

“Metamorphic evolution of relict eclogite-facies rocks in the Paleoproterozoic
Nagssugtoqidian Orogen, South-East Greenland”

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Abstract

The Nagssugtoqidian Orogen in the Tasiilaq region of South-East Greenland, is a roughly southeast-northwest trending, ~200 km wide Paleoproterozoic collisional orogen. It predominantly consists of a variety of Archean and Paleoproterozoic rocks, which were variably affected by several stages of deformation and metamorphism. The orogen has been subdivided into three different terranes. From north to south, these are the medium-pressure granulite-facies Schweizerland Terrane, the mainly high-pressure amphibolite-facies Kuummiut Terrane and the low-pressure amphibolite-facies Isertoq Terrane with the calc-alkaline Ammassalik Intrusive Complex. In the Kuummiut Terrane, variably retrogressed high-pressure mineral assemblages are preserved within mafic dykes, as well as boudins and boudinaged layers of mafic to ultramafic supracrustal rock in TTG gneiss. These mineral assemblages had previously not been investigated in much detail, but may provide important insight into the conditions of subduction, metamorphism and exhumation in the Paleoproterozoic, an era from which the geothermal regimes and nature of plate-tectonic processes are not well understood and high-pressure rocks are scarce. Within the framework of a larger expedition program to South-East Greenland, this thesis investigated the mineral textural evolution, PT-path and geochronology of variably retrogressed eclogite-facies rocks of the Kuummiut Terrane, via detailed petrological and mineral textural analysis, bulk-rock and mineral chemistry, conventional geothermobarometry, pseudosection modelling, and U-Pb isotopic dating.

Well-equilibrated high-pressure and amphibolite-facies mineral assemblages, with only minor replacement textures, were found in garnet-pyroxenite, garnet-amphibolite and garnet-kyanite schist. Retrogressed eclogite, in contrast, is characterized by complex mineral reaction textures and the formation of two chemically and mineralogically distinct domains, leading to domainal equilibration volumes. A clinopyroxene domain is dominated by a fine-grained, worm-like to globular, diopside-plagioclase symplectite, which is often intergrown with and partially replaced by a coarser-grained hornblende-plagioclase symplectite. The fine-grained symplectite is interpreted to have grown at the expense of omphacite, which is only preserved in a Na-rich retrogressed eclogite sample. In a garnet domain, coarse-grained garnet is surrounded and variably pseudomorphed by corona-textured plagioclase $\pm$ amphibole $\pm$ clinopyroxene $\pm$ orthopyroxene.

Geothermobarometry and pseudosection modelling, in combination with the various mineral textures, reveal evidence for four metamorphic stages along a clockwise PT-path. Ca-rich cores of large garnet grains in retrogressed eclogite are interpreted as prograde in origin and yield PT-conditions of 14-19 kbar and 600-750 °C (I), followed by eclogite-facies metamorphism at 17-19 kbar and 740-810 °C (II). The retrograde PT-evolution is initially characterized by near-isothermal decompression to high-pressure granulite-facies conditions of 13.8-15.4 kbar and 760-880 °C (III), with subsequent decompression and minor cooling to high-pressure amphibolite-facies conditions of

8.8-10.9 kbar and 660-840 °C (IV). The PT-path implies that the Kuummiut Terrane probably experienced an initially rapid, tectonically driven exhumation. Furthermore, a large degree in consistency in the PT-data for the variably retrogressed high-pressure rocks suggests that they underwent the same metamorphic history, with the degree of retrogression and type of replacement assemblage mainly being controlled by fluid activity.

LA-SF-ICP-MS U-Pb dating was carried out on zircon, monazite, titanite and rutile from retrogressed eclogite and the garnet-kyanite schist. A large range in $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 2634 ± 63 and 1617 ± 91 Ma has been obtained, of which zircon yields the oldest and rutile the youngest dates. Detrital zircon in garnet-kyanite schist gives Archean to Paleoproterozoic dates and confines the maximum deposition of the precursor to the metasediment at 2107 ± 21 Ma. In retrogressed eclogite, the oldest zircon dates at 2146 ± 63 to 2092 ± 22 Ma are proposed to reflect the age of dyke emplacement into the Archean TTG gneiss. Dyke emplacement and deposition of supracrustal rocks most likely occurred near-contemporaneously during Paleoproterozoic extension and basin formation, as indicated by overlapping dates for detrital and magmatic zircon. The remaining data characterize the metamorphic evolution of the Kuummiut Terrane, about 200 m.y. after dyke emplacement. Consistent metamorphic dates between 1891 ± 10 and 1882 ± 3 Ma are obtained from the majority of the zircon, monazite and titanite analyses. The REE and U-Pb systematics in zircon indicate decoupling during retrograde metamorphism, with the REE patterns typical of zircon growth at eclogite-facies conditions, whereas the U-Pb dates and microtextures indicate recrystallization at high-pressure amphibolite-facies conditions. Based on previous studies and a similar decoupling in the magmatic zircons, the dates are interpreted to reflect mineral growth and recrystallization during high-pressure amphibolite-facies metamorphism. Regional medium-pressure amphibolite-facies metamorphism, associated with the collision of the Rae and North Atlantic cratons, is reflected by the 1872 ± 70 to 1821 ± 31 Ma monazite and titanite dates. Afterwards, the Kuummiut Terrane experienced relatively slow erosion-controlled cooling, with only minor thermal perturbations, as identified via rutile cooling ages at 1793 ± 10 Ma, 1738 ± 14 to 1720 ± 12 Ma, 1645 ± 63 Ma and 1617 ± 91 Ma.

This thesis shows that a combination of different datasets from several variably retrogressed lithologies, rather than just examining eclogite-facies mineral assemblages, provides an invaluable tool for the characterization of the tectonometamorphic evolution of a Paleoproterozoic collisional orogen.

Kurzfassung

Das Nagssugtoqidian Orogen in der Tasiilaq Region von Südost-Grönland ist ein ungefähr Südost-Nordwest verlaufendes und ~200 km langes paläoproterozoisches Kollisionsorogen. Es besteht hauptsächlich aus einer Vielzahl archaischer und paläoproterozoischer Gesteine, die während verschiedener Deformations- und Metamorphosephasen unterschiedlich stark überprägt wurden. Das Orogen wird in drei Terrane unterteilt. Dies sind (von Norden nach Süden), das Mitteldruck-granulitfazielle Schweizerland Terran, das hauptsächlich Hochdruck-amphibolitfazielle Kuummiut Terran und das Niedrigdruck-amphibolitfazielle Isertoq Terran, welches den kalk-alkalinen Ammassalik Intrusiv Komplex enthält. Unterschiedlich stark überprägte Hochdruck-Mineralparagenesen sind in mafischen Gängen sowie in Boudins und boudinierten Lagen von mafischen und ultramafischen suprakrustalen Gesteinen in TTG Gneis erhalten. Diese Mineralparagenesen wurden bisher noch nicht ausführlich untersucht, jedoch könnten durch sie wichtige Erkenntnisse über Subduktion, Metamorphose und Exhumierung im Paläoproterozoikum gewonnen werden, einer Ära aus der die geothermischen Regime und der Charakter von plattentektonischen Prozessen nicht sehr gut verstanden werden und aus der Hochdruckgesteine schlecht erhalten sind. Im Rahmen eines größeren Geländeprogramms in Südost-Grönland untersucht diese Studie die Entwicklung der Mineral-Reaktionstexturen, den PT-Pfad, sowie die Geochronologie der unterschiedlich stark überprägten eklogitfaziellen Gesteine des Kuummiut Terran. Hierfür wurden folgende Methoden angewandt; petrologische und textuelle Untersuchungen, Gesamtgesteins- und Mineralchemie, konventionelle Geothermobarometrie, Pseudoschnitt-Modellierungen, und U-Pb Datierung an akzessorischen Mineralen.

Vollständig equilibrierte eklogit- und amphibolitfazielle Mineralparagenesen mit nur wenigen Reaktionstexturen wurden in Granat-Pyroxenit, Granat-Amphibolit und Granat-Disthen Schiefer entdeckt. Retrograd überprägter Eklogit besteht hauptsächlich aus zwei mineralogisch und chemisch unterschiedlichen Domänen, deren Bildung domänenspezifische Gleichgewichte zu Folge hatte. Eine Klinopyroxen-Domäne wird von feinkörnigem, wurmartig bis kugelförmigem Diopsid-Plagioklas Symplektit dominiert, der oft von grobkörnigem Hornblende-Plagioklas Symplektit umgeben ist und verdrängt wird. Der feinkörnige Symplektit stellt ein Verdrängungsprodukt nach Omphazit dar, der nur noch in einer Na-reichen Probe erhalten ist. In der Granat-Domäne wird grobkörniger Granat unterschiedlich stark von umgebendem Plagioklas $\pm$ Amphibol $\pm$ Klinopyroxen $\pm$ Orthopyroxen verdrängt.

Mittels konventioneller Geothermobarometrie, Pseudoschnitt-Modellierungen und den Reaktionstexturen konnte ein PT-Pfad für die metamorphe Entwicklung des Kuummiut Terran erstellt werden. Ca-reiche Granat-Kerne werden als prograde Bildung unter PT-Bedingungen von 14-19 kbar und 600-750 °C interpretiert (I), gefolgt von eklogitfazieller Metamorphose bei 17-19 kbar und 740-

810 °C (II). Darauf folgend kam es zu einer nahezu-isothermalen Dekompression unter Hochdruck-granulitfaziellen Bedingungen (13.8-15.4 kbar und 760-880 °C, III) und einer weiteren Dekompression und geringen Abkühlung unter Hochdruck-amphibolitfaziellen Bedingungen (8.8-10.9 kbar und 660-840 °C, IV). Der PT-Pfad zeigt an, dass das Kuummiut Terran anfangs wahrscheinlich eine relativ schnelle, tektonisch-kontrollierte Exhumierung erfuhr. Eine hohe Übereinstimmung in den PT-Daten lässt außerdem vermuten, dass die verschieden stark überprägten Hochdruckgesteine dieselbe metamorphe Entwicklung erfahren haben, wobei das Ausmaß der retrograden Überprägung und die retrograde Mineralparagenese hauptsächlich von der Fluid-Aktivität kontrolliert werden.

U-Pb Datierungen mittels LA-SF-ICP-MS wurden an Zirkon, Monazit, Titanit und Rutil aus retrograd überprägtem Eklogit und Granat-Disthen Schiefer durchgeführt. Die gewonnenen $^{207}\text{Pb}/^{206}\text{Pb}$ -Daten variieren zwischen 2634 ± 63 und 1617 ± 91 Ma, wobei Zirkon die ältesten und Rutil die jüngsten Daten gibt. Detritischer Zirkon aus dem Granat-Disthen Schiefer hat archaische bis paläoproterozoische Alter und begrenzt die Ablagerung des Metasedimentprotoliths auf 2107 ± 21 Ma. Die ältesten Zirkone aus retrograd überprägtem Eklogit (2146 ± 63 bis 2092 ± 22 Ma) stellen das Alter der Intrusion der mafischen Gänge in den TTG Gneis dar. Die Daten für detritische und magmatische Zirkone überlappen, was auf eine nahezu-zeitgleiche Intrusion und Sedimentation hindeutet. Die übrigen Altersdaten charakterisieren die metamorphe Entwicklung des Kuummiut Terran, ca. 200 Millionen Jahre nach der Intrusion der mafischen Gänge. Der Großteil der Zirkon-, Monazit- und Titanitkörner gibt konsistente Daten zwischen 1891 ± 10 und 1882 ± 3 Ma. Obwohl die SEE-Muster in Zirkon typisch für Wachstum unter eklogitfaziellen Bedingungen sind, ist Zirkon mit retrograden Mineralphasen assoziiert und seine Daten überlappen im Fehler mit denen von amphibolitfaziellstem Titanit. Ähnliche Beobachtungen wurden in früheren Studien und in magmatischem Zirkon gemacht und zeigen an, dass die U-Pb und REE Systeme während der retrograden Metamorphose voneinander entkoppelt wurden. Die $^{207}\text{Pb}/^{206}\text{Pb}$ Daten werden daher als Neuwachstum und Rekristallisation während der Hochdruck amphibolitfaziellen Metamorphose interpretiert. Titanit und Monazit Daten von 1872 ± 70 bis 1821 ± 31 Ma reflektieren regionale amphibolitfazielle Metamorphose-Bedingungen bei mittlerem Druck, die im Zuge der Kollision zwischen dem Rae- und Nordatlantik-Kraton vorherrschten. Im Anschluss erfuhr das Kuummiut Terran eine relativ langsame, erosions-kontrollierte Abkühlung mit einzelnen thermischen Ereignissen, die durch Rutil Abkühlalter bei 1793 ± 10 Ma, 1738 ± 14 bis 1720 ± 12 Ma, 1645 ± 63 Ma und 1617 ± 91 Ma belegt sind.

Diese Studie zeigt, dass die Kombination verschiedener Datenquellen von unterschiedlich retrograd überprägten Proben, anstatt nur eklogitfazielle Mineralparagenesen zu untersuchen, ein wichtiges Werkzeug zur Charakterisierung der tektonometamorphen Entwicklung von paläoproterozoischen Kollisionsorogenen darstellt.

List of Publications

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1 Introduction

1.1 State of the art

Modern orogenic systems are characterized by a widespread abundance of high-pressure and ultrahigh-pressure rocks (Chopin, 2003; Brown, 2008; Ernst and Liou, 2008), reflecting the stabilization of low geotherms since the Neoproterozoic (Brown, 2008), with subduction of crustal rocks to and rapid exhumation from mantle depths (Gebauer et al., 1997; O'Brien et al., 2001; Baldwin et al., 2004a). The change in density associated with the transformation from basalt (~ 3.0 g/cm³, or gabbro) to eclogite (3.2-3.6 g/cm³) during subduction results in an enhanced slab pull and consequently in higher rates of plate motions (Davies, 1992). The formation of eclogite may be regarded as one of the driving forces for plate tectonics on Earth (Ahrens and Schubert, 1975; Hacker, 1996; Stern, 2007), and its appearance in the rock record may tell us when a plate-tectonics mode of behavior (similar to today) was first adopted (Collins et al., 2004; Mints et al., 2010; Shirey and Richardson, 2011). Going back in time (>1 Ga), however, eclogite- and blueschist-facies rocks become increasingly rare (Brown, 2008; Brown and Johnson, 2018). The scarcity of high-pressure to ultrahigh-pressure (UHP) rocks in the Paleoproterozoic and Archean rock record has been interpreted as evidence of fundamentally different plate tectonic systematics. Some authors attribute the paucity of the high-pressure rocks to an absence of tectonic environments able to produce and/or exhume these rocks (Collins et al., 2004; Condie and Kröner, 2008; Anderson et al., 2012), while others suggest that higher mantle temperatures and a weak lithosphere inhibited deep subduction of crustal rocks (Burg and Ford, 1997; Chopin, 2003; Stern, 2007; Condie and Kröner, 2008; Bradley, 2011; Anderson et al., 2012; Loose and Schenk, 2018). The sparsity of eclogite and blueschist >1.8 Ga, however, may also simply be a preservation problem (Möller et al., 1995; Baldwin et al., 2004b; Collins et al., 2004; St-Onge et al., 2006; Brown, 2008; Glassley et al., 2014; Weller and St-Onge, 2017). In recent years, a growing number of studies have reported the presence of relict eclogite- to high-pressure granulite-facies mineral assemblages in Paleoproterozoic mafic dykes intrusive into Archean tonalite-trondhjemite-granodiorite (TTG) gneiss (Nutman and Friend, 1989; Nutman et al., 1992, 2008; Smelov and Beryozkin, 1993; Zhao et al., 2001; Guo et al., 2002; Baldwin et al., 2004b; Skublov et al., 2011; Tam et al., 2012; Imayama et al., 2017; Liu et al., 2017; Weller and St-Onge, 2017; Yu et al., 2017) or in Paleoproterozoic metamafic lithologies of mid-ocean ridge basalt (MORB) type (Möller et al., 1995; Sklyarov et al., 1998; Collins et al., 2004; Boniface et al., 2012; Loose and Schenk, 2018). These mineral assemblages allow insight into the conditions of subduction, metamorphism and exhumation during the Paleoproterozoic, an era from which the nature of plate-tectonic processes is not well understood but which records global subduction-related orogenic activity and that coincides with the formation of the supercontinent Nuna or Columbia (Hoffman,

1988; Rogers and Santosh, 2002; Zhao et al., 2002; Brown, 2007; Reddy and Evans, 2009; St-Onge et al., 2009; Mertanen and Pesonen, 2012). The relict eclogite- to high-pressure granulite-facies mineral assemblages usually yield PT-data for several stages along a clockwise, near-isothermal decompression PTt-path (Nutman and Friend, 1989; Möller et al., 1995; Zhao et al., 2001; Guo et al., 2002; Tam et al., 2012; Imayama et al., 2017; Liu et al., 2017; Yu et al., 2017). Such PT-paths are characteristic for sites of significant tectonic crustal thickening, such as a collisional orogenic setting, in which the maximum pressure was reached before the maximum temperature (England and Thompson, 1984; Thompson and England, 1984; Harley, 1989; Brown, 1993; Zhao et al., 2002; O'Brien and Rötzler, 2003). Further evidence of collisional orogeny is given, if the terranes containing high-pressure mineral assemblages are juxtaposed against terranes characterized by high-temperature, low-pressure metamorphism, as the one-sided subduction in convergent tectonic settings results in the development of dual thermal regimes or paired metamorphic belts (Miyashiro, 1973; Brown, 2008). These dual thermal regimes encompass a low geotherm in the subducting plate (low dT/dP around ~ 7-12 °C/km; Wang et al., 1989; Maekawa et al., 1993; Möller et al., 1995; John et al., 2003), and a high geotherm in the overriding plate (high dT/dP around >>20 °C/km; Perkins and Chipera, 1985; Schumacher et al., 1990; Bohlen, 1991; Dempster et al., 1991). The presence of such paired metamorphic belts is interpreted as one of the hallmarks of modern plate tectonics (Brown, 2008; Condie and Kröner, 2008). Identification of low geotherms, eclogite and/or contrasting types of metamorphism in Paleoproterozoic orogenic belts, implies that some form of plate tectonics similar to today may have already been present around 2 Ga (Möller et al., 1995; Zhao et al., 2001; Boniface et al., 2012; Imayama et al., 2017; Weller and St-Onge, 2017; Yu et al., 2017) or even earlier (Condie and Kröner, 2008; Brown and Johnson, 2018), prior to the transition to a modern plate tectonic regime in the Neoproterozoic (Brown, 2008).

1.2 Aim of this study

A poorly defined occurrence of relict high-pressure metamorphic rocks has been reported from the Paleoproterozoic Nagssugtoqidian Orogen in South-East Greenland (Chadwick et al., 1989; Nutman and Friend, 1989; Nutman et al., 2008).

Here, relict high-pressure mineral assemblages occur in Paleoproterozoic mafic dykes and boudins and boudinaged layers of mafic to ultramafic supracrustal rock in the orthogneiss-dominated core of the orogen, the so called Kuummiut Terrane (Wright et al., 1973; Chadwick et al., 1989; Nutman and Friend, 1989; Messiga et al., 1990; Nutman et al., 2008; Kolb, 2014). The southern lower-pressure Isertoq Terrane incorporates the high-temperature Ammassalik Intrusive Complex (Friend and Nutman, 1989; Kolb, 2014; Lebrun et al., 2018). The tectonic setting in which these contrasting types

of metamorphism occurred, as well as the processes that were operative during metamorphism, may be identified, and a model depicting the tectonometamorphic evolution constructed, by a combination of thermobarometric, geochronological and structural data from the high-pressure rocks of the Kuummiut Terrane (Zhao et al., 2001; Collins et al., 2004; Nutman et al., 2008). As the Nagssugtoqidian Orogen in South-East Greenland is regarded as part of an originally continuous orogenic belt stretching from North America through Canada, Greenland and Scotland to the Baltic shield (Wright et al., 1973; Hoffman, 1988; Bridgwater et al., 1990; Kalsbeek et al., 1993; Park, 1995; Van Gool et al., 2002; Zhao et al., 2002; St-Onge et al., 2009), this information may furthermore be used to infer local differences in geotectonic settings during the formation of the Nuna supercontinent. To date, however, the tectonometamorphic evolution of the high-pressure rocks of the Kuummiut Terrane has only poorly been studied.

In the Kuummiut terrane, a variety of mineral textures indicate exhumation from eclogite-facies depths, with retrogression at high-pressure granulite- and amphibolite-facies conditions (Nutman and Friend, 1989; Messiga et al., 1990). However, the maximum depth of subduction in earlier studies was only constrained to ca. 40 km (Nutman and Friend, 1989; Mengel et al., 1990; Messiga et al., 1990; Nutman et al., 2008). Furthermore, despite knowledge of a clockwise, near-isothermal decompression path (Nutman and Friend, 1989; Messiga et al., 1990; Nutman et al., 2008), the retrograde metamorphic evolution of the high-pressure rocks has not been investigated in detail, with unclear evidence for the PT-conditions, the trend of the PT-path and the number of retrograde stages (Messiga et al., 1990; Nutman et al., 2008; Nicoli et al., 2018). The prograde metamorphic evolution, in contrast, is unknown. The regional extent of eclogite-facies metamorphism is also not known or if differences in the tectonometamorphic evolution exist, which might be expected regarding the size of the high-pressure terrane (ca. 80×100 km). One goal of this study is thus to quantify the conditions of metamorphism, constrain the maximum depth of subduction, and determine local differences in the metamorphic evolution, in order to refine the previously constrained tectonometamorphic evolution of the high-pressure rocks of the Kuummiut Terrane.

Most geochronological studies in the Kuummiut Terrane focused on the felsic lithologies (Kalsbeek et al., 1993; Nutman et al., 2008; Baden, 2013; Kokfelt et al., 2016; Thrane et al., 2016; Nicoli et al., 2018), with only few constraints on the emplacement age of the mafic dykes and metamorphic evolution of the mafic dykes and supracrustal rocks (Kalsbeek et al., 1993; Nutman et al., 2008). Magma formation and dyke emplacement into the Archean TTG gneiss was previously recorded at 2400-2200 Ma (Sm-Nd model dates – T_{DM} ; Bridgwater et al., 1990; Kalsbeek et al., 1993) and 2015 ± 15 Ma (zircon U-Pb; Nutman et al., 2008), with crosscutting relationships implying that dyke formation and emplacement occurred prior to the emplacement of the Ammassalik Intrusive complex and eclogite-facies metamorphism (Chadwick et al., 1989; Kalsbeek and Taylor, 1989;

Nutman and Friend, 1989). Prior to this study it was unknown if these different age estimates indicate dyke emplacement over multiple stages (Kolb, 2014). A 1867 ± 28 Ma zircon U-Pb age from a mafic dyke has been interpreted to date the timing of eclogite-facies metamorphism, based on the inclusion content (garnet and clinopyroxene, no plagioclase) and trace element chemistry of zircon (depressed HREE abundances with no significant Eu anomaly; Nutman et al., 2008). This age, however, has a relatively large error and overlaps with some ages for retrograde metamorphism (1890 to 1820 Ma; Nutman et al., 2008; Baden, 2013; Nicoli et al., 2018). The large areal extent of the high-pressure terrane further questions how representative this age is. In addition, although the mineral textures recorded by Nutman et al. (2008) imply subduction to mantle depths, followed by rapid exhumation, the timing of prograde and retrograde metamorphism, as well as the rates of subduction and exhumation, had not been constrained. Furthermore, it was unknown whether the metamafic lithologies also retain evidence for several thermal events along the retrograde path, which was previously constrained from zircon U-Pb dates in felsic lithologies (Nutman et al., 2008; Baden, 2013; Nicoli et al., 2018). Therefore, another goal of this thesis is to obtain age constraints on the emplacement and metamorphic evolution of the metamafic rocks in the Kuummiut Terrane.

Previous studies imply that the Kuummiut Terrane was buried to mantle depths in a subduction to collision scenario (Nutman et al., 2008; Kolb, 2014). Rapid exhumation, inferred from the clockwise near-isothermal decompression PT-path, is typical of collisional orogenic settings and may be driven by extensional exhumation during synorogenic collapse (McClelland and Gilotti, 2003). Other mechanisms include channel flow and/or ductile extrusion, crust-mantle delamination, and slab break-off (Platt, 1993; Brun and Faccenna, 2008). Erosion, in contrast to these tectonic processes, is more typical for near-isobaric cooling with only minor decompression and a relatively slow process. The final goal of this thesis is to determine what the mechanisms of exhumation are, if there was a change in exhumation style during the retrograde evolution of the high-pressure rocks and along which structures the high-pressure rocks were exhumed.

1.3 Applied Methods

To resolve the aforementioned issues, a systematic sampling of metamafic lithologies from various areas in the Kuummiut Terrane was carried out in the framework of a larger expedition program to South-East Greenland in 2014, under the auspices of the Geological Survey of Denmark and Greenland (GEUS). In addition, several samples from the GEUS archives and some by Allen Nutman and Clark Friend, collected during previous field seasons in South-East Greenland, were studied. The fieldwork and archive samples represent the basis for a detailed petrological study. Petrographic investigations focused on the identification of mineral reaction textures as well as on the

characterization of prograde, relict high-pressure and retrograde mineral assemblages. These investigations were conducted via hand lens in the field, microscopy (transmitted and reflected light), BSE-imaging (Backscatter Electron) using an electron microprobe and compositional phase mappings obtained via QEMSCAN (Quantitative Evaluation of Minerals by SCANNing electron microscopy). Compositional mappings (Mg, Al, Ca, Mn, Fe, Na) and WDX- and EDX-analysis (Wavelength- and Energy-Dispersive X-ray analysis) at the electron microprobe, together with the BSE- and QEMSCAN-images, were used to determine the mineral chemistry and mineral chemical changes (especially in zoned minerals, i.e. garnet, plagioclase and clinopyroxene) during the metamorphic evolution of the high-pressure rocks of the Kuummiut Terrane. Bulk-rock major element analysis was conducted via X-ray fluorescence (XRF) on borate glass fusion pellets and used in pseudosection modelling of selected samples via the computer program THERMOCALC (Version 3.33; Powell and Holland, 1988; updated October 2009) and the November 2003 updated version of the internally consistent dataset from Holland and Powell (1998). Mineral chemical data was utilized in the modelling of compositional isopleths in the pseudosections (e.g. X_{Grs} of garnet). Furthermore, various conventional geothermobarometric methods (e.g. Ca in Opx thermometry and garnet-anorthite-diopside-quartz barometry) were applied to the mineral chemistry of suitable equilibrium mineral assemblages, to obtain PT-estimates independent of the bulk-rock approach. These conventional PT-estimates were compared to the results from pseudosection modelling, to test the applicability of this method in heavily retrogressed rocks. The PT-data from pseudosection modelling and conventional geothermobarometry were combined with the results from petrographic and mineral chemical analysis to determine the PT-history of the high-pressure rocks. Age constraints on the deposition of the supracrustal rocks, the emplacement of the mafic dykes and the metamorphic evolution of the high-pressure rocks, were obtained via LA-SF-ICP-MS (Laser Ablation – Sector Field – Inductively Coupled Plasma – Mass Spectrometry) U-Pb dating of texturally well-characterized accessory phases (zircon, monazite, titanite and rutile). Microtextural analysis, pseudosection modelling and zircon trace element data obtained from LA-SF-ICP-MS, were used to interpret and correlate the U-Pb dates to the magmatic and metamorphic evolution of the Kuummiut Terrane.

The individual methods are explained in more detail in Chapters 2, 3 and 4, as well as in the Appendices. Some more background information on the two main methods used, THERMOCALC pseudosection modelling and LA-ICP-MS U-Pb dating is given below.

1.3.1 Pseudosection modelling

THERMOCALC is a Pascal computer program (Powell and Holland, 1988) that uses an internally-consistent thermodynamic dataset (Holland and Powell, 1985; Powell and Holland, 1985) and a non-linear equation solver to undertake thermobarometry and phase diagram calculations for metamorphic rocks of varying composition. Extensive documentation, tutorials, publications, datafiles and software

regarding the THERMOCALC computer program can be found on the THERMOCALC resources website (www.metamorph.geo.uni-mainz.de/thermocalc/) that is housed at the Johannes Gutenberg University of Mainz.

The program and dataset are based on the principle of equilibrium in metamorphic rocks: during metamorphism, a mineral assemblage will change towards a state at which it is in equilibrium with its surroundings (e.g. PT-conditions, Powell and Holland, 2006). Such a change may be driven or strongly promoted at some scale by fluid, melt, deformation or rapidly changing PT-conditions (Brodie and Rutter, 1985; Rubie, 1986; Andersen et al., 1991; McFarlane et al., 2003; Powell and Holland, 2006; Putnis and Austrheim 2010). Depending on the amount of the catalyst, a precursor rock/mineral assemblage may either become fully re-equilibrated, with little to no replacement textures, or partially re-equilibrated, where evidence of both the precursor and reaction product are widespread (Powell and Holland, 2006). In the latter case, special care has to be taken in choosing the minerals that supposedly were in equilibrium once. Nevertheless, both fully and partially re-equilibrated rocks reflect a frozen-in stage of equilibrium for an assemblage of different minerals (on different scales), which corresponds to a certain set of PT-conditions along the metamorphic evolution of the rock (Powell and Holland, 1985). Their study, using calculations based on equilibrium thermodynamics, can help us to identify metamorphic processes and the PT-conditions of formation for certain rocks (Powell and Holland, 2006).

Two different applications for equilibrium thermodynamics exist (Powell and Holland, 1985):

Conventional or classic geothermobarometry essentially utilizes the deviation of an experimentally determined PT-curve for a specific chemical reaction, determined via mineral chemical analysis in the phases of interest to calculate PT-conditions. Good geothermometers have steep/near vertical slopes in PT-space and are temperature-sensitive (i.e. exchange and solvus thermometry), while good geobarometers have a gentle/near horizontal slope and are pressure-sensitive (i.e. net transfer reactions; Bucher and Frey, 2002).

The other method is known as phase or mineral equilibria modelling. It involves a database that allows the calculation of any equilibrium reaction, as long as the database encompasses reliable data for all phases in the reaction (Powell and Holland, 1985). Thermodynamic calculations using this method require that thermodynamic data for a group of mineral endmembers and the activity-composition relationships in the minerals are known (Powell and Holland, 1985). The former poses a principal problem, as calorimetrically-derived enthalpies of formation are not precise enough for calculations. As a solution, Holland and Powell (1985) used a least squares analysis of experimentally determined mineral equilibria (from high PT-assemblages) to generate an internally consistent thermodynamic dataset. This dataset allows the calculation of uncertainties on and correlations between the enthalpies of formation of end-members of minerals (Holland and Powell, 1985).

Furthermore, its power lies in performing calculations with all applicable reactions between mineral end-members and in determining the reliability of these calculations (Powell and Holland, 1988).

Two approaches exist in mineral equilibria modelling (Powell and Holland, 2006), both of which can be utilized with THERMOCALC:

(1) Inverse modelling – using the chemical composition of a mineral assemblage in equilibrium to determine the PT-conditions of formation via the Average PT-method. This method, which has also been termed “optimal thermobarometry” (Powell and Holland, 1994), may be modified to calculate P at T or T at P, with varying fluid content. In contrast to conventional geothermobarometry, the Average PT-approach is consistent with other methods of mineral equilibria modelling, allows the determination of uncertainties on the calculated PT-conditions, and identifies whether or not a specified equilibrium mineral assemblage will yield thermobarometric information at all (Baldwin, 2016).

(2) Forward modelling – using a certain bulk-rock composition to draw phase diagrams (with the program drawpd) and to study conditions and processes of rock formation. These phase diagrams include PT-projections, PT-, PX- and TX-pseudosections, compatibility diagrams and μ - μ diagrams (μ being the chemical potential). THERMOCALC also features the ability to calculate mineral mode and composition isopleths, which are frequently used to limit the stability field of a modeled mineral assemblage via comparison to the actual mineral modes and major element chemistry. The bulk composition used in phase diagram modelling in general is not equal to an actual rock composition, which in addition to the major oxides (e.g. SiO_2 , Al_2O_3 , Na_2O , K_2O , MgO , FeO) includes many minor and trace elements (Powell, 1991). The contribution of the latter to the phase equilibria cannot be unequivocally determined (Powell, 1991) and in many cases, thermodynamic data may not be available for certain elements and phases of interest. As such, phase diagram modelling in THERMOCALC is conducted in a well-defined model system, which is a set of user-specified common oxide components (e.g. K_2O - FeO - MgO - Al_2O_3 - SiO_2 - H_2O or KFMASH), to which the actual bulk composition is normalized. The number of components in the model system is chosen so that all petrological phases of interest can be adequately modeled (defined by the mineral chemistry of both major and minor minerals; Powell and Holland, 2006), that the modelling is as close to the real rock as possible (Powell, 1991), and that the addition of a component does not result in the addition of phases that are not actually present in the rock (Powell and Holland, 2006). In line with the first two points, phase diagram modelling is usually conducted in large model systems, especially concerning mafic rocks (e.g. Bruand et al., 2010; Groppo and Castelli, 2010; Dziggel et al., 2012; Yu et al., 2017).

PT-pseudosections show the fields of stability in PT-space of different equilibrium mineral assemblages for a single bulk-rock composition. In contrast to PT-projections, the limitation to a

single bulk-rock composition results in fewer reaction lines in the pseudosection, as the specific bulk-rock composition will not see certain reactions (Baldwin et al., 2016). Tx- and Px-pseudosections are drawn at constant P or T, respectively, and indicate how the equilibrium mineral assemblages change from one bulk composition to another. This change in bulk-rock composition may just involve one (i.e. X_{Fe}) or all of the components of the model system. Apart from giving PT-information for the equilibrium mineral assemblage of interest, pseudosections provide further invaluable thermobarometric information, such as mineral modes and compositions, as well as the stability fields for minerals at a certain bulk-rock composition (Powell and Holland, 2008; Powell et al., 2009). These stability fields may be used to differentiate between prograde, peak and retrograde mineral phases. Furthermore, including H_2O and melt in the calculations allows the determination of the solidus, while the evolution of the mineral composition in a zoned mineral can be used to deduce part of a PT-path.

One drawback/challenge to the pseudosection approach is that the bulk-rock composition employed in modelling should represent a volume of equilibration in the rock. In many cases, however, it is unknown how large the equilibration volume is, as metamorphic rocks contain zoned minerals (e.g. garnet), diffusional re-equilibration or melt loss may have changed mineral compositions during cooling, and mineral reaction textures imply disequilibrium (Powell and Holland, 2008; Powell et al., 2009). The retrograde evolution of eclogite, in particular, results in the formation of chemically and mineralogically distinct domains (see also Zhang et al., 2000; Zhang et al., 2003; Scott et al., 2013), and these domainal equilibration volumes are not accounted for by the bulk-rock geochemistry. Consequently, a bulk-rock composition obtained via XRF-analysis is most suitable in well-equilibrated samples (large volume of equilibration, e.g. high-temperature rocks), whereas in partially reacted rocks, a bulk-rock composition obtained from combined mineral chemical data and modes (from point counting or quantitative X-ray mappings, e.g. QEMSCAN) might be more suitable (Powell and Holland, 2008).

The above arguments show that pseudosection or phase diagram modelling in general, may yield many important thermobarometric information for metamorphic rocks, but that it is not straightforward and essentially depends on how well the model system replicates the actual rock, how well the used bulk composition mirrors equilibrium and how well the mineral textures and mineral compositional relationships can be understood in terms of equilibrium (Powell, 1991).

Perple_X is another commonly used program for calculation and display of phase diagrams, phase equilibria and thermodynamic data (Connolly, 1990; www.perplex.ethz.ch). It is much faster in its calculations than THERMOCALC, due to a Gibbs energy minimizer that models stability fields and not field boundaries (Powell and Holland, 2006). As such, it may produce entire diagrams in a couple

of minutes, while in THERMOCALC, the diagram has to be built up curve by curve (Powell and Holland, 2006). However, Perple_X uses pseudocompounds rather than solid solutions to calculate reactions between end-members (Connolly, 2005), which might result in the introduction of small artifacts or pixelization in the pseudosection and oversight of smaller features which are close to each other (Powell et al., 2009; Hirsch and Baldwin, 2016). Detailed comparisons between pseudosections constructed via Perple_X and THERMOCALC can be found on the Perple_X website at www.perplex.ethz.ch/perplex_thermocalc_comparison.html.

