system tends not to be as useful in the upper amphibolite to granulite facies, because rocks of this grade are commonly affected by low grade retrogression, resulting in anomalously low temperatures in seemingly equilibrated garnet-biotite pairs (Bucher and Frey, 1992). This is a factor that must be carefully
considered when interpreting any data from the Kapisillit pelites, given the high grade of metamorphism (upper amphibolite facies) in all samples.

The garnet-aluminosilicate-quartz-plagioclase (GASP) geobarometer is represented by the reaction:

$$Ca_2Al_2Si_3O_{12} + 2Al_2SiO_3 + SiO_2 = 3CaAl_2Si_2O$$

Grossular (Grt) + Al-silicate + Quartz = Plagioclase

The reaction involves two phases with solid-solutions and a third with a polymorphic reaction. This last attribute requires three versions of the barometer calculation (one calibrated for each aluminosilicate polymorph). The widespread occurrence of the garnet-aluminosilicate-plagioclase assemblage makes this the most widely used geobarometer for pelitic rocks in the amphibolite and granulite facies (Bucher and Frey, 2002). In order to obtain meaningful results, GASP pressure calculations require analyses from garnet and plagioclase in equilibrium. This may therefore present a problem if any phases have undergone retrogression or if zoning patterns (and therefore, composition) in garnet or plagioclase have been homogenised or altered in any way. It is consequently vital to consider the extent of alteration in a sample prior to interpreting pressure estimates. Errors on the GASP barometer are more easily quantifiable than the garnet-biotite thermometer, with the main component being geological error. However, additional errors are sourced from the garnet and plagioclase activity models, end-member calculations and temperature input errors (for absolute error, this is typically ±25°C). Older calibrations yielded high errors on this barometer of ±2.5kbar (Bucher and Frey, 2002); however, more recent calibrations by Holdaway (2001; 2004) give an absolute error of ±0.8kbar. This study uses the GASP calibration of Holdaway (2001).

The garnet-rutile-aluminosilicate-ilmenite-quartz (GRAIL) geobarometer is represented by the reaction:

$$Fe_3Al_2Si_3O_{12} + 3TiO_2 = 3FeTiO_3 + Al_2SiO_3 + 2SiO_2$$

Almandine (Grt) + Rutile = Ilmenite + Al-silicate + Quartz

In this system, ilmenite-haematite solid solution poses a particular problem when using electron probe analyses, due to the Fe$^{3+}$ component in haematite. Electron
probe analysis cannot detect ionic valencies and so the ferric iron component in haematite (as for garnet and biotite for other barometers) is calculated empirically on a site-filling basis (based on Droop, 1987). The activity models for garnet and ilmenite contribute to the error in this geobarometer; however error sourced from the input temperature is not considered to significantly affect the accuracy. The greatest source of error in the system is geological error, particularly in the identification of equilibrium assemblages, as Fe-Ti oxides may develop at varying points of the P-T curve. However a thorough textural study prior to analysis should largely eliminate this problem and Bohlen et al (1983b) suggest that inclusion assemblages in garnet should be targeted for best results. The GRAIL geobarometer is best suited to medium or high grade Barrovian metamorphic rocks and typically carries an absolute error of ±0.5kbar when a temperature input error of ±50°C is applied (Bohlen et al., 1983b).

The three geothermobarometric techniques described above all require a petrographic study prior to analysis, in order to establish the textural relationships between key phases in a sample. This is vital because the use of phases that are not in equilibrium will yield geologically meaningless results. The following textural features should be examined prior to geochemical analysis, in order to target those parts of a sample most likely to yield accurate and relevant data.

As discussed above, inclusion assemblages are targeted where possible, as minerals are more likely to be shielded from later re-equilibration if they are totally enclosed by a porphyroblast phases. In the case of GASP, GRAIL and Grt-Bt, the relevant porphyroblast is garnet. The presence of an inclusion assemblage also spatially constrains the portion of garnet that grew in equilibrium with the remaining assemblage.

Petrological evidence is also required to ascertain the part of the P-T path that is represented by each reaction. For example, ilmenite, rutile and quartz inclusions in garnet cores may represent the GRAIL assemblage during the early stages of metamorphism, whereas sillimanite inclusions in garnet rims may represent the GASP assemblage at peak or near-peak P-T conditions. For the GASP reaction, the correct aluminosilicate polymorph must be selected in samples exhibiting more than one Al-silicate phase. In addition, the correct polymorph must also be correlated with
the correct phase of garnet growth. Failure to do this will result in a meaningless pressure estimate. Similarly, pressure estimates in GASP that put the sample in the stability field of the wrong aluminosilicate must be rejected (i.e. the stable aluminosilicate polymorph must be the same as the polymorph named in the pressure equation), as such a result suggests an element of disequilibrium in one or more of the phases analysed (Holdaway, 2001).

Figure 6.64: SEM (BSE) image of two biotite textures in sample 478336. Garnet is spatially associated with the coarse biotite that forms the peak S2 gneissic fabric and in the absence of alteration in biotite may have yielded temperatures for peak/near peak metamorphism. The partially resorbed lower edge of the garnet is not in equilibrium with the intergrown lathes of biotite and plagioclase and cannot be used for thermometry under any circumstances.

It is crucial to texturally identify any form of metamorphic overprinting in a sample, as this may also affect P-T estimates. All geochemical data should be reviewed prior to thermobarometric modelling to ascertain the scale of partial overprints that may (or may not) have been visible using microscopy. Evidence of resorption (or other breakdown textures) involving one or more phases in a potential thermobarometric assemblage must be investigated as phases exhibiting mutual resorption textures did not exist in equilibrium so should not be used. An example of this may be the relationship between garnet and the two biotite textures (sections 6.4 to 6.7; Fig 6.64). In this case, garnet and coarse matrix biotite may have been used to calculate peak temperatures for the sample thermometry. However widespread partial chloritisation of matrix biotite prevented the use of Grt-Bt thermometry in this sample. The serrated, partially resorbed rims on the lower edge of the garnet are
mantled and replaced by intergrown biotite and plagioclase, showing the
development of biotite from the breakdown of garnet. Mineral geochemistry from
this texture would not be at equilibrium and so should not be used to give
temperature estimates.

With regards to the Kapisillit sample suite, many samples preserve evidence
of garnet breakdown to biotite and plagioclase, thus limiting the use of the Grt-Bt
geothermometer. Care should also be taken with Fe-Ti oxides, as these phases
commonly form in association with biotite breakdown and do not necessarily form
part of the GRAIL assemblage (Bohlen et al., 1983b).

The P-T history of a sample may also be investigated using assemblage
pseudosection analysis. These are multi-dimensional graphs of P-T space that show
all reactions possible for a rock with a specific bulk composition. Time constraints
prevented the construction of pseudosections for specific samples in the Kapisillit
dataset. Instead, the P-T histories of the three field areas were constrained by
comparing the mineralogy and P-T estimates with published pseudosections. As a
pseudosection is bulk-composition specific, it was necessary to identify
pseudosections for rocks with a similar bulk composition and mineralogy. For this
study, it was therefore necessary to find muscovite- and K-feldspar-absent models
and this severely limited the number of pseudosection studies available. Ultimately
the studies of Zeh (2001) were selected, which models Grt-Sil-(Ky)-Bt-Pl-Qtz
bearing samples where the only muscovite occurrences are retrograde. Pseudosections
were modelled using the KFMASH and MnKFMASH systems.

Another issue with comparing samples to published pseudosections is that
there will always be some discrepancy in bulk composition and mineral
geochemistry. This will affect the size of the multivariant fields occupied by specific
assemblages and the positioning/angle of the univariant lines that separate the fields.
For example, Mn stabilises garnet and Zn stabilises staurolite, therefore a sample
with high total Mn and Zn may cause the in-reactions of garnet and staurolite to be
stabilised to lower pressures and temperatures than those with lower bulk Mn and
Zn. Similarly, rocks with high Mg are more likely to see an increase in size of the
cordierite stability field at high temperatures. Because of this, the exact position of
the Kapisillit samples on the pseudosections of Zeh (2001) cannot be accurately
Considering the limitations to the use of thermobarometry discussed above, the following considerations were made regarding the selection of datapoints for P-T analysis. In the following sections, the garnet-biotite thermometer is only employed in samples showing minimal biotite alteration in BSE imaging, of which selected mineral analyses contained >9wt% K₂O and have wt% oxide totals (minus H₂O) >95%. Samples used for GASP and GRAIL barometry have X_{grs} ≥0.05 and X_{an} ≥0.17 according to the recommendations of Holdaway, (2001) and Bohlen, (1983b). The choice of garnet analyses was also based on textural constraints; only garnet cores containing rutile were selected for GRAIL analysis and garnet regions including sillimanite were selected for GASP. Garnet rims showing late stage diffusive zoning were avoided. Where no P-T estimates could be obtained via thermobarometry, conditions of metamorphism are estimated using pseudosections of Zeh (2001), calculated for muscovite and K-feldspar-poor pelitic rocks and the average pelite P-T grids of Spear (1993). A table of analyses used for thermobarometry is given in Appendix C; Table 5.

Table 6.2: Key reactions identified in pelitic units in Norsanna, Tummeralik and Aputitooq Mountain

<table>
<thead>
<tr>
<th>No.</th>
<th>Reaction</th>
</tr>
</thead>
<tbody>
<tr>
<td>4</td>
<td>Ctd + Qtz = Alm + St</td>
</tr>
<tr>
<td>5</td>
<td>Fe-St + Qtz = Alm + Als</td>
</tr>
<tr>
<td>6</td>
<td>Grt + Chl + Qtz = St + Bt + H₂O</td>
</tr>
<tr>
<td>7</td>
<td>Grt + Chl = Bt + Als</td>
</tr>
<tr>
<td>8</td>
<td>Grs + Sil + Qtz = An</td>
</tr>
<tr>
<td>9</td>
<td>Mg-Crd + phi +Sil = Mg-Chl</td>
</tr>
<tr>
<td>10</td>
<td>Grt + Sil + Qtz = Crd</td>
</tr>
<tr>
<td>11</td>
<td>Sil + Bt = Chl + Ms</td>
</tr>
</tbody>
</table>
Norsanna

The inclusion of rutile in garnet cores in samples 482402 and 482435 indicates an initial phase of garnet growth in equilibrium with rutile and quartz. The replacement of kyanite and staurolite by sillimanite also suggests these minerals to belong to an earlier assemblage. This is further supported by the orientation of former kyanite in the gneissic fabric, as it shows kyanite to pre-date the development of the biotite-sillimanite fabric. These early assemblages and textures are interpreted to represent an early metamorphic episode in Norsanna (termed hereafter MIN), which formed the assemblages Grt-Ky-Bt-Pl-Qtz ± Rut and Grt-St-Bt-Pl-Qtz in rocks of differing bulk compositions. Estimated temperatures based on this assemblage suggest conditions in the kyanite field above the terminal chloritoid reaction 4 at ca. 500°C but below to the terminal staurolite reaction 5 at ca. 660°C. A phase of resorption in garnet in 481439, which separates garnet cores from rims may be the result of garnet consumption reactions 6 and 7 (reactions shown in Table 6.2).

Zeh (2001) calculated pseudosections for pelitic rocks with a similar bulk composition to the Kapisillit pelites which contain the assemblage Ky-Sil-Grt-St-Crd-Bt-Chl-Ms-Pl-Qtz, where chlorite and muscovite are present only as a late phase. MIN is considered to lie in the KFMASH divariant Chl-St-Grt-Bt field (Fig. 6.65), where chlorite is consumed and replaced by staurolite moving towards higher pressure and temperature via reaction 6. It is reiterated here that staurolite and former kyanite are not encountered in the same sample. This discrepancy may result from differences in the XFe of the bulk composition in the samples, which produce Grt-St-Bt-Pl-Qtz (481439) and Grt-Ky-Bt-Pl-Qtz (482402 and 482435). In this scenario, it is suggested that a higher XFe in 481439 relative to 482402 and 482435 stabilises the staurolite field to higher PT, by changing the position of reaction 5 in P-T space. Alternatively, sample 481439 may record a marginally earlier portion of the P-T path. Application of the GRAIL barometer of Bohlen (1983a) using garnet and sparse ilmenite analyses from 482402 yields pressures between 6.49 and 6.99 kbar (± 500 bars) for the temperature interval 500-675°C. However comparisons with average pelite P-T grids using the calculated pressures suggest M1 temperatures to be <640°C, as sillimanite becomes the stable aluminosilicate above this temperature at the given pressure interval. Consequently a maximum temperature-pressure estimate.
for $M_1^N$ is given as 640°C and 6.98 kbar, in agreement with pseudosections (Zeh, 2001) and places $M_1^N$ in the mid-amphibolite facies.

The upper amphibolite facies assemblage Grt-Sil-Bt-Pl-Qtz is present in samples 481439, 482402, and 482435 and defines the $M_2^N$ event. The development of sillimanite during reaction 5 in 481439 suggests that the terminal staurolite reaction was crossed in the sillimanite field. Sillimanite inclusions in garnet rims in sample 482402 further supports a second garnet growth phase after the kyanite = sillimanite transition. The drop in $X_{grs}$ across the resorbed core boundary in 481439 may reflect the GASP reaction 8. However $X_{grs}$ in garnet rims in all samples was too low to apply the GASP barometer. The Grt-Sil-Bt-Pl-Qtz assemblage is considered to represent the maximum stable assemblage attained in Norsanna and is considered to exceed the ~640°C temperature estimate for $M_1^N$. The absence of significant migmatisation in all three samples suggests that partial melting in these units was not widespread, and consequently temperatures are not expected to far exceed ~670°C (Bucher and Frey, 2002). Estimations from P-T grids (Bucher and Frey, 2002) and pseudosections (Zeh, 2001) suggest that even up to temperatures of 700°C, maximum pressures will not exceed 7.5k-8kbar. Assuming a maximum temperature of 670°C thus gives an approximate pressure of 7kbar for $M_2^N$, not much greater than the GRAIL estimates for $M_1^N$, suggesting that crustal thickening in Norsanna had largely ceased by $M_2^N$. 

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The appearance of cordierite in sample 482435 defines the onset of M3N and forms the KFMASH divariant assemblage Grt-Sil-Crd-Bt (+Qtz, H2O), where cordierite forms from the consumption of sillimanite. This is observed as cordierite pseudomorphs after sillimanite in the matrix. The stabilisation of cordierite by Mg suggests that the cordierite field in 482435 may cover a wider P-T space than that presented in Fig. 6.65, due to lower MgO bulk compositions in the samples used by Zeh (2001) to construct the pseudosections.

The converse could be applied to 482402 and 481439, which are richer in Fe and would arguably result in a reduction of the P-T space occupied by the cordierite field, accompanied by a corresponding extension of the Grt-Sil-Bt field into lower P-T conditions. The low gradient of the cordierite-in reaction indicates a strong pressure dependence and suggests that a degree of decompression ± heating would be required to instigate the production of cordierite in M3N. Further evidence of a reduction in pressure is lent by the presence of coronae of plagioclase and biotite on garnet in 481440 and 482402, however the extent of this decompression is not constrained.
The variable chloritisation of biotite, breakdown of sillimanite and limited breakdown of cordierite in 482435 via reaction 9 are interpreted as a late episode of transitional greenschist-amphibolite retrogression (M4N). The overprint is also associated with the breakdown of sillimanite to white mica, suggesting re-equilibration in the KFMASH trivariant Sil-Chl-Ms field (Fig. 6.65) at temperatures of 550-580°C and pressure of 4-5kbar. Invasive fluids associated with M4N muscovite veining are considered to be the source of hydrous fluids that instigated the event. Potassium in muscovite may be derived from the breakdown of biotite via chloritisation.

The metamorphic evolution of Norsanna preserves four main phases which define a clockwise Barrovian P-T path (Fig. 6.65). M1N took place under moderate pressure mid-amphibolite facies conditions in the presence of kyanite and rutile at ≤640°C and 6.9kbar. Sample 481439 preserves staurolite but is kyanite-absent and so may be argued to represent marginally lower grade P-T conditions, however the sample still falls within a mid-amphibolite facies assemblage and so is correlated with M1N. Peak M2N conditions in the upper amphibolite facies were attained at pressures similar to M1N, suggesting a near isobaric heating path up to 650-700°C. Temperatures are not expected to greatly exceed these estimates due to low volumes of partial melt in all samples. M3N is characterised by a reduction in pressure at high temperatures, leading to the development of cordierite from sillimanite breakdown. M4N is defined by a partial overprint under transitional greenschist-amphibolite facies at 550-580°C and 4-5kbar. The event is considered to have been instigated by the infiltration of hydrous fluid-bearing veins which liberated K from biotite and led to the growth of retrograde muscovite.
Figure 6.66: a) Schematic KFMASH pseudosection and PT path for Tummeralik based on muscovite-poor pelitic rocks of Zeh (2001). The blue fields indicate the P-T space that would contain the assemblages associated with M1 and the blue dashed open circle reflects the range of estimated GRAIL conditions for M1. The dashed red arrow is a suggested clockwise P-T path for Tummeralik. Black circles refer to P-T estimates obtained qualitatively by assemblage and pseudosection analysis and so have no quantifiable errors. b) Schematic MnKFMASH pseudosection detailing the narrow Grt-Ky-St-Bt-Qtz-H2O-(PI) field and the approximate position of M1.

**Tummeralik and the Ice Lake reconnaissance area**

The inclusion assemblages preserved in garnet cores indicates initial (M1) garnet growth in the presence of biotite, rutile and quartz, with staurolite, ilmenite, zircon, monazite and plagioclase locally preserved. A single occurrence of chlorite in garnet cores is observed in samples 478325 in the Ice Lake reconnaissance area. The presence of sillimanite pseudomorphs after kyanite lying parallel to the biotite-sillimanite fabric suggests kyanite as an early aluminosilicate phase that formed prior to the peak fabric. Comparison with the pseudosections of Zeh (2001) suggests that M1 preserves assemblages in the KFMASH divariant Grt-St-Bt-Chl (+Qtz +H2O) field (Fig. 6.66). Moving towards higher pressure and temperature, chlorite is consumed via reaction 6 to form the KFMASH trivariant assemblage Grt-St-Bt (+Qtz +H2O). This assemblage is interpreted to represent a larger field in the Tummeralik samples than those calculated by Zeh (2001), due to the stabilisation of staurolite at lower P-T as a result of significantly higher Zn in staurolite in the Tummeralik samples (~0.15 Zn (Tummeralik) per formula unit versus 0.01 (Zeh, 2001); based on 24 oxygens). Kyanite-staurolite textures are not encountered in any samples, and as
such their coexistence during M1T cannot be substantiated. Pseudosections do suggest a small, highly temperature-dependent field at ca. 650°C, 6.5-8kbar where the KFMASH and MnKFMASH assemblage Grt-Ky-St-Bt-Qtz-(Pl)-H2O exists, before staurolite is consumed via reaction 5 to form the assemblage Grt-Ky-Bt-Qtz-(Pl)-H2O. Consequently, kyanite may have been stable with the staurolite-bearing M1T assemblage over a short P-T interval, but may equally marginally post-date staurolite. GRAIL barometry of ilmenite in garnet cores in samples 478323, 478339 and 492237 yields a spread in pressure estimates (Fig. 6.67). Multiple analyses in 478339 and 492237 estimate pressure between 5.3 and 6.3kbar over the temperature interval 400-725°C (Fig. 6.67). However comparisons with pseudosections suggest that sillimanite would be the stable aluminosilicate at these pressures. This suggests that the mineral chemistry of garnet and/or ilmenite in the analysed samples may not have retained GRAIL equilibrium compositions and as such, are not considered to record the pressure of the M1T event. A higher pressure of 7-8.3kbar was yielded for a single analysis in sample 478323, which is considered to more accurately represent pressure conditions of M1T, as it is the only sample to yield pressures within the kyanite stability field. However it is noted that this result comes from a single analysis and so cannot be tested for reproducibility. An approximate P-T estimate of 600-650°C and 6-8kbar is thus suggested for M1T based on textural relationships and pseudosection analysis.

The upper amphibolite facies assemblage Grt-Sil-Bt-Pl-Qtz defines the M2T assemblage, with no vestiges of staurolite preserved in the matrix, suggesting the attainment of P-T conditions past the terminal staurolite reaction (5) in the KFMASH trivariant Grt-Sil-Bt-Qtz-(Pl)-H2O field (Zeh 2001). The incorporation of sillimanite into garnet rims indicates garnet rim growth after the Ky = Sil transition. Peak P-T conditions were calculated in two samples using the combined garnet-biotite thermometry and GASP barometry of Holdaway (2001). Similar to GRAIL, estimates from Grt-Bt thermometry and GASP barometry are variable. Sample 478339 gave peak conditions of 760°C at 7.2kbar, whereas 492237 yielded a lower value of 692°C at 6.3kbar (± 25°C and ± 800bars). Although interpreted growth zoning profiles are preserved in 478339, the absence of significant partial melting suggests that temperatures have been overestimated, potentially due to re-equilibration of zoning in garnet and biotite. The pressure estimate for 492237 is
similarly treated with caution, again due to the potential for homogenisation in both garnet and biotite. However the calculated pressures from 492237 are in broad agreement with mineral textures and pseudosections, and thus are considered to reflect approximate P-T conditions for M2\text{T}.

![Figure 6.67: A plot of temperature vs. pressure illustrating the variation in GRAIL pressure estimates over a range of temperatures for the Tummeralik pelites. Coloured dashed lines delineate the maximum error on pressure estimates for each sample with ± 500 bars (absolute error) on each point. Blue, orange and green error margins relate to samples 478323, 478339 and 492237 respectively.](image)

The development of post-peak (M3\text{T}) cordierite in sample 478343, which pseudomorphs sillimanite, suggests the transition from the Grt-Sil-Bt field to Grt-Crd-Sil-Bt (both +Qtz +H₂O +Pl). The occurrence of garnet consumption textures suggests the operation of reaction 10. Coronitic textures in garnet are interpreted as decompression textures, and are compatible with the low gradient(s) of the cordierite-forming reaction(s). M3\text{T} is consequently interpreted as a phase of decompression from peak M2\text{T} pressures, however the extent of decompression remains poorly constrained.