1.3.2 LA-ICP-MS U-Pb dating

1.3.2.1 U-Pb Dating

Uranium-lead (U-Pb) dating is a long-known (Boltwood, 1907; Holmes, 1911), and well-refined (Begemann et al., 2001; Schoene et al., 2006; Mattinson, 2010) radiometric scheme, which power lies in the existence of two (or three, if ^{232}Th is included) parallel decay routes (Gehrels, 2012; Corfu, 2013a; Schoene, 2014). The method is based on the radioactive decay of the mother nuclides ^{235}U and ^{238}U , to their stable daughter isotopes ^{207}Pb and ^{206}Pb , via a series of alpha (and beta) decays (Corfu, 2013b; Schoene, 2014). Assuming the system remained closed after achieving isotopic equilibrium; the amount of radiogenic daughter isotopes in a given material is time-sensitive and can be determined via the isochron equations:

$$(^{206}\text{Pb}/^{204}\text{Pb}) = (^{206}\text{Pb}/^{204}\text{Pb})_0 + (^{238}\text{U}/^{204}\text{Pb}) \times (e^{\lambda_{238}t} - 1) \quad \text{and} \quad (1)$$

$$(^{207}\text{Pb}/^{204}\text{Pb}) = (^{207}\text{Pb}/^{204}\text{Pb})_0 + (^{235}\text{U}/^{204}\text{Pb}) \times (e^{\lambda_{235}t} - 1) \quad (2)$$

(Schoene, 2014). The subscript 0 denotes the amount of daughter isotope present at the time the system closed, commonly referred to as initial, non-radiogenic or common lead, while λ denotes the decay constants of the two U-isotopes. As with other dating methods, the isochron equations are given in normalized form, with ^{204}Pb representing the only non-radiogenic isotope of Pb, since during U-Pb analysis, isotope ratios are easier to measure than absolute concentrations of elements (Schoene, 2014).

Equations 1 and 2 can be used to directly determine a model date, if the amount of common Pb is known and corrected for or its contribution can be neglected (Schoene, 2014). The latter is applicable to minerals where U is readily incorporated into the crystal structure, while Pb is strongly rejected, resulting in considerably more radiogenic, than non-radiogenic Pb (i.e. zircon; Schoene, 2014). Common Pb can be estimated using a bulk Pb evolution model (i.e. Stacey and Kramers, 1975), from co-genetic low-U minerals (Chamberlain and Bowring, 2000), or from the y-intercept in an isochron diagram (usually $^{238}\text{U}/^{204}\text{Pb}$ to $^{206}\text{Pb}/^{204}\text{Pb}$), if the set of minerals form a linear array (Schoene, 2014). Utilization of just one of the decay schemes is referred to as the U-Pb isochron dating method.

The U-Pb dual decay system, however, also allows for a third dating method, the generation of a Pb-

Pb date, by dividing equation (2) by equation (1):

$$\frac{[(^{207}\text{Pb}/^{204}\text{Pb}) - (^{207}\text{Pb}/^{204}\text{Pb})_0]}{[(^{206}\text{Pb}/^{204}\text{Pb}) - (^{206}\text{Pb}/^{204}\text{Pb})_0]} = (^{235}\text{U}/^{238}\text{U}) \times \frac{[(e^{\lambda_{235}t} - 1)]}{[(e^{\lambda_{238}t} - 1)]} = (^{207}\text{Pb}/^{206}\text{Pb})^*, \quad (3)$$

with * denoting the ratio of radiogenic Pb (Schoene, 2014). This equation can only be solved iteratively via numerical or graphical methods for t.

In a closed system, all three dates would agree within error (Corfu, 2013b; Schoene, 2014). However, especially for metamorphic rocks, the U-Pb systematics are usually disturbed subsequent to the achievement of isotopic equilibrium. To test the validity of the closed-system assumption, and to extract further geochronologic information, U-Pb data are presented in graphical form, with the two most common graphical presentations being the Wetherill concordia plot and the Tera-Wasserburg diagram.

In the concordia diagram (Wetherill, 1956), $^{206}\text{Pb}^*/^{238}\text{U}$ is plotted against $^{207}\text{Pb}^*/^{235}\text{U}$ from the same analysis and the resulting data point is evaluated with respect to its vicinity to the concordia curve. The concordia curve represents the non-linear set of solutions for equations (1) and (2) at equal values of t (Wetherill 1956; Schoene, 2014). Consequently, in a closed system where all three dates are equal, the data points would lie on the concordia, i.e. they would represent concordant analysis (Wetherill, 1956). Those that do not fall on the concordia are termed discordant (Wetherill, 1956) and have experienced some form of open-system behavior (Schoene, 2014), such as Pb-loss due to incomplete re-equilibration/recrystallization during metamorphism (Timmermann et al., 2004) and diffusion (Mezger and Krogstad, 1997), or mixing of differently aged domains (Rowley et al., 1997). Further less common open-system behavior processes are Pb-gain, U-loss and U-gain (Schoene, 2014). Data points of minerals, which experienced the same Pb-loss event, plot along a straight line in the concordia diagram, the discordia. The intercepts of this discordia with the concordia curve usually give the age of the crystallization/equilibration event (upper intercept) and the age of the Pb-loss event (lower intercept; Corfu, 2013b; Schoene, 2014). However, large amounts of common Pb that were not corrected for may also lead to discordance (Schoene, 2014). The U/Pb-data of minerals that do not experience mobility of Pb and U after their formation, in contrast, move along the concordia with increasing age (Singh, 2001).

In the Tera-Wasserburg diagram, $^{238}\text{U}/^{206}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ are plotted on the x- and y-axes, respectively (Tera and Wasserburg, 1972a, b). The diagram combines the abilities of both the isochron method and concordia diagram to display initial Pb compositions and test for open system behavior (Schoene, 2014). The latter can be achieved by plotting initial Pb-corrected data for $^{238}\text{U}/^{206}\text{Pb}^*$ and $^{207}\text{Pb}^*/^{206}\text{Pb}^*$ into the diagram and then comparing the data to a concordia. A cogenetic suite of samples that plot along a line offset to the concordia, give their true age at the lower intercept and

their initial Pb composition at the y-intercept (Schoene, 2014). This process of initial Pb determination, however, is compromised if the sample experienced Pb-loss (Schoene, 2014). The Tera-Wasserburg diagram is preferred by some over the Concordia plot, as the error correlation between $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{238}\text{U}/^{206}\text{Pb}$ is much smaller than between $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$ (Corfu, 2013b).

Due to the long half-lives of the mother nuclides ($^{235}\text{U} \sim 704 \text{ Ma}$; and $^{238}\text{U} \sim 4.5 \text{ Ga}$; Holden, 1981), the U-Pb method is capable of dating events as early as the formation of the solar system to relatively recent events in the Pleistocene (Bindemann et al., 2001; Bouvier and Wadhwa, 2010; Schoene, 2014). For young samples (especially Phanerozoic), the $^{206}\text{Pb}/^{238}\text{U}$ isotope ratio yields the most precise ages, whereas for samples older than 1 Ga, $^{207}\text{Pb}/^{206}\text{Pb}$ ratios are typically quoted (Ludwig, 1998; Gehrels, 2012; Corfu, 2013b). The method is particularly popular among geochronologists, due to the abundance and weathering resistance of U-bearing minerals in most rock types (Weiss Frondel and Fleischer, 1950; Corfu, 2013a; Schoene, 2014). Owing to a high closure temperature, strong weathering resistance, and rejection of Pb in the crystal structure, zircon represents the most commonly utilized material in U-Pb dating (Mezger and Krogstad, 1997; Krogh et al., 1998; Hanchar and Hoskin, 2003). However, an ever-growing number of alternatives are published, with virtually any U-bearing material being datable, which might provide additional geo- and thermochronologic information (Schoene, 2014 and references therein). In general, U-Pb dating is conducted via Thermal Ionization Mass Spectrometry (TIMS; Krogh, 1973), Secondary Ion Mass Spectrometry/Sensitive High-Resolution Ion Microprobe (SIMS/SHRIMP; Compston et al., 1986) or Laser Ablation – Inductively Coupled Plasma – Mass Spectrometry (LA-ICP-MS; Zeh et al., 2013).

1.3.2.2 LA-ICP-MS analysis

LA-ICP-MS is a very sensitive, high-resolution analytical technique, typically employed in the field of mineralogy/petrology for rapid, quantitative determination of trace element and isotopic compositions (see below). Due to the application of a laser-ablation unit for sample introduction (typically a gas, i.e. ArF or solid state laser, i.e. Nd:YAG; Thomas, 2004; Schoene, 2014), the LA-ICP-MS method allows spatially resolved (μm -scale) analysis of solid samples (Simonetti et al., 2006; Koch and Günther, 2011). Such high-resolution analytical capabilities are required for the in-situ characterization of trace element and isotopic compositions (Simonetti et al., 2006; Gerdes and Zeh, 2009), as well as the chemical characterization of zoned and inclusion-rich minerals (Kooijman et al., 2010; Sindern et al., 2012), and melt and fluid inclusions (Heinrich et al., 2003). The normal ICP-MS method uses liquid samples, such as diluted acidic/digested solutions of well-homogenized bulk-rock powders (i.e. from fusion pellets) or mineral separates (after handpicking), for analysis (Thomas, 2004). Here, the spatial resolution of the LA-ICP-MS method is lost, consequently showing that the normal ICP-MS method is more suitable for bulk-rock analyses, while the LA-ICP-MS method is

very well-suited for micro-analytical applications (Longerich, 2008).

The underlying principle of the (LA-) ICP-MS method is the ionization of a sample aerosol in a high-temperature plasma, with subsequent time-resolved analysis of the ions in a mass spectrometer.

The high-temperature (~ 6000 - 10000 K; Taylor, 2001; Thomas, 2004) inductively coupled plasma is produced at the end of a torch that sits behind the specimen chamber of the ICP-MS system. The torch consists of three nested concentric quartz tubes, a center, a middle/intermediate and an outer tube (Thomas, 2004), through which the sample, auxiliary and plasma gases (mostly Ar) are introduced, respectively. At the start of the ICP, a high voltage electric spark is applied for a short period of time to ionize part of the Ar atoms in the plasma gas (separation into positively charged Ar ions and free electrons; Taylor, 2001; Thomas, 2004). The free electrons interact with the oscillating high radio-frequency (27 or 40 Mhz; Longerich, 2008) magnetic field of the grounded copper induction coil around the torch, resulting in an electrical current that heats up the Ar gas and causes further ionization (Taylor, 2001; Thomas, 2004). This process is continued, until a self-sustaining, balanced state is reached (Taylor, 2001), at which the rate of new electrons is nearly equal to the rate of ionized argon recombining with free electrons, resulting in a stable plasma cloud of high temperature. The plasma can be retained, because the high flow of plasma gas (~ 15 L/min; Thomas, 2004) keeps the plasma away from the walls of the torch, thus preventing melting. This outer argon gas flow is, as such, often referred to as cooling gas. Furthermore, auxiliary gas (~ 1 L/min) in the intermediate tube controls the spacing of the base of the plasma relative to the open end of the torch (Thomas, 2004).

The central tube transfers the sample gas (~ 1 L/min; Thomas, 2004) into the central region of the plasma. In order to maximize plasma temperature and ionization efficiency, the sample is introduced as a homogeneously sized aerosol (Thomas, 2004). Liquid samples are pumped via a peristaltic pump into an Ar-assisted nebulizer, with the resulting aerosol subsequently being cleansed from larger droplets in a cooled spray chamber, before being introduced into the plasma via a sample injector (Thomas, 2004). For solid samples, the sample is ablated via a focused laser system in an air-tight ablation cell, flushed out of the cell using He and/or Ar gas, and mixed downstream in the ablation funnel with Ar gas flowing to the plasma torch (Liu et al., 2013). In the central region of the plasma, the sample aerosol is dried, vaporized, atomized and finally ionized (Thomas, 2004; Longerich, 2008). In LA-ICP-MS machines, the laser system is influenced by many factors such as the ablation cell design, the laser wavelength and the pulse width (Schaltegger et al., 2015).

In the last step, the ions of the sample gas travel to the mass spectrometer, where their intensities are converted into detectable signals. In contrast to the torch, the mass spectrometer operates at low-pressure, since at atmospheric pressure, ions only travel for about $0.1 \mu\text{m}$ before they lose their charge or get deflected (Longerich, 2008). As such, at the interface between the plasma and mass

spectrometer, there are two or more cones with small orifices, which lower the pressure in steps, from atmospheric pressure at the plasma, to high vacuum in the mass spectrometer (Thomas, 2004; Longerich, 2008). Once a high vacuum is reached, the ion focusing system steers the ion beam into the mass analyzer, where the ions are separated based on their mass to charge ratio (Thomas, 2004). Several different types of mass analyzers exist, the three most commonly employed in ICP-MS systems being the sector field (SF), quadrupole (quad) and time of flight (TOF) analyzers (Longerich, 2008). The choice of the mass analyzer depends on the mass resolution, sensitivity and precision required (Koch and Günther, 2011).

Sector field instruments use an electrostatic (ESA) or magnetic sector analyzer (MSA), or a combination of the two (doubly-focusing system) in a specific geometric setting (i.e. Nier-Johnson geometry), to separate the sample ions into different beams (Longerich, 2008). The method is based on the principle that when high velocity charged ions pass through an electrostatic or magnetic field (applied perpendicular to their flight path), they are forced to travel in a curved path (i.e. they are deflected; Taylor, 2001; Longerich, 2008). Varying the magnetic field or acceleration potential applied to the ion beam and ESA, as such, allows the user to select ions with a specific mass to charge ratio and energy, based on their different flight paths (Thomas, 2004; Longerich, 2008). Sector field instruments either come as single or multi collector systems, the latter requiring a doubly focusing ESA-MSA geometry to spatially separate the ions, while the single collector only employs both sector fields for high-resolution applications (Longerich, 2008). Multi collector instruments allow simultaneous detection over a limited range of mass, e.g. ^{207}Pb to ^{238}U , due to nine or more detectors at the focal plane of the magnet (Thomas, 2004; Longerich, 2008). Further advantages include a high sensitivity with low uncertainties, due to higher count rates (as all masses are detected all the time), and low background noise. The main disadvantages are relatively slow measurement times when switching between masses and very high capital costs (Longerich, 2008). In contrast to Quad and TOF mass analyzers, sector field instruments have the advantage of low continuum backgrounds and high sensitivity, as well as the possibility to apply high resolution to resolve some interferences (though this results in lower sensitivity; Taylor, 2001; Thomas, 2004). Sector field mass analyzers, however, are more expensive than Quad mass analyzers and sometimes have slower measurement times (Thomas, 2004; Longerich, 2008).

Quad mass analyzers do not allow spatial separation of the ion beams, but rather only work as mass filters through a complex electrical field of alternating (AC) and direct current (DC) potentials applied to opposite pairs of an array of four metallic rods (Thomas, 2004; Longerich, 2008). By choosing a particular AC/DC potential, only ions of desired mass to charge ratio travel through the middle of the four rods, whereas the rest are deflected (Thomas, 2004). The most important advantages of Quad analyzers are their low cost and the small volume required for the analyzer. Furthermore, analyses are fast and relatively good sensitivities can be reached. However, high resolution cannot be achieved, and the Quad mass analyzer also lacks the background and uncertainty

reduction capabilities of sector field instruments (Longerich, 2008).

TOF mass analyzers work on the principle that when applying an equal kinetic energy (i.e. via an acceleration voltage) to a group of ions, their velocities will be different based on their masses (Taylor, 2001; Thomas, 2004). As such, in order to characterize the sample ions, the mass-dependent time it takes them to move from the ion source to the detector is measured (Thomas, 2004; Duwe and Neff, 2007). The timing of extraction from the ion source is controlled by either using a pulsed ionization method (orthogonal design), or rapid electric field switching (axial design), to create ion packets which travel down the flight tube to the detector (Thomas, 2004; Longerich, 2008). The TOF mass analyzer has the fastest scanning speed of all mass analyzers and encompasses the largest practical mass range (Thomas, 2004). The use of pulsed ionization or beam switching, however, results in low sensitivities, and depending on the instrument setup, high background signals may be present (Thomas, 2004; Longerich, 2008).

At the end of the ICP-MS system, the detector transforms the magnitude of the ion current into a readable number (counts/s) for computer acquisition (Thomas, 2004). Depending on the instrument, the number and type of detectors may vary, with older systems implementing both secondary electron multipliers and Faraday cups (Taylor 2001; Thomas, 2004; Simonetti et al., 2008). The latter are used for the direct measurement of high ion currents ($> 10^4$ ions/s; analog mode), whereas secondary electron multiplication is used for the amplification of low ion currents (up to 10^6 , pulse-counting mode; Taylor, 2001; Thomas, 2004; Longerich, 2008). Discrete dynode detectors in modern machines are capable of both modes of operation (Thomas, 2004). The mass to charge ratio identifies which element/isotope was measured and the concentration of the analyte can be determined via calibration or reference standards (Thomas, 2004).

Commercially available since the late 1980's (Gray, 1985), LA-ICP-MS was quickly established as a relatively fast and reliable analysis technique for the determination of in-situ trace element compositions (Perkins et al., 1991; Jackson et al., 1992; Jarvis and Williams, 1993). Initial demonstrations that the LA-ICP-MS method may yield U-Pb dates with sufficient precision and accuracy followed shortly thereafter (Feng et al., 1993; Fryer et al., 1993). Since then, many studies have used this analysis technique as a complementary geochronological tool to other established methods such as TIMS and SHRIMP (Hirata and Nesbitt, 1995; Horn et al., 2000; Li et al., 2001; Horstwood et al., 2003; Tiepolo, 2003; Jackson et al., 2004; Chang et al., 2006; Frei and Gerdes, 2009; Liu et al 2010). The main advantages of the LA-ICP-MS technique are a high sample throughput, with little to no sample preparation, ease of operation, and its relatively low purchasing and operating costs, resulting in widespread availability of analytical facilities (Feng et al., 1993; Jackson et al., 2004; Thomas, 2004; Chang et al., 2006; Koch and Günther, 2011; Kröner et al., 2014). However, the LA-ICP-MS method lacks the precision, accuracy and reproducibility of the TIMS

method, which is seen as the benchmark regarding high-quality U-Pb data (Horstwood et al., 2003; Schoene, 2014). Likewise, considering the spatial resolution, LA-ICP-MS machines lack behind a SHRIMP and also produce larger analytical pits (Kröner et al., 2014; Schaltegger et al., 2015). Further difficulty exists in the correction of the raw data, due to large and temporally variable fractionation of U and Pb, interference of Hg from the Ar plasma gas in the common Pb correction, and instrument induced mass discrimination (Jackson et al., 2004; Košler, 2008; Frei and Gerdes, 2009; Schaltegger et al., 2015). There is, however, constant research regarding optimization of the technique, i.e. the instrumental setup, standards and correction procedure (e.g. Sláma et al., 2008; Frei and Gerdes, 2009; McFarlane and Luo, 2012), as well as new applications (e.g. Gerdes and Zeh, 2006, 2009; Zack et al. 2011; Ito, 2014).

1.4 Thesis structure

Data and results presented in this thesis are mainly sourced from a GEUS report for the 2014 field activities in the Kuummiut Terrane of South-East Greenland (Chapter 2) and two peer-reviewed publications (Chapters 3 and 4).

Chapter 2 contains the petrological and structural observations of the 2014 fieldwork, which was published in a GEUS report in March 2018. During fieldwork, we observed that retrogressed eclogite-facies mineral assemblages and reaction textures are not limited to mafic dykes intrusive into Archean TTG gneiss, but also occur in mafic to ultramafic supracrustal rocks of the Helheim and Kuummiut units (Kolb, 2014). Furthermore, despite sampling a plethora of mafic rock lithologies, well-preserved garnet-omphacite assemblages were not found. However, well-equilibrated high-pressure mineral assemblages were identified in ultramafic garnet-pyroxenite of several localities. In addition, the field observations showed that the structural inventory of the Kuummiut Terrane is quite complex, in relation to several bounding shear zones and a complicated tectonic history. The report presented in Chapter 2 is a joint work by Annika Dziggel and Sascha Müller. Both authors were in the field and collected the samples and petrological data. Annika Dziggel collected and interpreted the structural data, plotted in stereographic projections by both authors. Sascha Müller extracted the field data for the report from an Android database and mainly wrote the petrographic and mineralogical paragraphs, with additional input from Annika Dziggel.

Chapter 3 reports the petrography, mineral textures, mineral chemistry and PT-conditions of metamorphism from the high-pressure rocks of the Kuummiut Terrane. This manuscript was published in “Lithos” in January 2018. One of the main results of this study is that the high-pressure rocks were variably retrogressed during decompression, with the degree of replacement and type of

replacement assemblage being controlled by fluid availability. Overall, the data yield a clockwise PT-path with near-isothermal decompression after eclogite-facies metamorphism, showing that exhumation was initially rapid and tectonically-controlled. The manuscript presented in Chapter 3 is a collaborative effort by Sascha Müller, Annika Dziggel, Jochen Kolb and Sven Sindern. Sascha Müller conducted the sample preparations for thin section and fusion pellet production. Sven Sindern collected the bulk-rock major element data via XRF. Sascha Müller analyzed the thin sections via microscopy, electron microprobe and QEMSCAN, and collected the mineral chemistry data. Furthermore, Sascha Müller, with assistance by Annika Dziggel, performed the thermobarometric calculations and prepared the PT-pseudosections and other diagrams. Jochen Kolb assisted in the drawing of the schematic cross-sections of the Nagssugtoqidian Orogen. Sascha Müller mainly wrote the manuscript, with assistance by Annika Dziggel, Jochen Kolb and Sven Sindern. All authors assisted in the data interpretation and discussion.

Chapter 4 contains a manuscript that was published in “Precambrian Research” in September 2018. This chapter focuses on the geochronology of the high-pressure rocks of the Kuummiut Terrane, investigated via LA-SF-ICP-MS U-Pb dating of zircon, monazite, titanite and rutile from retrogressed eclogite and garnet-kyanite schist. Ages for dyke emplacement, deposition of the precursor to the metasediment and for several stages during retrograde metamorphism are reported. Zircon and monazite experienced partial to complete recrystallization during retrogression, while titanite records formation and rutile cooling ages. Titanite and rutile data, in combination with the clockwise PT-path, indicate a change from an initially rapid, tectonically-driven exhumation, to a slow, erosion-controlled cooling. The manuscript presented in Chapter 4 is a collaborative work between Sascha Müller, Annika Dziggel, Sven Sindern, Thomas Find Kokfelt, Axel Gerdes and Jochen Kolb. Sascha Müller performed all the sample analyses (via microscopy, electron microprobe and QEMSCAN) and preparations (crushing, magnetic and density separation, handpicking) prior to the LA-ICP-MS work. Collection of U-Pb isotope data from the thin sections and grain mounts via LA-ICP-MS was conducted by Sascha Müller under the supervision of Axel Gerdes at Goethe-University Frankfurt (GUF) and by Thomas Find Kokfelt at the Department of Petrology and Economic Geology, Geological Survey of Denmark and Greenland (GEUS, Copenhagen). Axel Gerdes and Thomas Find Kokfelt processed the raw isotope data and provided the data sheets. Sascha Müller and Thomas Find Kokfelt at GEUS collected trace element data for zircon. Sascha Müller prepared the U-Pb concordia and other diagrams. The manuscript was mainly written by Sascha Müller, with the help and suggestions from all co-authors. All authors assisted in the data interpretation and discussion.

Chapter 5 summarizes the main results of this thesis, discusses problems that can arise during the study of incompletely retrogressed eclogite, and gives an outlook on further research that could be carried out to improve the present dataset. Data tables are given in Chapters 2-4 and in the

Appendices.

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2 Summary of the 2014 fieldwork carried out in the Kuummiut Terrane of the Paleoproterozoic Nagssugtoqidian Orogen, South-East Greenland

A. Dziggel, S. Müller

2.1. Preface

This report presents the results of fieldwork carried out during the 2014 field season in South-East Greenland in the framework of the joint GEUS-MMR ‘SEGMENT’-project (2009-2016), focusing on reassessing the geology and mineral potential of the area between 62°30’N and 66°30’N. The main results of the SEGMENT-project are reported in Kolb et al. 2016 [Kolb, J., Stensgaard, B.M. & Kokfelt, T.F. Danmarks Og Grønlands Geologiske Undersøgelse Rapport 2016/38, 157 pp.].

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2.2. Introduction

This report covers work done during the 2014 SEGMENT expedition to South-East Greenland by Annika Dziggel (ADZ) and Sascha Müller (SMU) (team 2) – both from RWTH Aachen University, Germany. Fieldwork was conducted during four weeks (14.07.14 – 10.08.14) throughout the northern part of the Nagssugtoqidian Orogen (Kuummiut Terrane) in seven different camp sites (Fig. 2.1), and included several reconnaissance stops.

The main interest of fieldwork lied on understanding the tectono-metamorphic evolution of the high-pressure rocks preserved in Paleoproterozoic mafic dykes that transect the Archean country rocks (TTG gneiss) of the Kuummiut Terrane (Andrews et al., 1973; Wright et al., 1973; Bridgwater et al., 1990; Kolb, 2014). Fieldwork and sampling therefore concentrated on metamafic dykes either containing eclogite-facies mineral assemblages (mainly garnet and omphacite) or assemblages that have not been affected severely by retrograde metamorphism (i.e. rocks in which relict eclogite-facies minerals or replacement textures are still recognizable). During fieldwork, however, it was noted that relict high-pressure mineral assemblages are also preserved in boudins and boudinaged layers of mafic to ultramafic supracrustal rock of the Kuummiut Terrane, and consequently these were also

sampled. Different camp sites were chosen after detailed literature research and planning in the field, in order to cover as many occurrences of eclogite and its retrogressed equivalents as possible. In addition, the different camp positions (Fig. 2.1) allowed for an investigation of potential differences in the metamorphic grade and PT-evolution of the Kuummiut Terrane.

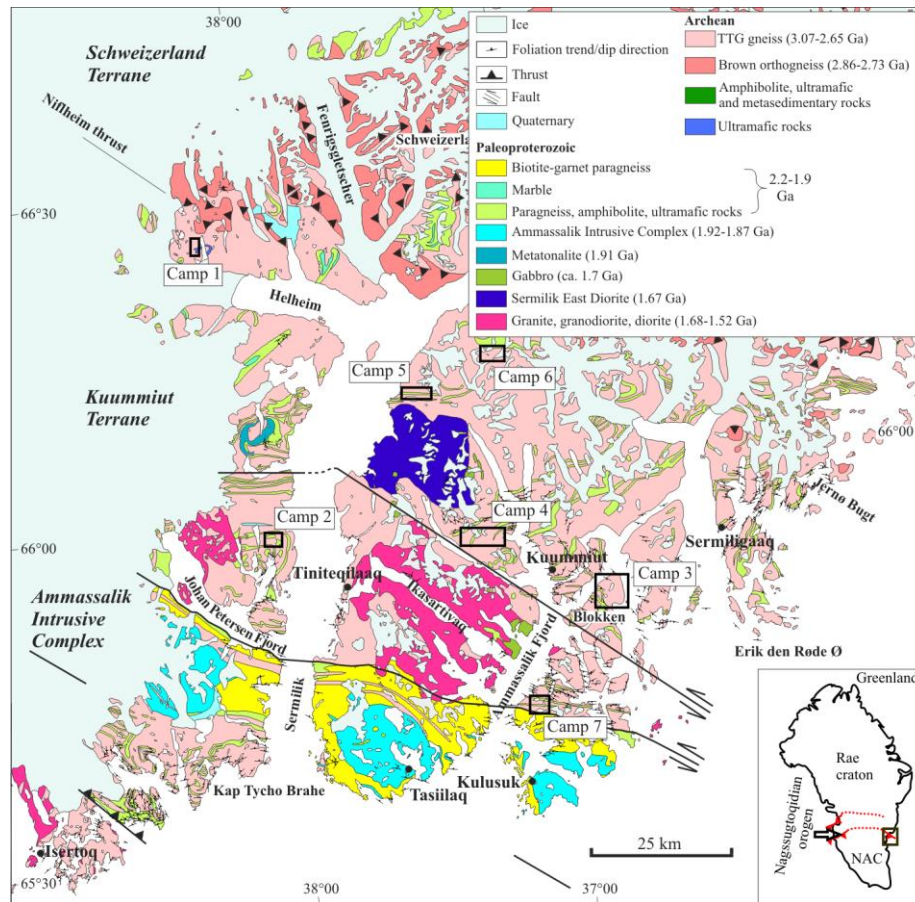


Fig. 2.1. Geological map of the northern Nagssugtoqidian Orogen in South-East Greenland (modified after Escher, 1990). Rectangles outline the investigated and sampled camp sites.

Three out of the seven camps were (at least for some time) joint camps with other teams. Camp 1 was shared with Kristoffer Szilas (KSZ) – Stanford University, USA, Jonas Tusch (JOT) – University of Cologne, Germany and Sam Weatherley (SMW) – GEUS from team 4, as well as Matti Nellesmann Petersen (MNP) – GEUS to help us with the use of the android device and aFieldwork app. In camp 2 we were visited by Jochen Kolb (JKOL) – GEUS and Anne Brandt Johannesen (ABJ) – University of Oulu, Finland from team 1. Vincent van Hinsberg (VIVH) – McGill University, Canada and Majken D. Poulsen (MADP) – GEUS of team 6 shared camp with us at camp 5.

A list of localities, including camp locations with working periods and GPS-data is given in Appendix 1. Over the course of the fieldwork, a total of hundred and twelve samples were collected, including gneiss, granite, diorite and carbonate rock. They, together with samples collected during earlier field

seasons, are listed in Appendix 2. Nevertheless, detailed petrographic and geochemical examinations, such as whole-rock major and trace element analysis, as well as EPMA (Electron Probe Micro-Analyzer) mineral-chemical analysis, mostly focus on the collected eclogitic and amphibolitic rocks. The following sections summarize our main observations during fieldwork for the seven camp sites (Chapter 2.3), as well as the three reconnaissance days (Chapter 2.4). Please note that because the structural geology is complex, it is not possible to correlate the different fabrics across the region, and the $D_{1,2,3}$ nomenclature used to document the sequence of deformation events in each locality only refers to the local structural evolution. More detailed information regarding the mineral textures, geochemistry and PT-history of the variably retrogressed eclogite-facies rocks of the Kuummiut Terrane is presented in Müller et al. (2018; Chapter 3).

2.3. Camps

2.3.1 Camp 1. North of Helheim glacier

Camp 1 covered an area north of the Helheim glacier, in which we worked from July 15 to July 17 2014. The camp area is dominated by dioritic to TTG-type (Trondhjemite-Tonalite-Granodiorite) gneiss and an up to 50 m thick east-northeast to west-southwest trending supracrustal belt made up of ultramafic lithologies and garnet amphibolite. The contact between the TTG gneiss and the mafic to ultramafic sequence is tectonic, primary intrusive contact relationships between the two rock types were not observed. The northern and southern margins of the supracrustal belt are marked by mylonitic shear zones.

The TTG gneiss is fine- to medium-grained, and typically consists of quartz, plagioclase, hornblende, biotite, and, locally, garnet. It contains a well-developed foliation that is defined by a fine layering and the preferred orientation of biotite and hornblende, as well as amphibolitic schlieren and plagioclase-rich leucosomes (Fig. 2.2a).

Garnet-amphibolite mainly occurs along the southern and northern margin of the supracrustal belt, and has a mineral assemblage of garnet, hornblende, plagioclase, quartz (Fig. 2.2b), and, locally, clinopyroxene and ferrous sulfides. Leucosomes with concentrations of peritectic garnet and clinopyroxene have also been observed in places (Fig. 2.2c). Clinopyroxene, however, has mostly been found in layers enriched in amphibole, and is locally associated with chalcopyrite (Fig. 2.2d). Retrograde epidote is mainly restricted to schistose to mylonitic amphibolite. Large garnet grains (up to 5 cm) in the amphibolite exhibit composition-related zoning and plagioclase coronas along their margins, indicating decomposition during uplift.

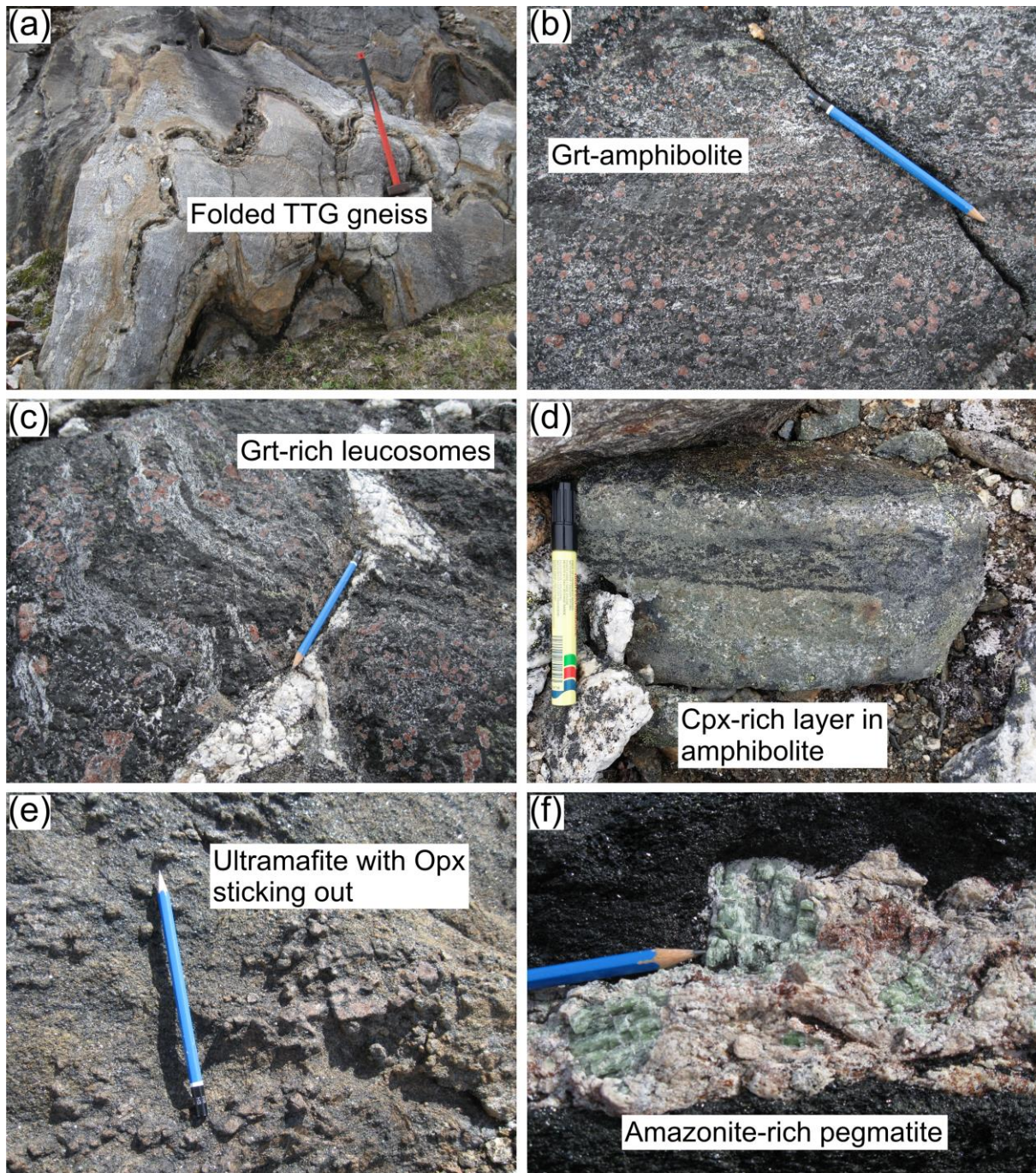


Fig. 2.2. Field photographs of various rock types and structures in the camp 1 area.

Ultramafic lithologies are the dominant rock type within the supracrustal belt. They either occur as schistose or massive units with a general mineral assemblage of olivine, plagioclase, garnet, and hornblende. Several ultramafic units also contain chromite and orthopyroxene, the later forming up to 1 cm large porphyroblasts that stick out due to being more resistant to weathering than the matrix minerals (Fig. 2.2e).

The whole sequence has been intruded by several generations of pegmatite dykes. Foliation parallel and locally amazonite-bearing pegmatites (Fig. 2.2f) have been observed as boudinaged layers in

mylonitic gneiss and epidote-bearing amphibolite schist. More commonly, pegmatites made up of garnet, biotite, plagioclase, hornblende, and quartz can be observed that crosscut all lithological units. The contact zone to TTG gneiss is marked by a hornblende selvage.