A later metamorphic event is defined in Tummeralik and the Ice Lake reconnaissance area by the breakdown of sillimanite and biotite via reaction 11. Chlorite rims around partially resorbed cordierite in 478343 also indicates the operation of reaction 6. This forms the greenschist-amphibolite facies transition assemblage Chl-Sil-Ms, which occupies the approximate P-T space 550-575°C, 4-
5kbar. This field appears to be largely stabilised by Mg, as the P-T space occupied by the field decreases with increasing X_{Fe} until aluminosilicate is replaced by staurolite at X_{Fe} >0.6 (Zeh, 2001). However such Fe-rich bulk compositions were not encountered in Tummeralik. The development of muscovite veins which cut the biotite-sillimanite foliation in samples 478325 and 478323 are interpreted to indicate source of the retrogressive fluids, which instigated retrograde metamorphism. Where present, the high angle between veins and the peak gneissosity suggest that M4_T is unrelated to the high grade M1-M3_T events.

The metamorphic evolution of Tummeralik preserves four main metamorphic episodes, which define a clockwise Barrovian P-T path (Fig. 6.66). M1_T took place under moderate staurolite to kyanite zone conditions in the presence of rutile at approximately 600-650°C and 6-8kbar. Peak M2_T conditions in the upper amphibolite facies were attained at ca. 690°C, 6.3kbar. Temperatures are not expected to far exceed 700°C, as significant partial melting is not recorded in the Tummeralik pelites. M3_T is characterised by decompression at high temperature, leading to the development of cordierite from sillimanite and garnet breakdown. M4_T is defined by a partial overprint under transitional greenschist-amphibolite facies at 550-575°C and 4-5kbar. The event is considered to have been instigated by the invasion of hydrous fluid-bearing veins which liberated K from biotite and led to the growth of retrograde muscovite. This final episode is thought to be unrelated to M1-M3_T metamorphism.

**Aputitooq Mountain**

The early evolution of Aputitooq Mountain is preserved in garnet cores in samples 492202 and 492219, which suggest the M1_A assemblage Grt-St-Bt-Rut-Ilm-Qtz. Unlike pelites in Norsanna and Tummeralik, there is no evidence of an early kyanite phase. However kyanite has been observed near Kapisillit village to the northeast of Aputitooq Mountain, which led (Nutman and Friend (2007) to suggest a prograde path in the kyanite stability field. The GRAIL reaction was employed, but estimates were rejected because simulations with kyanite as the aluminosilicate phase produced pressures in the sillimanite stability field. Consequently a P-T estimate for M1_A is proposed based on pseudosections (Zeh 2001), and is constrained
to the KFMASH trivariant assemblage Grt-St-Bt (+Qtz +H₂O +Pl) (Fig. 6.68). This places M₁ₐ at approximately ~640°C and 6.5-8kbar. Despite the similar Xₚₑ of sample SK1 (Zeh, 2001) and 492219, it is expected that the Grt-St-Bt field would occupy a larger PT space, due again to high Zn in staurolite (0.18-0.2 Zn p.f.u. (Aputitooq Mountain) versus 0.00-0.0.01 (Zeh, 2001); based on 24 oxygens).

Attempts were also made to constrain the conditions of M₂ₐ, which produced the assemblage Grt-Sil-Bt-Pl-Qtz from the breakdown of staurolite via reaction 6. Garnet-biotite thermometry yielded temperatures of ca.650°C, with data from the two viable samples (492219, 492202) yielding temperatures 648°C and 653°C respectively. However calculated pressures from the GASP barometer are not in agreement and yield lower pressures than would be expected (4.9 and 2.7 kbar at 648 and 653°C respectively). These temperature estimates are lower than those proposed by Nutman and Friend (2007), which suggested metamorphic temperatures of ca. 700°C, however the samples used in their study are cordierite-bearing. Consequently it is possible that the sample suite of Nutman and Friend (2007) records a later part
of the P-T path nearer the thermal maximum. Alternatively, the results obtained from garnet-biotite and GASP may reflect the temperature and pressure of re-equilibration of the high grade assemblage, rather than the initial prograde equilibrium. Biotites in Aputitooq Mountain are the least chloritised of any pelitic rocks encountered in the three field areas. However garnet in the area is interpreted to have been (partially) homogenised, altered and retrogressed, based on flat growth zoning profiles and ubiquitous reaction textures. The re-equilibration of garnet may thus account for the inconsistencies in the P-T estimates for different samples. The production of cordierite post-dates the biotite-sillimanite fabric and defines the M3A event. Coronitic textures on both sillimanite and garnet in 492710 indicate the operation of reaction 10 in the presence of a fluid as suggested by the symplectic intergrowths of quartz and cordierite in garnet coronae. The complete absence of plagioclase in sample 492710 sets it apart from all other samples in the Kapisillit sample suite and makes it particularly difficult to apply to the pseudosections of Zeh (2001). However the replacement of garnet and sillimanite by cordierite implies decompression from high temperature and as such is interpreted to record M3A decompression at transitional amphibolite-granulite facies conditions, also in agreement with Nutman and Friend (2007).

Pelitic assemblages in the Aputitooq Mountain area indicate a clockwise path that can be divided into three parts (Fig. 6.68). However all three stages are poorly constrained. The prograde path (M1A) preserved in garnet cores suggests staurolite-zone conditions at ~640°C and 6.5-8kbar. Peak metamorphism (M2A) took place at upper amphibolite facies at ≥650°C, which was followed by post-peak decompression (M3A) and potentially some heating to transitional granulite facies, resulting in the widespread development of coronitic rims on garnet and the development of cordierite in plagioclase-absent bulk compositions.
6.11 Summary of P-T paths in the three field areas

Combined mineral assemblage, thermobarometry and pseudosection analysis show that all three field areas have undergone upper amphibolite to transitional granulite facies metamorphism. Norsanna and Tummeralik record four phases of high grade metamorphism, whereas Aputitooq Mountain preserves strong evidence for three metamorphic episodes. The key conclusions from sections 6.9 and 6.10 are summarised below:

1. The absence of prograde/peak muscovite and K-feldspar in most samples is interpreted as the result of high Fe and Mg in bulk compositions. This prevented the use of Thompson projections and limited the number of published pseudo sections available for comparative P-T modelling.

2. Widespread homogenisation and partial retrogression in garnet, plagioclase and biotite prevented the widespread use of the GRAIL, GASP and Grt-Bt geothermobarometers.

3. M1 is associated with metamorphism in the mid-amphibolite facies in the staurolite/kyanite zone. All geobarometers yield pressure estimates between 6 and 7 kbars, with corresponding temperatures of 600-650°C, suggesting moderate P-T conditions in all three field areas.

4. Peak P-T conditions in all field areas was attained during M2 and reached the upper amphibolite facies (sillimanite zone). Pressures of 6.5-7 kbars are not thought to be markedly higher than M1 pressure estimates. An absence of widespread migmatisation in all samples suggests temperature were not much greater than ~650-670°C.

5. M3 is characterised by the breakdown of sillimanite and biotite to cordierite, suggesting high temperature (upper amphibolite-transitional granulite) decompression below ~6 kbar. This is the last recorded metamorphic event in Aputitooq Mountain.

6. M4 in Norsanna and Tummeralik took place under transitional greenschist-amphibolite facies conditions at ~550-575°C and ~4 kbar. Discordant relationships between peak fabrics and retrograde veins suggest M4 is unrelated to the high grade M1-M3.


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CHAPTER 7
THE NEOARCHAEOAN TECTONOTHERMAL EVOLUTION OF THE SE NUUK REGION

7.1 Overview

Structural and geochronological data from Norsanna and Tummeralik were presented in Chapters 2, 3 and 4. Initial interpretations discussed in Chapter 5 suggested a shared post-Dn structural evolution in Norsanna and Tummeralik. This interpretation needed further substantiation from metamorphic analysis. Chapter 6 presented petrological data and defined P-T paths for all pelitic samples obtained from Norsanna, Tummeralik and the Ice Lake area, with additional P-T studies of metapelites from Aputitooq Mountain in the northern Kapsillit region. This chapter will integrate the results and interpretations from all areas to propose a Neoarchaean tectonothermal model for the Kapsillit region and will consider the model in the context of the Nuuk region as a whole.

7.2 2720-2700Ma metamorphism in Norsanna and Tummeralik

Correlations of lithological, structural and geochronological data discussed in Chapter 5 suggest that D1 and D2 represent correlative events in Norsanna and Tummeralik. In addition, these events, which involve initial km-scale isoclinal folding followed by F1-coaxial attenuation and high strain zone development, are considered to be progressive elements of a single tectonothermal event. Dating of late-D2 leucosomes has constrained shearing to ca. 2720Ma in Norsanna and the event has been traced as far east as Ameralla Fjord, where metamorphic zircons of this age occur in Mesoarchaeon gneisses interpreted to belong to the Kapsillik terrane.

Re-equilibration of key phases and low Ca in garnet and plagioclase in a number of samples limited the use of thermobarometry in the majority of pelitic samples in all field areas. However detailed petrography of pelitic assemblages in Norsanna, Tummeralik and the Ice Lake area (which is hereafter grouped with
Tummeralik, as the Ice Lake area has been correlated with Tummeralik based on mapping and metamorphic assemblages) identified very similar pressure-temperature paths. In addition, key mineral assemblages are often characterised by the same mineral textures. Common petrological and geochemical features of pelites in all field areas and their interpreted P-T paths are summarised below:

1. The early mid-amphibolite facies M1 assemblage Grt-Bt-Pl-Qtz ± Ky ± St ± Rut ± Ilm is preserved in garnet cores and as rare relict matrix kyanite. Kyanite, where present, is blocky and has been replaced by subgrained sillimanite.

2. Garnet inclusion patterns are similar in virtually all samples. Staurolite forms inclusions in garnet cores and is often associated with rutile. Cordierite-bearing samples exhibit distinctive 3-domain inclusion patterns in garnet (inclusion-free core – Chl-Bt-Rut-Qtz-Mnz rings – inclusion-free rim), where a significant phase of resorption separates domains 2 and 3. All garnets in sillimanite-bearing lithologies contain clusters or annuli of fibrolitic sillimanite in their rims.

3. Growth zoning in garnet is largely absent from most samples. However retrograde zoning in Fe, Mg and Mn is very common at grain rims and often encroaches into sillimanite-bearing annuli. Garnet in most samples contains $X_{grs}$ between 0.03 and 0.05.

4. The dominant gneissic fabric in all aluminosilicate-bearing samples is defined by coarse, intergrown biotite-sillimanite, which wraps garnet in all cases. The peak M2 mineral assemblage is the upper amphibolite facies assemblage Grt-Sil-Bt-Pl-Qtz ± Ilm ± Rut ± Zrc ± Mnz.

5. Development of cordierite during M3 at the expense of garnet and sillimanite and is not associated with the development of a new fabric.

6. Biotite in all Norsanna and Tummeralik samples has undergone some degree of chloritisation during M4 greenschist facies retrogression, which is most pronounced in samples within or adjacent to $D_{4N}$ or $D_{3T}$ fault zones.
7. A distinct M4 event was not identified in the Aputitooq Mountain pelites. Biotite in some samples has undergone some chloritisation, albeit to a lesser degree than in Norsanna and Tummeralik.

8. Muscovite is only developed as a retrograde phase and is associated with M4 and the D4N or D3T events.

The preservation of former kyanite in the matrix and staurolite in garnet cores during M1 suggests early metamorphism at moderate pressures and is in keeping with the GRAIL estimate of ~6.98 kbar at ~640°C in Norsanna. It may be argued that the (M1) kyanite and staurolite assemblages reflect an early (Dn) tectonothermal event. The orientation of Dn kyanite parallel to S1,2 in both areas may then be explained by porphyroblast rotation during the development of the peak M2 fabric. In Norsanna, the presence of ≥3560Ma metamorphism in Eoarchaean rocks (the polyphase orthogneiss and associated lithologies) presents the possibility that kyanite may have developed during Dn. However, the kyanite- and staurolite-bearing pelitic gneisses in Norsanna are exclusively hosted by the homogeneous orthogneiss. They also contain detrital cores with ages between 2800 and 3035Ma and exhibit a single metamorphic age of ca. 2720Ma (Nutman and Friend, 2007; Nutman et al., 2004), suggesting kyanite and staurolite formation during D1 rather than Dn.

The timing of M1 metamorphism in Tummeralik and the Ice Lake area remains ambiguous as it is not established whether the development of kyanite is the result of an early phase of metamorphism associated with Dn deformation or simply associated with the prograde P-T segment of a single tectonothermal event. Evidence for Dn deformation in Tummeralik is suggested by the presence of isoclinally folded mafic enclaves in the tonalitic orthogneiss that are truncated by the host material. However, as comparable structures were not observed in the large supracrustal amphibolitic bodies or the pelitic units hosted therein no correlations are drawn between the highly deformed amphibolitic material in the tonalitic orthogneiss and the large supracrustal bodies. The maximum age for the Dn isoclinal folding in the mafic enclaves is constrained to 3046 ± 10 Ma by the magmatic age of the host tonalitic orthogneiss (Chapter 5). However this does not constrain the age of metamorphism in the supracrustal sequences as the contact relationships between these and the tonalitic orthogneisses are invariably highly strained. Detrital zircon
studies of the Tummeralik pelites are required to provide upper age limits on the
tectonothermal events that have affected these supracrustals.

The M1 kyanite-staurolite-rutile bearing assemblages in Norsanna and
Tummeralik are considered to represent the early stages of the upper amphibolite
facies event, whose later stages have been dated at 2720Ma. This is interpreted to
have occurred contemporaneously with the D1 phase of km-scale folding, which
formed the km-scale Norsanna and Tummeralik Isoclines. D1 is the first recorded
phase of folding associated with the 2720Ma event and as such, the development of
eyearly kyanite and staurolite during M1 is compatible with this initial phase of crustal
thickening.

Peak (M2) metamorphism in the sillimanite field is recorded by the
assemblage Grt-Sil-Bt-Pl-Qtz in all aluminosilicate-bearing samples. The attainment
of upper amphibolite facies M2 conditions is considered to marginally predate the
onset of D2 attenuation and high strain zone development, based on the rotation of
sillimanite lineations by D2 in both Norsanna and Tummeralik. However conditions
in the sillimanite field were maintained during D2. This is suggested by the
development of the second, subordinate sillimanite lineation, which is oriented
approximately dip-parallel to the NHSZ and is considered to reflect the direction of
motion on the shear zone during D2.

The model preferred here suggests that M1 and M2 in Norsanna and
Tummeralik represent prograde and peak metamorphism in a single tectonothermal
event. Evidence presented in Chapter 5 also indicates that Norsanna and Tummeralik
may have a shared deformational history in terms of the D1 and D2 events.
Metamorphic assemblage analysis from the two areas tends to support this
interpretation, based on the observation that the M1 and M2 events occupy
approximately the same P-T field (Fig. 7.1) and occur at the same point in the
deformational history; M1 is considered to be contemporaneous with D1 isoclinal
folding, with M2 conditions being established prior to D2 attenuation. The major
problem with correlating peak metamorphism in the two areas lies with the difficulty
in obtaining unequivocal age data in Tummeralik. Geochronological data from
Norsanna (this study; Nutman et al., 2004) can confirm that D2 had occurred by ca.
2720 Ma, however a leucosome interpreted to record late-D2 intrusion in Tummeralik

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gives a significantly younger age of 2659 Ma. It has been noted that the 2659 Ma age was not reproducible in the other Tummeralik geochronology sample, and it is emphasised that a single leucosome (regardless of level of alteration) is not sufficient to interpret the presence or absence of a 2659 Ma event in Tummeralik. Consequently, further detailed study of the timing of metamorphism in this area is vital. Despite the discrepancies in the metamorphic zircon data in Tummeralik, the observation of ca. 2720 Ma metamorphic populations in orthogneisses and felsic leucosomes in Norsanna and Ameralla Fjord is consistent with the presence of a shared ca. 2720 Ma event across the Færingehavn, Tre Brødre and southern Kapisillik terranes.

Figure 7.1: P-T-t-deformation path for Norsanna and Tummeralik.
7.3 2650-2600Ma metamorphism in Aputitooq Mountain

The Aputitooq Mountain metamorphic zircon population is dominated by age populations between 2650 and 2600Ma in supracrustal rocks which experienced upper amphibolite facies metamorphism followed by decompression (Bennett, Pers.Comm.; Bennett and Heiss, Pers.Comm.). Metamorphic assemblages in the Aputitooq Mountain study area show pelitic rocks to have followed a mid-amphibolite facies prograde path (Fig. 7.2).

![Diagram of P-T-t-deformation path Aputitooq Mountain](image)

Although it is not certain whether prograde metamorphism took place in the kyanite stability field in this area, units to the northeast near Kapisillit village (Fig. 7.3) contain relict kyanite and are interpreted by Nutman and Friend, (2007) to represent 'high-pressure' metamorphism. Previous work has suggested that 'high pressure granulite facies' metamorphism at 2650Ma was instigated by the thrusting of the Færingehavn-Tre Brødre terranes over the Kapisillik terrane in the approximately north-south trending belt that runs parallel to the east coast of Itilleq Fjord and includes Aputitooq Mountain (Friend and Nutman, 2005; Nutman et al., 2007a). The development of low-angle break thrusts with associated mesoscale isoclinal folding supports the interpretation that Aputitooq Mountain is an area of large scale crustal
interleaving. Bennett and Heiss (Pers. Comm.) also note that the supracrustal rocks apparently show more complex metamorphic histories than the host orthogneisses, and which are separated from the orthogneisses by shear zones. This further supports the argument for an intensely tectonised zone comprising a number of unrelated lithologies. However the differences in apparent metamorphic grade may also be an artefact of contrasting bulk compositions and as such, this interpretation would need to be confirmed by detailed metamorphic study of all units.

Figure 7.3: A sketch map of the Kapisillit Mapsheet showing major lithologies, structural trends and key localities in the Kapisillit region. Structures in Norsanna and Tummeralik that have been correlated are shown in dark red and blue.
Although prograde and peak (M1 and M2) metamorphism has been correlated with the \( D_{n+1} \) event of Bennett and Heiss (Pers. Comm), the timing of this part of the PT curve is difficult to constrain confidently. Nutman and Friend, (2007) have dated metamorphic segregations in amphibolites and pelites (garnet-clinopyroxene and kyanite-bearing respectively) at 2650Ma in the area around Kapisillit village. In addition, metamorphic rims on detrital zircons from pelite samples 492202, 492206 and 492219 suggest a polymetamorphic history prior to 2650Ma. All three samples show major metamorphic zircon growth and recrystallisation between 2635 and 2650Ma, with smaller \( ^{207}\text{Pb}/^{206}\text{Pb} \) metamorphic peaks between 2680 and 2720Ma. The best defined pre-2650Ma metamorphic population is yielded by sillimanite-bearing sample 492202, which exhibits well-defined metamorphic rims at 2700Ma (Bennett, Pers.Comm.). This would seem to suggest that this part of the ‘Kapisillik’ terrane has seen high grade metamorphism at this time. However caution is exercised here as this is the only pelitic sample from the Aputitooq sample suite that does not exhibit Mesoarchaean inheritance and metamorphism, which is typical of units from the Kapisillik and Ivisaartoq supracrustal belts (Nutman et al., 2004; Polat et al., 2007).

The single inheritance age of ca. 2824Ma is more compatible with the supracrustal units in Norsanna dated by Nutman et al., (2004). This is consistent with the sample belonging to an unrelated supracrustal package that has been tectonically juxtaposed against the heterogeneous supracrustal units in Aputitooq Mountain. The remaining pelitic samples containing pre-2650Ma metamorphic zircon show only minor peaks in \( ^{207}\text{Pb}/^{206}\text{Pb} \) at ca. 2720Ma. Consequently no definite age constraints can be placed upon prograde and peak metamorphism based on current knowledge, however the possibility of a high grade metamorphic event at ca. 2700Ma in the Aputitooq Mountain study area should be considered in future investigations.

In a regional context, high grade metamorphism and deformation at ca. 2650Ma appears to be a definitive event in the Kapisillik terrane (Bennett, Pers.Comm.; Friend and Nutman, 2005; Nutman et al., 2007a). The structures observed in and around Aputitooq Mountain (and extending south towards Ameralla Fjord) are consistent with an interleaved zone separating the ‘Færingehavn-Tre Brødre’ units to the west and ‘Kapisillik’ units to the east, with the latter at the
bottom of the structural pile. Considering the presence of 2720Ma events on both sides of this boundary, the juxtaposition of the two blocks is considered to predate 2720Ma.

In this interpretation, the 2650 Ma event may be the result of overprinting or reworking of a 2720Ma or older boundary between the purported Færingehavn-Tre Brødre and Kapisillik terranes. The southerly extent of the 2650Ma event remains uncertain due to ambiguous geochronological data in the central Kapisillit region. However the increasing absence of 2650Ma ages south of Kapisillit village suggests that the influence of the 2650Ma event did not extent far south of Tummeralik (Fig. 7.4). The non-record of 2650 Ma metamorphism in the southern Kapisillik terrane is not an isolated occurrence in the region. No metamorphic event of this age is recognised south of Ameralik and Ameralla Fjords, in the area interpreted as Tasiussarsuaq terrane. Similarly, the Ivisaartoq supracrustal belt to the northeast of Kapisillit does not preserve zircons at 2650Ma, despite being interpreted as part of the Kapisillik terrane (Friend and Nutman, 2005). The absence of 2650 Ma metamorphism in this part of the Kapisillik terrane led Nutman and Friend, (2007) to suggest that the Ivisaartoq belt occupied a higher structural level at this time and so avoided the high pressure metamorphic overprint. This work suggests that the high grade metamorphism in Aputitooq Mountain at ca. 2650Ma corresponds to a separate P-T path to that followed by Norsanna, Ameralla Fjord and Tummeralik. This event reached upper amphibolite facies grade in the area around Aputitooq Mountain and Kapisillit village, but high grade metamorphism appears not to affect lithologies in the southern Kapisillik terrane. This suggests that the predominant high grade metamorphic event in the northern Kapisillik terrane at 2650Ma did not extend to the southern terrane. This may be a result of distance from the cause of metamorphism or due to the southern Kapisillik terrane occupying a higher structural level at this time.
Figure 7.4: Representative analyses from the Kapisillit Mapsheet geochronological database showing the distribution of metamorphic ages in the Kapisillit region.
7.4 *The timing of M3: Decompression vs. low pressure metamorphic overprint*

The development of cordierite defines the M3 event in Norsanna, Tummeralik and Aputitooq Mountain. It has been suggested that cordierite growth south of Kapisillit village took place just after the ‘high-pressure’ (garnet-kyanite-rutile in metapelites) metamorphic event at ca. 2650Ma, with metamorphic $^{207}\text{Pb} / ^{206}\text{Pb}$ ages of $2643 \pm 10$ and $2638 \pm 6$Ma yielded by SHRIMP analysis of metamorphic zircon and monazite respectively (Nutman and Friend, 2007). The identification of extensional shear zones and partial melt ponded in boudin necks at ca. 2650Ma suggest that M3 in Aputitooq Mountain may have formed in an extensional environment. The small time gap between M2 and M3 in Aputitooq Mountain and Kapisillit village may be compatible with an interpretation of orogenic collapse following at thermal peak at ca. 2650Ma.