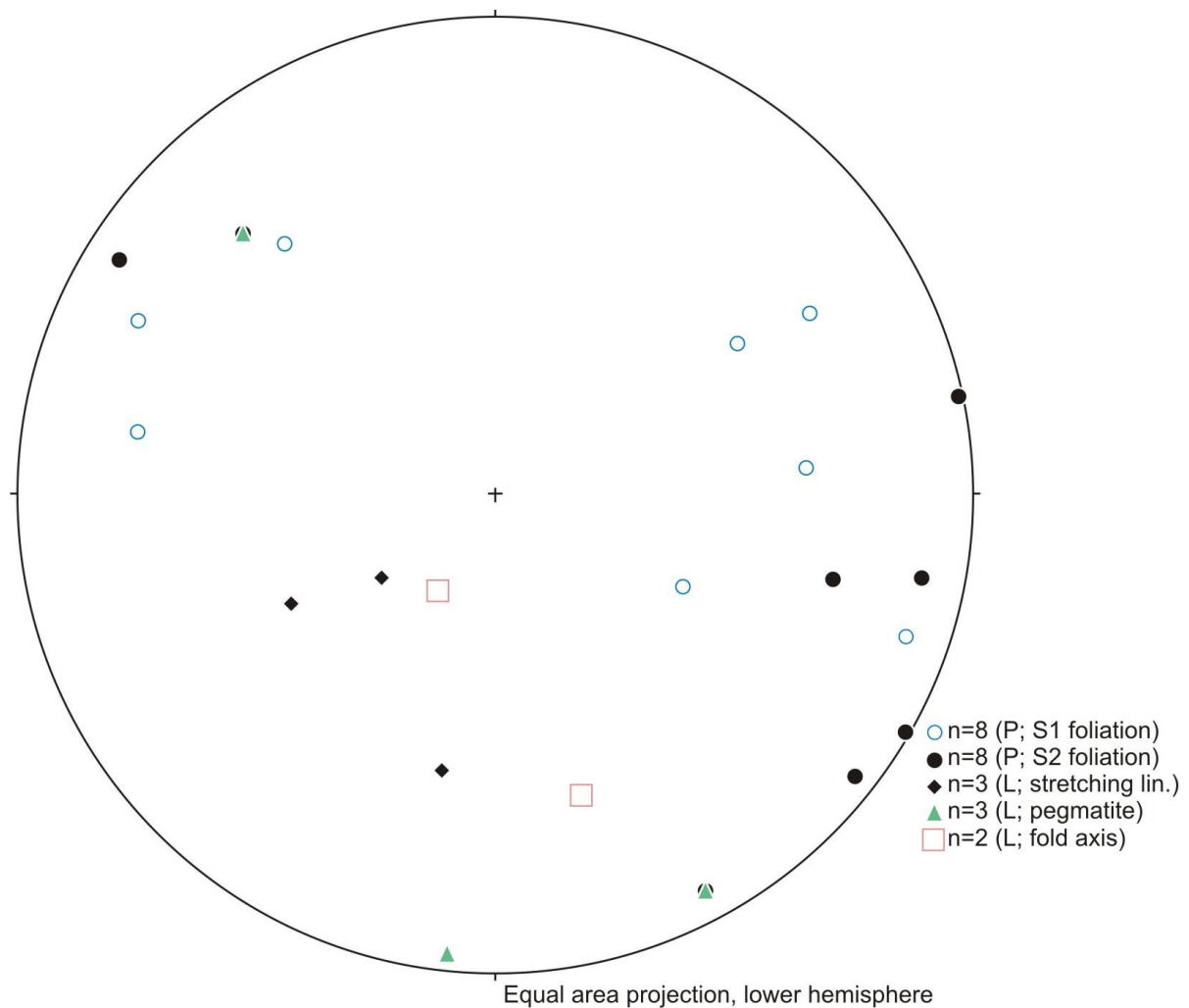


Fig. 2.3. *Stereographic projection of structural data collected at camp 1.*

The earliest fabric preserved in the camp 1 area is a variably developed S_1 foliation that is parallel to the lithological layering, and that dips at moderate to steep angles to the W/NW and E/SE (Fig. 2.3). The associated mineral stretching lineation plunges at moderate to steep angles to the SW. S-C fabrics and garnet sigma clasts point to an oblique reverse sense of movement that was broadly to the NE. Open to close F_2 folds refold the S_1 foliation. The fold axes plunge at moderate to steep angles to the S and SW, more or less parallel to the mineral stretching lineation. A second foliation (S_2) overprints the S_1 foliation, and is best developed in mylonitic shear zones along the northern and southern margin of the supracrustal belt. Along fold limbs, the S_2 foliation is more or less parallel to S_1 , but is usually steeper than the former (Fig. 2.3). The S_2 foliation is also synchronous with the intrusion of up to

several m wide pegmatite dykes that locally crosscut the S_1 foliation.

2.3.2 Camp 2. North of Johan Petersen Fjord

Camp 2 is located west of the Sermilik Fjord and north of the Johan Petersen Fjord (Fig. 2.1). We stayed here from July 18 to July 22 2014. Rock types in this area include TTG gneiss, as well as a variety of mafic and ultramafic lithologies that form elongate, up to 100 m wide and folded supracrustal belts, or occur as boudins and boudinaged layers of variable size within the TTG gneiss. In addition, mafic dykes intrusive into TTG gneiss have been observed. In general, the mineral assemblages in all these rock types point to amphibolite-facies conditions, even though mineral textures in some cases indicate a higher-pressure origin.

TTG gneiss is the most dominant type of rock, often occurring as a leucocratic and strongly foliated rock, with a mineral assemblage of quartz, plagioclase, biotite, and, locally, garnet and hornblende (Fig. 2.4a). Along the contact with mafic and especially ultramafic lithologies, the gneiss is hydrothermally altered and rusty (Fig. 2.4b).

Gneiss within these rust zones is rich in aluminum and contains high amounts of Al-rich silicates such as garnet, plagioclase, biotite, muscovite, sillimanite, and kyanite. The garnet is purple in color and may have gem-quality, as it has a high clarity and is generally free of inclusions; with a grain size of up to several cm. Kyanite forms up to several mm large crystals that are translucent to transparent and pale blue in color.

The ultramafic schist in contact with the rusty gneiss is composed of randomly orientated white tremolite, hornblende, olivine, garnet, and an unidentified dark green mineral, most likely clinopyroxene (Fig. 2.4c). The rock has a garbenschiefer-like texture and is often crosscut by pegmatite dykes.

However, pegmatite dykes are generally widespread and crosscut all rock types including the garnetiferous alteration zones, indicating that the hydrothermal overprint is most likely unrelated to their emplacement. Pegmatite dykes throughout the camp 2 area mostly consist of quartz, plagioclase, biotite and K-feldspar, except for one instance, where a pegmatite with open space quartz growth, muscovite and chlorite has been recorded. Apart from pegmatite, epidote and quartz veins crosscutting the different rock types were also observed.

The mineralogy and composition of mafic and ultramafic rocks is variable. Green, clinopyroxene-rich and up to several m large ultramafic boudins made up of clinopyroxene, quartz, plagioclase, and pyrite are locally present in TTG gneiss. The mineral assemblage in these ultramafic boudins is most likely of igneous origin, as indicated by the ophitic texture defined by plagioclase. Other ultramafic

rocks are composed of olivine, hornblende, and garnet, and locally contain veinlets of garnet, clinopyroxene and plagioclase.

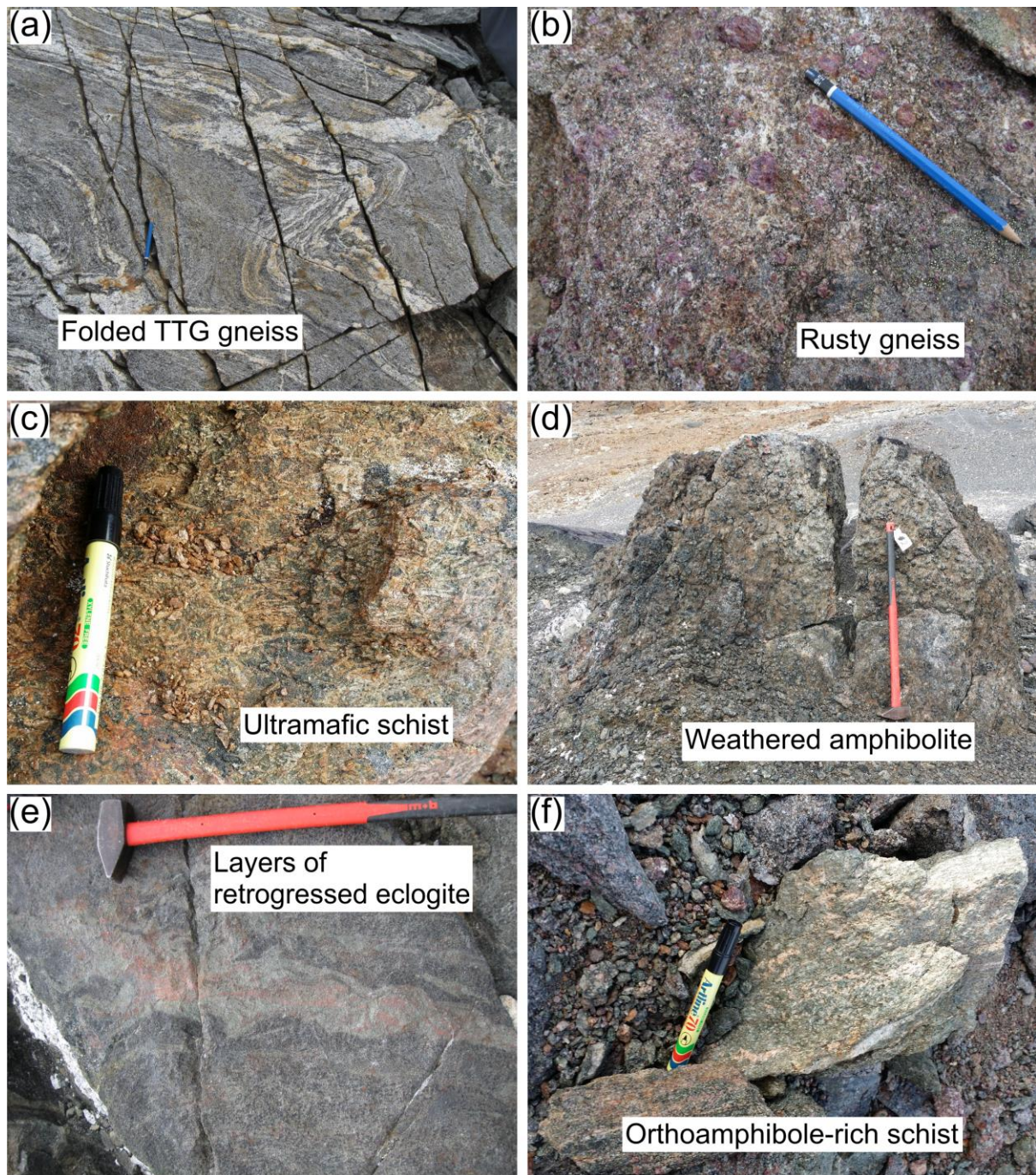


Fig. 2.4. Field photographs of various rock types and structures in the camp 2 area.

Several types of mafic rock can be distinguished based on mineral assemblages and textural characteristics. The most common type is a schistose to massive, medium- to coarse-grained amphibolite, consisting of garnet, quartz, plagioclase, hornblende, and, locally, biotite (Fig. 2.4d). Garnet in this rock is commonly rimmed by plagioclase coronas, indicative of decompression. Some

heavily weathered outcrops with honeycomb textures and large garnet sigma clasts contain foliation-parallel epidote layers. Due to a locally intense weathering, biotite within this rock appears green and flaky.

The second type of mafic rock is massive, medium- to coarse-grained and unfoliated, and occurs both within the supracrustal belts and in the center of boudinaged mafic dyke (Fig. 2.4e). This type of rock mainly consists of garnet and clinopyroxene that have been variably replaced by amphibolite-facies minerals such as hornblende and plagioclase. Replacement textures indicate that garnet and clinopyroxene represent relict high-pressure minerals, and this type of rock is therefore referred to as retrogressed eclogite (see Chapter 3).

The third type of mafic rock is a green amphibolite schist that is dominated by orthoamphibole and that contains minor amounts of magnetite and graphite (Fig. 2.4f).

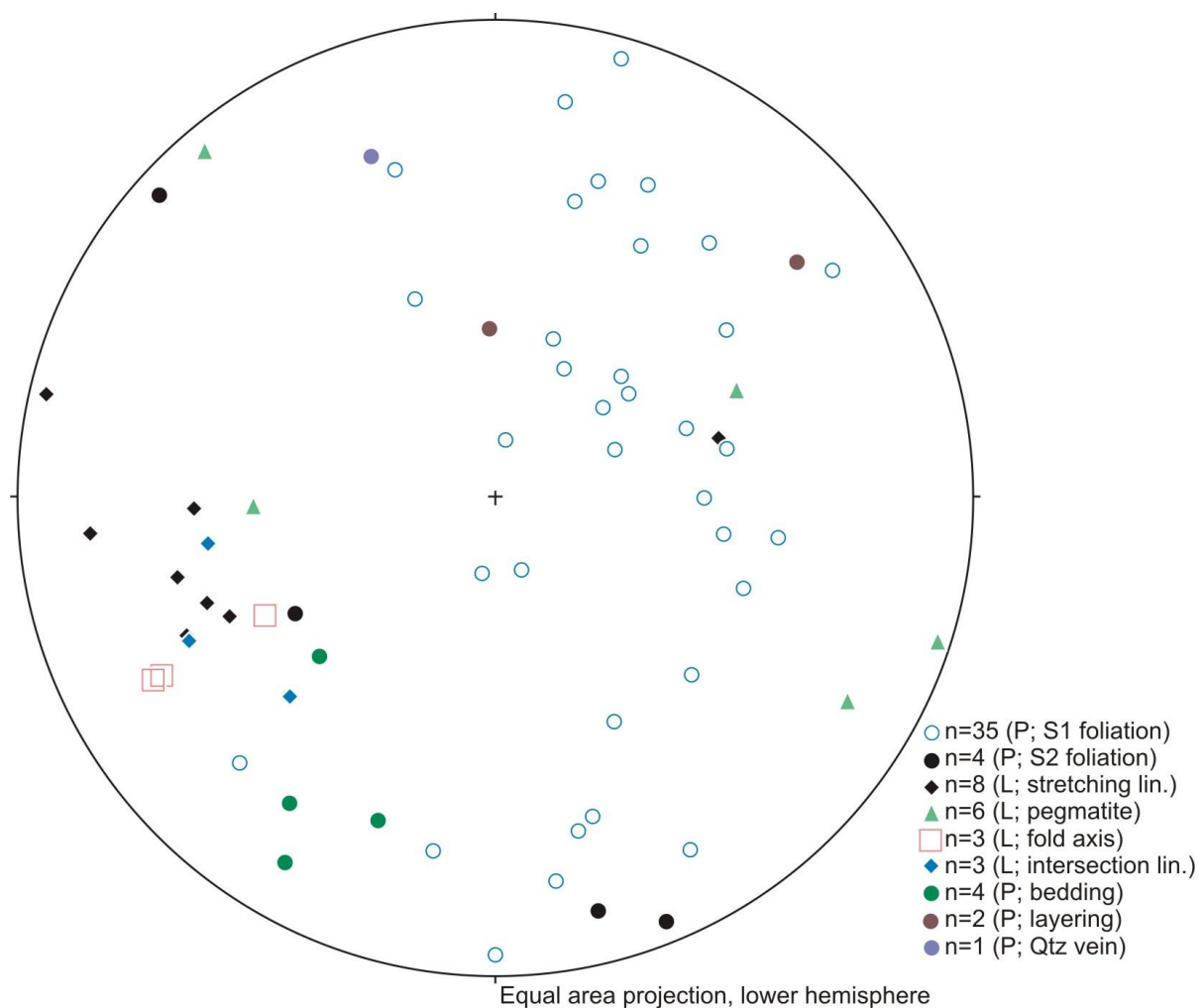


Fig. 2.5. Stereographic projection of structural data collected at camp 2.

Due to its position in the hinge zone of a major D₂ fold, the orientation of structures in the camp 2 area is highly variable (Fig. 2.5). The most prominent fabric is a pervasive S₁ foliation in TTG gneiss

and supracrustal lithologies that is parallel to lithological layering, and that is either subhorizontal, or dips at low to moderate angles mainly to the SW and NW. A locally developed mineral stretching lineation plunges at low to moderate angles mainly to the WSW. The S_1 foliation is folded by open to close F_2 folds. The fold axes, as well as the associated intersection lineations, plunge at low to moderate angles to the SW, broadly parallel to the L_1 lineation. A locally developed S_2 foliation overprints the earlier fabric, and dips at moderate to steep angles to the N and NE and to the SE (Fig. 2.5). Shear sense indicators, such as garnet sigma clasts, show a top to the west sense of shear. A conjugate set of pegmatite dykes crosscuts all earlier fabrics, and dips at moderate to steep angles to the NW and SE. It is parallel to S_2 in an SE dipping orientation, where shear sense indicators point to a sinistral sense of shear.

2.3.3 Camp 3. Blokken island

The camp 3 area (23.07.14-26.07.14) is situated in the north-eastern part of the island Blokken (Fig. 2.1). This island is mostly comprised of Archean TTG gneiss, the so-called Blokken gneiss (Bridgwater and Myers, 1979), Paleoproterozoic mafic dykes containing variably retrogressed eclogite, as well as supracrustal belts made up of amphibolite and kyanite-garnet schist.

The Blokken gneiss is schistose to mylonitic, and appears as a white (Fig. 2.6a) or rusty (Fig. 2.6b) unit with a common mineral assemblage of quartz, plagioclase, biotite, hornblende and locally garnet. Foliation-parallel quartz veins and pegmatitic layers were observed in a few localities (Fig. 2.6c), as well as epidote veins of different orientations in others. The pegmatitic layers in the Blokken gneiss contain, in addition to quartz, plagioclase, and hornblende, garnet, and K-feldspar.

Boudinaged and locally eclogitic mafic dykes intrusive into the Blokken gneiss are more common than in other areas. Retrogressed eclogite north of the camp site occurs as green-red, fine- to medium-grained, massive to weakly foliated bodies within the Blokken gneiss (Fig. 2.6d). The rock has a mineral assemblage of garnet, clinopyroxene, hornblende, plagioclase and quartz. To the south, the retrogressed eclogite locally also contains biotite. The boudins are frequently crosscut by pegmatite dykes, which are made up of quartz, plagioclase and biotite.

The garnet-amphibolite occurs as fine- to medium-grained, boudinaged and foliation-parallel layers within gneiss (Fig. 2.6e). Apart from clinopyroxene, it has a similar mineral assemblage as the retrogressed eclogite and often contains leucosomes. However, in contrast to the mafic dykes, the amphibolite is strongly foliated and associated with kyanite-garnet-schist. Garnet within the amphibolite shows plagioclase coronas and exhibits leucosome tails showing a top to the NE sense of movement. The kyanite-garnet schist is made up of garnet, kyanite, plagioclase, quartz, biotite, muscovite, and sillimanite (Fig. 2.6f).



Fig. 2.6. Field photographs of various rock types and structures in the camp 3 area.

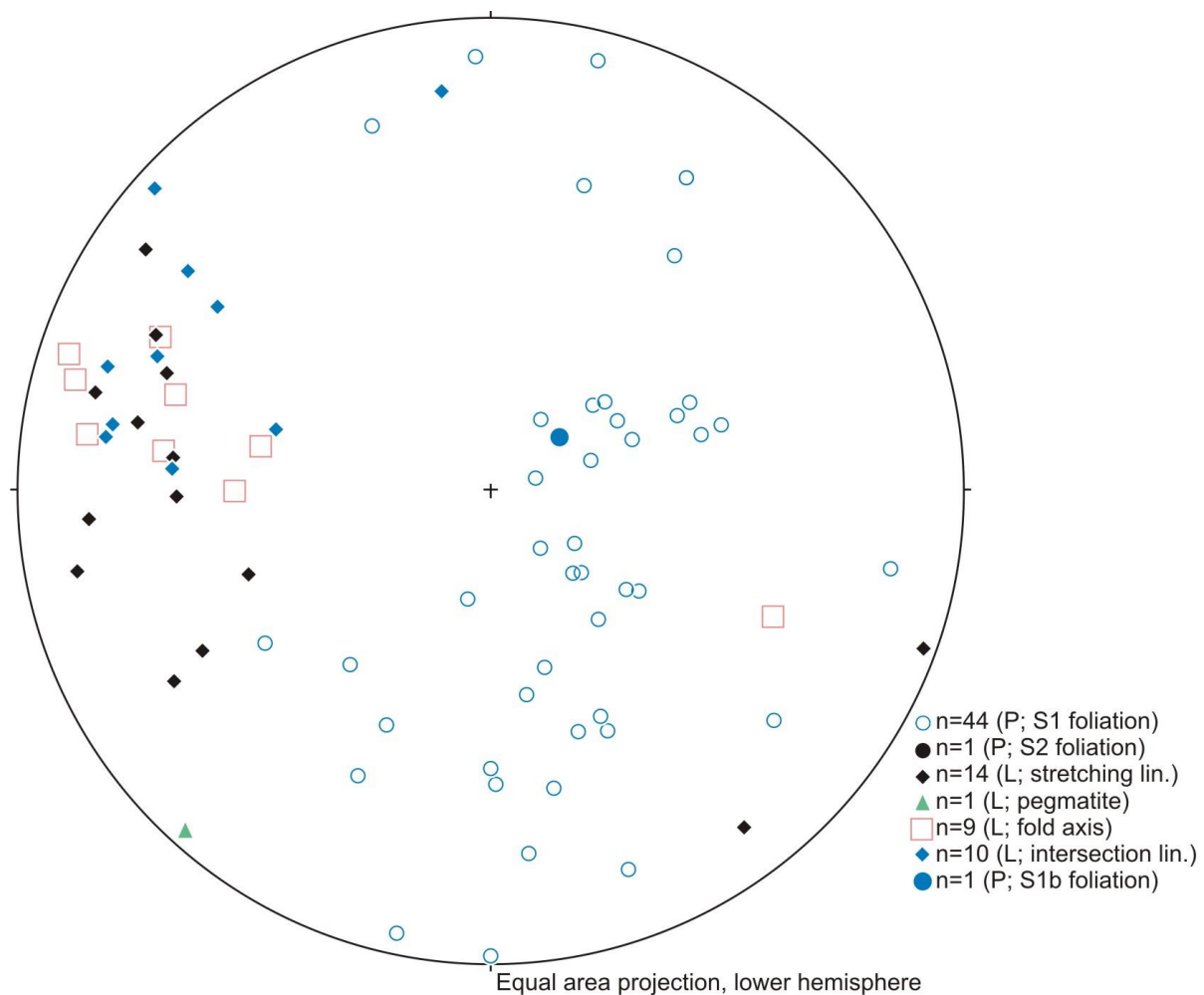


Fig. 2.7. *Stereographic projection of structural data collected at camp 3.*

The S_1 foliation in the camp 3 area dips at low to moderate angles mainly to the NW and SW (Fig. 2.7). The associated mineral stretching lineation is near downdip, and mainly plunges at low to moderate angles in a westerly direction, or to the SE. Shear sense indicators, such as S-C fabrics and leucosome tails around garnet point to a top-to-the east and northeast sense of movement. A locally developed S_{1b} foliation dips at shallow angles to the SW; S-C fabrics again indicate thrusting towards the NE (Fig. 2.7). The fold axes of the open to close F_2 folds, as well as intersection lineations, plunge to the WNW, and are either parallel or slightly oblique to the mineral stretching lineation (Fig. 2.7). A pegmatite dyke with a steeply NE dipping S_2 foliation was observed crosscutting an eclogite dyke at one locality. S-C fabrics indicate a dextral sense of shear during S_2 shearing.

2.3.4 Camp 4. Valley west of basecamp

The camp 4 area is located northwest of camp 3, in a valley west of the basecamp at the settlement of Kuummiut. Here, we conducted fieldwork between July 27 and July 29 2014. Contacts between different lithological units are mainly tectonic, and are generally very well exposed. Intrusive contacts

are only preserved around late- to post-tectonic, granitic and dioritic intrusions next to orthogneiss.

The dominant rock type in the camp 4 area is Archean TTG gneiss, which consists of biotite, garnet, plagioclase, quartz, and hornblende, and commonly contains leucosomes (Fig. 2.8a). In addition, pegmatite and amphibolite occur as boudins and foliation-parallel layers within the often layered TTG gneiss.



Fig. 2.8. Field photographs of various rock types and structures in the camp 4 area.

Supracrustal rocks are mainly preserved in up to 100 m wide east-west trending supracrustal belts. The supracrustal belts are dominated by amphibolite (Fig. 2.8b, made up of hornblende, plagioclase, quartz, and, locally, garnet) and garnet-kyanite schist of the general composition garnet, kyanite, (locally up to 20 cm large), biotite, quartz, and hornblende (Fig. 2.8c). Garnet-bearing amphibolite contains plagioclase coronas around garnet. Biotite is the main dark mineral in the garnet-kyanite schist, though its contents may vary from outcrop to outcrop and very biotite-garnet rich rocks locally occur. Some garnet-kyanite schist also contains plagioclase and sillimanite, either as matrix minerals or bright rims around garnet and kyanite, indicative of decompression. The schist has a rusty color in contrast to the bright gneiss and often contains foliation-parallel quartz veins and leucosomes. The garnet-kyanite schist and amphibolite are interlayered with TTG gneiss.

In the main valley of the camp 4 area close to a glacier stream, a highly silicic, layered rock consisting of almost pure quartz with traces of biotite and sulfides has been detected (Fig. 2.8d). The rock has a gossan-like weathering color, and may reflect a laminated quartz vein or highly silicified layered rock.

Eclogitic mafic dykes are locally preserved in TTG gneiss, and are best exposed in steep cliffs in the western and eastern parts of the camp 4 area. The dykes are boudinaged and oriented more or less parallel to the main foliation, indicating that they intruded pre- to syn-tectonically. Retrogressed eclogite sampled from gravel below a steep cliff with a large, subvertical mafic dyke hosted by TTG gneiss (Fig. 2.8e), shows a mineral assemblage of garnet, clinopyroxene, quartz, plagioclase, and hornblende. Further high-pressure mineral assemblages were observed in the core of a garnet-pyroxenite boudin hosted by TTG gneiss (Fig. 2.8f), which is mainly composed of garnet, orthopyroxene, clinopyroxene and hornblende. Along the contact with TTG gneiss, amphibolite-facies mineral assemblages have been identified.

Additional intrusive rocks in the camp 4 area comprise pegmatite dykes (Fig. 2.8f), as well as late- to post-tectonic diorite and granite. The pegmatite dykes are syn-tectonic, and either occur as foliation-parallel and commonly boudinaged layers within the orthogneiss, or crosscut the earliest fabric, and locally brecciate more competent lithologies such as mafic and ultramafic boudins. The youngest record of igneous activity is marked by late- to post-tectonic diorite and granite intrusions that crosscut the orthogneiss and pegmatite dykes. Within one intrusive body, a graduation from diorite to granite was observed, suggesting that they formed contemporaneously.

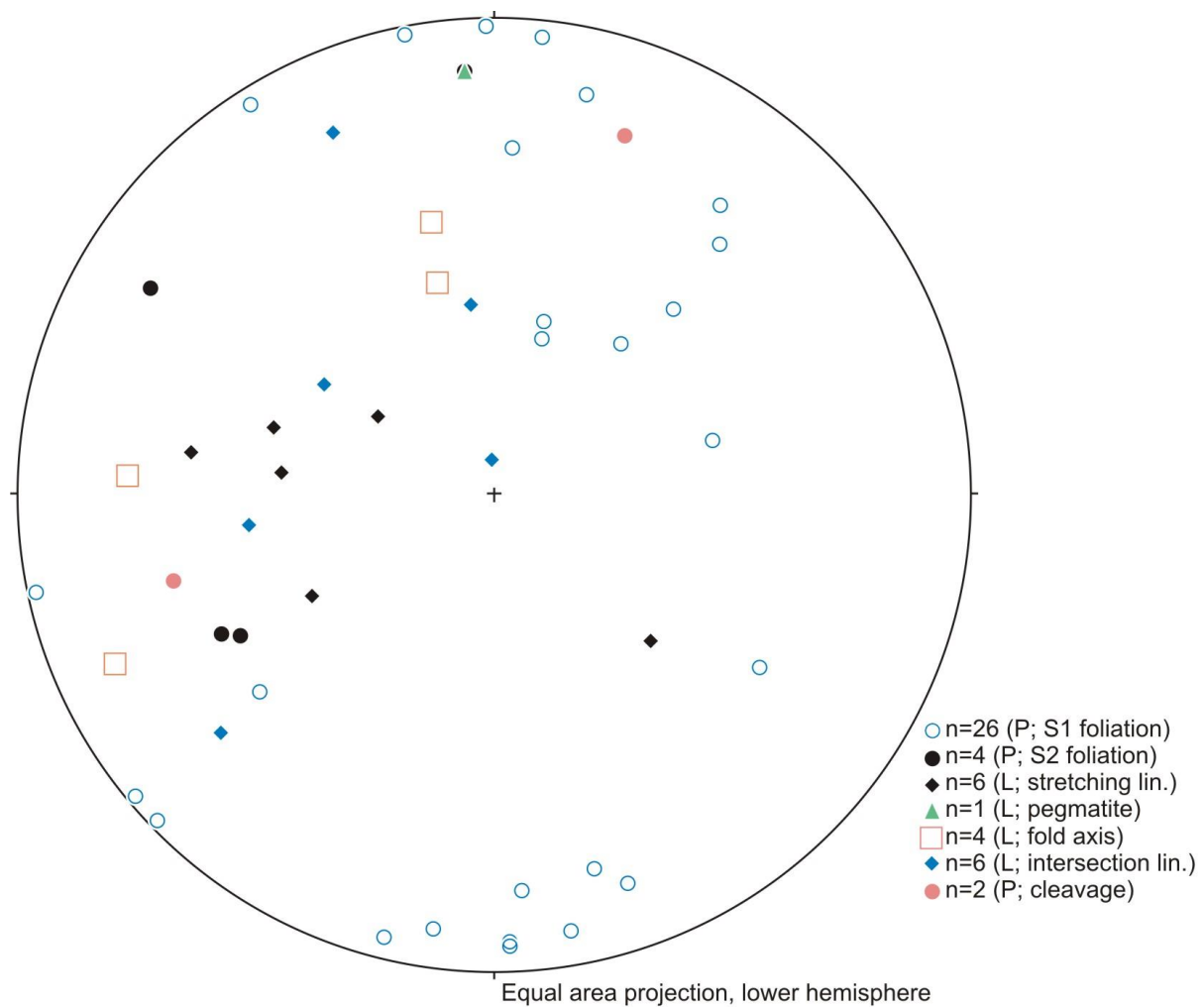


Fig. 2.9. *Stereographic projection of structural data collected at camp 4.*

The structural inventory of the camp 4 area can be explained in terms of three deformation events (Fig. 2.9). The earliest fabric is an E-W striking S_1 foliation, which is parallel to lithological layering, and which dips at moderate to steep angles mainly to the N and S (Fig. 2.9). This early foliation is locally associated with rootless isoclinal F_1 folds, and the associated mineral stretching lineation plunges at moderate angles mainly to the W (Fig. 2.9). Shear sense indicators, such as S-C fabrics point to a north-block-up sense of movement, with a dextral strike-slip component. The S_1 foliation has been folded by close to tight F_2 folds (Fig. 2.9), which fold axes plunge at moderate angles mainly to the W, parallel to the intersection lineations. At some localities, especially in the north-western part of the camp 4 area, the S_1 foliation dips at moderate to high angles to the NW and SE, and the fold axes plunge to the N and NW. Locally, the S_1 foliation is crosscut by a moderately to steeply E and S dipping S_2 foliation, which was sometimes observed to be synchronous with pegmatite intrusion.

2.3.5 Camp 5. North of the Sermilik East Diorite

Camp 5 is situated to the north of the Sermilik East Diorite (Fig. 2.1). Fieldwork was carried out from July 30 to August 01 2014. The area is dominated by TTG gneiss that encloses several east-west trending supracrustal belts made up of aluminous gneiss, marble and calcsilicate rock, as well as amphibolite schist.

The TTG gneiss is white to dark grey in color (Fig. 2.10a). The rock is characterized by an mm- to cm-scale intercalation of felsic and mafic layers, and primarily composed of biotite, quartz, plagioclase and hornblende. Epidote stringers were observed in one locality. The TTG gneiss often contains foliation-parallel and/or crosscutting pegmatite dykes and veinlets, as well as boudins and boudinaged layers of amphibolite.

Pegmatite dykes and veinlets are generally composed of quartz, plagioclase and biotite, but granitic compositions with K-feldspar have also been observed.

The aluminous gneiss (Fig. 2.10b) is more homogeneous than TTG gneiss, and mainly occurs at the contact between TTG gneiss and marble. The rock has a rusty weathering color, and a mineral assemblage of biotite, plagioclase, quartz, sillimanite, kyanite, muscovite, and garnet. Garnet is often replaced by plagioclase and may occur as lens-shaped porphyroblasts that stick out due to weathering.

Marble is massive to schistose and layered, and occurs in relatively thick sequences (sometimes more than 30 m) of interlayered calcareous and siliceous rock units. It has a bluish color on fresh surfaces (Fig. 2.10c) and is mainly composed of calcite and dolomite, with minor amounts of siderite and ankerite. Locally, marble is monomineralic and entirely composed of calcite. Monomineralic marble is interlayered with a darker type of carbonate rock, which contains variable amounts of white mica and biotite. Growth of siderite usually occurs in certain layers and on foliation planes along lithological contacts (Fig. 2.10d). Thin layers of calcsilicate rock predominantly occur at the contact between the marble and rusty gneiss. The calcsilicate rock mainly consists of quartz, carbonate, and, locally, clinopyroxene and/or wollastonite, and is locally associated with foliation-parallel laminated quartz-veins (Fig. 2.10e).

The amphibolite schist is fine-grained to mylonitic, and often contains mm-wide leucocratic layers that possibly represent leucosomes (Fig. 2.10f). It is usually composed of garnet, plagioclase, quartz, and hornblende, and locally contains boudinaged pegmatites and quartz veins, as well as plagioclase coronas and leucosome tails around garnet.

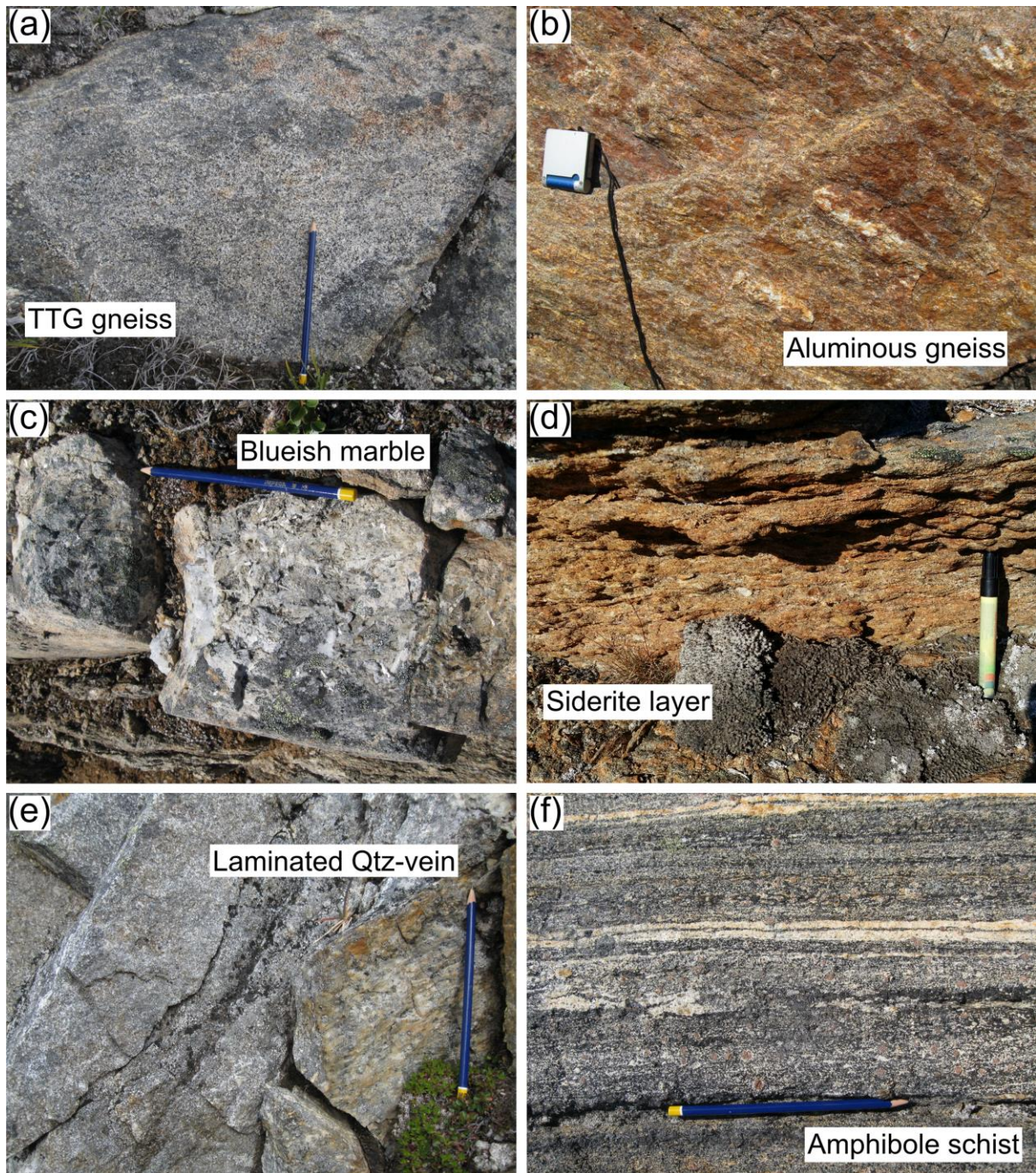


Fig. 2.10. *Field photographs of various rock types and structures in the camp 5 area.*

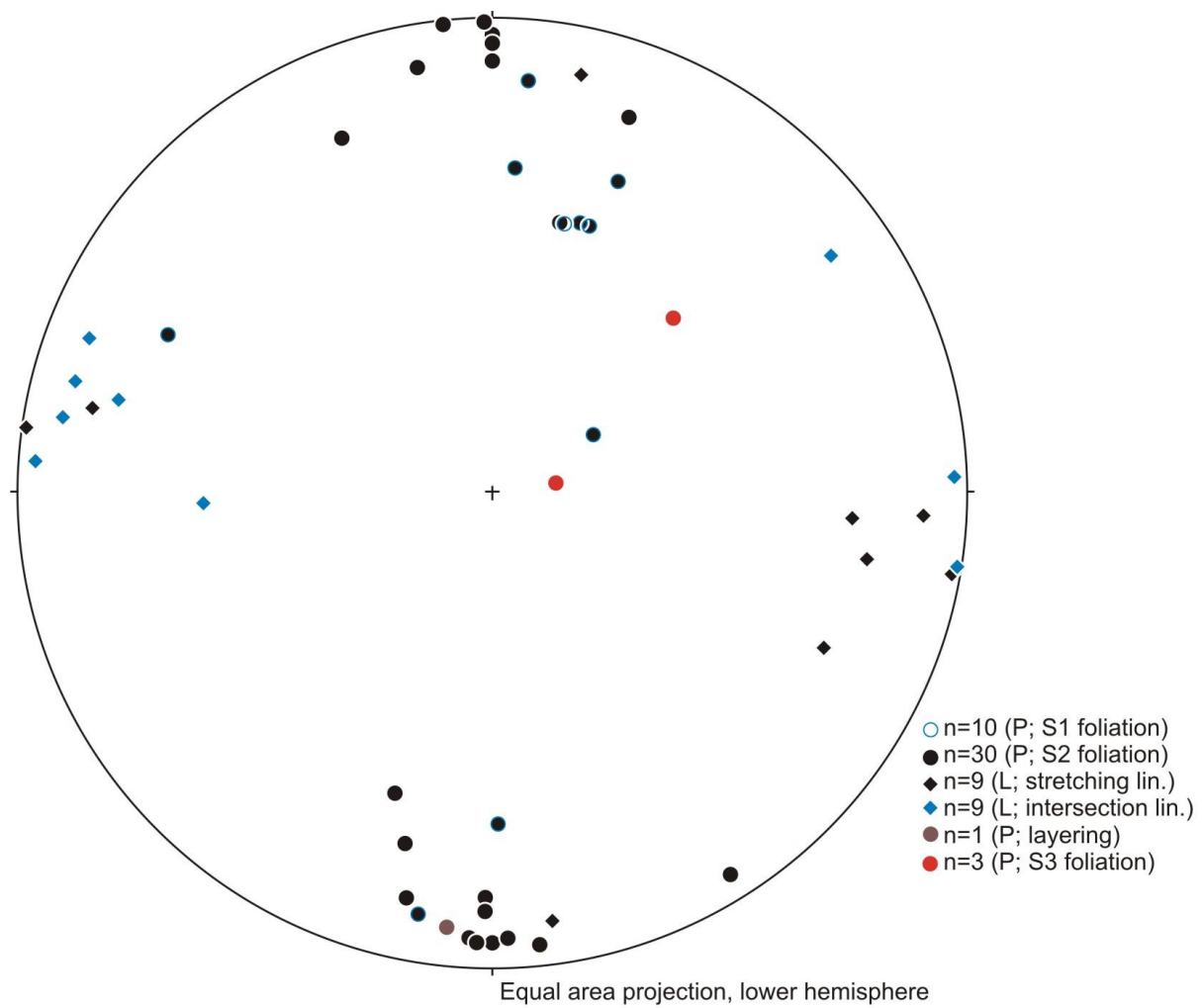


Fig. 2.11. *Stereographic projection of structural data collected at camp 5.*

The camp 5 area is dominated by an EW trending and up to 1 km wide strike-slip shear zone. Within this shear zone, the foliation dips subvertically to the N and S (Fig. 2.11), whereas outside the shear zone moderate dips are recorded. Based on its orientation and the presence of subhorizontal mineral stretching lineations, we interpret the shear zone to have formed during the regional D₂ deformation (Kolb, 2014). Away from the shear zone, the dominant fabric preserved in the gneiss is less steep and locally contains an early mineral stretching lineation. This fabric is therefore referred to as an S₁ foliation. Within the shear zone, the S₁ foliation is only preserved in the hinges of close to tight F₂ folds, and is otherwise transposed into the S₂ foliation. Locally, two different foliations that are slightly oblique to each other have been recorded. The L₂ lineations are subhorizontal, and plunge at low angles to the WNW and ESE, parallel to the intersection lineations (Fig. 2.11). Shear sense indicators variably point to a dextral or sinistral sense of shear, the latter of which seems to dominate.

2.3.6 Camp 6. South of the Niflheim thrust

Camp 6 (02.08.14-05.08.14) is situated in an east-west trending valley a few km south of the Niflheim thrust that forms the contact between the Kuummiut and Schweizerland Terranes (Fig. 2.1). Glacial gravel with boulder sizes of up to several meters occupies large areas where the glaciers have retreated. In contrast to all other camp sites, the gravel not only includes TTG gneiss, amphibolite or retrogressed eclogite, but also granulite-facies gneiss (clinopyroxene + orthopyroxene + leucosomes), which is likely derived from the Archean Schweizerland Terrane to the north.

The camp 6 area is comprised of four dominant rock types, including TTG gneiss, ultramafic rock, aluminous gneiss, and garnet amphibolite. Mafic dykes intrusive into the TTG gneiss also occur, but could not be sampled as they are only exposed in steep cliffs of the surrounding mountains.

The TTG gneiss is strongly layered, and has a general mineral assemblage of biotite, plagioclase, quartz, and garnet (Fig. 2.12a), locally with hornblende. The layering is defined by foliation-parallel leucosomes, and is particularly well-developed in garnet-rich varieties. In addition, the rock contains boudins and layers of amphibolite, ultramafic rock, and pegmatite. Crosscutting and foliation-parallel quartz veins are locally abundant.

The ultramafic rock appears as an up to several 100 meters thick, coarse-grained and massive, unfoliated to foliated unit, ranging in color from greenish to brown and locally black (Fig. 2.12b). In some cases, a rusty weathering color was observed. Tremolite, hornblende, clinopyroxene, olivine, orthopyroxene and plagioclase constitute the dominant mineral assemblage. Except for local occurrences of magnetite, the ultramafic rock is non-magnetic. Along the contact of intruding pegmatite dykes and other felsic intrusive rocks, the ultramafic rock is rich in biotite.

The aluminous gneiss has a rusty weathering color, and consists of quartz, plagioclase, muscovite, biotite and, locally, garnet and sillimanite (Fig. 2.12c). Quartz veins either crosscut the foliation, or occur as foliation-parallel stringers and veinlets.

The garnet amphibolite is schistose to mylonitic, and mainly consists of hornblende, plagioclase, quartz, and garnet (Fig. 2.12d). Garnet is often rimmed by plagioclase coronas and in some cases partially pseudomorphed by plagioclase-hornblende symplectites. The rock contains up to 15 cm thick layers rich in garnet and clinopyroxene and is crosscut by pegmatite and quartz veins. Garnet and clinopyroxene in amphibolite of the camp 6 area are associated with leucosomes, suggesting partial melting at upper amphibolite-facies conditions and a peritectic origin of garnet and clinopyroxene. The leucosomes form foliation-parallel stringers and are locally associated with pegmatite seams (Fig. 2.12e), indicating that at least some of the pegmatite intruded during high-temperature metamorphism.

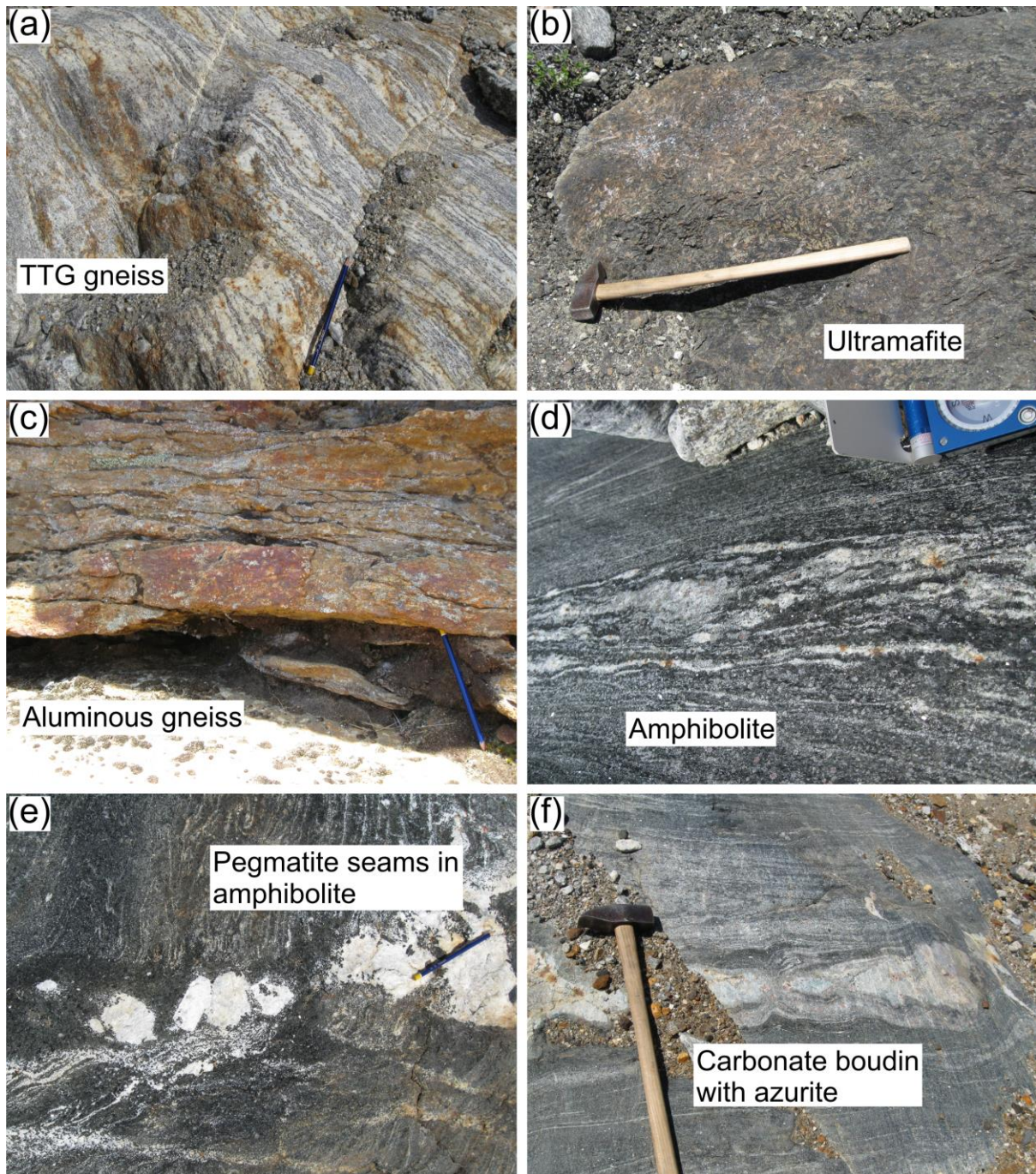


Fig. 2.12. *Field photographs of various rock types and structures in the camp 6 area.*

More commonly, however, the up to one-meter thick pegmatite dykes crosscut the foliation in the amphibolite. At one locality, up to several decimeters large carbonate and calcsilicate boudins hosted by amphibolite have been observed (Fig. 2.12f). The boudins are mainly made up of calcite and dolomite, and locally contain azurite along their margins.

Pegmatite crosscutting all units is generally composed of quartz, plagioclase and biotite, as well as hornblende in some localities.

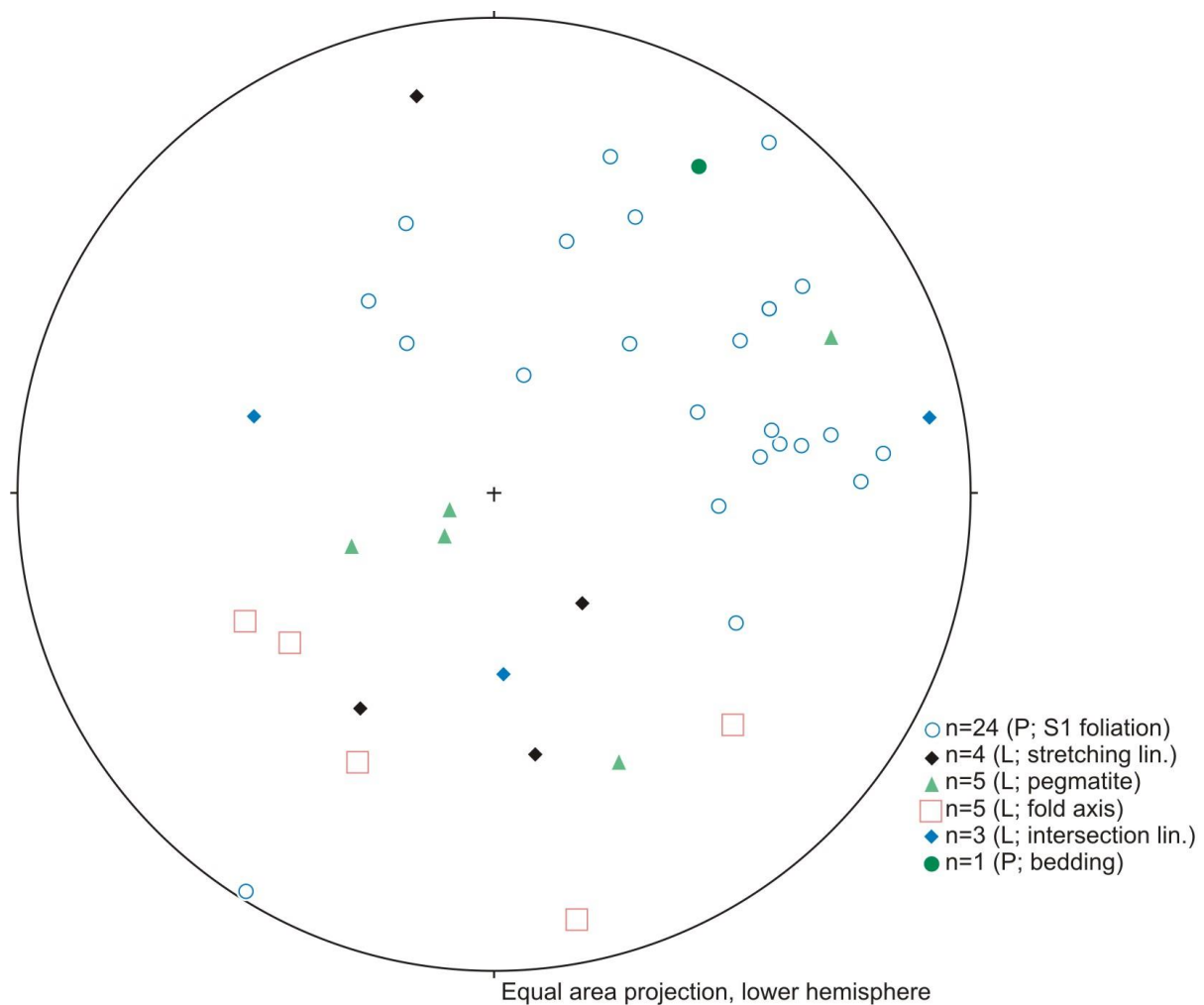


Fig. 2.13. *Stereographic projection of structural data collected at camp 6.*

The dominant fabric in the camp 6 area is a moderately, mainly W to SW dipping S_1 foliation (Fig. 2.13). Where present, the associated mineral stretching is either downdip, or plunges at low angles to the N. Shear sense indicators in the western part of the camp 6 area point to a top-to-the east and north-east sense of movement, whereas in the east, a north- or east-block-up sense of movement (i.e. extensional shearing) is indicated. The L_1 lineation is more or less parallel to the axes of close to tight F_1 (or F_2) folds. A conjugate set of crosscutting pegmatite dykes dips at low to moderate angles to the NE, NW, and SW (Fig. 2.13). The NW dipping pegmatite has a weak foliation along its margin and S-C fabrics point to a dextral shear sense, whereas the foliation in the NE dipping pegmatites indicates a sinistral shear sense.

2.3.7 Camp 7. Contact between the Kuummiut Terrane and Ammassalik Intrusive Complex

The last camp (05.08.14-07.08.14) is situated close to the contact between the Ammassalik Intrusive Complex and the Kuummiut Terrane east of the Ammassalik Fjord (Fig. 2.1). In this camp, we mainly

examined TTG gneiss and ultramafic rock of the Kuummiut Terrane.

The TTG gneiss is layered and has a typical mineral assemblage of quartz, garnet, biotite, plagioclase, and hornblende. The garnet content within the TTG gneiss varies, from very garnet-rich gneiss (Fig. 2.14a), to gneiss in which garnet is either absent (Fig. 2.14b) or restricted to certain layers. Rusty colored gneiss, interlayered with light grey to white gneiss, is more aluminous and contains sillimanite, as well as a higher amount of garnet. Where present, garnet is locally rimmed by biotite or hornblende and rich in inclusions. It is also often associated with leucosomes and shows leucosome tails (Fig. 2.14c). In highly sheared units, garnet is largely weathered out and leaves a honeycomb texture. The TTG gneiss contains amphibolitic (garnet, hornblende plagioclase, quartz), gabbroic (biotite, hornblende, plagioclase, olivine, pyroxene), and ultramafic layers and lenses of variable size.

As in other camp areas, the gneiss in the camp 7 area commonly contains both foliation parallel and crosscutting pegmatite dykes. They have a mineral composition of plagioclase, quartz, biotite, garnet and K-feldspar.

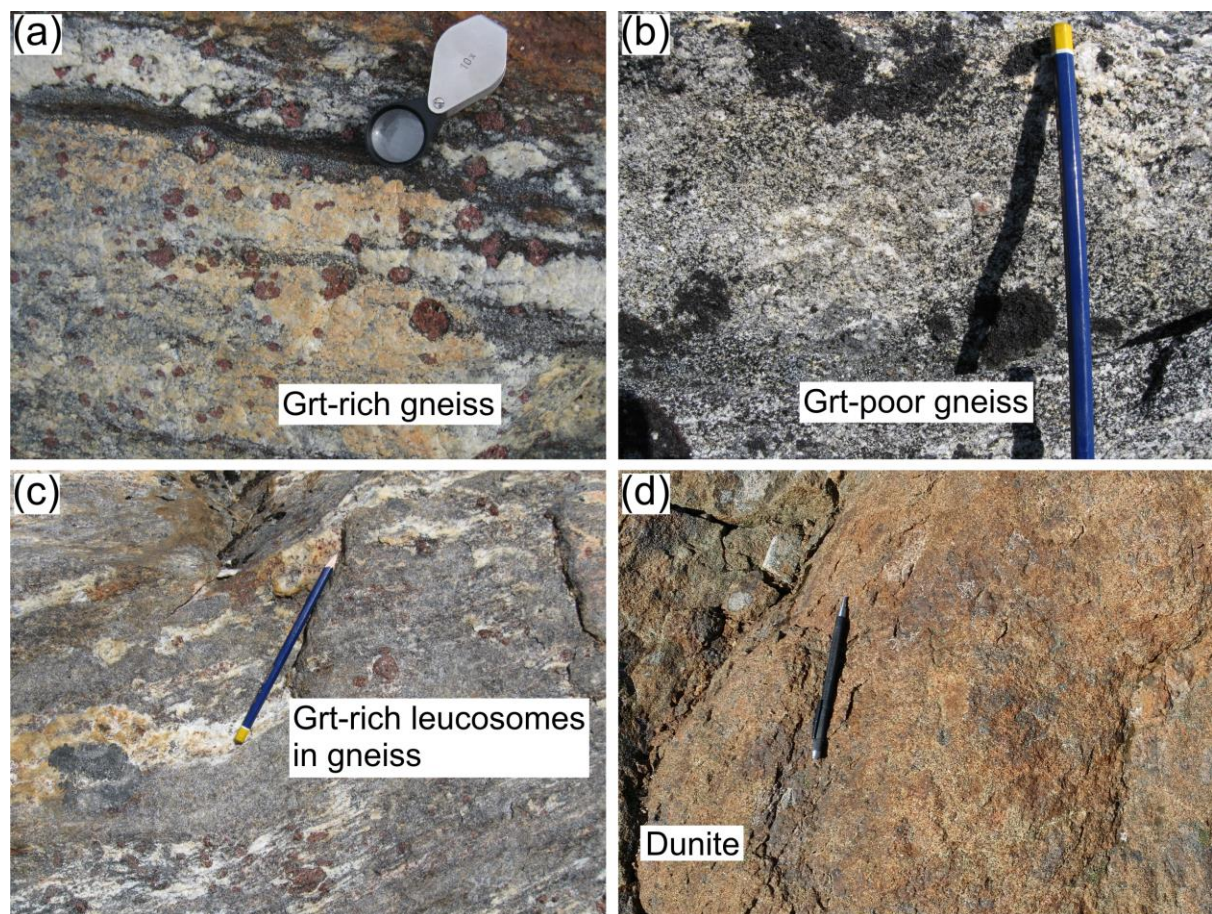


Fig. 2.14. Field photographs of various rock types and structures in the camp 7 area.

Ultramafic rock occurs as elongate, up to 100 m long bodies or as variably sized boudins within the gneiss (Fig. 2.14d). Most of the ultramafic rock is strongly weathered, with a brown, gossan-like weathering color. It has a lherzolitic mineral assemblage of olivine, orthopyroxene, garnet, clinopyroxene, and biotite, though olivine and garnet contents vary. Some of the ultramafic boudins are crosscut by quartz veins and pegmatite dykes. Along the margins of the dykes, the ultramafic rock is more mafic in composition, as indicated by the presence of quartz and a higher garnet abundance. More mafic compositions were also observed at the contact to the surrounding gneiss.

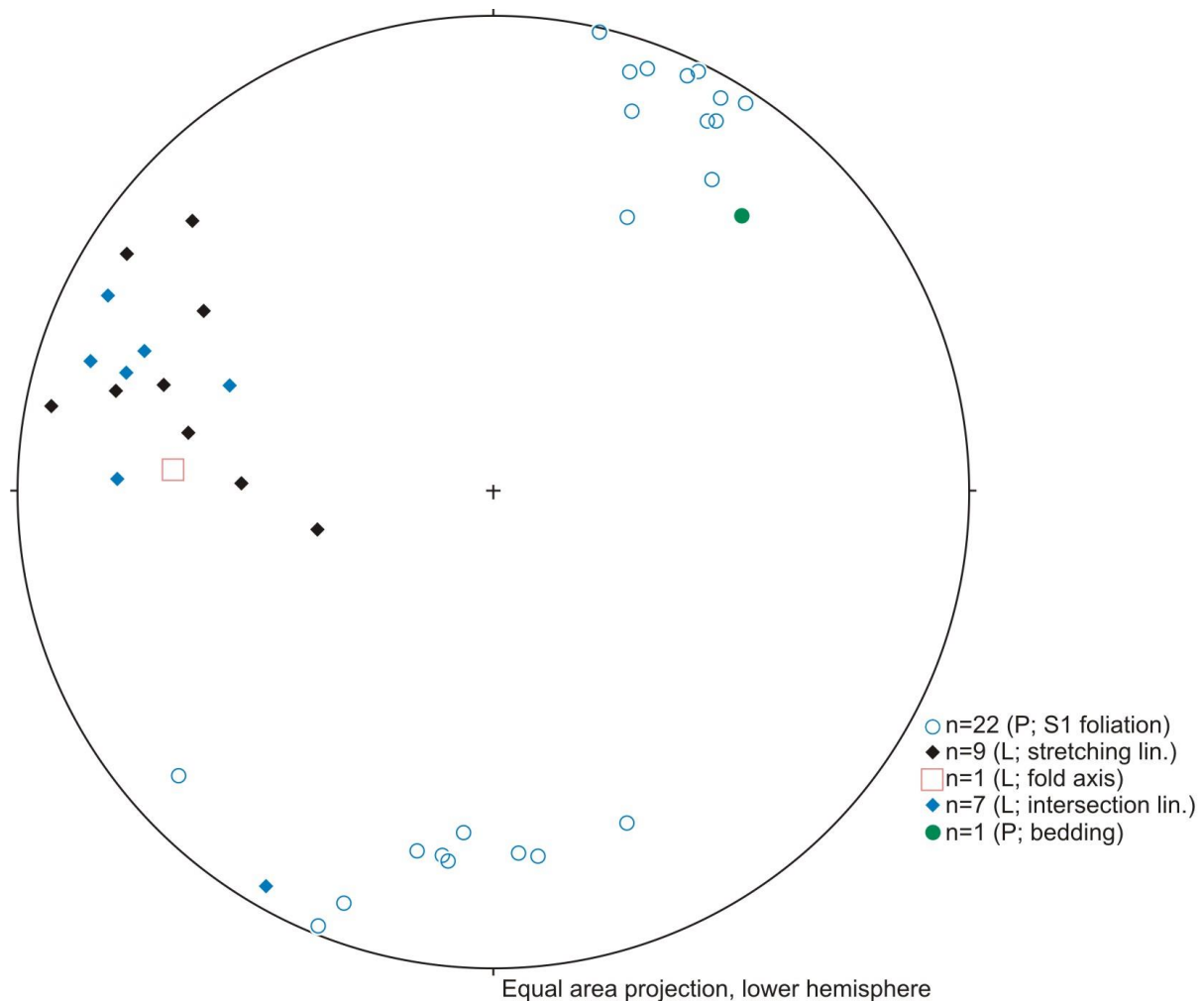


Fig. 2.15. Stereographic projection of structural data collected at camp 7.

The structural inventory of the camp 7 area is very similar to that of the camp 5 area (Fig. 2.15). The dominant foliation (here termed S₁) strikes W/NW to E/SE, and dips mainly at steep angles to the N/NE and S/SW (Fig. 2.15). The associated mineral stretching lineation plunges at low angles to the W/NW, parallel to the fold axes and intersection lineations of close to tight F₂ folds. Shear sense indicators, such as S-C fabrics and sigma clasts, point to an oblique-sinistral to oblique dextral sense of shear.

2.4. Reconnaissance stops

2.4.1 Reco day 1

The first day of reconnaissance work was carried out on July 21 2014 together with JKOL and ABJ (team 1), and focused on the contact relationships between the Ammassalik Intrusive Complex and the Kuummiut Terrane in the south-western part of the Kuummiut Terrane, in an area east of Sermilik Fjord (Fig. 2.1). The gneiss there is composed of quartz, plagioclase, hornblende, garnet, and K-feldspar. It is rich in garnet in the south and shows an oblique-reverse sinistral shear sense, consistent with the moderately westerly plunging mineral stretching lineations. Towards the north, the gneiss becomes less garnet-rich and more ultramylonitic, indicating increasing strain intensities during retrogression. Amphibolitic relicts within this northern gneiss are hence dominated by hornblende and plagioclase. Garnet reappears further north as small grains that are always associated to leucosomes. The various gneiss units are commonly intruded by pegmatite, which locally forms relatively thin (<20 cm) conjugate vein sets.

2.4.2 Reco day 2

On July 23 2014, we joined Leon Bagas (LBA) – School of Earth and Environment, University of Western Australia, Australia and Nanna Rosing-Schow (NRS) – University of Copenhagen, Denmark (team 3), as well as Bo Møller Stensgaard (BMST) – GEUS, on a reco day in the Auppalluttoq area west of the Sermilik Fjord.

Our first stop was at a sheared diorite with small leucosome veinlets and a hornblende-plagioclase-quartz-garnet assemblage. Garnet in the diorite is always associated with the leucosomes, indicating a peritectic origin. The S_1 foliation in the diorite dips at low angles to the S, and contains a well-developed mineral stretching lineation that plunges at low angles to the SE. The diorite is crosscut by a moderately easterly dipping pegmatite dyke with a mineral assemblage of quartz, K-feldspar, garnet, biotite, and plagioclase.

At next, we examined a contact between massive and weakly foliated amphibolite and layered TTG gneiss to the south of the last locality. The amphibolite mainly consists of hornblende, plagioclase, and quartz, and locally contains garnet, which is mostly replaced by plagioclase. Boulders of aluminous gneiss in the vicinity of the contact contain sillimanite and kyanite. The S_1 foliation dips at moderate angles to the S. An oblique S_2 foliation crosscuts the S_1 foliation and the pegmatite dyke, and dips at moderate angles to the SSW.

The last stop was at another amphibolite/gneiss contact. The gneiss here, however, is quite rusty and strongly foliated with a mineral assemblage of garnet, biotite, plagioclase, and quartz. The amphibolite schist has a mineral assemblage of hornblende, plagioclase, quartz, and garnet, and is interlayered with graphite schist. The foliation dips at moderate angles to the S, and the associated mineral stretching lineation plunges at low angles to the W. Shear sense indicators in the graphite

schist point to an oblique reverse sinistral sense of shear.

2.4.3 Reco day 3

During the third day of reconnaissance work on July 30 2014, we were accompanied by VIVH and MADP (team 6), as well as BMST. This reco day focused on some rusty gneiss and retrogressed eclogite localities around and near our campsite 2 (Fig. 2.1), with the observations for the rusty gneiss localities given in Chapter 2.3.2.

Some 5 km to the west of camp 2 (Fig. 2.1), we examined a sequence of highly-weathered garnet amphibolite, in which some layers preserve relict eclogite-facies mineral assemblages and well-developed, coarse-grained reaction textures between garnet and clinopyroxene (Fig. 2.16).

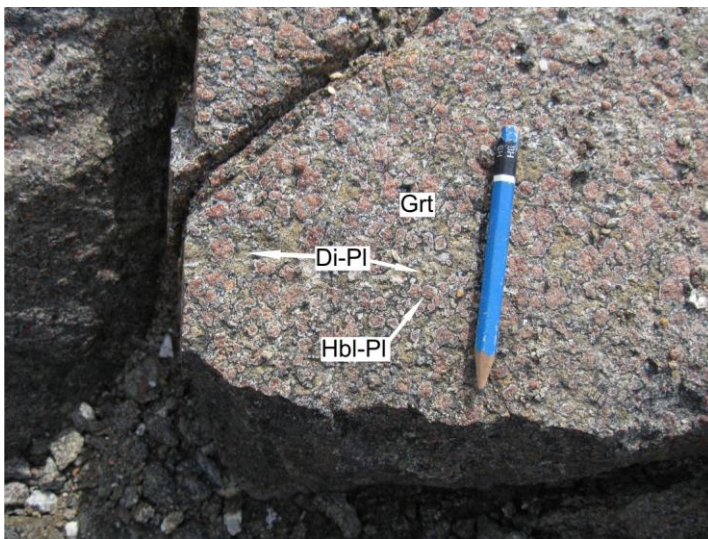


Fig. 2.16. Field photograph of mineral reaction textures in a mafic supracrustal rock (retrogressed eclogite) collected on Reco day 3.

Garnet in this mafic supracrustal rock is rimmed by coronitic plagioclase, and clinopyroxene has been partially replaced by hornblende. The amphibolite and retrogressed eclogite are intruded by up to 5 m thick pegmatite dykes. The foliation in the amphibolite and surrounding TTG gneiss dips at moderate to steep angles to the N and S, and the associated mineral stretching lineation plunges at moderate values to the W. A second foliation is developed in the pegmatite dyke, and dips at moderate values to the W.

Our next stop was at a ca. 200x300 m large ultramafic boudin at the northern end of the fjord to the east of the last locality (Fig. 2.1). This body, which is not shown in the geological maps, mainly consists of tremolite, magnetite, olivine, and orthopyroxene, and was too magnetic for structural readings. Towards the contact with the surrounding gneiss, the rock is more mafic in composition and contains garnet and clinopyroxene.

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3 Mineral textural evolution and PT-path of relict eclogite-facies rocks in the Paleoproterozoic Nagssugtoqidian Orogen, South-East Greenland

S. Müller, A. Dziggel, J. Kolb, S. Sindern

3.1 Abstract

The Nagssugtoqidian Orogen in South-East Greenland is a deeply eroded, Paleoproterozoic collision orogen. It consists of a variety of Archean and Paleoproterozoic rocks, most notably TTG gneiss, a variety of supracrustal rocks and basic dykes. This study aims at providing new insight into the geodynamic processes and subduction depth of this orogen by investigating the metamorphic evolution of garnet pyroxenite, retrogressed eclogite and amphibolite-facies rocks that are exposed within the Kuummiut Terrane of the Nagssugtoqidian Orogen. The garnet-pyroxenite has a dominant mineral assemblage of garnet, orthopyroxene, clinopyroxene and hornblende, while garnet-amphibolite and garnet-kyanite schist are made up of garnet, hornblende, plagioclase and quartz, and garnet, kyanite, biotite and quartz, respectively. Relicts of, and pseudomorphs after, eclogite-facies mineral assemblages are frequently found within basic metavolcanic rocks and Paleoproterozoic discordant basic dykes. In the retrogressed eclogite, the retrograde mineral reactions ceased prior to completion, resulting in the formation of two domains. A clinopyroxene domain consists of diopside-plagioclase symplectites, which are interpreted to have grown at the expense of omphacite. The symplectites are surrounded and partly replaced by hornblende and plagioclase. Omphacite (X_{Jd} 25–42) is preserved in a Na-rich sample, where it occurs in the core of large clinopyroxene and as inclusion in garnet and hornblende. In a garnet domain, garnet is variably replaced by an inner corona of plagioclase and an outer corona of amphibole $\pm$ orthopyroxene and clinopyroxene. The degree of retrogression as well as the type of the retrograde assemblage in both domains appears to be dependent on fluid activity. Large garnet grains preserve Ca-rich cores, interpreted as prograde in origin, while Mg-rich garnet rims formed during eclogite-facies metamorphism and later re-equilibration. Pseudosection modelling combined with conventional geothermobarometry reveals a clockwise PT-evolution, involving eclogite-facies conditions of 17–19 kbar and 740–810 °C, followed by near-isothermal decompression to high-pressure granulite-facies conditions (13.8–15.4 kbar, 760–880 °C) and subsequent decompression with minor cooling to high-pressure amphibolite-facies grades (8.8–10.9 kbar, 660–840 °C). These data show that rocks of the Kuummiut Terrane were exhumed from 70 to about 30 km into the mid- and lower crust. The PT-path implies that exhumation initially was rapid and tectonically-controlled.