The timing of M3 in Norsanna and Tummeralik is more difficult to constrain. Section 7.2 discussed the possibility of cordierite formation at ca. 2720Ma in association with normal motion on the NHSZ. A major problem arises in the determination of sense of shear on the NHSZ, due to the apparent lack of shear indicators along the length of the shear zone. Evidence from a single $F_2$ shear fold in the hinge area of the km-scale $F_{3N}$ fold indicates a normal-sense top-to-S motion. If this fold is representative, and the NHSZ formed initially as an extensional shear zone, it is possible that the zone formed as an orogenic collapse structure in response to thermal softening during peak metamorphism. $D_2$ collapse was initiated after the onset of M2 and may have instigated the production of M3 cordierite through decompression. This proposed structural link between exhumation and the NHSZ also suggests that the relatively undeformed late-$D_2$ felsic leucosomes may represent decompressional melting. However the rare shear indicators denoting normal-sense motion could equally be formed by the (re)folding of a thrust plane (during or after motion) and would therefore, not necessarily relate to a phase of exhumation. Such examples of folded thrust planes are documented in the Swiss Alps (Homewood et al., 1986; Pfiffner, 1986). This scenario is suggested based on the observation that the NHSZ is folded by the $F_{3N}$ syncline and the amount of rotation associated with
folding is difficult to ascertain, as the pre-D$_{3N}$ orientation of rock units are not known.

The above scenario implies that M3 followed directly on, and was instigated by thermal collapse following the thermal maximum attained during M2. Similar ages for cordierite growth following peak metamorphism are observed at Qilanngaarsuit, where cordierite forms from the breakdown of sillimanite and garnet in Palaeoarchaean supracrustal units at ca. 2712Ma (Nutman and Friend, 2007). However no evidence of decompression immediately following M2 is described and the low pressure event at 2712Ma may reflect a lower pressure overprint that is not necessarily associated with thermal collapse and decompression.

The distance between Qilanngaarsuit, Norsanna and Tummeralik (~50 and 90km respectively) may also pose problems for correlating between the cordierite-forming events, as it is unlikely that the three areas occupied the same structural level during metamorphism/deformation. Instead, it is suggested that M3 in Norsanna and Tummeralik may be correlative of M3 in Aputitooq Mountain. This is the preferred interpretation for the trigger for M3 because of the frequency of 2650-2620Ma ages in monazite data from Norsanna. In this case, both M2 and D$_2$ took place in Norsanna and Tummeralik at ca. 2720Ma, but movement on the NHSZ did not instigate M3. In this scenario, the NHSZ may have a thrust or normal sense of motion. The M2 peak assemblages were subsequently overprinted by a lower pressure event at 2650Ma that is not related to M2 or D$_2$ but may be responsible for resetting 2720Ma ages in monazite in Norsanna, in part resulting in the large spread in $^{207}$Pb/$^{206}$Pb ages between 2720 and 2650Ma.

7.5 Post-2650Ma thermal pulses in the Kapisillit region

Recent geochronology suggests the Kapisillit region was subject to a number of thermal overprints following M3 metamorphism in Aputitooq Mountain, Norsanna and Tummeralik. A common characteristic of all post-2650Ma thermal events in the zircon record in Kapisillit and Norsanna is that the only apparent new mineral growth is observed in accessory minerals. However it is suggested that repeated high temperature thermal pulses in Norsanna, Aputitooq Mountain and
Tummeralik may have contributed to homogenisation of biotite and plagioclase in a large number of samples. In addition, homogenisation and/or re-equilibration of garnet rims may also have been affected by these repeated thermal pulses.

As many as four separate thermal overprints in zircon have been identified between 2650 and 2550Ma (Friend and Nutman, 2005; Hollis et al., 2004; Nutman et al., 2007a; Nutman and Friend, 2007), the most notable of which occurred at ca. 2635Ma and 2550Ma (Fig. 7.5). Ages of 2650 and 2550Ma are manifest around Kapisillit Fjord as pegmatite veins and melt segregations that pond in extensional shears (Nutman and Friend, 2007). The presence of these events in Norsanna is supported by metamorphic populations in monazite, which identify peaks in
Numerous thermal pulses have also been identified in the Aputitooq Mountain pelites between 2530 and 2600 Ma, with the best defined peak at 2580 Ma (Bennett, Pers. Comm.). Metamorphism at ca. 2635 Ma has been correlated with upper amphibolite facies metamorphism on the island of Storø (Fig. 7.5) to the northwest of Kapisillit (Nutman et al., 2007). Whilst the connection between monazite growth and Neoarchaean tectonism requires further study, it is likely that monazite growth and/or modification in Norsanna between 2650 and 2600 Ma is either the result of high-grade metamorphism at 2650 Ma or a distant effect of the reworking of the Kapisillit-Akia terrane boundary on the island of Storø (Friend and Nutman, 2005; Nutman et al., 2007a). Age populations in zircon at 2580 Ma may be associated with granitic veins that are intruded near Kapisillit village around this time. The large (150 km long, ≥2000 m thick), dominantly post-tectonic Qôrqt granite forms an elongate sheeted complex that crops out in the northwestern corner of the Kapisillit Mapsheet (Appendix D). The granite intrudes Færingehavn, Tre Brødre and Kapsillik rocks over much of central Godthåbsfjord (Brown et al, 1991; Friend et al, 1985) and so demonstrably post-dates any accretionary tectonism in these three terranes. Although the body is largely post-tectonic, recent studies (Nutman and Friend, Pers. Comm.) suggest that the earliest stages of intrusion may have overlapped with waning tectonism in the Nuuk region. Good structural and geochronological constraints on the Qôrqt granite mean the unit has long been used as both a relative and absolute time marker in the Nuuk region and the majority of recent studies have focussed on geochronology of the body. The emplacement age is currently taken as ca. 2560 Ma (e.g. Windley and Garde, 2008), but intrusion is considered to have taken place in a number of pulses between 2580 and 2500 Ma (Moorbath et al, 1981). As the earliest stages of Qôrqt granite intrusion may have occurred as early as 2580 Ma, it is suggested that granitic vein intrusion in Aputitooq Mountain between 2580 and 2500 Ma may be associated with the Qôrqt granite (Nutman and Friend, Pers. Comm.).
7.6 Greenschist facies retrogression in the Nuuk region

The existence of major greenschist facies overprints in the Nuuk region as a whole is considered minor, with previous outcrops of greenschist assemblages confined to late high strain zones a few metres wide (Chapters 2 and 3, this study; Bennett and Heiss, Pers. Comm.). Although this is the case at outcrop scale, this study has shown that partial greenschist facies overprints are not confined to these zones, and may affect a much wider area than traditional mapping techniques can recognise. This is because chloritisation in the majority of samples is characterised by geochemical trends in biotite, which can only be identified using optical or scanning electron microscopy or mineral geochemistry. Visible chlorite is only present in samples within the immediate vicinity of an M4 shear. The greatest impact of this interpretation is the effect of low grade re-equilibration on high grade metamorphic assemblages, which has serious implications for the calculation of peak metamorphic pressure and temperatures using both traditional thermobarometry and thermodynamic modelling techniques. As a result, it is recommended that future P-T calculations take into consideration the possibility of (multiple phases of) re-equilibration when studying pelitic units, certainly in the Kapisillit region, but also for any samples that show polyphase metamorphic histories. Pseudosection calculations based on bulk compositions should also be corrected for late muscovite introduced through greenschist facies vein systems and the consequent changes in bulk rock chemistry.

The absence of cross-cutting relationships between the \( D_{4N}\text{Norsanna} \) and \( D_{3T}\text{Tummeralik} \) \( (D_{4N} \) and \( D_{3T} \) respectively) has inhibited attempts to correlate greenschist facies structures in both areas. Chapter 2 discussed the possible relationship between \( D_{4N} \) and movement on the Kobbefjord fault system between 2000-1800Ma (Hollis et al., 2004; Smith and Dymek, 1983). Field evidence suggests a Proterozoic age for \( D_{3T} \) – probably post-2200Ma based on mafic dyke truncations (Hollis et al. 2004; Garde 2003). NW-directed thrusts have been tentatively correlated with NE-striking high strain zones in the Bjørneøen Greenstone Belt, however these tend to be high grade (amphibolite facies) and are thought to be associated with the main terrane amalgamation event during the Neoarchaean (Hollis et al. 2004). A number of NW-striking fault structures are recognised in the Kapisillit region (see 1:100,000 map...
Appendices) but have yet to be correlated within the Nuuk region evolutionary framework. The M4 greenschist facies overprint is not as marked in Aputitooq Mountain as in other areas, which is attributed to the absence of major greenschist facies mylonite zones in the vicinity. Relative timing constraints are not available for the brittle fracture zones, however some correlation could arguably be drawn with the NW-striking D₃T fault zone, which is also associated with pyrite mineralisation. The elevated regional temperatures that were associated with greenschist retrogression may also be counted amongst the 'post-2650Ma thermal pulses' discussed in section 7.5. These events are considered to have played an important role in the re-equilibration of peak mineral assemblages and partial resetting of isotopic systems in zircon and monazite across the Kapisillit region.

7.7 Neoarchaean metamorphism and deformation in the SE Nuuk region: Correlations with regional events

Structural, metamorphic and geochronological data presented in this study, combined with data from other sources suggests a complex Neoarchaean tectonothermal evolution for the central and northern Kapisillit region. New metamorphic and geochronological evidence suggests that 2720Ma metamorphism can be extended into the southern Kapisillit terrane along Ameralla Fjord. A revised summary of the tectonothermal evolution of the Nuuk region, based on established and new data, is presented in Table 7.1. The following section places the interpretations of this study in a regional context and discusses the implications for the evolution of the Nuuk region.