3.2 Keywords

Eclogite, Nagssugtoqidian Orogen, Paleoproterozoic, Pseudosection modelling, Symplectites

3.3 Introduction

The Paleoproterozoic era (2.5–1.6 Ga) was a time of marked change in plate dynamics and plate-configuration (Rogers and Santosh, 2002; Zhao et al., 2002; St-Onge et al., 2009). Subduction to collision orogenic systems were presumably fundamentally different from today, owing to higher mantle temperature and a weak lithosphere, as seen in the lack of blueschist and ultrahigh-P (UHP) rocks preserved from this and earlier timespans (Bradley, 2011; Anderson et al., 2012). Other authors attribute the paucity of high- and ultrahigh-pressure rocks to a lack of preservation (Möller et al., 1995; Baldwin et al., 2004; Collins et al., 2004; St-Onge et al., 2006; Brown, 2008) or to the lack of tectonic environments able to produce and exhume such high-pressure rocks (Collins et al., 2004; Anderson et al., 2012). That some form of plate tectonics did exist even prior to the Paleoproterozoic is indicated by the rare presence of Meso- to Neoarchean eclogite, high-pressure granulite and paired metamorphic belts, which have been reported from e.g. the North Atlantic craton in Greenland (Tappe et al., 2011; Dziggel et al., 2014, 2017), the Belomorian Mobile Belt in Russia (Mints et al., 2014), the Udachnaya kimberlite in Siberia (Jacob et al., 1994), the Kaapvaal Craton in South Africa (Shirey et al., 2001) and the Lewisian complex in Scotland (Zirkler et al., 2012). Nevertheless, near the end of the Paleoproterozoic, the Archean tectonic style changed to a Proterozoic plate tectonic regime (Brown, 2008), coinciding with the amalgamation of continental crust, a major period of eclogite- and high-pressure granulite-facies metamorphism, widespread global orogenic activity and the formation of the oldest and relatively well-established supercontinent Nuna (also known as Columbia or Hudsonland; Hoffman, 1988; Rogers and Santosh, 2002; Zhao et al., 2002; Reddy and Evans, 2009; St-Onge et al., 2009; Mertanen and Pesonen, 2012). Our current understanding of the geothermal regimes and style of subduction in the Paleoproterozoic is poor. Collisional orogens of this age usually represent collision between two Archean provinces (Hoffman, 1988; Windley, 1995). Most of these orogens are dominated by TTG gneiss and often contain cross-cutting basic dykes of various metamorphic grade and age (Nutman and Friend, 1989; Zhao et al., 2001; Guo et al., 2002; Baldwin et al., 2004; Vuollo and Huhma, 2005; Wang et al., 2007; Hou et al., 2008; Nutman et al., 2008; Peng et al., 2009; French and Heaman, 2010). It is especially these basic dykes that have been shown to contain relict eclogite- and high-pressure granulite-facies mineral assemblages (Nutman and Friend, 1989; Zhao et al., 2001; Guo et al., 2002; Baldwin et al., 2004; Tam et al., 2012). To date, only few studies of Paleoproterozoic high-pressure metamorphic rocks have been carried out (Nutman and Friend, 1989; Messiga et al., 1990; Möller et al., 1995; Sklyarov et al., 1998; Zhao et al., 2001; Guo et

al., 2002; Baldwin et al., 2004; Nutman et al., 2008; Tam et al., 2012). These studies show that the high-pressure rocks are often intensely retrogressed at amphibolite-facies conditions during exhumation. Nevertheless, the relict high-pressure assemblages and retrograde reaction textures provide important insight into the conditions of high-pressure metamorphism and the exhumation of these rocks in Paleoproterozoic collisional orogenic belts. In this study, we undertake a more detailed investigation of high-pressure rocks in the Nagssugtoqidian Orogen of South-East Greenland (Wright et al., 1973; Chadwick et al., 1989; Dawes et al., 1989a; Nutman and Friend, 1989; Messiga et al., 1990; Nutman et al., 2008). In order to provide new insights into the geodynamic processes and subduction depths of this Paleoproterozoic orogen, we combine detailed textural and mineral chemical analysis with conventional geothermobarometry and pseudosection modelling.

3.4 Geological setting

In the Tasiilaq region of South-East Greenland, the Nagssugtoqidian Orogen is a deeply eroded, roughly southeast-northwest trending, ~200 km wide, Paleoproterozoic collisional orogen (Fig. 3.1, Wright et al., 1973; Bridgwater, 1976; Myers, 1987; Chadwick et al., 1989).

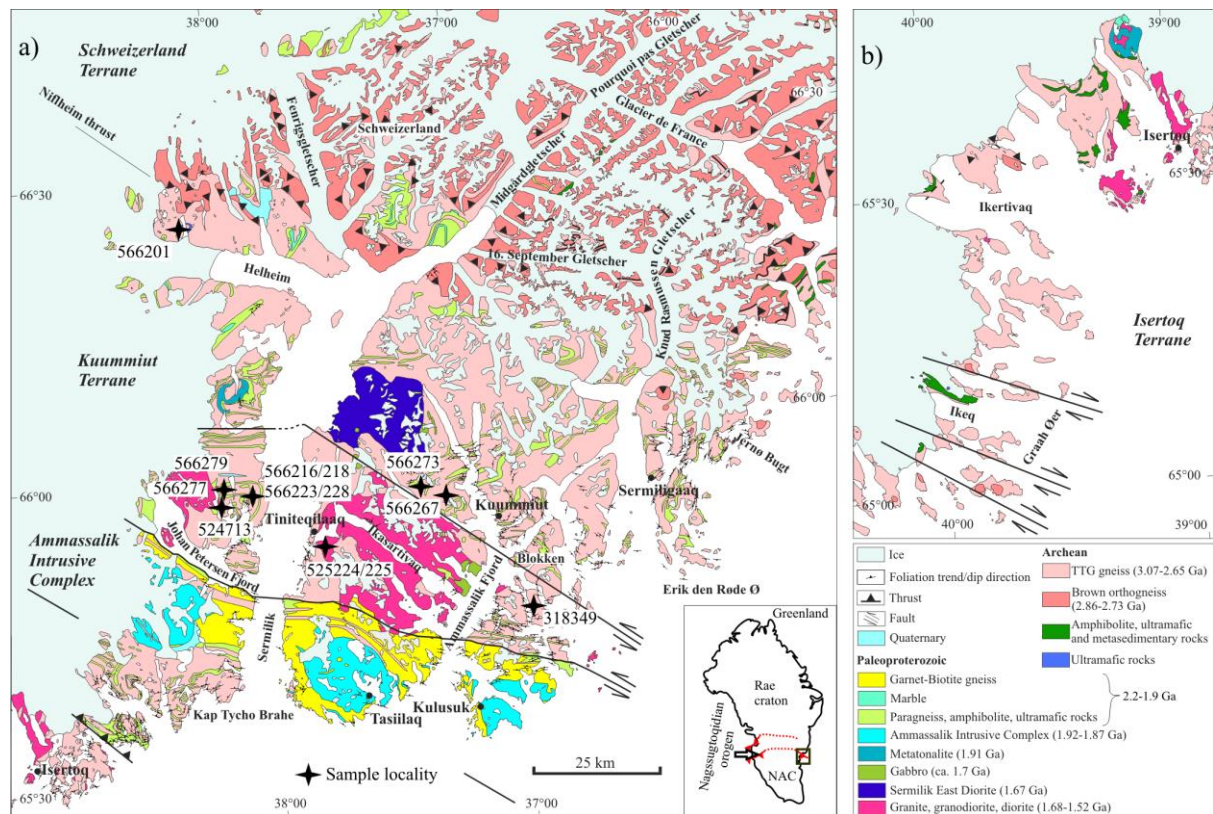


Fig. 3.1. Geological map of the (a) northern and (b) southern Nagssugtoqidian Orogen (modified after Escher, 1990). Stars in (a) represent sample sites.

This orogen is bound by two Archean cratons, including the Rae Craton to the north and the North Atlantic Craton to the south (Bridgwater, 1976; Chadwick et al., 1989; Kolb, 2014).

The boundary zones with the Archean cratons are either defined by near-vertical, dextral strike-slip shear zones (Myers, 1987) or in the south, by a zone of north-dipping shear zones in the Umivik area (Escher et al., 1989). The collisional orogen consists of a variety of Archean and Paleoproterozoic rocks that were affected by several stages of deformation and high-grade metamorphism (Andrews et al., 1973; Wright et al., 1973; Bridgwater et al., 1976; Nutman et al., 2008; Kolb, 2014). Based on differences in stratigraphy and tectono-metamorphic history, three different terranes can be identified, which from north to south are the Schweizerland, the Kuummiut and the Isertoq terranes, the latter including the Ammassalik Intrusive Complex (Fig. 3.1; Kolb, 2014).

The Schweizerland Terrane is characterized by Archean medium-pressure granulite-facies orthogneiss, which was variably retrogressed to hornblende-biotite gneiss towards the Niflheim thrust (Fig. 3.1a; Dawes et al., 1989b). Minor mafic rocks occur in narrow bands in the orthogneiss (Bridgwater and Myers, 1979; Myers, 1987; Dawes et al., 1989b). The orthogneiss yields zircon U-Pb ages of 2.86–2.73 Ga (Kalsbeek et al., 1993; Nutman et al., 2008; Kokfelt et al., 2016a). Granulite-facies metamorphism is interpreted to have occurred at ca. 2.72 Ga (Nutman et al., 2008). The orthogneiss was subsequently intruded by pegmatites during retrograde amphibolite-facies metamorphism, dated at ca. 2.63 Ga (Pedersen and Bridgwater, 1979).

The Kuummiut Terrane is a mainly high-pressure amphibolite-facies terrane dominated by Neoarchean quartzo-feldspathic gneiss and bands of supracrustal rocks (Fig. 3.1a). The latter were previously assigned to the early Proterozoic Síportôq Supracrustal Association (SSA, Chadwick et al., 1989; Hall et al., 1989) and more recently recognized as geologically and geochronologically different units (Kolb, 2014). The fine- to medium-grained Archean gneiss is of granodioritic to tonalitic (TTG) composition, with a common mineral assemblage of quartz + biotite + plagioclase ± garnet ± hornblende ± K-feldspar (Dawes et al., 1989a). It contains a well-developed foliation, which is defined by compositional banding and the preferred grain orientation of biotite and hornblende (Chadwick et al., 1989; Dawes et al., 1989a). Locally, the regional orthogneiss is migmatitic, with polyphase leucosomes and pegmatite bands (Dawes, 1989). Zircon U-Pb dating gives an age range of 3.08–2.71 Ga for the Archean gneiss (Kalsbeek et al., 1993; Nutman et al., 2008; Kokfelt et al., 2016a), with granulite-facies metamorphism dated at ca. 2.72 Ga using zircon rims (Nutman et al., 2008). Later metamorphic and magmatic activity is reflected by a metamorphic zircon population in an Archean gneiss dated at ca. 2.68 Ga (Kokfelt et al., 2016a), a monzogranite collected at the Niflheim thrust (ca. 2.65 Ga, Kokfelt et al., 2016a) and an orthogneiss from east of Sermilik Fjord (ca. 2.64 Ga, Kalsbeek et al., 1993). Supracrustal rocks occur in up to 100 m wide, foliation-parallel belts or as boudins and boudinaged layers of variable size within the TTG gneiss. Paragneiss, quartzite and biotite schist constitute the Kuummiut unit and represent the dominant Paleoproterozoic rocks north of Tasiilaq and east of Sermilik Fjord (Kolb, 2014; Kolb et al., 2016), whereas metadiorite,

amphibolite, marble, calc-silicate rock, quartzite, kyanite/sillimanite-garnet schist, ultramafic rock and aluminous gneiss west and north of the fjord constitute the Helheim unit (Kolb, 2014; Kolb et al., 2016). Primary intrusive contact relationships between the TTG gneiss and supracrustal belts are not preserved, indicating large-scale imbrication (Wright et al., 1973). In some instances, the contacts are marked by hydrothermal alteration with gossan-like weathering (Bridgwater et al., 1976; Dawes, 1989). Kyanite-bearing and other metasedimentary rock units yield U-Pb zircon ages of 2.60 to 1.89 Ga for detrital grains, with complete recrystallization and new growth of zircons in the Paleoproterozoic (1.87–1.66 Ga, Nutman et al., 2008; Thrane et al., 2016). Swarms of east-west to east-northeast-west-southwest striking, discordant basic dykes occur as boudins or boudinaged layers within the TTG gneiss (Andrews et al., 1973; Wright et al., 1973; Chadwick et al., 1989; Dawes et al., 1989a; Bridgwater et al., 1990). These dykes are oriented sub-parallel or at an angle to the foliation, indicating that they intruded the Archean rocks prior to Paleoproterozoic deformation (Andrews et al., 1973; Bridgwater et al., 1973; Bridgwater, 1976; Bridgwater and Myers, 1979; Chadwick et al., 1989), consistent with a U-Pb zircon age of ca. 2.02 Ga for prismatic to jagged anhedral zircon grains from the dykes (Nutman et al., 2008). The cores of larger dykes locally preserve medium- to coarse-grained, relict high-pressure mineral assemblages that are variably retrogressed and that contain clinopyroxene-plagioclase symplectites after omphacite (Wright et al., 1973; Chadwick et al., 1989; Nutman and Friend, 1989; Messiga et al., 1990; Nutman et al., 2008). Smaller dykes or the rims of larger dykes typically represent fine- to medium-grained garnet amphibolite or amphibolite (Bridgwater et al., 1976; Chadwick et al., 1989; Nutman and Friend, 1989; Messiga et al., 1990). Homogenous to sector-zoned equant zircon grains from the basic dykes yield a $^{207}\text{Pb}/^{206}\text{Pb}$ age of ca. 1.87 Ga. This age is interpreted as the age of high-pressure eclogite-facies metamorphism, based on the absence of plagioclase and abundance of garnet and clinopyroxene inclusions in zircon and the lack of a significant negative Eu anomaly and depressed HREE abundances in zircon (Nutman et al., 2008). Samarium-Nd garnet-clinopyroxene whole-rock data of the dykes defines an isochron and yields an age of ca. 1.82 Ga (Kalsbeek et al., 1993). This age was attributed to a subsequent retrograde, lower-pressure stage along a high-temperature decompression path, in concordance with the observed mineral reaction textures (Nutman et al., 2008).

The calc-alkaline Ammassalik Intrusive Complex (AIC) to the south of the Kuummiut Terrane (Fig. 3.1a) consists of three ovoid intrusion centers, namely the Johan Petersen, Tasiilaq and Kulusuk intrusive centers (Kokfelt et al., 2016b; Lebrun et al., 2018). Their emplacement into the Isertoq Terrane was dated between 1.92 and 1.87 Ga (Hansen and Kalsbeek 1989; Kalsbeek et al., 1993; Nutman et al., 2008; Lebrun et al., 2018). Nutman et al. (2008) propose that the emplacement occurred prior to eclogite-facies metamorphism in the Kuummiut Terrane. In conjunction with earlier studies (Kalsbeek et al., 1993; Nutman et al., 2008), whole-rock major and trace element geochemistry, as well as Lu-Hf, Sm-Nd and O isotope data recently collected by Lebrun et al. (2018) indicate that the AIC most likely formed in a continental arc setting with mixing of melts from

multiple sources. The AIC is bordered to the north by a ~7 km wide mylonitic shear zone (Friend and Nutman, 1989; Nutman et al., 2008; Kolb, 2014).

South of the AIC, the Isertoq Terrane is dominated by low-pressure Archean amphibolite-facies gneiss with supracrustal rocks and a variety of younger intrusive rocks cropping out around Isertoq (Fig. 3.1b; Escher et al., 1989). One orthogneiss sample close to Isertoq gave a Sm–Nd whole rock model age of 3.05 Ga (Kalsbeek et al., 1993). More recent U–Pb zircon data from several gneiss samples around Isertoq reveal multiple age populations, the oldest between 2.82 and 2.72 Ga being interpreted as protolith ages, whereas ages between ca. 1.86 and 1.68 Ga most likely reflect the Nagssugtoqidian Orogeny and later heating by post-collisional intrusions (Kokfelt et al., 2016a). The Isertoq Terrane contains two major sets of basic dykes. Foliation-parallel, east-west striking boninitic dykes are undeformed in the southern part of the Isertoq Terrane but are progressively more deformed and reworked towards the north (Escher et al., 1989; Klausen et al., 2016). The dykes are Proterozoic in age, with Sm–Nd model ages of ca. 2.17 Ga (Kalsbeek et al., 1993). These older dykes are crosscut by roughly north-south striking tholeiitic dykes, which were presumed to be Tertiary in age (Escher et al., 1989) but more recently reinterpreted as late-orogenic in origin (Klausen et al., 2016).

Major post-tectonic complexes of granite, diorite and gabbro, along with several generations of pegmatite dykes, were dated by zircon ID-TIMS at ca. 1.68 Ga (Kalsbeek et al., 1993) and more recently by LA-SF-ICP-MS at 1.67–1.52 Ga (Thrane et al., 2016). They intrude and crosscut almost all lithological units within the Nagssugtoqidian Orogen and mark the end of metamorphic and igneous activity (Andrews et al., 1973). Contacts between the older rocks and these younger intrusives are rather sharp and show narrow contact aureoles with low-pressure metamorphic assemblages.

3.5 Analytical methods

Around 150 samples from various localities in the Kuummiut Terrane were collected during several fieldtrips, with the oldest ones collected by Allen Nutman and Clark Friend in 1986 (Nutman and Friend, 1989) and the most recent samples collected by the authors in 2014 (see Fig. 3.1a for sample sites and Appendix 2 for a full sample list). The JEOL JXA-8900 R Microprobe at the Institute of Applied Mineralogy and Economic Geology at RWTH Aachen University (equipped with a tungsten filament and five wavelength dispersive spectrometers) and the JEOL JXA-8530 F Hyperprobe at the Institute of Mineralogy at WWU Münster (equipped with a field emission gun and five wavelength-dispersive spectrometers) were used to acquire backscattered electron (BSE) images and mineral chemical data for 13 polished thin sections (Table 3.1). Electron microprobe analysis was carried out using an acceleration voltage of 15 kV and a probe current of either 23 (Aachen) or 15 nA (Münster). The electron beam was fully focused (1–2 μm) during measurements for most phases, except for plagioclase, muscovite and biotite, which were analyzed with a defocused (10 μm) beam. Both,

natural and synthetic mineral standards were used for calibration. Representative compositions of clinopyroxene, orthopyroxene, plagioclase, amphibole and garnet are given in Tables 3.2–3.5. Ferric and ferrous iron contents for ferromagnesian minerals were determined by charge balance after the method of Droop (1987). Mineral abbreviations used in this paper follow the scheme of Whitney and Evans (2010, and Appendix 3).

Table 3.1

Mineral assemblages and replacement textures in the samples observed.

Sample	HP-assemblage or relict HP-phases	Relict HT-phases	Retrograde assemblage	Remarks
Garnet-pyroxenite				
566273	Di + Opx + Hbl + Grt + Ilm + Mag	None	None	Free of replacement textures
566279	Di + Opx + Hbl + Grt + Ilm	None	None	Free of replacement textures
Retrogressed Eclogite				
318349	Omp + Grt _a + Rt	Opx + Ilm + Mag	Di + Pl + Hbl + Grt _b	Worm-like Di-Pl-Hbl symplectites, Hbl-Pl-Ilm-mag symplectites after and Pl-Hbl-Pl coronas around Grt, Omp as inclusion in Grt and Hbl and in cores of large Cpx
566218	Grt _a + Rt + Qz	Aug + Opx + Ilm	Di + Pl + Hbl + Grt _b + Ttn	Worm-like Di-Pl-Hbl symplectites, Hbl-Pl coronas around Grt, Ttn coronas around Rt and Ilm, Aug + Opx + Am coronas between Qz/Grt
566216	Grt _a + Rt + Qz	Aug + Ilm	Di + Pl + Hbl + Grt _b + Ttn	Worm-like Di-Pl-Hbl symplectites, Hbl-Pl coronas around Grt, Ttn coronas around Rt and Ilm, Aug + Am coronas between Qz/Grt
566277	Grt _a + Rt + Qz	Ilm	Di + Pl + Hbl + Grt _b + Ttn	Worm-like Di-Pl-Hbl symplectites, Hbl-Pl symplectites after and Hbl-Pl coronas around Grt, Ttn coronas around Rt and Ilm, Act corona between Qz/Grt
525225b	Grt _a + Qz	Aug + Opx + Ilm	Di + Pl + Hbl + Grt _b + Bt	Globular Di-Pl-Hbl symplectites, Hbl-Pl symplectites after and Hbl-Pl coronas around Grt, Aug + Opx + Am coronas between Qz/Grt
525224	Rt + Qz	Opx + Ilm	Di + Pl + Hbl + Grt _b + Bt	Globular Di-Pl-Hbl symplectites, Hbl-Pl symplectites after and Hbl-Pl coronas around Grt, Opx-Hbl intergrowths between Qz/Grt
524713	Rt	Ilm	Di + Pl + Hbl + Grt _b + Ttn	Di-Pl-Hbl symplectites, Hbl-Pl symplectites after and Hbl-Pl coronas around Grt, Ttn coronas around Rt and Ilm
Garnet-amphibolite				
566201	Rt	None	Pl + Hbl + Grt _b + Ilm + Bt + Di	Hbl-Di intergrowths, leucosome-bearing
566223	None	Di	Pl + Hbl + Qz + Grt _b + Bt + Rt	Hbl-Di intergrowths
566228	None	None	Pl + Hbl + Qz + Grt _b + Bt + Ilm + Rt	Bt-Pl-Hbl symplectites after and Bt-Pl-Hbl coronas around Grt
Garnet-kyanite schist				
566267	None		Grt + Bt + Ms + Ky + Qz + Pl + Rt + Ilm	Qz-vein and leucosome-bearing, free of replacement textures

Table 3.2

Representative electron microprobe data for clinopyroxene and orthopyroxene.

Grt-Py – Garnet-pyroxenite, Retr – Retrogressed eclogite, Grt-Amp – Garnet-amphibolite.

Sample Nr.	566279	566279	566279	566279	524713	566216	566216	566216	318349	318349	318349	566218	525224	566201
Sample type	Grt-Py	Grt-Py	Grt-Py	Grt-Py	Retr	Retr	Retr	Retr	Retr	Retr	Retr	Retr	Retr	Grt-Amp
Mineral	Di	Di	Opx	Opx	Di	Di (Cpx-domain)	Di (Grt-Cor)	Di (Large grains)	Omp (Grt-Inc)	Di	Omp (Cpx-domain)	Aug	Opx	Di
<i>Composition (wt%)</i>														
SiO ₂	53.58	53.09	54.50	53.47	53.88	52.90	52.42	52.72	55.02	52.73	52.74	49.39	54.81	52.86
TiO ₂	bdl	bdl	bdl	bdl	0.05	0.10	0.18	0.17	0.12	0.03	0.22	0.56	0.04	bdl
Al ₂ O ₃	0.86	1.38	1.01	2.41	0.47	2.47	2.28	2.13	10.05	2.37	9.39	6.46	0.41	1.47
Cr ₂ O ₃	0.21	0.23	0.12	0.18	bdl	bdl	0.04	bdl	0.04	0.05	bdl	bdl	bdl	bdl
FeO	4.91	4.04	15.74	16.37	5.45	6.67	7.49	6.85	6.14	8.13	8.24	7.87	25.04	9.09
MnO	0.21	0.18	0.39	0.37	0.19	0.12	0.11	0.26	bdl	0.04	0.04	0.04	0.28	0.23
MgO	16.17	16.68	27.83	27.19	15.51	14.23	13.63	14.67	9.05	13.79	8.95	14.17	18.80	13.30
NiO	bdl	bdl	bdl	bdl	0.03	0.06	bdl	bdl	nm	bdl	0.03	bdl	nm	bdl
CaO	23.64	24.25	0.29	0.36	24.32	22.98	23.33	22.61	13.29	21.15	15.24	20.04	0.74	22.93
Na ₂ O	0.25	0.19	bdl	bdl	0.13	0.63	0.51	0.58	5.77	1.15	4.64	0.85	0.06	0.35
K ₂ O	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	0.11	bdl	bdl
Total	99.83	100.04	99.88	100.35	100.03	100.16	99.99	99.99	99.48	99.44	99.49	99.49	100.18	100.23
<i>Atoms per formula unit (4 cations, 6 oxygens)</i>														
Si	1.967	1.937	1.963	1.920	1.984	1.948	1.943	1.943	1.982	1.955	1.925	1.822	2.049	1.968
Al ^{IV}	0.033	0.059	0.037	0.080	0.016	0.052	0.057	0.057	0.018	0.045	0.075	0.178	0.000	0.032
Al ^{VI}	0.004	0.000	0.005	0.022	0.005	0.055	0.042	0.036	0.409	0.059	0.329	0.103	0.018	0.032
Fe ³⁺	0.041	0.074	0.029	0.052	0.018	0.036	0.041	0.053	0.005	0.065	0.061	0.110	0.000	0.026
Ti	0.000	0.000	0.000	0.000	0.001	0.003	0.005	0.005	0.003	0.001	0.006	0.016	0.001	0.000
Cr	0.006	0.007	0.003	0.005	0.000	0.000	0.001	0.000	0.001	0.001	0.000	0.000	0.000	0.000
Fe ²⁺	0.110	0.049	0.445	0.439	0.150	0.169	0.192	0.158	0.180	0.187	0.190	0.133	0.783	0.257
Mg	0.885	0.907	1.494	1.456	0.851	0.781	0.753	0.806	0.486	0.762	0.487	0.779	1.048	0.738
Mn	0.007	0.006	0.012	0.011	0.006	0.004	0.003	0.008	0.000	0.001	0.001	0.001	0.009	0.007
Ca	0.930	0.948	0.011	0.014	0.960	0.907	0.926	0.893	0.513	0.840	0.596	0.792	0.030	0.914
Na	0.018	0.013	0.000	0.000	0.009	0.045	0.037	0.041	0.403	0.083	0.328	0.061	0.004	0.025
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.005	0.000	0.000
Total	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	3.943	4.000
X _{Fe}	0.11	0.05	0.23	0.23	0.15	0.18	0.20	0.16	0.27	0.20	0.28	0.15	0.43	0.26
X _{Jd}	0.00	0.00	0.00	0.00	0.01	0.05	0.04	0.04	0.40	0.06	0.33	0.07	0.00	0.03
X _{Wo}	0.45	0.44	0.00	0.00	0.47	0.43	0.44	0.42	0.25	0.39	0.26	0.32	0.01	0.44

X_{Fe} = Fe²⁺ / (Fe²⁺ + Mg), X_{Jd} = (Na + K if Na + K > Al^{IV}, otherwise Al^{IV}), X_{Wo} = (Ca-Ca for Ts-components)/2; nm – not measured, bdl – below detection limit.

Due to the presence of two chemically distinct domains in retrogressed eclogite, as well as a large variety in mineral compositions and replacement textures, a FEI Quanta650F QEMSCAN was used to acquire compositional phase mappings of 6 thin sections (Appendix 4). Detailed measurement procedures are described in Farber et al. (2016). The samples were analyzed in the field scan mode to ensure full coverage of the polished section. The spatial resolution was 10 μm and a spot size of 5.7 μm was applied. Two thousand counts/s were detected via two Bruker 133 eV EDX spectrometers, using an acceleration voltage of 15 kV and a probe current of 10 nA that is readjusted in a 2-h cycle.

Table 3.3

Representative electron microprobe data for amphibole.

Grt-Py – Garnet-pyroxenite, Retr – Retrogressed eclogite, Grt-Amp – Garnet-amphibolite.

Sample Nr.	566273	566279	566216	566216	566216	524713	524713	318349	318349	566223	566223	566228
Sample type	Grt-Pyr	Grt-Pyr	Retr	Retr	Retr	Retr	Retr	Retr	Retr	Grt-Amp	Grt-Amp	Grt-Amp
Mineral	Hbl	Hbl	Hbl	Act	Ath	Hbl (Cpx-domain)	Prg (Grt-Cor)	Ts	Hst	Hbl	Act	Ath
<i>Composition (wt%)</i>												
SiO ₂	46.34	46.74	44.90	52.89	54.74	44.87	41.15	41.19	40.02	42.98	54.90	54.53
TiO ₂	0.19	0.16	1.46	0.45	0.03	1.41	1.36	0.24	1.55	1.00	bdl	bdl
Al ₂ O ₃	11.95	12.12	10.85	4.70	0.52	12.71	15.04	16.57	16.61	14.23	2.59	0.81
Cr ₂ O ₃	0.83	0.70	0.08	bdl	bdl	0.06	0.04	bdl	0.05	0.03	0.13	bdl
FeO	6.40	6.05	13.11	8.81	20.49	9.28	12.55	14.21	14.06	14.66	7.39	20.35
MnO	0.06	0.08	0.11	0.17	0.47	0.18	0.21	0.06	0.03	0.18	bdl	0.50
MgO	17.79	17.26	12.66	17.76	20.11	14.75	11.20	12.07	11.24	11.01	19.14	19.89
NiO	0.04	0.04	bdl	bdl	0.04	bdl	0.12	bdl	bdl	0.03	bdl	0.03
CaO	12.00	12.23	11.42	11.78	0.86	11.96	11.91	10.51	11.02	11.47	12.60	1.07
Na ₂ O	1.57	1.71	1.45	0.55	0.05	1.98	2.50	3.04	2.75	1.95	0.22	0.11
K ₂ O	0.08	0.04	0.47	0.13	bdl	0.20	0.26	0.14	0.65	0.57	0.08	bdl
Total	97.21	97.09	96.51	97.24	97.27	97.40	96.22	98.03	97.98	98.08	97.05	97.26
<i>Atoms per formula unit (15 cations, 23 oxygens)</i>												
Si	6.557	6.635	6.661	7.510	7.910	6.487	6.160	6.052	5.932	6.337	7.749	7.886
Al ^{IV}	1.443	1.365	1.339	0.490	0.089	1.513	1.840	1.948	2.068	1.663	0.251	0.114
Al ^{VI}	0.550	0.663	0.557	0.296	0.000	0.652	0.813	0.922	0.834	0.810	0.180	0.024
Fe ³⁺	0.592	0.346	0.000	0.000	0.071	0.000	0.000	0.081	0.000	0.000	0.025	0.059
Ti	0.020	0.017	0.163	0.048	0.003	0.153	0.153	0.027	0.173	0.111	0.000	0.000
Cr	0.093	0.079	0.009	0.000	0.000	0.007	0.005	0.000	0.006	0.003	0.015	0.000
Fe ²⁺	0.165	0.372	1.626	1.046	2.404	1.122	1.571	1.665	1.743	1.808	0.847	2.402
Mg	3.753	3.653	2.800	3.759	4.332	3.179	2.499	2.644	2.484	2.420	4.028	4.288
Mn	0.007	0.010	0.014	0.020	0.058	0.022	0.027	0.007	0.004	0.022	0.000	0.061
Ca	1.819	1.860	1.815	1.792	0.133	1.853	1.910	1.655	1.750	1.812	1.906	0.166
Na	0.431	0.471	0.417	0.151	0.014	0.555	0.726	0.866	0.790	0.557	0.060	0.031
K	0.014	0.007	0.089	0.024	0.000	0.037	0.050	0.026	0.123	0.107	0.014	0.000
Total	15.445	15.478	15.490	15.137	15.014	15.580	15.753	15.892	15.907	15.651	15.075	15.031
X _{Fe}	0.17	0.16	0.37	0.22	0.36	0.26	0.39	0.40	0.41	0.43	0.18	0.36

X_{Fe} = Fe²⁺ / (Fe²⁺ + Mg); bdl – below detection limit.

Bulk-rock major element geochemical data used for pseudosection modelling were obtained by X-ray fluorescence spectroscopy on fusion pellets after the method described in Kramm et al. (2017), using a SPECTRO X-LAB 2000 energy-dispersive spectrometer. For the analyses, the samples were crushed and milled in an agate mortar, before being dried to determine the loss on ignition (LOI). Borate glass fusion pellets were prepared by melting (1150 °C) a homogenized powder of 0.5 g

sample and 5 g Dilithiumtetra-Lithiummetaborate (FLUXANA) as a fluxing agent. Surface tension in the fusion pellets was achieved by inserting an ammonium iodide pill into the oven at the start of the smelting. All bulk-rock data, as well as a more thorough description of the method is given in Appendix 6.5.

3.6 Petrography and mineral chemistry

3.6.1 Garnet-pyroxenite

3.6.1.1 Petrography

Garnet-pyroxenite (samples 566273 and 566279) was sampled from two localities in the Kuummiut Terrane (Fig. 3.1a). This type of rock occurs as massive black to brownish boudins within TTG gneiss (Fig. 3.2a) and is coarser-grained than the garnet-amphibolite and retrogressed eclogite. The dominant mineral assemblage is texturally well equilibrated and consists of diopside + orthopyroxene + hornblende + garnet + ilmenite $\pm$ magnetite (Fig. 3.2b). Some outcrops show felsic veins crosscutting the boudins but these are mainly concentrated at the boudin rims, where greenish amphibolite surrounds the pyroxenite core. Garnet-pyroxenite is devoid of any replacement textures.

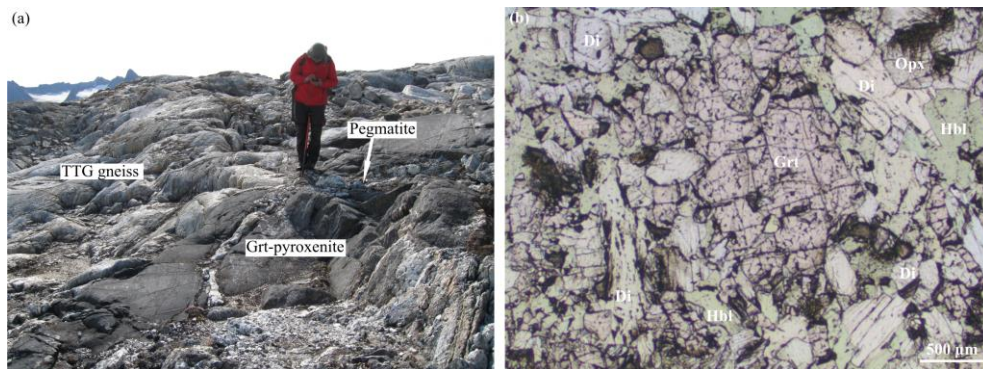


Fig. 3.2. (a) Field photograph and (b) photomicrograph of garnet-pyroxenite sample 566273.

3.6.1.2 Mineral chemistry

The X_{Fe} in diopside ranges from 0.04 to 0.12. The X_{Jd} is <0.01 and the grains record elevated Cr concentrations (Table 3.2). Orthopyroxene is also rich in Mg, with X_{Fe} values varying from 0.19 to 0.25 (Table 3.2). Both types of pyroxene show Al-zoning (Table 3.2), with Al-rich compositions recorded in pyroxene inclusions in garnet or pyroxene grains near the contact with garnet. Hornblende ranges in composition from tschermakite to magnesio-hornblende, with X_{Fe} values ranging between 0.15 and 0.21 (Table 3.3). In both samples, the garnet grains record a slight enrichment in Fe along the outermost rims, but are otherwise unzoned and have a core composition of $Alm_{38-41}Grs_{17-19}Prp_{37-41}Sps_3$ and a rim composition of $Alm_{41-44}Grs_{13-19}Prp_{34-41}Sps_{3-4}$ (Table 3.4).

Table 3.4

Representative electron microprobe data for garnet.

Grt-Py - Garnet-pyroxenite, Retr – Retrogressed eclogite, Grt-Amp – Garnet-amphibolite, Sed-Metasediment.

Sample Nr.	566279	566279	566216	566216	566216	525224	525224	566223	566223	566201	566201	566228	566228	566267
Sample type	Grt-Py	Grt-Py	Retr	Retr	Retr	Retr	Retr	Grt-Amp	Grt-Amp	Grt-Amp	Grt-Amp	Grt-Amp	Grt-Amp	Sed
Mineral	Grt	Grt	Grt _a	Grt _b	Grt-Rim	Grt-Core	Grt-Rim	Grt-Core	Grt-Rim	Grt-Core	Grt-Rim	Grt-Core	Grt-Rim	Grt
<i>Composition (wt%)</i>														
SiO ₂	38.77	38.87	38.33	38.94	38.64	38.59	38.05	39.27	38.70	38.07	38.21	37.96	38.32	37.41
TiO ₂	bdl	bdl	0.07	bdl	0.04	bdl	bdl	bdl	0.03	0.06	bdl	0.05	bdl	bdl
Al ₂ O ₃	20.88	21.04	21.13	21.73	21.42	21.57	21.60	21.94	22.25	21.26	21.38	21.26	21.28	21.20
Cr ₂ O ₃	1.20	1.16	0.05	0.03	bdl	bdl	bdl	bdl	0.05	bdl	bdl	0.15	0.13	bdl
FeO	21.23	19.81	23.48	21.58	23.56	23.54	26.55	21.89	24.26	22.92	25.19	25.37	24.48	35.05
MnO	1.64	1.40	2.25	1.10	1.38	0.46	0.96	0.45	0.55	1.92	1.75	1.96	1.43	1.70
MgO	9.48	10.17	3.75	7.83	7.10	6.62	4.85	8.60	8.34	3.60	4.62	4.71	5.63	3.16
ZnO	0.03	bdl	nm	nm	nm	0.06	bdl	bdl	0.03	0.03	bdl	bdl	bdl	bdl
CaO	6.43	6.75	10.95	8.90	7.89	9.11	7.95	7.87	5.81	11.99	9.06	8.35	8.58	1.47
Na ₂ O	bdl	bdl	nm	nm	nm	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl
K ₂ O	bdl	bdl	nm	nm	nm	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl
Total	99.63	99.20	100.01	100.11	100.03	99.89	99.96	100.02	99.99	99.82	100.21	99.81	99.85	99.99
<i>Atoms per formula unit (8 cations, 12 oxygens)</i>														
Si	2.968	2.969	3.001	2.974	2.977	2.977	2.976	2.990	2.966	2.981	2.981	2.977	2.983	3.006
Al ^{IV}	0.032	0.031	0.000	0.026	0.023	0.023	0.024	0.010	0.034	0.019	0.019	0.023	0.017	0.000
Al ^{VI}	1.852	1.863	1.950	1.929	1.923	1.937	1.968	1.959	1.975	1.942	1.946	1.942	1.935	2.008
Fe ³⁺	0.107	0.097	0.038	0.095	0.096	0.086	0.056	0.051	0.053	0.070	0.073	0.066	0.074	0.000
Ti	0.000	0.000	0.004	0.000	0.002	0.000	0.000	0.000	0.002	0.004	0.000	0.003	0.000	0.000
Cr	0.073	0.070	0.003	0.002	0.000	0.000	0.000	0.000	0.003	0.000	0.000	0.009	0.008	0.000
Fe ²⁺	1.252	1.168	1.500	1.283	1.423	1.432	1.681	1.343	1.502	1.431	1.571	1.597	1.520	2.355
Mg	1.082	1.158	0.438	0.891	0.816	0.761	0.566	0.976	0.953	0.420	0.537	0.551	0.653	0.379
Mn	0.106	0.091	0.149	0.071	0.090	0.030	0.064	0.029	0.036	0.127	0.116	0.130	0.094	0.116
Ca	0.527	0.552	0.918	0.728	0.651	0.753	0.666	0.642	0.477	1.006	0.757	0.702	0.716	0.127
Na	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	7.990
X _{Fe}	0.54	0.50	0.77	0.59	0.64	0.65	0.75	0.58	0.61	0.77	0.75	0.74	0.70	0.86
X _{Alm}	0.42	0.39	0.50	0.43	0.48	0.48	0.56	0.45	0.51	0.48	0.53	0.54	0.51	0.79
X _{Grs}	0.18	0.19	0.31	0.24	0.22	0.25	0.22	0.21	0.16	0.34	0.25	0.24	0.24	0.04
X _{Prp}	0.36	0.39	0.15	0.30	0.27	0.26	0.19	0.33	0.32	0.14	0.18	0.18	0.22	0.13
X _{Sps}	0.04	0.03	0.05	0.02	0.03	0.01	0.02	0.01	0.01	0.04	0.04	0.04	0.03	0.04

X_{Fe} = Fe²⁺ / (Fe²⁺ + Mg), X_{Alm} = Fe²⁺ / (Fe²⁺ + Ca + Mg + Mn), X_{Grs} = Ca / (Ca + Fe²⁺ + Mg + Mn), X_{Prp} = Mg / (Mg + Fe²⁺ + Ca + Mn), X_{Sps} = Mn / (Mn + Fe²⁺ + Ca + Mg). nm – not measured, bdl – below detection limit.

3.6.2 Retrogressed eclogite

3.6.2.1 Petrography

Retrogressed eclogite (samples 318349, 524713, 525224, 525225b, 566216, 566218 and 566277) was collected from various localities in the Kuummiut Terrane (Fig. 3.1a). These relatively fine-grained rocks are mostly present in the Paleoproterozoic basic dykes (Fig. 3.3a) or in boudins of meta-volcanic rock (Fig. 3.3b).

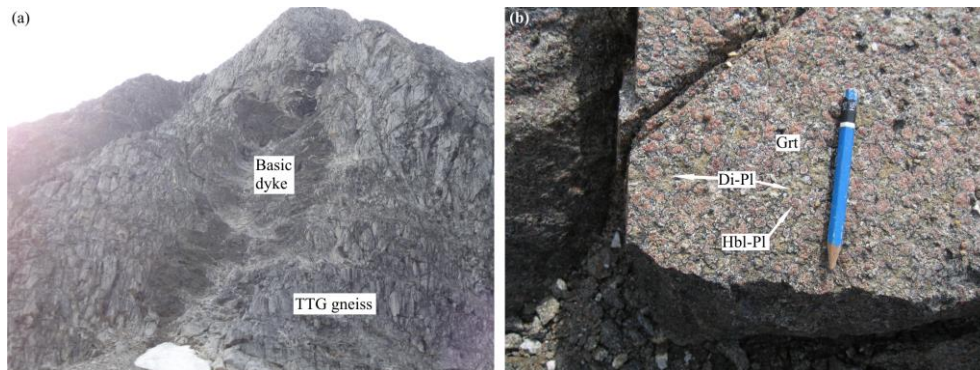


Fig. 3.3. (a) Field photograph of a basic dyke in TTG gneiss. (b) Hand specimen of the basic supracrustal rock sample 566277, showing mineral reaction textures.

As noted in previous studies (Nutman and Friend, 1989; Messiga et al., 1990; Nutman et al., 2008), the retrogressed eclogite contains complex mineral reaction textures and assemblages (Table 3.1), with a dominant mineral assemblage of clinopyroxene + plagioclase + hornblende + garnet $\pm$ quartz $\pm$ ilmenite $\pm$ biotite $\pm$ orthopyroxene (Figs. 3.3b and 3.4). Rutile and titanite occur as minor components throughout the rock. Accessory phases, usually present as inclusions in garnet, are zircon, magnetite and monazite. Most retrogressed eclogite samples show a uniform distribution of the four main minerals over two different domains (Figs. 3.4–3.6).

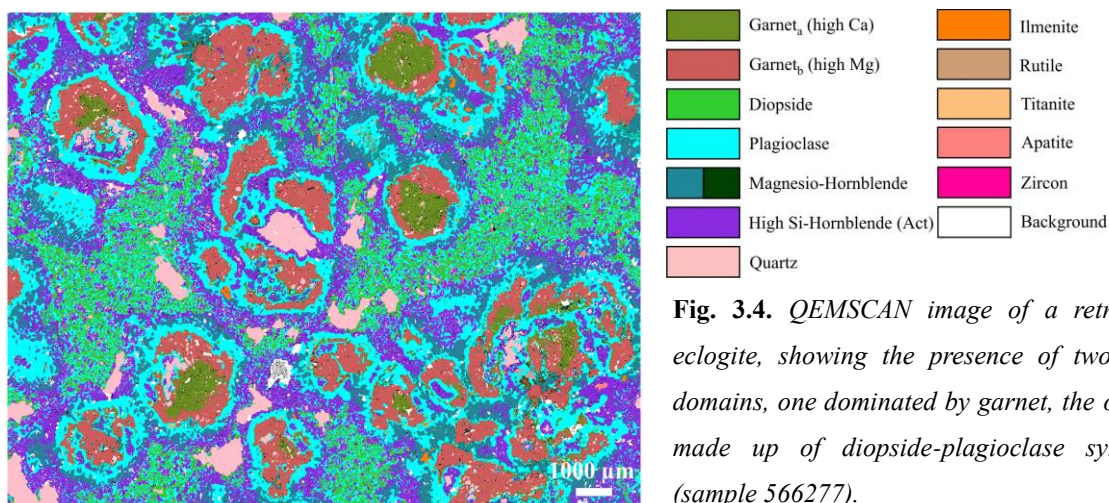


Fig. 3.4. QEMSCAN image of a retrogressed eclogite, showing the presence of two distinct domains, one dominated by garnet, the other one made up of diopside-plagioclase symplectite (sample 566277).

Table 3.5

Representative electron microprobe data for plagioclase.

Retr – Retrogressed eclogite, Grt-Amp – Garnet-amphibolite, Sed-Metasediment.

Sample Nr.	566277	566277	566277	566277	525225b	566216	566216	566223	566201	566201	566201	566267	566267
Sample type	Retr	Retr	Retr	Retr	Retr	Retr	Retr	Grt-Amp	Grt-Amp	Grt-Amp	Grt-Amp	Sed	Sed
Mineral	Pl (Grt-Cor)	Pl (Cpx-domain)	Pl (Inc)	Pl (Inc)	Pl (Grt-Symp)	Pl	Pl	Pl	Pl (Grt)	Pl (Am)	Pl (Inc)	Pl	Pl
<i>Composition (wt%)</i>													
SiO ₂	57.43	62.05	57.47	61.51	54.43	57.59	60.52	55.68	49.80	54.23	60.51	62.57	65.15
TiO ₂	bdl	0.06	bdl	bdl	bdl	bdl	bdl	bdl	bdl	bdl	0.07	bdl	0.04
Al ₂ O ₃	27.11	23.83	27.10	24.28	29.73	26.97	24.68	28.78	32.38	29.36	24.94	24.01	22.22
FeO	0.25	0.10	0.50	0.07	0.08	0.13	0.09	0.14	0.05	0.03	0.12	0.06	0.29
MnO	bdl	bdl	bdl	bdl	0.03	bdl	bdl	bdl	bdl	bdl	bdl	bdl	0.03
CaO	9.06	5.10	8.80	5.81	11.28	7.92	6.41	10.17	15.04	11.27	5.95	4.59	2.51
BaO	0.05	bdl	bdl	bdl	0.07	0.06	0.14	bdl	bdl	0.10	bdl	0.03	bdl
SrO	bdl	bdl	bdl	0.04	bdl	0.22	0.17	bdl	bdl	bdl	bdl	bdl	bdl
Na ₂ O	6.43	8.96	6.58	8.49	4.76	6.87	7.77	5.55	3.16	4.90	7.79	8.60	9.73
K ₂ O	0.04	0.07	0.03	0.06	0.10	0.11	0.14	0.03	bdl	0.08	0.03	0.11	0.05
Total	100.37	100.11	100.48	100.22	100.48	99.65	99.75	100.35	100.43	99.97	99.34	99.97	99.98
<i>Atoms per formula unit (5 cations, 8 oxygens)</i>													
Si	2.567	2.750	2.567	2.726	2.442	2.585	2.700	2.493	2.263	2.447	2.702	2.766	2.864
Al	1.428	1.245	1.427	1.268	1.572	1.427	1.298	1.519	1.734	1.561	1.313	1.251	1.151
Fe	0.009	0.004	0.019	0.003	0.003	0.005	0.003	0.005	0.002	0.001	0.004	0.002	0.011
Mn	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001
Ba	0.001	0.000	0.000	0.000	0.001	0.001	0.002	0.000	0.000	0.002	0.000	0.001	0.000
Ca	0.434	0.242	0.421	0.276	0.542	0.381	0.306	0.488	0.732	0.545	0.285	0.217	0.118
Na	0.557	0.770	0.570	0.730	0.414	0.598	0.672	0.482	0.278	0.429	0.674	0.737	0.829
K	0.002	0.004	0.002	0.003	0.006	0.006	0.008	0.002	0.000	0.005	0.002	0.006	0.003
Total	4.999	5.015	5.005	5.006	4.982	5.003	4.991	4.989	5.009	4.989	4.980	4.980	4.977
X _{An}	0.44	0.24	0.42	0.27	0.56	0.39	0.31	0.50	0.72	0.56	0.30	0.23	0.12
X _{Ab}	0.56	0.76	0.57	0.72	0.43	0.61	0.68	0.50	0.28	0.44	0.70	0.77	0.87
X _{Or}	0.00	0.00	0.00	0.00	0.01	0.01	0.01	0.00	0.00	0.00	0.00	0.01	0.00

X_{An} = Ca / (Ca + Na + K), X_{Ab} = Na / (Ca + Na + K), X_{Or} = K / (Ca + Na + K).

The first domain, hereafter referred to as the clinopyroxene domain, mainly consists of symplectitic intergrowths of diopside and plagioclase. The symplectites mostly occur as either very fine-grained (~10 µm) lamellar (Fig. 3.5a) or fine-grained (~100 µm) globular (Fig. 3.5b) intergrowths, which are often intergrown with and partially replaced by plagioclase + amphibole symplectites (Fig. 3.4).

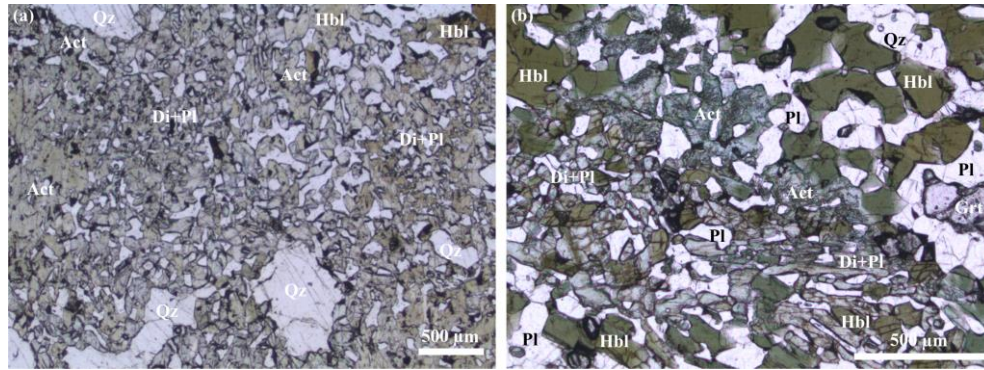


Fig. 3.5. Photomicrographs of the clinopyroxene domain showing different textures: **(a)** Fine- to very fine-grained, worm-like intergrowths of diopside and plagioclase intergrown with and surrounded by amphibole and plagioclase symplectites (sample 566277). **(b)** Globular intergrowth of diopside, plagioclase, hornblende and quartz, surrounded by actinolite (sample 525224).

The second domain, hereafter referred to as the garnet domain (Fig. 3.4), is characterized by corona textures around coarse-grained garnet (Fig. 3.6a–d). In general, plagioclase constitutes the inner corona, followed by either one or two amphibole coronas towards the contact with the clinopyroxene domain. In the latter case, magnesio-hornblende represents the second corona, whereas more Si-rich amphibole (actinolite or actinolitic hornblende) forms the third corona and separates the garnet domain from the clinopyroxene domain. Garnet grains that experienced an only minor to moderate replacement are well rounded and almost inclusion-free (Fig. 3.6a). Further replacement results in the formation of small indentations and fractures in garnet, which are often filled with plagioclase ± hornblende ± ilmenite (Fig. 3.6b). Garnet pseudomorphs are characterized by a symplectitic replacement assemblage of plagioclase + hornblende ± Fe-Ti-rich phases (ilmenite, titanite, pyrite, magnetite, Fig. 3.6c). Garnet near large quartz grains lacks the coarse-grained magnesio-hornblende corona and is rimmed by thin (< 100 µm) coronas of actinolite ± diopside ± orthopyroxene (Figs. 3.4 and 3.6d). All three coronas are at least partially surrounded by magnesio-hornblende next to plagioclase. In addition, the two pyroxenes are frequently intergrown with Si-rich amphibole of varying composition (Table 3.3). Amphibole-free, orthopyroxene-plagioclase coronas around garnet (such as those observed by Nutman and Friend, 1989) were not identified in our samples. Diopside can also be found as large grains next to garnet (100–300 µm), where it is partially surrounded by thin hornblende and actinolite coronas (Fig. 3.6e).

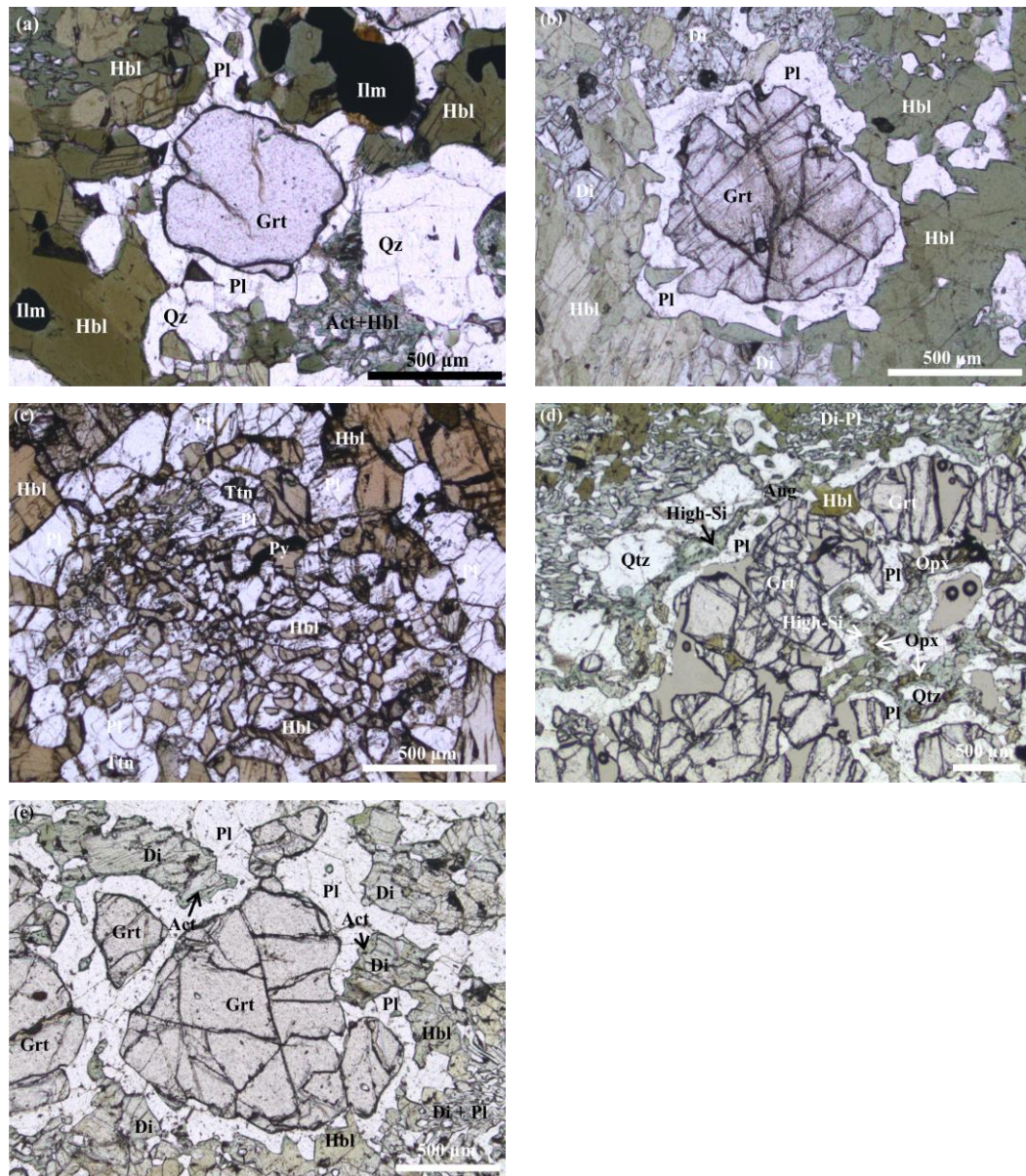


Fig. 3.6. Photomicrographs showing different degrees of garnet replacement and different symplectitic replacement assemblages after garnet in retrogressed eclogite. **(a)** Garnet grain surrounded by an inner plagioclase and an outer hornblende corona (sample 525224). **(b)** Increasing replacement of garnet by plagioclase and hornblende, resulting in the formation of indentations and fractures in garnet (sample 525225b). **(c)** Symplectitic intergrowths of plagioclase, hornblende, titanite and pyrite forming pseudomorphs after garnet. Coronitic textures are still preserved (sample 524713). **(d)** Thin two-pyroxene + high-Si amphibole coronas between quartz and garnet (sample 566218). **(e)** Large diopside grains with thin hornblende and actinolite coronas next to garnet (sample 566216).

A Na-rich retrogressed eclogite sample shows another sequence of coronas around garnet (Fig. 3.7). In this sample, inner plagioclase coronas around garnet are either not present or discontinuous, with small patches of hornblende in between. More commonly, garnet is surrounded by an inner hornblende corona, followed by a plagioclase corona towards the contact with the

clinopyroxene domain. Garnet in this sample is variably pseudomorphed by a symplectitic replacement assemblage of hornblende + ilmenite $\pm$ magnetite $\pm$ plagioclase (Fig. 3.7).

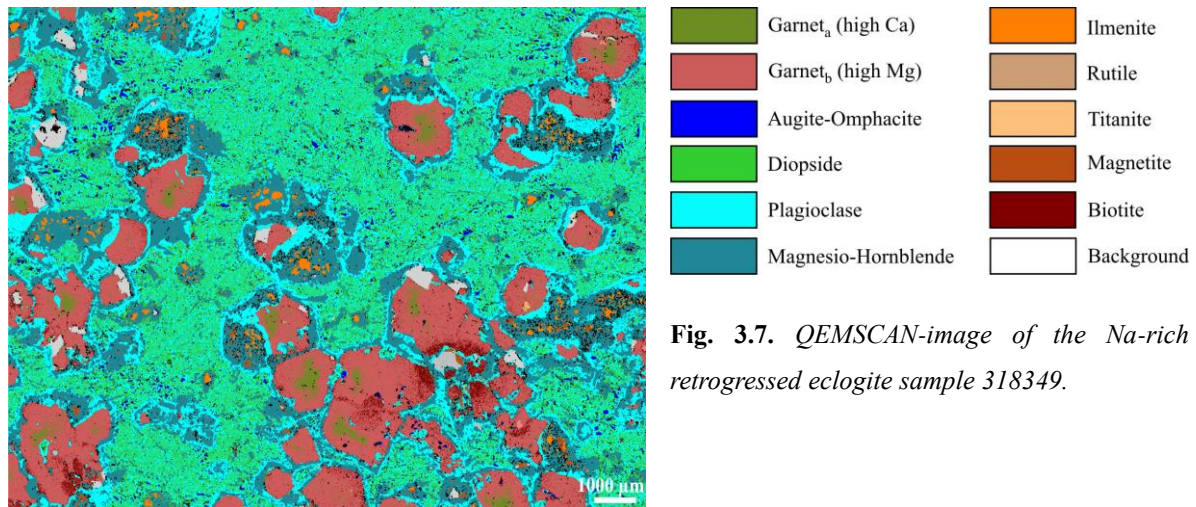


Fig. 3.7. *QEMSCAN-image of the Na-rich retrogressed eclogite sample 318349.*

Garnet grains in the retrogressed eclogite samples contain different types of inclusions, either appearing as small individual grains (mostly plagioclase, hornblende and rutile, rarely epidote) or in grain clusters (quartz, monazite and zircon). Omphacite inclusions in garnet, often surrounded by plagioclase, were found in the Na-rich retrogressed eclogite sample (Fig. 3.7).

Ilmenite, rutile and titanite, in decreasing order of abundance, are distributed randomly throughout the two domains. The grain size (from ~ 10 – several $100 \mu\text{m}$) and modal amount (<1 – 10%) vary considerably and the three Ti-phases are often intergrown with each other (Fig. 3.8). Virtually all rutile grains contain lamellae of ilmenite and are intergrown with large ilmenite grains. Titanite forms coronas around both ilmenite and rutile.

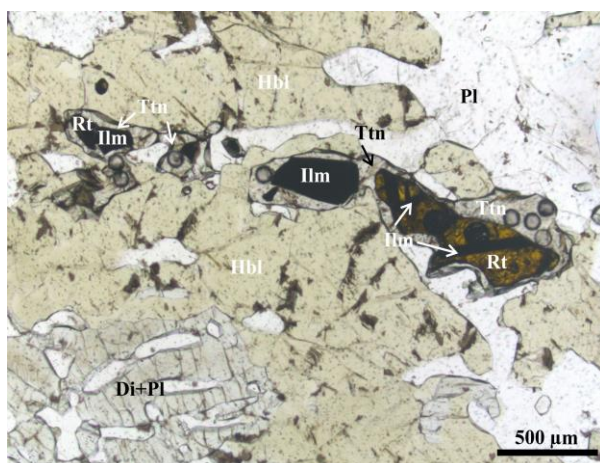


Fig. 3.8. *Photomicrograph of titanite coronas around rutile-ilmenite intergrowths (sample 566216). Ilmenite occurs as both large grains in association with rutile and exsolution lamellae within the rutile grain.*

3.6.2.2 Mineral chemistry

Three different types of clinopyroxene were identified in the retrogressed eclogites (Table 3.2). The majority of the clinopyroxene is diopsidic in composition, with X_{Jd} ranging from 0 to 0.05 and X_{Fe} values varying between 0.12 and 0.30 (Table 3.2). There is no compositional difference between the diopside in symplectites in the clinopyroxene domain and the diopside grains and coronas around garnet (Table 3.2; sample 566216). A few retrogressed eclogite samples additionally contain augite, both in coronas around garnet and in the clinopyroxene domain (Table 3.2). Augite is relatively poor in Ca, enriched in Al and slightly more sodic than diopside (X_{Jd} 0.03–0.06), with X_{Fe} values ranging from 0.15–0.24. The strongest variation in clinopyroxene composition was documented in the Na-rich retrogressed eclogite (Table 3.2; sample 318349). In this sample, the cores of larger clinopyroxene grains are omphacite enriched in Na and Al, with X_{Jd} varying from 0.25 to 0.38 and X_{Fe} values of 0.18–0.29 (Fig. 3.9a, c; Table 3.2). Rims of larger grains and very fine-grained lamellae of clinopyroxene (<10 μm) are Ca-rich and can be classified as diopside to augite, with X_{Jd} varying from 0.06 to 0.25 and X_{Fe} from 0.21–0.26 (Table 3.2).

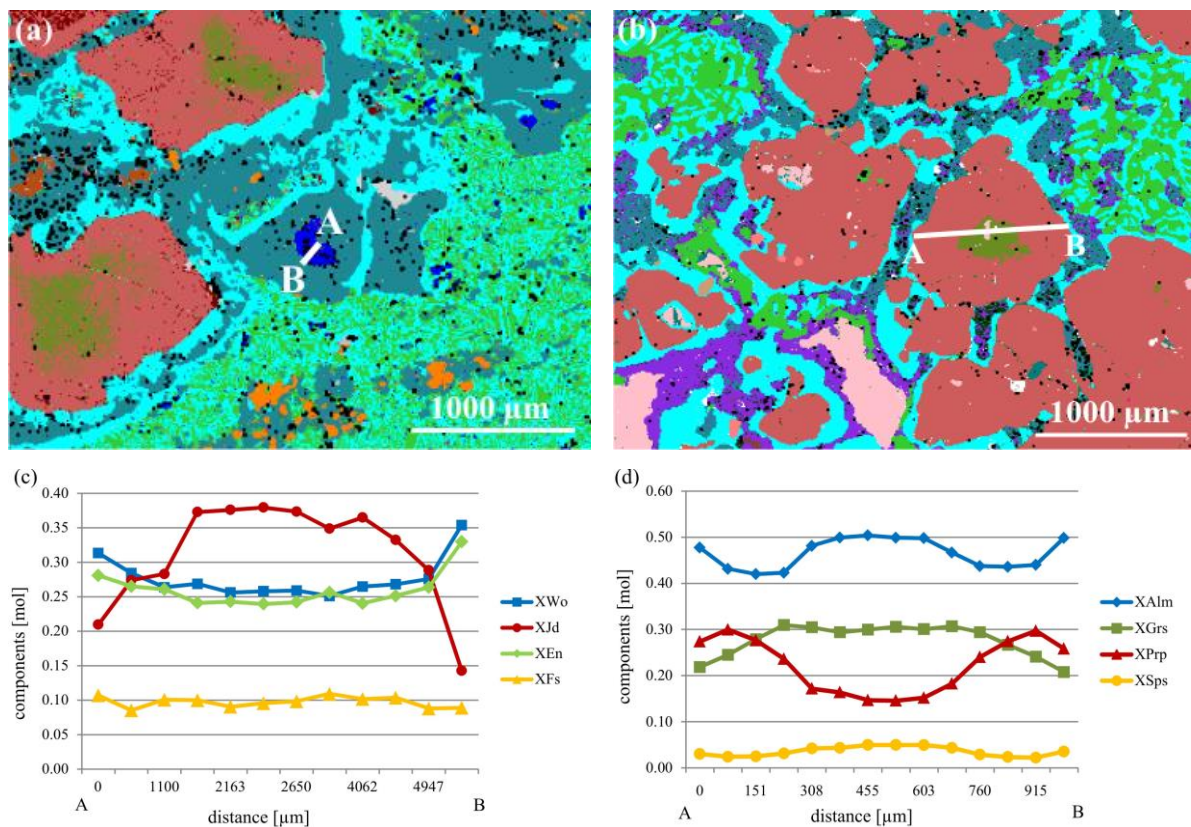


Fig. 3.9. Chemical zoning in the retrogressed eclogites: **(a)** QEMSCAN-image of the Na-rich sample 318349 (legend see Fig. 3.7) and **(b)** sample 566216 (legend see Fig. 3.4). White lines indicate position of line measurements shown in **(c and d)**. **(c)** Zoning profile of a clinopyroxene inclusion in hornblende. **(d)** Garnet zoning profile in a large garnet grain. See text for discussion.

In addition, omphacite ($X_{\text{Jd}} \sim 0.4$) occurs as inclusion in garnet and amphibole (Table 3.2). Orthopyroxene shows uniform compositions for all samples with X_{Fe} values of 0.41–0.43 (Table 3.2).

Plagioclase varies in composition in the retrogressed eclogite samples (Table 3.5). Plagioclase in diopside-plagioclase symplectites is relatively enriched in Na (An_{24-30} ; Table 3.5), whereas plagioclase in contact with garnet is more Ca-rich (An_{32-44} ; Table 3.5). Plagioclase inclusions in garnet range from Na-rich plagioclase often intergrown with quartz (An_{27} , Table 3.5) to more Ca-rich plagioclase (An_{42} , Table 3.5). The composition of these grains is very similar to plagioclase grains in contact with either garnet or diopside. Plagioclase coronas around garnet are commonly zoned, ranging from Ca-rich andesine (An_{44}) at the contact with garnet or hornblende to Na-rich andesine (An_{32}) in the center of the corona (Table 3.5; sample 566216). The most Ca-rich plagioclase (An_{56} ; Table 3.5) occurs in symplectitic intergrowths of fully pseudomorphed garnet grains.

Garnet is generally zoned. Large garnet grains (Fig. 3.9b, d; Table 3.4; sample 566216) locally preserve a Ca-rich core (hereafter called Grt_a), which locally has higher Fe and/or Mn concentrations ($\text{Alm}_{50}\text{Grs}_{31}\text{Prp}_{15}\text{Sps}_5$). Towards the rim, garnet is Mg-rich (hereafter called Grt_b ; $\text{Alm}_{43}\text{Grs}_{24}\text{Prp}_{30}\text{Sps}_2$), followed by a thin ($\sim 10 \mu\text{m}$) Mn- and Fe-rich rim ($\text{Alm}_{48}\text{Grs}_{22}\text{Prp}_{27}\text{Sps}_3$). The remaining grains (Table 3.4; sample 525224) either show a weaker zoning or, more frequently, are similar in composition to the Mg-rich zone ($\text{Alm}_{48}\text{Grs}_{25}\text{Prp}_{26}\text{Sps}_1$) with a sharp transition towards the thin Mn- and Fe-rich rim ($\text{Alm}_{56}\text{Grs}_{22}\text{Prp}_{19}\text{Sps}_2$).

Compositionally, three different types of amphibole can be distinguished for the majority of the retrogressed eclogites (Table 3.3). Most dominant is relatively Al-rich magnesio-hornblende, which occurs in symplectites of the clinopyroxene domain and in coronas around garnet, with X_{Fe} values varying between 0.20 and 0.37. In samples showing extensive garnet replacement (Table 3.3; sample 524713), hornblende varies from more Ca-rich compositions in the clinopyroxene domain (X_{Fe} 0.26–0.30), to Al- and Fe-rich compositions in coronas around garnet (X_{Fe} 0.33–0.39). Actinolite or actinolitic hornblende forms coronas separating garnet from quartz and the clinopyroxene domain. It is often partially surrounded by magnesio-hornblende towards plagioclase and shows much higher Si and Mg contents than magnesio-hornblende, with X_{Fe} varying from 0.18 to 0.35. Rare Si- and especially Fe-rich anthophyllite replaces orthopyroxene in thin coronas around garnet (sample 566216; Table 3.3). Hornblende in the Na-rich retrogressed eclogite has X_{Fe} values between 0.33 and 0.43 and is relatively enriched in Ti where intergrown with ilmenite (Table 3.3; sample 318349).

3.6.3 Amphibolite

3.6.3.1 Petrography

Amphibolite investigated in this study (samples 566201, 566223 and 566228) was sampled from the area close to the Niflheim thrust and in the southern part of the Kuummiut Terrane (Fig. 3.1a). The rocks are medium- to coarse-grained and massive to schistose (Fig. 3.10a). Most samples are dominated by hornblende with varying amounts of bleb-like plagioclase and quartz (Fig. 3.10b).

Garnet, biotite, diopside, titanite, ilmenite and rutile are locally present (Table 3.1). Garnet-bearing amphibolite often contains leucosomes in association with garnet (Fig. 3.10a), with some garnet grains having leucosome tails. Locally, actinolite replacing diopside, rutile-ilmenite-titanite intergrowths and biotite-hornblende-plagioclase symplectites after garnet can be observed. Large garnet grains contain numerous inclusions of plagioclase, quartz, diopside and hornblende and more commonly show replacement textures than smaller garnet grains.

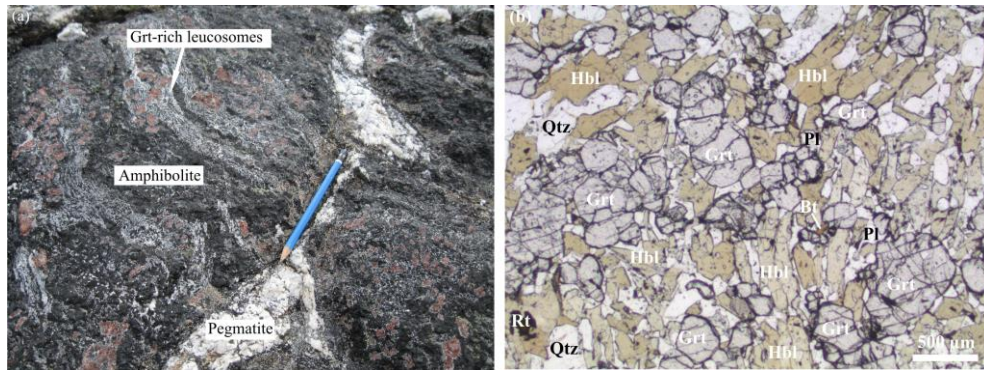


Fig. 3.10. (a) Field photograph of a leucosome-bearing garnet-amphibolite. (b) Photomicrograph showing the dominant mineral assemblage of amphibolite sample 566223.

3.6.3.2 Mineral chemistry

Plagioclase shows variable compositions, depending on the type of garnet present in the sample (Table 3.5). Samples containing small garnet grains show uniform plagioclase compositions (An_{48-54} ; Table 3.5; sample 566223). In samples containing large garnet grains (Table 3.5; sample 566201), plagioclase ranges from Ca-rich compositions around garnet (An_{67-77}) to more Na-rich compositions towards amphibole (An_{54-57}). Even more Na-rich plagioclase was found as inclusion in garnet (An_{30}). Similar to the retrogressed eclogite samples, three different amphibole types can be distinguished, which are similar in composition to those from the retrogressed eclogites. The most common amphibole is magnesio-hornblende with only minor compositional variation and X_{Fe} values varying from 0.41 to 0.43 (Table 3.3). Actinolite with X_{Fe} values of 0.18–0.22 replaces diopside and occurs in intergrowths with anthophyllite in sample 566228 (Table 3.3). The latter most Si-rich amphibole has X_{Fe} values of 0.36–0.39. Garnet is slightly zoned (Table 3.4). Small garnet grains (sample 566223) surrounded by hornblende show thin Fe-rich rims (from $Alm_{45}Grs_{22}Prp_{33}Sps_1$ to $Alm_{51}Grs_{16}Prp_{32}Sps_1$). Large garnet grains in sample 566201 show Ca- and Mn-rich cores ($Alm_{48}Grs_{33}Prp_{15}Sps_6$) and Mg- and Fe-rich rims ($Alm_{51}Grs_{28}Prp_{18}Sps_4$). In sample 566228, large garnet grains show decreasing Fe and Mn concentrations from the core towards the rim in contact with the biotite-plagioclase-hornblende symplectite (from $Alm_{53}Grs_{22}Prp_{20}Sps_4$ to $Alm_{51}Grs_{25}Prp_{22}Sps_3$). Diopside in the amphibolite shows a similar range in composition as in the retrogressed eclogite, with X_{Fe} values varying from 0.25 to 0.32 (Table 3.2).