The amphibolite facies metamorphism and deformation at ca. 2720Ma in the southern part of the Nuuk region has been attributed to the northward thrusting of the Tasiussaruaq terrane over the already amalgamated Færingehavn and Tre Brødre terranes in a number of papers (Crowley, 2002; Friend and Nutman, 1991; Friend and Nutman, 2005; Friend et al., 1996; Friend et al., 1987; McGregor et al., 1991; Nutman et al., 1989; Nutman et al., 2004). However a high grade metamorphic event at 2720Ma has not previously been recognised in the Kapisillik terrane. This prompted a number of studies to suggest that the two events are mutually exclusive (Friend and Nutman, 2005; Nutman et al., 2007a; Nutman and Friend, 2007). New
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Orthogneiss emplacement

Deposition of supracrustal sequence

Metamorphism/deformation

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evidence from this study suggests the presence of ca. 2720Ma upper amphibolite facies metamorphism in the south of the Kapisillik terrane, the effect of which apparently decreases moving northwards towards Kapisillit village. Preliminary U/Pb data from the granulite area immediately south of Tummeralik (Fig. 7.5) yields metamorphic ages of ca. 2795Ma (Solgevik, 2005), which would be compatible with the age of granulite facies metamorphism in the Tasiusarsuaq terrane (Crowley, 2002; Friend and Nutman, 2001), suggesting the close proximity of the purported northern edge of the Tasiusarsuaq terrane. It has been established that M1-D1 and M2-D2 reflect metamorphism at ca. 2720Ma, which is correlated with Tasiusarsuaq thrusting at ca. 2720Ma (see above references). This interpretation implies that the Færingehavn-Tre Brødre and Kapisillik blocks had docked some time prior to ca. 2720Ma but post ca. 2825Ma, in agreement with the interpretations of (Nutman and Friend, 2007; Polat et al., 2007) and in line with the correlations between Norsanna, Ameralla Fjord and Tummeralik presented in Chapter 5. The timing of juxtaposition between the Færingehavn-Tre Brødre and Kapisillik blocks has yet to be identified, however Nutman and Friend (2007) suggest vestigial evidence for a 2767Ma metamorphic event in Inner Ameralik. In terms of structures, the approximately E-W trending folds and fabrics in Tummeralik are compatible with the N-S convergence associated with the Tasiusarsuaq collision. It may also be argued that the thrusting of the Tasiusarsuaq terrane over pre-amalgamated Færingehavn-Tre Brødre and Kapisillik terranes could also explain the difficulty in tracing the boundary south of Ameralik Fjord. This last suggestion would however require further work to substantiate.

The high grade metamorphism and tectonism associated with the 2720Ma event is considered to represent the final juxtaposition of terrane blocks in the Nuuk region (Friend et al., 1987; Friend et al., 1988). However a large amount of evidence exists to suggest that tectono-thermal activity did not end at this time. Upper amphibolite facies metamorphism between 2650 and 2635Ma is well documented in the Kapisillik terrane. 2635Ma metamorphism is most noted on the island of Storø, where upper amphibolite facies shear zones rework the purported southern boundary of the Akia terrane (Garde, 1997; Garde et al., 2000; Kelly and Frei, 2004; Nutman et al., 2007b). Later thermal pulses recorded in the zircon and monazite record are also
interpreted to reflect post-amalgamation reactivation, reworking and deformation along previously established boundaries between juxtaposed crustal blocks. Such continuing tectonism is considered a viable interpretation, given the significant level of terrane boundary reworking observed in modern-day terrane complexes, such as the North American Cordillera. In addition, the mylonitic terrane boundaries common in the Nuuk region may arguably represent lower crustal analogues of the strike-slip faults that accommodate the oblique collision that forms the driving force for terrane migration (Coney et al., 1980; Monger et al., 1982). If this were the case, then the mylonitic terrane boundaries may not represent the original collisional boundaries associated with terrane accretion, but rather reflect the approximate position of the original boundaries, which were subsequently overprinted by strike-slip motion. Such an analogue may be proposed for the southern boundary of the Akia terrane, whose original position is thought to reside in a high grade strain partitioned deformation zone on the islands of Bjørneøen and Semitsiaq (Kelly and Frei, 2004). However this boundary is overprinted by the 2635Ma Storø High Strain Zone (Nutman et al., 2007b) and even later by the Proterozoic Ivinnguit Fault (Hollis et al., 2004), which has been linked to the greenschist facies faulting and retrogression in the Kapisillit region.


CHAPTER 8
CONCLUSIONS AND FURTHER WORK

8.1 Conclusions

The project goals set out in Chapter 1 aimed to investigate the tectonothermal evolution of the Kapisillit region by integrating structural, metamorphic and geochronological data from three field areas spread across the entire Kapisillit region. This has been achieved through a multidisciplinary study involving field mapping, detailed mineralogy and petrology, geochemistry and zircon and monazite geochronology.

Chapters 2 and 3 described detailed field relations and the structural evolution of Norsanna, Tummeralik and the Ice Lake reconnaissance area. Chapter 4 confirmed previous interpretations that Norsanna and Tummeralik contain orthogneisses with emplacement ages compatible with the Amitsaq-Ikkattoq gneisses (Færingehavn-Tre Brødre) and Kapisillik gneisses respectively. In addition, metamorphic zircon was shown to develop in units belonging to the Færingehavn, Tre Brødre and Kapsillik terranes in the vicinity of Ameralik and Ameralla Fjords. Younger metamorphic ages were obtained from Tummeralik but treated as minimum ages of metamorphism due to significant metamictisation and alteration in zircons from these samples. Preliminary monazite data from Norsanna also suggested the presence of at least 2 thermal pulses at \( \leq 2650 \) Ma. Data from chapters 2, 3 and 4 suggest that Norsanna and Tummeralik share a common metamorphic evolution at 2720 Ma. At this time both areas underwent progressive km-scale isoclinal folding and shearing and attenuation, after which Norsanna was affected by a phase of open folding that did not extend to Tummeralik. The structural history of Aputitooq Mountain is markedly different to Norsanna and Tummeralik, showing isoclinal folding and thrusting at upper amphibolite facies, followed by SE-directed folding. In addition, the age of metamorphism in Aputitooq Mountain has been constrained at ca. 2650 Ma (Bennett, Pers. Comm.), indicating a different P-T-t-deformation history for the northern Kapisillit region.
Chapter 6 described the metamorphic evolution of pelitic units from Norsanna, Tummeralik and Aputitooq Mountain. Pelitic samples are invariably depleted in K and/or enriched in Fe and Mg, causing the non-development of prograde and peak muscovite and K-feldspar. This feature is considered a function of protolith chemistry, where the relative depletion on K with respect to Fe and Mg is caused by input from a mafic (possibly volcanic) source and is in keeping with the active margin environment inherent in a terrane accretion model.

A well defined clockwise P-T-deformation path at 2720Ma was deduced for samples from Norsanna and Tummeralik, with prograde metamorphism in the kyanite zone interpreted to occur contemporaneously with isoclinal folding. Peak metamorphism in the upper amphibolite facies (650-700°C at ca. 7kbar) was attained prior to shearing and attenuation and followed by the development of cordierite. The tectonic environment that led to the development of post-peak cordierite in these two areas remains uncertain, as apparent normal sense motion on the NHSZ may indicate exhumation and orogenic collapse following the thermal maximum. However cordierite may equally be associated with later high grade metamorphic events in the Kapisillit region. Ca. 2720Ma metamorphism and deformation in the central Kapisillit region is correlated with the Tasiusarsuaq collision from the (present-day) south, which is associated with regional upper amphibolite facies metamorphism in the southern Færingehavn and Tre Brødre terranes. This event has been previously suggested to represent the final phase of collision in the amalgamation of the terrane collage and is the preferred interpretation for this work. However this study extends the influence of the 2720Ma event into the Kapisillik terrane, suggesting that the Færingehavn-Tre Brødre and Kapisillik terranes were juxtaposed prior to 2720Ma.

The prograde P-T-deformation path in Aputitooq Mountain took place in the kyanite field and was also followed by sillimanite-zone peak metamorphism at ca. 2650Ma and post-peak cordierite growth. The development of cordierite in all three areas is considered a result of metamorphism at ca. 2650Ma. In Aputitooq Mountain, this may be a response to decompression. In Norsanna and Tummeralik it is interpreted as a later lower pressure overprint that is not related to peak metamorphism at 2720Ma. Evidence for significant thermal overprints in Norsanna at ≤2650Ma are suggested by polygenetic monazite ages however this hypothesis
requires further testing. The multiple tectonothermal events in the Kapisillik (and wider Nuuk) regions post-date the interpreted final amalgamation event at 2720Ma. Tectonism clearly continued after 2720Ma, perhaps due to reworking of pre-established boundaries. The continuing strike slip motion on the North American Cordillera is cited as an example of post-amalgamation tectonism in a terrane complex.

Greenschist facies retrogression is present in all areas but most pronounced in Norsanna and Tummeralik. Chlorite grade retrogression is focussed around mylonite zones that are roughly contemporaneous with the intrusion of doleritic to picritic dykes that are correlated with a Proterozoic dyking event that affects a large proportion of West Greenland. Greenschist facies shear zones in Norsanna are correlated with the Kobbefjord Faulting event at 2000-1800Ma. Retrograde muscovite in pelitic samples forms from the greenschist facies retrogression of biotite and sillimanite, where K is liberated from the breakdown of biotite to chlorite. Previous mapping had identified chlorite grade retrogression in the immediate vicinity of greenschist shear zones, however this study has shown retrogression to be present on microscopic scales in virtually all the pelitic samples. In addition, the severe metamictisation and alteration of zircon samples is Tummeralik is attributed to the sample localities’ proximities to a major greenschist facies lineament.

The work presented in this thesis provides a more detailed study of the 2720Ma and 2650Ma events in the Kapisillit region. It also provides insight into the initial bulk compositions of pelitic units and their possible origin as sourced from mafic volcanics or other similar mafic source. The initial bulk composition depleted in K-Al bearing phases is also cited to suggest that reactions involving the breakdown of muscovite to K-feldspar are not applicable to these samples. This exemplifies the importance of detailed petrology prior to the selection of suitable thermobarometric techniques. Similarly, a good understanding of the P-T history of an area is also important in the interpretation of geochronological data in high grade terrains.
8.2 Recommendations for further work

Pseudosections were not constructed during this study due to time constraints. As such, the P-T histories of pelitic samples were estimated using comparisons with published P-T grids and pseudosections. However the atypical pelitic bulk compositions in the Norsanna, Tummeralik and Aputitooq Mountain set severe limits on the applicability of average pelite P-T grids and pseudosections. In addition, the absence of prograde and peak K-Al phases also limits the number of suitable published pseudosections and as such there were fewer grids against which to compare samples. The estimation of metamorphic conditions was also hampered by the inapplicability of GASP and GRAIL in a number of samples due to inappropriate mineral assemblages, low-Ca or subsequent alteration. It is thus apparent that the P-T history of the Kapisillit region metasediments cannot be accurately constrained without constructing pseudosections for the individual samples.

Major element bulk rock analyses exist for all aluminosilicate and orthoamphibole-bearing assemblages studied in this work. In addition to pseudosection construction, further geochemical analysis of the samples may be undertaken to establish the source of the K-depletion and Fe-Mg enrichment. New powders may also be processed for trace element analysis and compared with studies of Garde (2007) and Polat et al. (2007) for the Qussugg and Ivissartoq supracrustal belts respectively.