3.6.4 Garnet-kyanite schist

3.6.4.1 Petrography

The garnet-kyanite schist (sample 566267) was sampled from the south-eastern part of the Kuummiut Terrane (Fig. 3.1a). It is intercalated with the TTG gneiss and unlike other, rusty-colored metasedimentary rocks, grey in color and massive. In addition to a well-developed foliation, which is defined by the preferred grain orientation of biotite and kyanite, abundant leucosomes and foliation-parallel quartz-veins can be identified (Fig. 3.11a). The medium- to coarse-grained rock is dominated by biotite, kyanite, quartz and garnet, together with muscovite and plagioclase of smaller grain size (Fig. 3.11b). Minor phases are monazite, zircon, rutile and ilmenite, which almost always occur as inclusions in kyanite, biotite and garnet (Table 3.1). Except for thin muscovite coronas around biotite and kyanite, the garnet-kyanite schist is free of replacement textures.

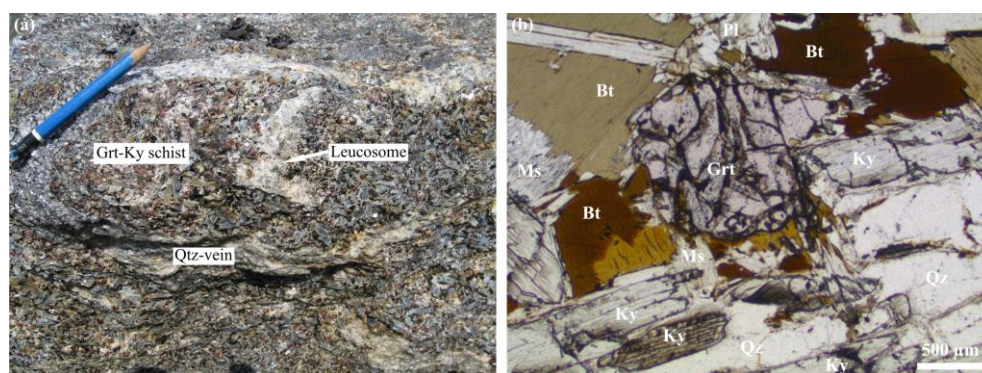


Fig. 3.11. Field photograph (a) and (b) photomicrograph of the garnet-kyanite schist sample 566267.

3.6.4.2 Mineral chemistry

Plagioclase grains are often altered (Fig. 3.11b) and record two different chemical compositions; brownish patches with high Na contents (An_{10-17}) and bright zones with more Ca-rich compositions (An_{20-25} ; Table 3.5). Biotite is Al-rich where it occurs as an inclusion in garnet and more Ti-rich in the matrix. X_{Fe} values for both biotite types, however, range from 0.57 to 0.64. Muscovite yields similar X_{Fe} values, varying from 0.51 to 0.62. Garnet is mostly almandine in composition and does not record any marked compositional zoning ($Alm_{79}Grs_4Prp_{13}Sps_4$; Table 3.4).

3.7 PT-conditions of metamorphism

After identifying relict high-pressure and lower-pressure retrograde mineral assemblages via careful petrographic observations (Table 3.1), the PT-conditions of metamorphism were estimated using a combination of pseudosection modelling and conventional geothermobarometry (Table 3.6).

Conventional thermometry was carried out using the Mg-Fe exchange between clinopyroxene

and orthopyroxene (Taylor, 1998) for garnet-pyroxenite and the Ca-in-orthopyroxene thermometer (Brey and Köhler, 1990) and enstatite-in-clinopyroxene thermometer (Nimis and Taylor, 2000) for garnet-pyroxenite and two of the retrogressed eclogite samples. The hornblende-plagioclase thermometer (Holland and Blundy, 1994) was used for retrogressed eclogite and amphibolite, and the garnet-biotite thermometer (Bhattacharya et al., 1992) for the garnet-kyanite schist. Conventional barometry was carried out using the garnet-orthopyroxene barometer (Harley, 1984) for garnet-pyroxenite, the GAES (garnet-anorthite-enstatite-quartz) and GADS (garnet-anorthite-diopside-quartz) barometers (Newton and Perkins, 1982) for retrogressed eclogite, the garnet-hornblende-plagioclase-quartz barometer (Kohn and Spear, 1990) for amphibolite and the GASP barometer (garnet-kyanite-quartz-anorthite; Koziol and Newton, 1988) for the garnet-kyanite schist.

PT-pseudosections were constructed from bulk-rock major element data using THERMOCALC version 3.33 (Powell and Holland, 1988; updated October 2009) and the November 2003 updated version of the internally consistent dataset from Holland and Powell (1998) (file *tc-ds55.txt*). The pseudosection modelling was carried out in the NCFMASHTO and NCKFMASHTO model systems. Bulk-rock compositions were converted into the model systems (mol. %) by ignoring small amounts of Cr_2O_3 , MnO and in some cases K_2O (Table 3.7). For the modelling of basic and ultrabasic bulk compositions, in which ilmenite is the main oxide phase, approximately 20% of total Fe was converted to Fe^{3+} (Diener and Powell, 2010). The garnet-kyanite schist contains low amounts of ilmenite and lacks hematite and magnetite, suggesting very low contents of Fe^{3+} . For this sample, the Fe^{3+} content was thus estimated using a T- X_{O} section, and 0.1% of the total Fe was converted to Fe^{3+} . The following phases and activity composition models were chosen for modelling: hornblende, actinolite, gedrite (Diener et al., 2007; updated by Diener and Powell, 2012), diopside and omphacite (Green et al., 2007; updated by Diener and Powell, 2012), biotite, garnet and melt (White et al., 2007), muscovite (Coggon and Holland, 2002), plagioclase and orthoclase (Holland and Powell, 2003), epidote, cordierite and staurolite (Holland and Powell, 1998), chlorite (Holland et al., 1998), orthopyroxene and magnetite (White et al., 2002) and ilmenite–hematite (White et al., 2000). Rutile, titanite, quartz, albite, forsterite, the aluminosilicates and aqueous fluid (H_2O) are considered as pure end-member phases. As melt was not included in the modelling of basic and ultrabasic bulk compositions, pseudosections for most samples were constructed assuming water-saturated conditions. For the garnet-pyroxenites, this approach is consistent with the fact that the rocks do not show any evidence of partial melting, in which case the water content of the residuum would be controlled by the conditions at which melting occurred (e.g., Dziggel et al., 2012).

The PT-conditions obtained from pseudosection modelling of garnet pyroxenite and amphibolite are also in good agreement with the results of conventional geothermobarometry on fluid-independent mineral equilibria (Table 3.6), consistent with water-saturated conditions. The garnet-kyanite schist shows clear evidence of partial melting, and the fluid content for the suprasolidus part of this section was constrained from the wet solidus (White et al., 2001).

Table 3.6

PT-results based on conventional geothermobarometry (with errors) and THERMOCALC pseudosection modelling.

Sample	Two-Pyroxene ($\pm$ 50 °C)	Ca in Opx ($\pm$ 19 °C)	Enstatite in Cpx ($\pm$ 30 °C)	Hbl-Pl ($\pm$ 40 °C)	Grt-Bt	Grt-Opx ($\pm$ 2.5 kbar)	GAES ($\pm$ 1.5 kbar)	GADS ($\pm$ 1.6 kbar)	Grt-Hbl-Pl-Qz ($\pm$ 0.5 kbar)	GASP ($\pm$ 0.65 kbar)	Pseudosection modelling
Garnet-Pyroxenite											
566273	805-905 °C	787-864 °C	808-921 °C			20.29-23.44 kbar					760-810 °C, 17.0-18.8 kbar
566279	700-847 °C	777-815 °C	692-851 °C			17.45-19.21 kbar					740-790 °C, 17.0-18.9 kbar
Retrogressed Eclogite											
318349				775-830 °C (Grt) and 700-750 °C (Di)							760-880 °C, 13.8-15.4 kbar
566218		731-895 °C		784-932 °C (Grt) and 681-727 °C (Di)			9.87-11.93 kbar	8.15-11.27 kbar			890-950 °C, 10.0-11.6 kbar
566216				702-762 °C (Grt) and 700-730 °C (Di)				10.80-12.22 kbar			760-840 °C, 10.9-11.8 kbar
525224			666-877 °C	705-749 °C (Grt) and 679-706 °C (Di)			11.25-11.35 kbar				860-890 °C, 12.5-13.0 kbar
Garnet-Amphibolite											
566201				661-744 °C					8.93-9.28 kbar		
566223				684-772 °C					8.64-9.81 kbar		700-840 °C, 10.1-10.9 kbar
566228				660-740 °C					9.29-9.43 kbar		
Garnet-Kyanite Schist											
566267					584- 710 °C					8.91-8.93 kbar	665-735 °C, 8.8-9.5 kbar

Table 3.7

Bulk-rock compositions (in mol. %) for samples used in pseudosection modelling.

Grt-Py - Garnet-pyroxenite, Retr – Retrogressed eclogite, Grt-Amp – Garnet-amphibolite, Sed – Metasediment.

Sample No.	566273	566279	318349	566218	566216	525224	566223	566267
Sample type	Grt-Py	Grt-Py	Retr	Retr	Retr	Retr	Grt-Amp	Sed
Figure	3.12a	3.12b	3.13a	3.13b	3.13c	3.13d	3.14a	3.14b
SiO ₂	49.46	50.02	48.62	50.35	53.05	52.85	55.86	65.91
TiO ₂	0.10	0.15	0.60	0.40	1.09	1.22	0.46	0.67
Al ₂ O ₃	1.34	0.63	9.07	13.72	9.65	10.28	9.49	18.40
FeO	6.85	5.75	12.65	6.47	8.45	11.43	10.53	8.81
MgO	25.09	25.63	12.25	11.06	11.78	10.14	12.74	3.34
CaO	16.39	17.14	11.74	14.88	12.40	10.46	8.69	0.32
Na ₂ O	0.76	0.67	5.07	3.00	3.42	2.93	1.91	0.42
K ₂ O	-	-	-	0.12	0.16	0.70	0.33	2.12
O	0.69	0.57	1.26	0.65	0.85	1.14	1.05	0.004
H ₂ O	excess	excess	excess	excess	excess	excess	excess	5.47

K₂O has been excluded from the modelling for sample 566273, 566279 and 318349.

3.7.1 Garnet-pyroxenites

The pseudosections for two garnet-pyroxenite samples were calculated in a PT-range of 4–23 kbar and 600–950 °C. Plagioclase and quartz are not stable in the modelled range for sample 566273 (Fig. 3.12a) and are only stable at very high temperatures (> 900 °C) in sample 566279 (Fig. 3.12b). The eclogite-facies assemblage of amphibole, diopside, garnet, orthopyroxene, ilmenite and magnetite in sample 566273 is stable at 13.6–21.3 kbar and 750–950 °C (Fig. 3.12a). In sample 566279, the eclogite-facies assemblage of amphibole, diopside, garnet, orthopyroxene and ilmenite occurs in a narrow phase field at conditions of 16.4– 19.7 kbar and 720–800 °C (Fig. 3.12b). Towards lower temperatures, chlorite is added to the assemblages in both samples, whereas towards lower-pressures, garnet is lost. Rutile, which is absent in the sample 566273, is introduced at high temperatures to the assemblage in sample 566279. The compositional isopleths for X_{Grs} in garnet and X_{Fe} in diopside and garnet are shown in Fig. 3.12a and b. The best agreement between the calculated and analyzed mineral compositions is at conditions of 17.0–18.8 kbar and 760–810 °C for sample 566273 and 17.0–18.9 kbar and 740–790 °C for sample 566279. These data are consistent with the results from conventional geothermobarometry (Table 3.6).

3.7.2 Na-rich retrogressed eclogite

The Na-rich retrogressed eclogite sample 318349 contains relict omphacite and Ca-rich garnet cores, which may potentially reflect eclogite-facies metamorphic conditions. The pseudosection (Fig. 3.13a) is dominated by two large high variance phase fields involving hornblende, omphacite, ilmenite and garnet, and hornblende, omphacite, ilmenite and plagioclase, respectively. These fields are separated by the hornblende-omphacite-garnet-plagioclase-ilmenite stability field at around ~14

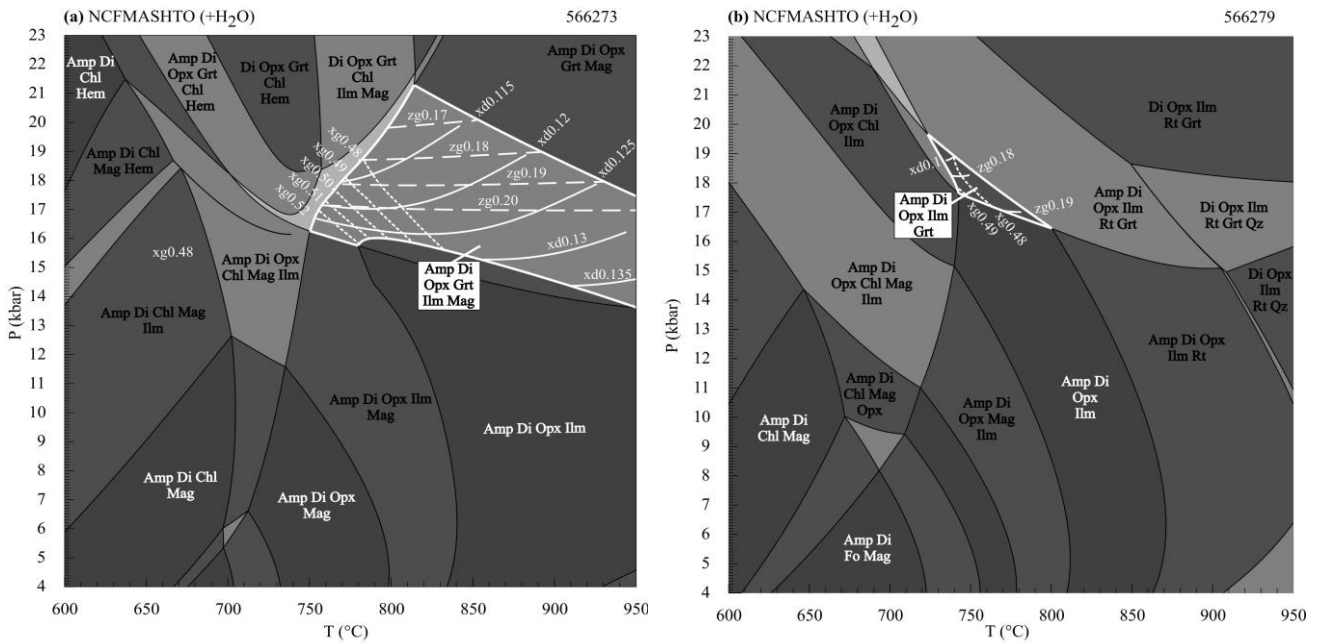


Fig. 3.12. *P-T pseudosections for (a) garnet-pyroxenite sample 566273 and (b) garnet-pyroxenite sample 566279. Bold white lines mark the stability fields of the peak assemblage. Compositional isopleths for X_{Grs} (zg), $X_{\text{Fe(Grt)}}$ (xg) and $X_{\text{Fe(Di)}}$ (xd) are also shown.*

kbar. At the eclogite-facies conditions estimated using the garnet pyroxenites, the pseudosection predicts a high-pressure assemblage of omphacite, garnet, ilmenite and hornblende. Towards lower temperatures (~ 17 kbar, 660°C) chlorite is added to the assemblage. The compositional isopleths show that at eclogite-facies conditions, X_{Jd} values in omphacite are predicted to be ~ 0.46 and X_{Grs} values in garnet are $0.21\text{--}0.23$. However, the X_{Grs} and X_{Fe} isopleths show that the garnet core compositions are stable at lower-temperature conditions of $14\text{--}19$ kbar and $610\text{--}700^\circ\text{C}$, suggesting a prograde metamorphic origin of the garnet cores. In addition, the most Na-rich omphacite found in sample 318349 occurs as an inclusion in the Mg-rich garnet rim (Grt_b) and has an X_{Jd} value of 0.42 , whereas omphacite inclusions in hornblende yield a maximum X_{Jd} value of 0.38 . According to the pseudosection, omphacite of these compositions is stable at pressures < 15 kbar and in plagioclase-bearing assemblages. Grossular values for eclogite-facies garnet overlap with those measured for Grt_b ($0.19\text{--}0.23$). However, the composition of garnet in equilibrium with omphacite with an X_{Jd} value of 0.42 is also consistent with that of Grt_b , yielding PT-conditions of $13.8\text{--}15.4$ kbar and $760\text{--}880^\circ\text{C}$ (Fig. 3.13a). The intersecting isopleths correspond to a mineral assemblage of omphacite, garnet, hornblende, ilmenite and plagioclase and indicate high-pressure granulite-facies PT-conditions during early decompression.

Ongoing decompression after the granulite-facies metamorphic overprint is indicated by the replacement of garnet by hornblende, plagioclase, ilmenite and magnetite, as well as the Al- and Na-zoning in clinopyroxene towards plagioclase. However, the retrogression resulted in the formation of domainal equilibration volumes that are not reflected by the bulk-rock chemical data used to construct

the pseudosections. The modelling thus cannot be used to provide further constraints on the retrograde PT-evolution of this sample. Hornblende-plagioclase thermometry in the clinopyroxene domain, applied to mineral pairs that are interpreted to either have formed or equilibrated during further retrogression, yields temperature conditions of 700–750 °C. These results point to amphibolite-facies conditions.

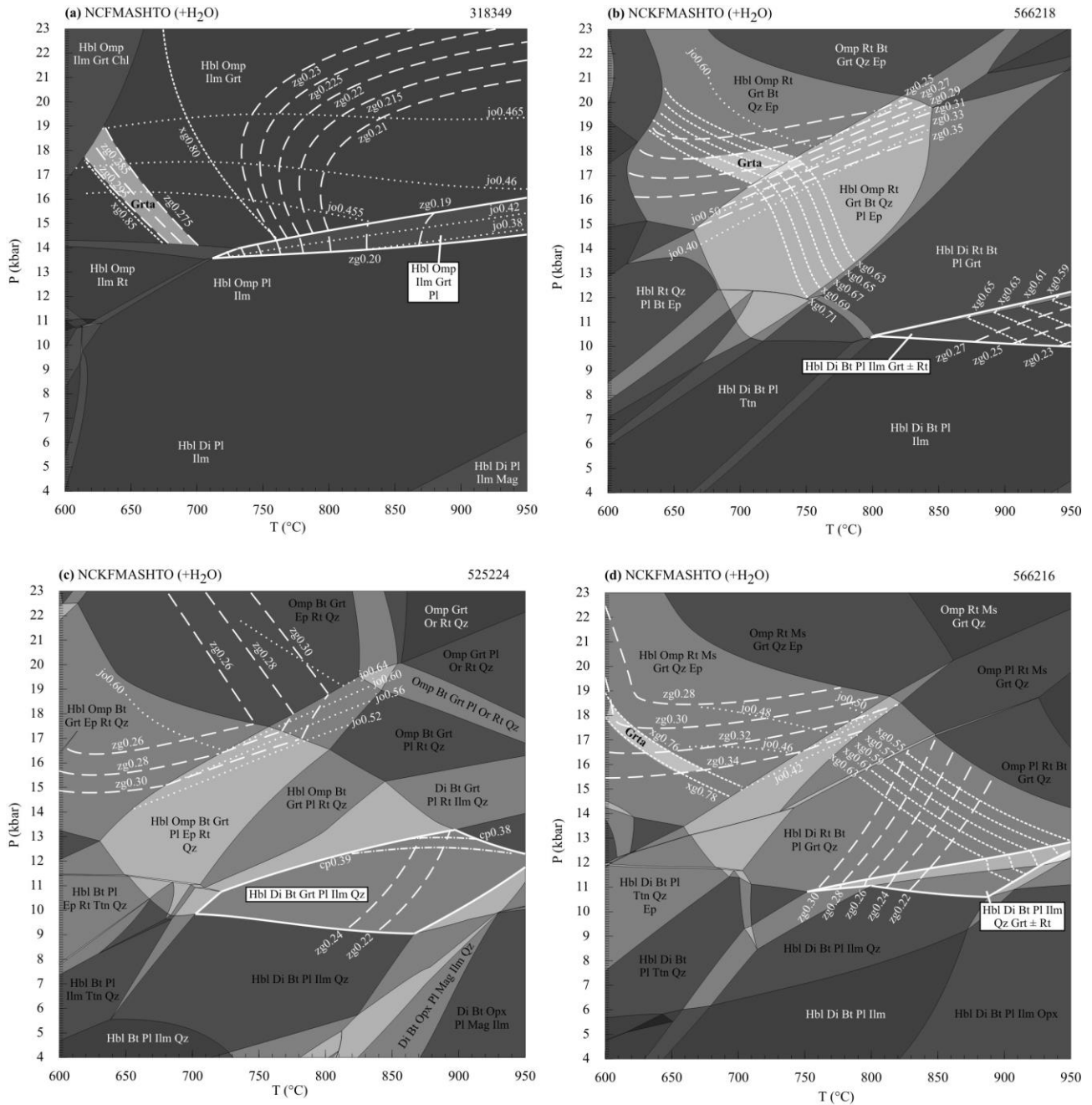


Fig. 3.13. *P-T pseudosections for (a) the Na-rich retrogressed eclogite sample 318349, (b) the retrogressed eclogite sample 566218, (c) the retrogressed eclogite sample 566216 and (d) the retrogressed eclogite sample 525224. Bold white lines mark the stability field of the dominant retrograde mineral assemblage. Compositional isopleths for X_{Grs} (zg), $X_{Fe(Grt)}$ (xg), X_{Jd} (jo) and X_{An} (cp) are also shown.*

3.7.3 Other retrogressed eclogites

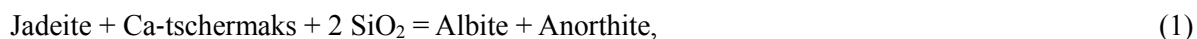
The retrogressed eclogite samples 566218 and 525224 contain orthopyroxene in coronas around garnet, whereas in sample 566216 the former presence of orthopyroxene is indicated by pseudomorphs of anthophyllite. The pseudosections (Fig. 3.13b–d), calculated in the NCKFMASHTO model system, predict an eclogite-facies metamorphic mineral assemblage of omphacite, garnet, quartz, rutile $\pm$ epidote $\pm$ plagioclase $\pm$ hornblende $\pm$ biotite/muscovite. Similar to the Na-rich sample 318349, garnet isopleths predict the Ca-rich Grt in samples 566216 and 566218 (Fig. 3.13b, d) to be stable at low-temperature, high-pressure conditions (16.6–18.1 kbar and 665–750 °C for sample 566218, 15.7–18.6 kbar and 600–675 °C for sample 566216). The dominant retrograde mineral assemblage of hornblende, diopside, garnet, plagioclase, ilmenite, biotite $\pm$ quartz $\pm$ rutile is predicted to be stable at conditions of 10.6–13.0 kbar and 700–950 °C (Fig. 3.13b–d). For samples 566218 and 566216 (Fig. 3.13b, d), it is unclear whether rutile is part of the retrograde assemblage or preserved as a relict higher-pressure phase, and therefore the stability field of the retrograde assemblage both above and below the rutile-in boundary are considered. Other than compositional isopleths of X_{Grs} in garnet, those of X_{An} in plagioclase are only partly consistent with the mineral compositions analyzed. The compositional isopleths of X_{Grs} in garnet further constrain the PT-window to 10.0–11.6 kbar and 890–950 °C in sample 566218 (Fig. 3.13b), 12.5–13.0 kbar and 860–890 °C in sample 525224 (Fig. 3.13c), and 10.9–11.8 kbar and 760–840 °C in sample 566216 (Fig. 3.13d). Hornblende-plagioclase thermometry applied to hornblende-plagioclase pairs in coronas around garnet in sample 566218 yields temperature conditions of 784–932 °C. These values are consistent with the results of the Ca in orthopyroxene and enstatite in clinopyroxene geothermometers and with the calculated stability fields from pseudosection-modelling (Table 3.6), even though the errors are relatively large. However, as the Na-rich sample, these samples are characterized by domainal equilibration volumes, which are not reflected by the pseudosections calculated based on whole-rock geochemical data. In addition, hornblende-plagioclase thermometry applied to hornblende-plagioclase pairs in the clinopyroxene domain in sample 566218 and for both domains in samples 566216 and 525224 reveals distinctly lower-temperature conditions of 681–727 °C, 700–762 °C and 679–749 °C (Table 3.6). Pressures calculated via the GAES and GADS barometers using pyroxenes in coronas around garnet and associated plagioclase are consistent in all three samples and range between 8.2 and 12.2 kbar (Table 3.6).

3.7.4 Amphibolite

The pseudosection for the garnet-amphibolite sample 566223 was calculated for 4–15 kbar and 600–950 °C (Fig. 3.14a). The amphibolite-facies mineral assemblage of hornblende, plagioclase, quartz, garnet, biotite and rutile is stable at conditions of 8.0–15.0 kbar and 680–870 °C. In this rock, garnet is only stable at pressures above 8 kbar, whereas ilmenite is stabilized towards higher-temperature conditions. The best agreement between the modelled and analyzed mineral compositions

and Friend, 1989). This worm-like to globular (Joanny et al., 1991; Agbossoumonde et al., 2001; Lardeaux et al., 2001) intergrowth is a typical symplectite after omphacite, which is moved out of its stable PT-range during decompression, unmixes and subsequently exsolves plagioclase lamellae within its crystal structure (Mysen, 1972; Boland and Van Roermund, 1983; O'Brien, 1989; Joanny et al., 1991).

The plagioclase symplectites formed from omphacite after the reaction

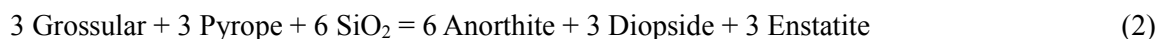


with Si (among other, minor components) being supplied by other metamorphic phases in the vicinity, such as quartz and garnet (Mysen, 1972; Boland and Van Roermund, 1983; O'Brien, 1989; Joanny et al., 1991). With increasing degree of retrogression, the ongoing consumption of jadeite and Ca-tschermaks by plagioclase results in the formation of Na-poor omphacite, sodic augite or diopside (Mysen, 1972). Although in most retrogressed eclogite samples, omphacite was fully replaced by reaction 1, the quartz-absent Na-rich retrogressed eclogite sample 318349 still preserves omphacite as inclusion in garnet and amphibole or in the core of larger clinopyroxene grains. These grains only show less sodic clinopyroxene compositions at their rims (Fig. 3.9a) and probably did not develop plagioclase lamellae, perhaps due to the high Na content in the rock or due to the consumption of quartz in this quartz-free sample prior to completion of reaction (1). Pseudosection modelling and conventional geothermobarometry (especially in sample 318349, Fig. 3.13a) show that omphacite decomposition was initiated at high-pressure granulite-facies conditions, with plagioclase contents increasing during decompression and clinopyroxene and plagioclase becoming more calcic due to the breakdown of omphacite and garnet. The fine-grained symplectites in the clinopyroxene domain contain a dense network of grain-boundaries, along which fluid might have infiltrated the rock. Such networks represent sites along which retrograde reactions progress the furthest and fastest (Thompson and England, 1984; Nutman and Friend, 1989; Kaneko et al., 2000; Brewer et al., 2003).

Contemporaneous with and coupled to the omphacite unmixing, garnet also became unstable. However, the stability field of garnet in metabasic rocks is much larger than that of omphacite (Fig. 3.13b–d), and garnet is shielded by its coronas, limiting permeability and thus retrogression. Full replacement of garnet is rare and more common for small grains, whereas large grains only show extensive partial replacement (Figs. 3.4 and 3.7), preserve prograde and retrograde chemical zoning (Figs. 3.4 and 3.9c) and contain inclusions of both eclogite-facies and retrograde mineral phases (Fig. 3.7). The symplectitic and coronitic lower-pressure reaction products that form after garnet are divided into dry and hydrous assemblages:

- (1) Dry replacement assemblages, preserved as thin coronas adjacent to quartz (Fig. 3.6d), are rare and characteristic of retrogression at granulite-facies conditions. These assemblages are made up of albitic plagioclase, orthopyroxene and low-Na clinopyroxene (Nutman and Friend, 1989; O'Brien, 1989; Vannucci et al., 1989; Messiga and Bettini, 1990; Zhao et al., 2001; Baldwin et al.,

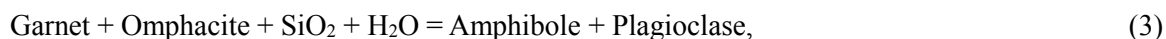
2004; Qu et al., 2011; Štípská et al., 2014), with plagioclase always constituting the innermost corona around garnet, followed by clinopyroxene and orthopyroxene adjacent to quartz. In a few samples, clinopyroxene is also present in large (100–300 µm) grains next to garnet. Their relationship to the granulite-facies replacement assemblage after garnet is unclear as they are not always associated with quartz. This suggests that the clinopyroxene may have consumed all quartz in the area and that elements for the pyroxene and plagioclase formation were supplied by garnet and quartz, after a reaction such as



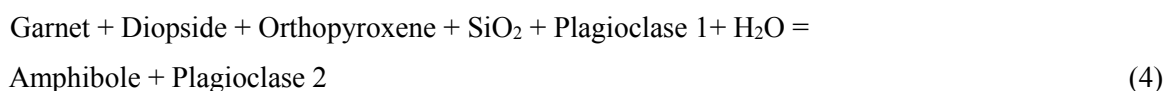
(O'Brien, 1989).

- (2) Hydrous replacement assemblages after garnet are characterized by amphibole + plagioclase, typically forming thick coronas between garnet and the clinopyroxene domain (Nutman and Friend, 1989; Messiga et al., 1990; Zhang et al., 1995). Different amphibole types, ranging from anthophyllite to magnesio-hornblende (Table 3.3), also occur associated with clino- and orthopyroxene, especially in thin coronas around garnet. The conditions and relative timing of the formation of amphibole + plagioclase coronas around garnet, and their relationship to other replacement products, remains ambiguous. Some authors favor a concurrent formation of amphibole + plagioclase and orthopyroxene + plagioclase assemblages (Nutman and Friend, 1989; Zhang et al., 1995; Baldwin et al., 2004; Štípská and Powell, 2005), with the type of replacement assemblage for garnet being dependent on the availability of water and granulite-facies assemblages being preserved due to local variations in H₂O activity. Others argue that amphibole grew at the expense of the earlier, dry orthopyroxene + clinopyroxene + plagioclase replacement assemblages and whatever is left of the high-pressure mineral phases omphacite and garnet (Nutman and Friend, 1989; Vannucci et al., 1989; Messiga and Bettini, 1990; O'Brien and Vrána, 1995; Zhao et al., 2001), implying that amphibole formation represents a distinct stage during retrogression. Due to the presence of omphacite inclusions in hornblende (Fig. 3.9a), the stability of hornblende at eclogite-facies conditions for all retrogressed eclogite samples (Fig. 3.13a–d), the common presence of intergrowths between amphibole and the pyroxenes in both domains, as well as the absence of pyroxene coronas in the majority of the samples, we favor a different model: Amphibole formation in coronas around garnet was most likely initiated during the onset of retrogression, concurrent with diopside + plagioclase formation after omphacite and later with orthopyroxene + clinopyroxene corona formation around garnet at high-pressure granulite-facies conditions, with the type of retrograde assemblage being controlled by fluid-influx. Ongoing retrogression at progressively lower-pressure conditions caused the hornblende + plagioclase pairs to re-equilibrate, with more complete re-equilibration in the clinopyroxene domain due to the presence of a large grain boundary network that facilitated fluid infiltration. This may explain the different temperature estimates for hornblende + plagioclase pairs in the two domains (Table 3.6).

Eventually, if the rock remains fluid-saturated during further retrogression, hornblende-plagioclase assemblages will replace the dry symplectite assemblages to form hornblende-rich amphibolite with varying amounts of plagioclase (see also Baker, 1986; Smelov and Beryozkin, 1993; Sajeev et al., 2010a; Faryad et al., 2013). This model suggests amphibole + plagioclase formation after the reactions



and



(Zhao et al., 2001; Baldwin et al., 2004; Sajeev et al., 2009; Butler et al., 2013). The mineral chemistry of amphibole in this context is interpreted to be mainly controlled by the textural setting rather than the PT-conditions of formation.

However, retrogressed eclogite represents a rock type in which reactions at amphibolite-facies conditions ceased prior to completion (i.e. due to limited fluid availability). In these samples, the hornblende-plagioclase coronas around garnet effectively shield the two domains from each other and hinder the transport of fluid and chemical components, which drive the coupled replacement reactions, across the domains (Messiga et al., 1990). This results in the formation of domainal equilibration volumes (as described by Zhang et al., 2000; O'Brien and Rötzler, 2003; Scott et al., 2013), with different mineral compositions between the two domains (see also Tubia and Gil Ibarguchi, 1991; Dachs and Proyer, 2001; Baldwin et al., 2004; Štípská and Powell, 2005; Sajeev et al., 2010b; Doukkari et al., 2014) and locally (as observed for garnet, hornblende and plagioclase) between domains of the same type.

In the samples studied, large garnet grains preserve two distinct compositions: Ca-rich Grt_a formed at high-pressure (14–19 kbar) and low-temperature (600–750 °C) conditions and probably represents garnet formed on the prograde path (Davoudian et al., 2007). The scarcity of other such features hints to thermal re-equilibration (both texturally and chemically) of the tectonically buried rocks during peak metamorphism (e.g. Faulhaber and Raith, 1991). The Mg-rich Grt_b, on the other hand, most likely formed at eclogite-facies conditions and equilibrated during retrograde metamorphism, with Ca-loss being attributed to plagioclase and hornblende formation (Guo et al., 2002; Rötzler et al., 2004). Thin Fe- and Mn-rich rims around garnet were produced during retrograde diffusive reactions (O'Brien, 1989; Markl and Bucher, 1997; Guo et al., 2002; Page et al., 2003; Rötzler et al., 2004).

In conjunction with omphacite and garnet, rutile also becomes unstable during decompression. At granulite-facies conditions, the excess Fe produced during the decomposition of garnet and omphacite results in the formation of ilmenite after rutile (Okrusch et al., 1991; Carswell

and O'Brien, 1993; O'Brien, 1997; Rötzler et al., 2004; Yang, 2004; Doukkari et al., 2014).

3.8.2 Metamorphic evolution and tectonic implications

Results from pseudosection modelling and conventional geothermobarometry show a large degree of consistency between rocks of different bulk composition and mineralogy, implying that all rock units of the Kuummiut Terrane underwent the same metamorphic history. Furthermore, the field relations show that there are no differences in the intrusive (enclaves in TTG gneiss) and alteration relationships (altered amphibolite-facies mineral assemblage in rims, cores with high-pressure or relict high-pressure mineral assemblage) between garnet-pyroxenite and retrogressed eclogite. As the high-pressure metamorphic event must have occurred after dyke emplacement (Nutman et al., 2008), this requires that the dykes, ultrabasic rocks and the surrounding TTG gneiss experienced the same metamorphic evolution. Combining the PT-results (Table 3.6) with the various textural observations yields evidence for at least four stages within the metamorphic evolution of the Kuummiut Terrane, including a prograde stage (I), an eclogite-facies stage (II), a retrograde high-pressure granulite-facies stage (III) and a retrograde, high-pressure amphibolite-facies stage (IV). The clockwise PT-path (Fig. 3.15) is typical of collisional orogenic settings, in which maximum P was reached before maximum T (England and Thompson, 1984; Thompson and England, 1984; Harley, 1989; Brown, 1993), consistent with earlier studies on the Kuummiut Terrane (Nutman and Friend, 1989; Messiga et al., 1990; Nutman et al., 2008).