Further geochronology of orthogneisses and leucosomes are vital in Tummeralik and the surrounding areas. The relationship between the tonalitic orthogneiss in Tummeralik and the Ameralla Fjord orthogneiss needs to be established, as this relationship may either confirm or negate the correlations between Norsanna and Tummeralik at 2720Ma. Targeted sampling of leucosome material in the central and eastern Kapisillit area should be undertaken to characterise the spatial distribution of the 2720 vs. 2650Ma metamorphic events and to establish the dominant metamorphic event in Tummeralik. It is also recommended that detailed dating studies of the area surrounding the greenschist facies lineaments in Tummeralik should be carried out, in order to ascertain the effect of greenschist retrogression on the U/Pb systematics in zircon.
It is suggested that detailed interpretations should be carried out on the preliminary monazite age data presented in this work, as chemical dates from monazite may be used to accurately constrain a point on the P-T path of Norsanna. The Tummeralik samples did not appear to yield as much monazite in the pelitic sample, however it is noted that monazite in Norsanna developed in aluminosilicate-absent rocks. It is consequently suggested that if monazite work is extended across the Kapisillit region (and the wider Nuuk region), sampling should perhaps be focussed on semi-pelitic samples. Ultimately, accurate timing constraints on P-T paths from monazite could be vital in correlating metamorphic events across the Kapisillik and wider area, and so it is recommended that more emphasis should be placed on monazite geochronology in future investigations.

Appendix A

Analytical Techniques and Methodologies
LA-SF-ICP-MS methodology: U/Pb zircon geochronology

Zircon U-Pb age dating was carried out at the Geological Survey of Denmark and Greenland (GEUS) in Copenhagen using laser ablation – magnetic sector field – inductively coupled plasma – mass spectrometry. The LA-SF-ICP-MS system employed at GEUS consists of a NewWave Research/Merchantek UP213 laser ablation system equipped with a frequency quintupled Nd-YAG laser emitting at a wavelength of 213 nm coupled to an Element2 (ThermoFinnigan, Bremen) single-collector double focusing magnetic sector field ICP-MS equipped with a fast field regulator for increased scanning speed.

The nominal pulse width of the laser is 5 ns with a pulse-to-pulse stability of 2 % RSD. The laser was operated at a repetition rate of 10 Hz and a nominal energy output of 44 %, corresponding to a laser fluency of 8J cm⁻². All data were acquired with a single spot analysis on each individual zircon grain with a beam diameter of 30 µm. For this spot diameter and the ablation time used (40 s) the ablated masses of zircon were approximately 200-500ng. Samples and standards were held in the standard ablation cell delivered with the UP 213 system. Helium was used to flush the sample cell and was mixed downstream with the Ar sample gas of the massspectrometer. The washout time for this configuration is approximately 3 min. All sample mounts were rigorously cleaned before introduction into the sample cell to remove surface Pb contamination.

The total acquisition time for each analysis was 70 s with the first 30 s used to measure the gas blank, followed by 40 s ablation on each zircon grain. The mass spectrometer was tuned to give large, stable signals for the \(^{206}\text{Pb}\) and \(^{238}\text{U}\) peaks, low background count rates (typically < 50 counts per second for \(^{207}\text{Pb}\) and < 10 counts per second for \(^{238}\text{U}\)) and low oxide production rates (\(^{238}\text{U}^{16}\text{O}/^{238}\text{U}\) below 2.5 %). All measurements were performed in low resolution mode using electrostatic scanning (E-scan) with the magnetic field resting at mass \(^{202}\text{Hg}\). The following masses were measured: \(^{202}\text{Hg}, ^{204}(\text{Pb + Hg}), ^{206}\text{Pb}, ^{207}\text{Pb}, ^{208}\text{Pb}, ^{232}\text{Th}, ^{235}\text{U}, \text{and} ^{238}\text{U}\). All Data were acquired on four samples per peak with a sampling and a settling time of 1 ms.
for each isotope. Mass $^{202}\text{Hg}$ was measured to monitor the $^{204}\text{Hg}$ interference on $^{204}\text{Pb}$ (using a $^{202}\text{Hg}/^{204}\text{Hg}$-ratio of 4.36). Only if the net intensities for mass $^{204}\text{Pb}$, corrected for $^{204}\text{Hg}$, are significantly above the limit of detection a common Pb-correction was performed. The laser induced elemental fractionation and the instrumental mass bias on measured isotopic ratios was corrected by matrix-matched external standardisation using the GJ-1 zircon standard (Jackson et al. 2004). Samples were analysed in sequences where three standards are analysed initially, followed by ten samples, again three standards, and so on. The Plisovice zircon standard (ID-TIMS age 338 ± 1 Ma; Aftalion et al. 1989, provided by Jan Košler, Charles University, Prague) was analysed as an unknown regularly. The results are consistently concordant at 340 ± 2 Ma.

The raw data were exported in ASCII format and processed using in-house data reduction spreadsheets. Final age calculations were also done using in-house age-calculation spreadsheets. The computer program IsoplotEx v. 3.0 (Ludwig 2003) was used to carry out the final computation of ages.

Electron microprobe: Mineral geochemistry and X-ray mapping

A number of different instrument conditions were used to produce both quantitative and qualitative data. Polished thin sections (2x3cm) of all pelitic samples and the amphibole-bearing samples were analysed using EPMA analysis. The mineralogy and textures present in all samples were examined using optical and SE microscopy prior to analysis. Mineral analyses of selected silicate and oxide samples were obtained using the Cameca SX100 electron microprobe facility at the School of Geosciences, University of Edinburgh. The ten major elements plus P were analysed for major and selected accessory (ilmenite and rutile) minerals, using as standards jadeite (NaKα), spinel (MgKα, AlKα), wollastonite (SiKα, CaKα), orthoclase (KKα), rutile (TiKα), apatite (PKα) and pure manganese (MnKα), iron (FeKα) and chromium (CrKα). A separate setup file was used for hydrous minerals, which analysed the same elements plus FKα and CIKα, with standard values measured on BaF₂ and NaCl respectively. An accelerating voltage of 15KeV and beam current of 20nA were used, with a static beam size of 10μm. Biotite and plagioclase used a rastered beam to avoid ion migration during analysis. Geochemical data was
interpreted in Excel using the method described in Deer et al. (1992), with $\text{Fe}^{2+}$ estimated for garnet using the equation from Droop (1987). Biotite end members were calculated from Powell (1978), with $\text{Fe}^{3+}$ estimated on a site filling basis modified from Droop (1987).

Representative garnets (or sections of garnets) from each pelitic sample were selected for analysis using the qualitative X-ray mapping function, in addition to the qualitative analysis described above. Two square or rectangular maps (measuring 512*512 or 512*384 pixels) were run for each selected garnet, producing a major element plus Y map to establish internal zoning and growth patterns. A second map was run to identify the principal inclusion minerals and their spatial distribution within the garnet. The major element setup analysed Ka on Ca, Mg, Fe, Mn and Y, with an accelerating voltage of 15KeV and beam current of 200nA. Dwell times were typically 0.05sec per pixel with a step size of ~5µm where, possible, although this varies from sample to sample depending on the size of the analysed garnet (stage constraints required a larger step size on large maps). Inclusion maps analysed for Si (Kα), Ce (La), Ti (Kα), Al (Kα) and Zr (La), and identified quartz, monazite, rutile (differentiated from ilmenite by comparison with the Fe map), aluminosilicate and zircon respectively. The inclusions maps were run over the same area as the major element maps and used the same column and beam conditions.

Large X-ray maps measuring approximately 4cm$^2$ were obtained from all aluminosilicate-bearing samples in an attempt to identify K-feldspar and muscovite and, where present, to assess the textural equilibrium of the two minerals. The setup was thus required to be able to differentiate between muscovite, K-feldspar, plagioclase and biotite. The elements Al (Kα), Fe(Kα), K(Kα), Mg(Kα) and Na(Kα) were analysed using an accelerating voltage of 15KeV and 50nA on a dwell time of 0.04 seconds. The large size of the maps required a 2*2 mosaic square (256*256 pixels per square) at a step size of 35µm,
Suitable grains were first identified using BSE imagery on the Philips XL30CP SEM at the School of Geosciences, University of Edinburgh. X-ray maps of selected monazite grains were obtained prior to quantitative analysis, in order to identify internal zoning patterns and target specific areas for analysis. X-ray maps analysed for Ce (La), Th (Ma), U (Mβ), Pb (Mβ) and Y (La), with an accelerating voltage of 20KeV and beam current of 200nA. Maps were acquired using a dwell time of 0.1 seconds with a step size of 2μm on a 512×512 pixel grid.

Standards used for quantitative analysis were Spinel (AlKα), Durango apatite (PKα), hematite (FeKα), La-REE glass (LaLα), Ce-REE glass (CeLα), Pb-silicate (SiKα), Pb-silicate (PbMβ), celestite (SKα, SrLα), K-tantalite (KKα), Y-oxide (Ya), Pr-REE glass (Pr Lβ), Th-oxide (ThMα), U-oxide (UMβ), Nd-REE glass (NdLβ), Sm-REE glass (SmLβ), Gd-REE glass (GdLβ), Dy-REE glass (DyLβ) and Yb-REE glass (YbLα). Two column conditions were set, with Al, P, Ca, Fe, La, and Ce analysed with an accelerating voltage of 20KeV and beam current of 20nA, whereas Si, Pb, S, K, Sr, Pb, Y, Pr, Th, U, Nd, Sm, Gd, Dy and Yb were analysed with an accelerating voltage of 20KeV and a beam current of 100nA.

Monazite compositions were calculated using Anders and Grevasse (1989), with chemical ages calculated using a version of AgeCalc (Berry, University of Tasmania) with modifications by David Steele at the University of Edinburgh. Error calculations for chemical ages used the combined methods of Pyle et al. (2005) and Reed (1986).


Appendix B

U/Pb zircon geochronology dataset
Table 1: U/Pb analytical data for all analyses obtained from LA-ICPMS dating of zircon. Red text denotes rejected analyses that were not used in age calculations but are displayed on concordia diagrams as black, unfilled ellipses.

<table>
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<tr>
<th>Sample</th>
<th>U (ppm)</th>
<th>Pb (ppm)</th>
<th>$T^k$</th>
<th>$U$</th>
<th>$T^f$</th>
<th>$T^g$</th>
<th>$T^i$</th>
<th>$T^j$</th>
<th>$T^k$</th>
<th>$T^l$</th>
<th>$T^m$</th>
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<th>$T^v$</th>
<th>$T^w$</th>
<th>$T^x$</th>
<th>$T^y$</th>
<th>$T^z$</th>
<th>% Conc.</th>
</tr>
</thead>
</table>

482426 - Polyphase orthogneiss, Norsanna (UTM 22W 513880E, 7124443N)

Population 1 - Magmatic: Oscillatory zoned rounded cores or grains with pyramidal end terminations. Strong CL response.

| 4 | 78 | 38 | 0.55 | 0.00004 | 37.75 | 1.3 | 0.78230 | 1.2 | 0.93 | 0.35000 | 0.5 | 3713 | 99 | 3725 | 92 | 3707 | 14 | 100.3 |
| 9 | 414 | 0.49 | 0.00021 | 12.31 | 8.4 | 0.31020 | 8.4 | 1.00 | 0.28790 | 0.5 | 2629 | 442 | 1742 | 292 | 3406 | 16 | 33.7 |
| 12 | 203 | 80 | 0.44 | 0.00004 | 29.76 | 1.8 | 0.69369 | 1.6 | 0.91 | 0.31112 | 0.7 | 3479 | 122 | 3397 | 108 | 3526 | 23 | 2.4 |
| 17 | 45 | 21 | 0.73 | 0.00010 | 35.83 | 2.4 | 0.78450 | 2 | 0.85 | 0.33130 | 1.3 | 3662 | 173 | 3733 | 149 | 3623 | 38 | 102 |
| 20 | 52 | 21 | 0.38 | 0.00067 | 31.13 | 2.6 | 0.67050 | 2.5 | 0.95 | 0.33680 | 0.8 | 3523 | 185 | 3308 | 166 | 3648 | 24 | 93.9 |
| 21 | 334 | 74 | 0.87 | 0.00025 | 13.46 | 2.5 | 0.33042 | 2.4 | 0.95 | 0.29540 | 0.8 | 2712 | 138 | 1840 | 89 | 3446 | 25 | 32.1 |
| 24 | 204 | 84 | 0.23 | 0.00005 | 31.08 | 1.3 | 0.70522 | 1.2 | 0.97 | 0.31959 | 0.3 | 3521 | 90 | 3440 | 86 | 3568 | 9 | 2.3 |

Population 2 - Metamorphic/altering magmatic: Patchy to sector zoned sub-ovoid grains or rims on Population 1. Moderate to weak CL response.

| 5 | 246 | 100 | 0.26 | 0.00000 | 30.25 | 1.3 | 0.69940 | 1.2 | 0.97 | 0.31370 | 0.3 | 3495 | 89 | 3418 | 84 | 3539 | 10 | 97.8 |
| 6 | 279 | 0.25 | 0.00000 | 29.38 | 1.3 | 0.67340 | 1.2 | 0.96 | 0.31650 | 0.4 | 3466 | 90 | 3319 | 83 | 3553 | 12 | 95.7 |
| 7 | 235 | 60 | 0.2 | 0.00002 | 12.52 | 1.4 | 0.48820 | 1.3 | 0.96 | 0.18600 | 0.4 | 2644 | 72 | 2563 | 67 | 2707 | 13 | 96.9 |
| 8 | 339 | 84 | 0.18 | 0.00000 | 12.19 | 1.7 | 0.47460 | 1.7 | 0.98 | 0.18630 | 0.4 | 2561 | 99 | 2504 | 85 | 2710 | 12 | 95.6 |
| 10 | 466 | 130 | 0.55 | 0.00009 | 19.06 | 2.2 | 0.48178 | 2.2 | 0.97 | 0.28700 | 0.5 | 3045 | 137 | 2535 | 111 | 3401 | 16 | 83.2 |
| 11 | 269 | 70 | 0.35 | 0.00000 | 12.63 | 2.4 | 0.49200 | 2.3 | 0.99 | 0.18590 | 0.4 | 2563 | 125 | 2583 | 120 | 2707 | 13 | 97.4 |
| 12 | 202 | 80 | 0.44 | 0.00004 | 29.76 | 1.8 | 0.69370 | 1.6 | 0.91 | 0.31110 | 0.7 | 3479 | 122 | 3397 | 108 | 3526 | 23 | 97.6 |
| 13 | 271 | 115 | 0.4 | 0.00001 | 31.04 | 1.6 | 0.70190 | 1.5 | 0.94 | 0.32070 | 0.5 | 3520 | 113 | 3428 | 104 | 3573 | 16 | 97.4 |
| 18 | 103 | 38 | 0.33 | 0.00000 | 26.12 | 1.5 | 0.64310 | 1.2 | 0.82 | 0.29460 | 0.9 | 3351 | 102 | 3201 | 97 | 3442 | 27 | 95.5 |
| 19 | 291 | 94 | 0.23 | 0.00005 | 22.46 | 3.9 | 0.58683 | 3.8 | 0.98 | 0.27672 | 0.7 | 3204 | 249 | 2984 | 228 | 3345 | 22 | 93.1 |
| 22 | 246 | 100 | 0.17 | 0.00000 | 30.39 | 1.3 | 0.71400 | 1.2 | 0.96 | 0.30870 | 0.4 | 3499 | 89 | 3473 | 84 | 3514 | 11 | 99.3 |
| 24 | 204 | 84 | 0.23 | 0.00005 | 31.08 | 1.3 | 0.70520 | 1.2 | 0.97 | 0.31960 | 0.3 | 3521 | 90 | 3440 | 86 | 3568 | 9 | 97.7 |
| 26 | 382 | 100 | 0.97 | 0.00026 | 16.71 | 5.4 | 0.43322 | 5.3 | 0.99 | 0.27972 | 0.8 | 2918 | 316 | 2320 | 248 | 3361 | 26 | 79.5 |

Population 3 - Recrystallised magmatic: Stubby to acicular grains with faint patchy zoning. Moderate to strong CL response.

| 23 | 263 | 107 | 0.35 | 0.00018 | 28.87 | 2.0 | 0.70034 | 1.8 | 0.91 | 0.29899 | 0.8 | 3449 | 137 | 3422 | 124 | 3465 | 25 | 99.2 |
| 25 | 349 | 123 | 0.31 | 0.00018 | 26.31 | 1.7 | 0.61160 | 1.5 | 0.9 | 0.31200 | 0.7 | 3358 | 112 | 3076 | 93 | 3531 | 23 | 91.6 |
482424 - Homogeneous orthogneiss, Norsanna (UTM 22W 514137E, 7142704N)

Population 1 - Magmatic/ altered magmatic: Stubby to acicular oscillatory zoned grains. Zoning may be blurred of locally recrystallised. No CL data.

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<th>Pb (ppm)</th>
<th>( T_i ) (Ma)</th>
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<th>( 20^\text{e} ) %</th>
<th>( 1^\sigma ) %</th>
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481418 - Tonalitic/granodioritic orthogneiss, Ameralla Fjord (UTM 22W 526575E, 7119594N)

Population 1 - Magmatic cores: Oscillatory zoned cores or equant to acicular grains with rounded pyramidal end terminations. Strong CL response.

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Population 2 – Metamorphic: Faint concentric (potentially relict oscillatory) zoning in stubby to elongate whole grains or mantles on Population 1. Weak to moderate CL response.
Inherited core: Blurred and convoluted oscillatory zoned cores with some patchy zoning. Strong CL response.

Sample | Population | 3 | 7 | 44 | 59 | 60 | 1 | 18 | 4 | 403 | 649 | 354 | 379 | 106 | 123 | 94 | 0.03 | 0.0020 | 13.81 | 1.6 | 0.53083 | 1.1 | 0.68 | 1.8686 | 1.1 | 2737 | 85 | 2745 | 58 | 2731 | 62 | 100.3

482444 - Garnet-bearing felsic leucosome, Norsanna (UTM 22W 517988E. 7122573N)


Sample | Population | 36 | 121 | 39 | 0.67 | 0.0000 | 15.62 | 1.7 | 0.55988 | 1.6 | 0.94 | 0.20239 | 0.6 | 2854 | 95 | 2866 | 89 | 2846 | 9 | 100.4

Population 2 – Subovoid: Sector zoned and planar banded stubby to sub-ovoid grains. Weak to moderate CL response.

Sample | Population | 30 | 824 | 217 | 0.02 | 0.00000 | 13.58 | 1.2 | 0.51933 | 1.2 | 0.98 | 0.18966 | 0.3 | 2721 | 67 | 2696 | 65 | 2739 | 4 | 99.1

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<th>p</th>
<th>207</th>
<th>1 σ</th>
<th>207/235</th>
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| 481438 - Garnet-bearing felsic leucosome, Norsanna (UTM 22W 515251E, 7121289N) |

Population 1 – Cores: Faint planar banded stubby to elongate cores. Weak CL response.
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**Population 2 – Inner rims:** Planar or sector zoned mantles on Population 1. Stubby to elongate morphology. Moderate CL response.

**Population 3 – Outer rims:** Faint planar banded rims and pyramidal terminations on Population 2. Weak CL response.
496831 - Tonalitic orthogneiss, Tummeralik (UTM 22W 561167E, 7125753N)


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Population 2 - Recrystallised magmatic/metamorphic: Planar banded or patchy zoned stubby to elongate mantles on Population 1. Weak to very weak CL response.
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<th>Pb (ppm)</th>
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<th>$^{204} \text{Pb} / ^{206} \text{Pb}$</th>
<th>$^{206} \text{Pb} / ^{238} \text{U}$</th>
<th>$^{207} \text{Pb} / ^{206} \text{Pb}$</th>
<th>$^{206} \text{Pb} / ^{235} \text{U}$</th>
<th>$\rho$</th>
<th>$^{207} \text{Pb} / ^{206} \text{Pb}$</th>
<th>$^{208} \text{Pb} / ^{206} \text{Pb}$</th>
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**Population 3 - Recrystallised rims:** Faint planar banded rims with sinuous fronts overgrowing Population 2. Weak to moderate CL response.

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**478334 - Garnet-bearing felsic leucosome, Tummeralik (UTM 22W 560155E, 7124217N)**

**Population 1 - Leucosome crystallisation:** Faint planar banding or patchy zoning in ovoid to stubby grains. Very weak CL response.

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<tr>
<td>9</td>
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Note: Small negative values for $^{204}\text{Pb}/^{206}\text{Pb}$ ratios are an artefact of overcompensation arising from the peak interference of $^{204}\text{Hg}$ on $^{204}\text{Pb}$. In such cases, the amount of $^{204}\text{Pb}$ in an analysis is less than the magnitude of $^{204}\text{Hg}$ interference, resulting in an apparent negative $^{204}\text{Pb}$ value.
Appendix C

Mineral geochemistry and Geothermobarometric datasets
Table 1: Comparative assemblages of framework minerals pelitic and semi-pelitic samples in Norsanna, Tummeralik and Aputitooq Mountain. p = porphyroblast; m = matrix; l = leucosome; inc_c = inclusion in garnet core; inc_r = inclusion in garnet rim; ps = (partially) pseudomorphed phase; rsil = replacing sillimanite (or other mineral); v= mineral in vein/fracture; acc = accessory mineral; symord = forming symplectic intergrowth with cordierite.

<table>
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<th>Aputitooq Mountain</th>
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Table 2: Comparative assemblages of accessory minerals pelitic and semi-pelitic samples in Norsanna, Tummeralik and Aputitoq Mountain. p = porphyroblast; m = matrix; l = leucosome; inc_e = inclusion in garnet core; inc_r = inclusion in garnet rim; ps = (partially) pseudomorphed phase; rsil = replacing sillimanite (or other mineral); v = mineral in vein/fracture; acc = accessory mineral in garnet and matrix; sym_interd = forming symplectic intergrowth with cordierite.

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Table 3a: Representative EPMA analyses of garnet from high grade pelitic rocks in the Kapisillit region.

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Table 3b: Representative EPMA analyses of plagioclase from high grade pelitic rocks in the Kapisillit region.

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Table 3b cont: Representative EPMA analyses of plagioclase from high grade pelitic rocks in the Kapisillit region.

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Table 3c: Representative analyses of biotite from high grade pelitic rocks in the Kapisillit region.

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Table 3c cont.: Representative analyses of biotite from high grade pelitic rocks in the Kapisillit region.

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Table 4: Bulk compositional XRF data for samples from Norsanna, Tummeralik and Aputitooq Mountain. A/AFM is projected from muscovite.

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Table 5: Garnet and ilmenite analyses used for GRAIL barometry (after Bohlen, 1983)

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Appendix D

Kapisillit Mapsheet

Permission to include the 2007 draft of the Kapisillit Mapsheet has been granted by the Head of Geological Mapping at the Geological Survey of Denmark and Greenland (GEUS). A copy of the map is included in the back sleeve.

Compiled by J.A. Hollis, S. Schmidt and E. Rehnström.