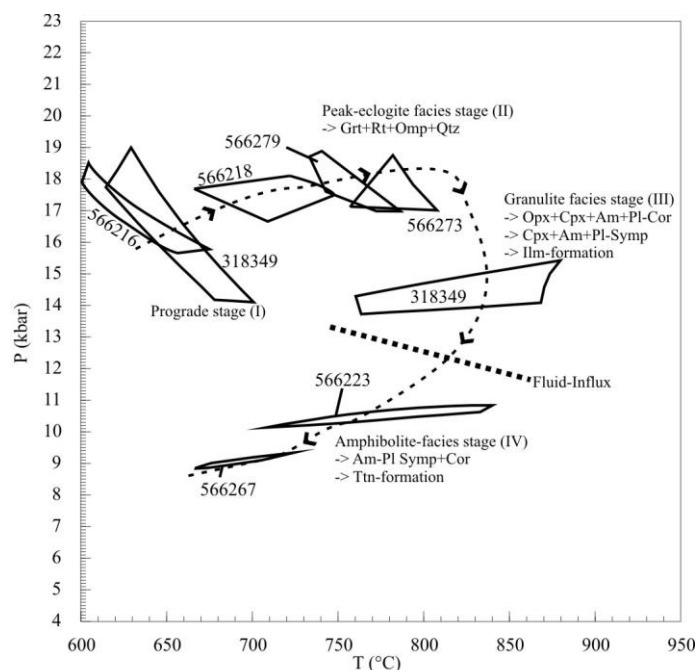


Fig. 3.15. *Metamorphic evolution of the Kuummiut Terrane as inferred from pseudosection modelling. See text for discussion.*

During collision or accretion, the Kuummiut Terrane was subducted to ~70 km depth, eventually forming Ca-rich Grt_a at conditions of 14–19 kbar and 600–750 °C (I) before being metamorphosed at eclogite-facies conditions of 17–19 kbar and 740–810 °C (II). Subsequent exhumation was associated with near-isothermal decompression into mid- to lower crustal levels of ~30 km and encompasses

several stages of retrogression (as also noted in Nutman and Friend, 1989; Mengel et al., 1990; Messiga et al., 1990; Nutman et al., 2008). Decompression initially involved the replacement of the eclogite-facies assemblage by either dry or wet symplectitic assemblages at high-pressure granulite-facies conditions of 13.8–15.4 kbar and 760–880 °C (III). The final stage of retrogression is characterized by further decompression, in combination with minor cooling, at high-pressure amphibolite-facies conditions of 8.8–10.9 kbar and 660–840 °C (IV), consistent with PT-conditions of 8–11 kbar and 660–780 °C for the decompressed assemblage by Nutman and Friend (1989) and Mengel et al. (1990).

In contrast to the basic and ultrabasic rock units, the TTG gneiss preserves no evidence for high-pressure metamorphism, as is commonly observed in many high-grade metamorphic terranes (Cooke and O'Brien, 2001). Two models exist to explain the occurrence of high-pressure rocks in a low-pressure terrane (Peacock and Goodge, 1995). In a foreign eclogite model (Erdmer and Helmstaedt, 1983; Peacock and Goodge, 1995), TTG gneiss never experienced eclogite-facies PT-conditions and the high-grade metamorphic rocks represent tectonically interleaved slivers. Such a model is at odds with the timing of dyke emplacement and high-pressure metamorphism in the Kuummiut Terrane (Nutman et al., 2008). Furthermore, the geometry of the basic dykes is often still recognizable and no major shear zones are developed at their margins. The in-situ eclogite model suggests a common PT-evolution for the high-pressure rocks and the host gneiss, with differences in recorded metamorphic grade due to a variable overprint. This variable overprint may be due to metastability of typical amphibolite-facies assemblages in TTG gneiss beyond prograde high-pressure amphibolite-facies metamorphism due to previous dehydration (Young and Kylander-Clark, 2015). However, we consider this explanation as unreasonable for the Kuummiut Terrane, as it would suggest localized hydration of the metabasic rock units and is also at odds with the presence of biotite in TTG gneiss. Other studies show that TTG gneiss may rarely preserve an eclogite-facies mineral assemblage in low-strain zones, where it is able to escape late-orogenic deformation and recrystallization (Krabbendam and Wain, 1997; Wain, 1998; Cooke and O'Brien, 2001; Van Gool et al., 2002). This is consistent with TTG gneiss being less resistant to deformation than eclogite (Peacock and Goodge, 1995; Wain, 1998), consequently re-equilibrating much faster during ductile deformation, and our own observations in the Kuummiut Terrane, where TTG gneiss preserves high-strain fabrics, whereas metabasic boudins are at low strain. Another factor that explains only local preservation of eclogite-facies mineral assemblages is varying fluid activity (Nutman and Friend, 1989; Elvevold and Gilotti, 2000; Cooke and O'Brien, 2001; Sajeew et al., 2009), with the TTG gneiss representing a sink for the fluids, consequently preventing pervasive fluid-rock interaction in the basic and ultrabasic rocks (Sajeew et al., 2009) and only allowing for extensive retrogression along boudin margins (Nutman and Friend, 1989; Elvevold and Gilotti, 2000). As such, TTG gneiss and basic-ultrabasic rock boudin margins are usually completely overprinted by lower-grade assemblages, while boudin cores preserve high-pressure mineral assemblages (Nutman and Friend, 1989; Cuthbert et al.,

2000; Elvevold and Gilotti, 2000; Tsai and Liou, 2000; Zhang et al., 2003; Nutman et al., 2008; Sajejev et al., 2009). We propose that a combination of ductile deformation and fluid-influx resulted in the variable overprint of the high-grade assemblages in TTG gneiss and rims of metabasic boudins in the Kuummiut Terrane.

The lack of evidence for partial melting as well as consistency between pseudosection modelling and conventional geothermobarometry indicate that the pro- and retrograde evolution of the garnet-pyroxenite and amphibolite occurred at water-saturated conditions. In contrast, retrogressed eclogite was shielded from widespread fluid-influx during decompression, although hornblende stability at eclogite-facies conditions suggests that the system was never completely dry (Fig. 3.13). Nevertheless, the progressive, regional-scale re-equilibration at amphibolite-facies conditions implies that intense rehydration occurred during exhumation, with TTG gneiss being more severely affected. The large amount of hornblende formed during retrogression of the metabasic rocks shows that fluid infiltration most likely dominated during the transition from stage III to IV, with the fluid most probably being derived from an external source.

In concordance with earlier work by Nutman et al. (2008) and Kolb (2014), we interpret the Isertoq Terrane with the AIC as the upper and the Kuummiut Terrane as the lower plate in a subduction- to collision scenario (Fig. 3.16a).

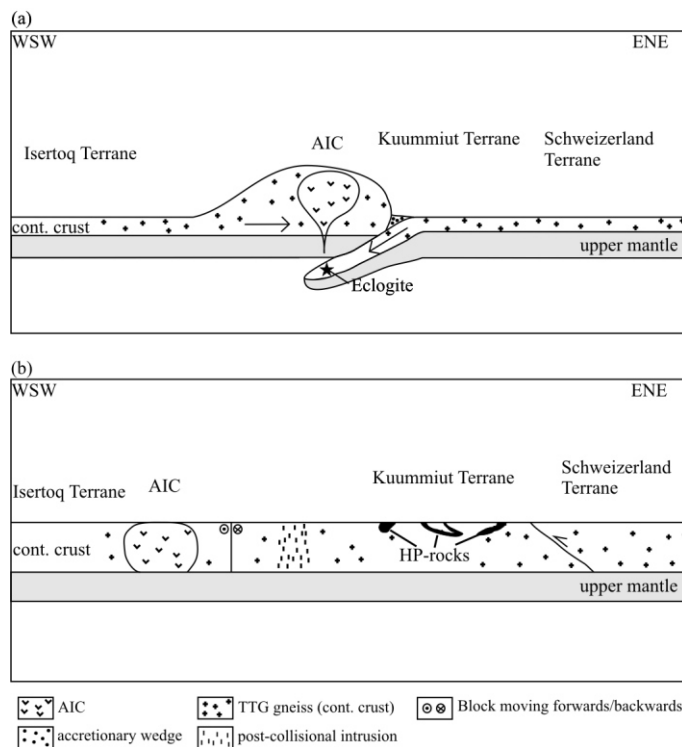


Fig. 3.16. Schematic cross-section of the tectonometamorphic evolution of the Nagssugtoqidian Orogen in South-East Greenland. (a) Burial stage, (b) Present day. For discussion see text.

Initial subduction of oceanic crust was followed by subduction of the continental crust of the Rae Craton (Kuummiut Terrane) underneath the North Atlantic Craton (Isertoq Terrane). This part of the

tectonic evolution is represented by the prograde metamorphic path of the high-pressure rocks of the Kuummiut Terrane leading to eclogite-facies metamorphism. Further subduction stopped at a certain stage due to processes such as delamination, mechanical stacking of the two plates or due to a lack of density differences between the upper and lower plate. The clockwise, near-isothermal decompression path determined for the Kuummiut Terrane, suggests an at least initially rapid exhumation, as the unroofing of the deeply subducted rocks had to be faster than the rate of thermal relaxation and cooling (Harley, 1989; Ring et al., 1999; Zhao et al., 2001; Guo et al., 2002; Baldwin et al., 2004). The fast exhumation of the subducted rocks in the Kuummiut Terrane was controlled by tectonic processes first, while later being controlled by a combination of erosional and tectonic processes (e.g. Rubatto and Hermann, 2001), leading to slow cooling with minor decompression and exposure of high-grade metamorphic rocks at the surface (Fig. 3.16b).

Following Brown (2007), the Kuummiut Terrane may be classified as a medium-temperature, eclogite-, high-pressure granulite-facies (E-HPG) metamorphic belt, formed in a subduction- to collision orogenic system. The metamorphic evolution of the Kuummiut Terrane is remarkably similar to a number of other Paleoproterozoic high-pressure metamorphic belts, such as the Usagaran and Ubende belts in Tanzania (Möller et al., 1995; Sklyarov et al., 1998) and the Trans-North China Orogen (Zhao et al., 2001). A similar clockwise PT-path with somewhat higher-temperature eclogite-facies conditions (920–1000 °C and 18–20 kbar) was also recorded from the Snowbird tectonic zone in Canada (Baldwin et al., 2004). Other Paleoproterozoic high-pressure metamorphic belts record peak high-pressure granulite- to amphibolite-facies conditions (e.g. Nutman et al., 1992; Smelov and Beryozkin, 1993; Smithies and Bagas, 1997; Guo et al., 2002; Zeh et al., 2004; St-Onge et al., 2007; Block et al., 2015), but it is unclear whether these lower-pressure assemblages are a result of a high-temperature overprint of eclogite-facies mineral assemblages or if they were never exposed to higher pressures. That high-pressure metamorphic conditions may vary considerably even within one orogenic system has recently been shown in the western part of the Nagssugtoqidian Orogen in West Greenland, where some rocks record evidence for UHP metamorphic conditions (ca. 70 kbar and 975 °C; Glassley et al., 2014). The range from UHP (ca. 70 kbar) to E-HPG (17–19 kbar) conditions may be taken as evidence for laterally varying depths of subduction, but may also be due to local variations in exhumation rates and processes, which are far from being well understood. Ring et al. (1999) and Glassley et al. (2014) suggest that a high degree of tectonic interleaving is required to juxtapose rocks of such different pressure regimes, which may either reflect mixing processes in a subduction channel, tectonic imbrication during collision, tectonic slicing during shear zone development or all three together. According to Ring et al. (1999), these different processes operate at different levels in the lithosphere. Interestingly, there is little difference between the eclogite-facies metamorphic pressures and inferred depths of subduction in collisional (e.g. Nagssugtoqidian and North China Orogen) and accretionary (e.g. Usagaran and Ubende belts) Paleoproterozoic orogenic systems, as indicated by eclogites (Nagssugtoqidian, this study; 17–19 kbar vs. Usagaran belt; Möller

et al., 1995; 18–19 kbar) and high-pressure granulites (North China Orogen; Zhao et al., 2001; 13.5–14.5 kbar vs. Ubende belt; Sklyarov et al., 1998; 12–14 kbar), respectively. However, the number of available PT-estimates is limited, and it is currently unknown how representative the published data are. In addition, our study shows that eclogite-facies mineral assemblages in Paleoproterozoic basic dykes of the Kuummiut Terrane are extremely rare. Therefore, failure to quantify eclogite-facies metamorphic conditions in Paleoproterozoic orogens may simply be related to a lack of preservation due to the strong retrograde metamorphic overprint.

3.9 Conclusions

- Eclogite-facies rocks of the Kuummiut Terrane in the Nagssugtoqidian Orogen were variably retrogressed during decompression, with the degree of replacement and the type of replacement assemblage being controlled by fluid availability.
- Relatively unaltered garnet-pyroxenites represent texturally well-equilibrated eclogite-facies samples that constrain the eclogite-facies conditions to 17–19 kbar and 740–810 °C.
- Replacement reactions in retrogressed eclogite ceased prior to completion, causing formation of clinopyroxene- and garnet-dominated domains, with domainal equilibration volumes and preservation of relict higher-pressure phases.
- Pseudosection modelling, geothermobarometry and mineral texture analysis in retrogressed eclogite predicts that omphacite and garnet were replaced by dry (clinopyroxene + orthopyroxene + plagioclase) and wet (hornblende + plagioclase) symplectitic assemblages during high-pressure granulite- (13.8–15.4 kbar, 760–880 °C) and amphibolite-facies (8.2–12.2 kbar, 680–750 °C) retrograde metamorphic stages. The latter is consistent with PT-conditions determined from texturally well-equilibrated garnet-amphibolite and garnet-kyanite schist.
- The results yield a clockwise PT-path, with exhumation from ~70 to about 30 km depth. Exhumation was initially rapid and controlled by tectonic processes.

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4 Age and temperature-time evolution of retrogressed eclogite-facies rocks in the Paleoproterozoic Nagssugtoqidian Orogen, South-East Greenland: constrained from U-Pb dating of zircon, monazite, titanite and rutile

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4.1 Abstract

LA-ICP-MS U-Pb dating of zircon, monazite, titanite and rutile was carried out to investigate the temperature-time evolution of eclogite-facies rocks in the Kuummiut Terrane of the Paleoproterozoic Nagssugtoqidian Orogen in South-East Greenland. The terrane is dominated by Archean TTG gneiss and a variety of supracrustal rocks; basic dykes intruded the gneiss during Paleoproterozoic deformation. Detrital zircon in garnet-kyanite schist gives Archean to Paleoproterozoic dates and confines the maximum deposition of the metasediment precursor to 2107 ± 21 Ma. Intrusion of the metabasic dykes occurred at 2146 ± 63 and 2092 ± 22 Ma, within error of the detrital zircon date, possibly indicating near-contemporaneous dyke emplacement and sedimentation. About 200 m.y. after dyke emplacement, the Kuummiut Terrane underwent a clockwise PT-evolution, involving eclogite-facies metamorphism and subsequent exhumation into the mid crust. The majority of zircon, monazite and titanite give metamorphic dates between 1891 ± 10 and 1882 ± 3 Ma. Although the REE patterns in metamorphic zircon reflect growth at eclogite-facies conditions, the zircons are associated with retrograde mineral assemblages and their dates are indistinguishable from amphibolite-facies titanite. This may be interpreted to indicate that the timing of eclogite- and high-pressure amphibolite-facies metamorphism overlap within error, consistent with rapid and tectonically-controlled exhumation. However, previous studies have shown that the REE and U-Pb systematics in zircon may be decoupled during retrograde metamorphism, and the range in dates is thus best interpreted to reflect mineral growth and recrystallization during high-pressure amphibolite-facies retrogression. Monazite and titanite dates between 1872 ± 70 and 1821 ± 31 Ma reflect regional medium-pressure amphibolite-facies metamorphism and mark the final stages of compressional deformation in the Kuummiut Terrane. The subsequent thermal evolution was associated with titanite growth until 1738 ± 61 Ma, a time where the majority of rutile cooled below its closure temperature. The youngest rutile dates at 1645 ± 63 and 1617 ± 91 Ma correlate with the emplacement of post-tectonic intrusive complexes. Collectively, the data show that after an initial tectonically-controlled exhumation, the Kuummiut Terrane experienced relatively slow, erosion-controlled cooling with only minor thermal perturbations during the waning stages of metamorphic and magmatic activity.

4.2 Keywords

Eclogite, South-East Greenland, U-Pb dating, Retrogression, Paleoproterozoic

4.3 Introduction

In the last few decades, a growing number of studies reported the presence of eclogite- to high-pressure granulite-facies mineral assemblages in Paleoproterozoic collisional belts. The high-pressure assemblages are particularly common in metabasic dykes intrusive into Archean TTG gneiss, e.g. in the Nagssugtoqidian Orogen in South-East Greenland (Nutman and Friend, 1989; Nutman et al., 2008; Müller et al., 2018), the Aldan Shield of Siberia (Nutman et al., 1992; Smelov and Beryozkin, 1993), the North China Craton (Zhao et al., 2001; Guo et al., 2002; Tam et al., 2012), the Snowbird tectonic zone in the western Canadian Shield (Baldwin et al., 2004) and the Belomorian Mobile Belt in the Kola Peninsula (Skublov et al., 2011; Imayama et al., 2017; Liu et al., 2017; Yu et al., 2017). In other Paleoproterozoic collisional belts, the high-pressure assemblages occur in metabasic lithologies of MORB affinity, e.g. in the Usagaran and Ubende belts of Tanzania (Möller et al., 1995; Sklyarov et al., 1998; Collins et al., 2004; Boniface et al., 2012) and the Eburnian-Transamazonian orogen of southern Cameroon (Loose and Schenk, 2018). The Paleoproterozoic high-pressure mineral assemblages document an era of global subduction-related orogenic activity during the formation of the supercontinent Nuna or Columbia (Hoffman, 1988; Rogers and Santosh, 2002; Zhao et al., 2002; Brown, 2008; Reddy and Evans, 2009; St-Onge et al., 2009; Mertanen and Pesonen, 2012; Müller et al., 2018). The accurate and precise dating of the high-pressure metamorphic stage and other stages in the PT-evolution of these rocks is not straightforward and remains a challenge due to an often strong retrograde overprint or the lack of datable minerals. Bulk-rock or multigrain dating methods such as Rb-Sr, Sm-Nd and K-Ar may be complicated by the low closure temperatures of the geochronometers in question (e.g. Harrison, 1981; Mezger et al., 1992; Christensen et al., 1994), low element concentrations (Thöni, 2006) or incomplete trace element and isotopic equilibration during metamorphism (Rubatto and Hermann, 2003; Xie et al., 2004; Thöni, 2006). A more promising method to gain robust age data may be U-Pb dating of accessory phases with relatively high closure temperatures for Pb diffusion, such as zircon, monazite, titanite and rutile (Möller et al., 2000; Parrish, 2001; Ayers et al., 2002; Corfu et al., 2003a; Brewer et al., 2003; Timmermann et al., 2004). Zircon is the most commonly dated accessory phase and a robust tool for determining metamorphic ages due to its high U concentrations and low-Pb diffusivity over a large range of crustal conditions (Cherniak et al., 1997; Rubatto et al., 1999). In addition, it commonly preserves complex growth histories, with inherited magmatic and prograde, peak and retrograde metamorphic ages (Rubatto et al., 1999; Hermann et al., 2001; Ayers et al., 2002; Rubatto and Hermann, 2003; Zhang et al., 2005; Zheng et

al., 2005; Nutman et al., 2008; Liu et al., 2017). Closure temperatures are estimated at >900 °C (Cherniak et al., 1997; Lee et al., 1997; Cherniak and Watson, 2000). Monazite incorporates larger amounts of actinides than zircon (Cherniak, 2010) but shows similar characteristics, in that it has very low Pb diffusivity (Cherniak et al., 2004) and is able to record complex P-T-t histories (Ayers et al., 2002; Mahan et al., 2006; Williams et al., 2007; Sindern et al., 2012). Closure temperatures have been estimated between 500-750 °C (Black et al., 1984; Copeland et al., 1988; Parrish, 1990; Suzuki et al., 1994; Smith and Giletti, 1997) and 900 °C (Bingen and Van Breemen, 1998; Braun et al., 1998; Kalt et al., 2000; Cherniak et al., 2004; Cherniak and Pyle, 2008). However, due to their accessory nature, an unequivocal correlation of zircon and monazite dates to a certain metamorphic stage on the PT-path is only possible via trace element data or inclusion content (Lanzirotti and Hanson, 1996; Hermann et al., 2001; Hermann and Rubatto, 2003; Rubatto and Hermann, 2007; Nutman et al., 2008). In contrast to monazite and zircon, titanite and rutile take part in metamorphic reactions, and their stability in PT-space can be determined via phase diagram modelling; however, ambiguity exists in the closure temperature estimates for both minerals. Early field-based studies indicated closure temperatures of around 450-650 °C for titanite (Mattinson, 1978; Heaman and Parrish, 1991; Mezger et al., 1991), consistent with subsequent experimental results suggesting closure around ca. 600 °C for slow cooling rates of ~5 °C/Ma and diffusion radii <500 µm (Cherniak, 1993). Other studies estimated higher closure temperatures of ~ 700 °C based on inheritance in titanite (Schärer et al., 1994; Scott and St-Onge, 1995; Pidgeon et al., 1996; Zhang and Schärer, 1996). Following from more recent studies (Kohn and Corrie, 2011; Gao et al., 2012; Spencer et al., 2013), Kohn (2017) suggested a closure temperature of around 800 °C, even at diameters of 10 µm. U-Pb dating of rutile has generally received less attention due to the often extremely low U content of rutile (Zack et al., 2011). Closure temperatures for Pb diffusion are estimated between 400 and 700 °C, depending on grain size and cooling rate (Mezger et al., 1989; Cherniak, 2000; Vry and Baker, 2006; Kooijman et al., 2010; Warren et al., 2011). Where successfully measured, rutile usually gives cooling ages (Connely et al., 2000; Li et al., 2003; Baldwin et al., 2004; Zack et al., 2011).

Based on U-Pb isotope data of zircon, monazite, titanite and rutile, we examine the geochronology of supracrustal deposition, dyke emplacement and metamorphism in a variety of retrogressed eclogite-facies rocks from the mainly high-pressure amphibolite-facies Kuummiut Terrane in the Paleoproterozoic Nagssugtoqidian Orogen of South-East Greenland (Wright et al., 1973; Chadwick et al., 1989; Dawes et al., 1989; Nutman and Friend, 1989; Messiga et al., 1990; Nutman et al., 2008; Kolb, 2014; Müller et al., 2018; Dziggel and Müller, 2018). These rocks were recently investigated in terms of their mineral textural evolution and PT-history, yielding four metamorphic stages on a clockwise PT-path (Müller et al., 2018). In order to provide constraints on the thermal evolution of the Kuummiut Terrane, we combine the PT-evolution with new and previously published age data (Bridgwater et al., 1990; Kalsbeek et al., 1993; Nutman et al., 2008; Thrane et al., 2016; Nicoli et al., 2018). Although the age of eclogite-facies metamorphism could not

be constrained unequivocally, our results are consistent with an initially rapid, tectonically-controlled exhumation followed by slow, erosion-controlled cooling.

4.4 Geological setting

The Nagssugtoqidian Orogen in South-East Greenland is situated in the Tasiilaq area (Fig. 4.1; Andrews et al., 1973; Wright et al., 1973; Bridgwater et al., 1976; Chadwick et al., 1989; Nutman et al., 2008; Kolb, 2014). This deeply eroded and roughly southeast-northwest trending Paleoproterozoic collisional orogen is bound by the Rae Craton to the north and the North Atlantic Craton to the south (Bridgwater, 1976; Chadwick et al., 1989; Kolb, 2014). The orogen has been subdivided into three different terranes, which from north to south are the Schweizerland, Kuummiut and Isertoq Terranes, the latter of which includes the Ammassalik Intrusive Complex (Fig. 4.1; Kolb, 2014; Lebrun et al., 2018). The available structural, geochronological and petrological data show that during the Paleoproterozoic collision, the continental crust of the Rae Craton (Kuummiut Terrane) was subducted underneath the North Atlantic Craton (Isertoq Terrane), leading to eclogite-facies metamorphism in the lower plate and the formation of the Ammassalik Intrusive Complex in the upper plate (Kolb, 2014; Müller et al., 2018; Lebrun et al., 2018).

The oldest and most dominant rock unit in the Kuummiut Terrane is fine- to medium-grained, Archean TTG gneiss (Fig. 4.1a), with a common amphibolite-facies mineral assemblage of quartz + biotite + plagioclase $\pm$ garnet $\pm$ hornblende (Chadwick et al., 1989; Dawes et al., 1989; Dziggel and Müller, 2018). Samarium-Nd model dates (T_{DM}) and zircon U-Pb data give roughly consistent intrusion ages for the protolith of the gneiss at 3020 to 2780 Ma (Kalsbeek and Taylor, 1989; Kalsbeek et al., 1993) and 3076 ± 14 to 2707 ± 15 Ma (Kalsbeek et al., 1993; Nutman et al., 2008; Kokfelt et al., 2016a), respectively. Archean to Paleoproterozoic supracrustal rocks are in tectonic contact with the Archean gneiss and occur as boudins, boudinaged layers and up to hundred-meter-wide foliation-parallel belts (Wright et al., 1973; Hall et al., 1989; Kolb, 2014; Dziggel and Müller, 2018). Two separate units are distinguished (Kolb, 2014). The Kuummiut unit occurs in an area north of Tasiilaq and east of Sermilik Fjord (Fig. 4.1a), and mainly comprises paragneiss, quartzite and schist, locally with marble, calc-silicate rocks, amphibolite and ultramafic rocks. The Helheim unit is exposed west and north of the Sermilik Fjord, and is made up of paragneiss, schist, amphibolite, marble, calc-silicate rocks, quartzite and ultramafic rocks. Kyanite- and sillimanite-bearing paragneiss from both units shows significant input of Archean detritus, as indicated by Sm-Nd model dates of ca. 2880 to 2750 Ma (T_{DM} ; Kalsbeek and Taylor, 1989) and detrital zircon U-Pb dates between ca. 2800 and 2600 Ma (Nutman et al., 2008; Thrane et al., 2016). The deposition of the precursor to the paragneiss has been constrained to between 2600 and 1900 Ma (Kalsbeek and Taylor, 1989; Kalsbeek et al., 1993; Nutman et al., 2008; Thrane et al., 2016). Several intrusive events occurred in the

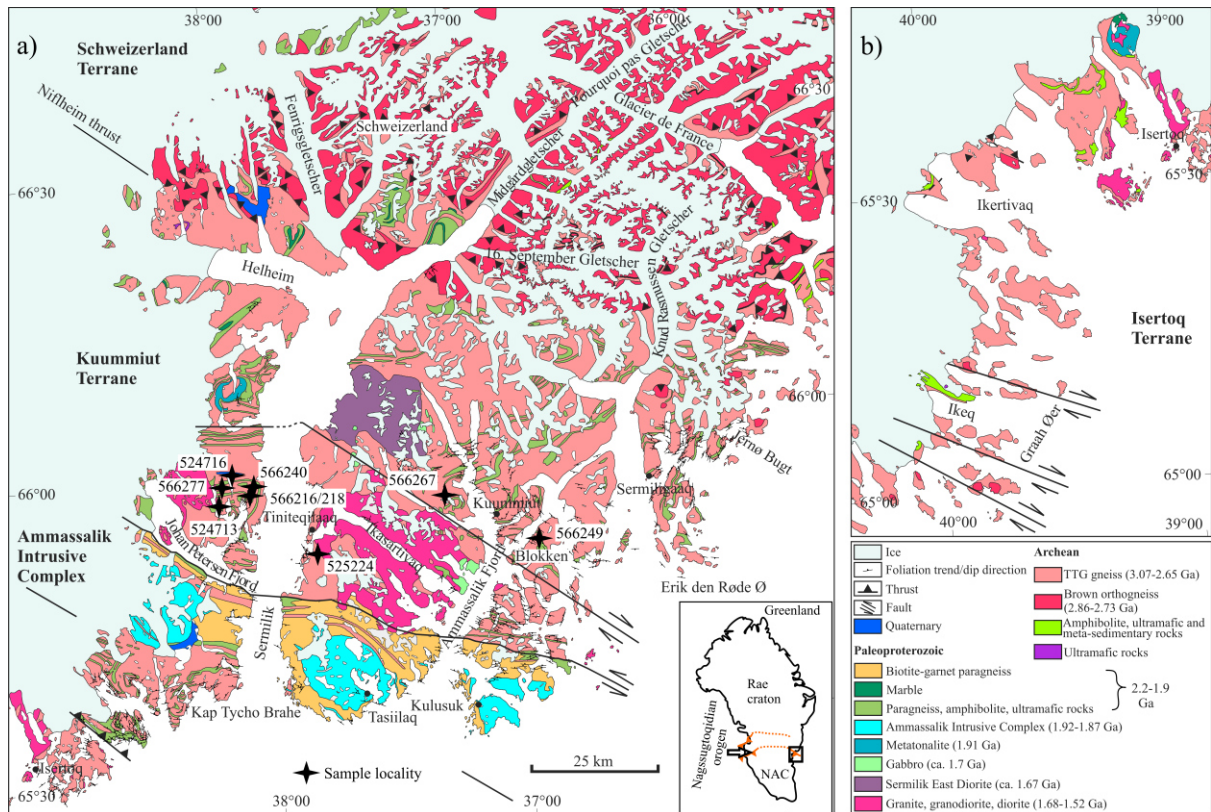


Fig. 4.1. Geological map of the northern (a) and (b) southern Nagssugtoqidian Orogen (modified after Escher, 1990). Stars in (a) represent sample sites for U-(Th)-Pb analysis.

Kuummiut Terrane during the Paleoproterozoic. Swarms of discordant, east-west to east-northeast-west-southwest striking basic dykes occur as boudins within the TTG gneiss (Andrews et al., 1973; Wright et al., 1973; Chadwick et al., 1989; Dawes et al., 1989; Bridgwater et al., 1990; Müller et al., 2018; Dziggel and Müller, 2018). Dyke emplacement has been interpreted to have occurred prior to eclogite-facies metamorphism at 1867 ± 28 Ma (Nutman et al., 2008), based on crosscutting relationships (Chadwick et al., 1989; Kalsbeek and Taylor, 1989; Nutman and Friend, 1989). Samarium-Nd model dates (T_{DM} 2400 to 2200 Ma; Bridgwater et al., 1990; Kalsbeek et al., 1993) and zircon U-Pb data (2015 ± 15 Ma; Nutman et al., 2008) from metabasic dykes in the northern part of the Kuummiut Terrane suggest magma formation and dyke emplacement prior to eclogite-facies metamorphism and the possibility of multiple dyke generations (Kolb, 2014). A tonalite west of Sermilik Fjord yields a zircon U-Pb age of 1901 ± 9 Ma (Kalsbeek et al., 1993; Nutman et al., 2008). The Sermilik East diorite intruded at 1670 ± 5 Ma (zircon U-Pb; Kokfelt et al., 2016b), consistent in age with other major post-tectonic granite, diorite and gabbro complexes, as well as several generations of pegmatite dykes (zircon U-Pb - 1680 ± 9 to 1545 ± 15 Ma; Kalsbeek et al., 1993 and references therein; Thrane et al., 2016).

The Isertoq Terrane to the south of the Ammassalik Intrusive Complex is also dominated by Archean TTG gneiss (Fig. 4.1b), here characterized by low-pressure amphibolite-facies mineral

assemblages (Escher et al., 1989; Kalsbeek et al., 1993; Kokfelt et al., 2016a). Intrusion ages for the protolith to the gneiss are given at 3050 Ma (Sm-Nd model age; Kalsbeek et al., 1993) and 2822 ± 8 to 2724 ± 4 Ma (zircon U-Pb; Kokfelt et al., 2016a). A deformed pegmatite on an island east of the settlement of Isertoq has been dated at 2742 ± 6 Ma (zircon U-Pb; Kokfelt et al., 2016a). A paragneiss sample near Isertoq yields detrital zircon dates from 2850 to 1900 Ma, indicating Archean detritus and final deposition during the Nagssugtoqidian orogeny (Thrane et al., 2016). During the Paleoproterozoic, the Isertoq Terrane was intruded by several, variably sized igneous intrusions. Foliation-parallel and east-west striking boninitic dykes show increasing degree of deformation from south to north and are Paleoproterozoic in age (<2.17 Ga; Escher et al., 1989; Kalsbeek et al., 1993; Klausen et al., 2016). The Ammassalik Intrusive Complex, in the northern part of the Isertoq Terrane, comprises three calc-alkaline magmatic intrusive centers, which were emplaced between 1911 ± 7 and 1872 ± 8 Ma (zircon U-Pb; Hansen and Kalsbeek, 1989; Kalsbeek et al., 1993; Nutman et al., 2008; Lebrun et al., 2018). North-south striking tholeiitic dykes crosscut the older Paleoproterozoic dyke generation and are interpreted as late-orogenic in origin (Klausen et al., 2016). Granite and diorite north of Isertoq yield zircon U-Pb intrusion ages of 1667 ± 4 to 1523 ± 12 Ma (Thrane et al., 2016).

4.5 Metamorphic evolution

Rocks of the Nagssugtoqidian orogen mainly record amphibolite- to granulite-facies conditions (Escher et al., 1989; Escher and Hall, 1989; Nutman and Friend, 1989; Mengel et al., 1990; Messiga et al., 1990; Nutman et al., 2008; Kolb, 2014; Nicoli et al., 2018; Müller et al., 2018). Detailed petrological, structural and geochronological studies reveal a complex tectono-thermal history, with metamorphism in the Archean and Paleoproterozoic (Kalsbeek et al., 1993; Nutman et al., 2008; Kolb, 2014; Müller et al., 2018). Archean TTG gneiss in the Kuummiut Terrane records a Neoproterozoic high-grade event at 2723 ± 49 to 2635 ± 20 Ma (zircon U-Pb; Kalsbeek et al., 1993; Nutman et al., 2008; Kokfelt et al., 2016a), which might be related to granulite-facies metamorphism in the Schweizerland Terrane (~ 2720 Ma; Nutman et al., 2008). Different degrees of Paleoproterozoic reworking are observed in the Kuummiut and Isertoq terranes (Kolb, 2014). Low-pressure amphibolite-facies assemblages are characteristic of the Isertoq Terrane away from the Ammassalik Intrusive Complex (Escher et al., 1989), with Archean gneiss only locally preserving partly retrogressed granulite-facies assemblages and textures (Escher et al., 1989; Bridgwater et al., 1990; Kolb, 2014). Rocks in the immediate vicinity of the complex record medium-pressure granulite-facies conditions (Nutman and Friend, 1989). In contrast, the Kuummiut Terrane is characterized by variably retrogressed eclogite-facies mineral assemblages that are preserved within the cores of larger metabasic dykes and basic supracrustal rocks (Wright et al., 1973; Chadwick et al., 1989; Nutman and Friend, 1989; Messiga et al., 1990; Nutman et al., 2008; Müller et al., 2018; Dziggel and Müller, 2018). Smaller dykes or

margins of larger dykes are characterized by amphibolite-facies mineral assemblages (Bridgwater et al., 1976; Chadwick et al., 1989; Nutman and Friend, 1989; Messiga et al., 1990; Dziggel and Müller, 2018). Recently, evidence for four different metamorphic stages on a clockwise PT-path has been obtained from combined mineral textural analysis, conventional geothermobarometry and pseudosection modelling on the variably retrogressed high-pressure rocks (Müller et al., 2018). In retrogressed eclogite, early garnet growth occurred during the prograde metamorphic evolution at conditions of 14-19 kbar and 600-750 °C (stage I, obtained from crosscutting X_{Grs} and X_{Fe} isopleths for Ca-rich garnet cores), followed by eclogite-facies metamorphism at 17-19 kbar and 740-810 °C (stage II, defined by well-equilibrated garnet-pyroxenite). Eclogite-facies conditions are also indicated by the presence of omphacite inclusions in garnet, hornblende and large clinopyroxene grains in a Na-rich retrogressed eclogite (Müller et al., 2018). The exhumation of the Kuummiut Terrane was associated with high-pressure granulite-facies conditions of 13.8-15.4 kbar and 760-880 °C (stage III, determined via isopleths for garnet and omphacite) and subsequent re-equilibration at high-pressure amphibolite-facies conditions of 8.8-10.9 kbar and 660-840 °C (stage IV, obtained from well-equilibrated garnet-amphibolite and garnet-kyanite schist). Other studies report further regional retrogression at medium-pressure amphibolite-facies conditions of ~5-7 kbar and 600-700 °C (Nutman et al., 2008; Baden, 2013; Kolb, 2014; Nicoli et al., 2018). An U-Pb zircon date of 1867 ± 28 Ma from a retrogressed eclogite has been interpreted as the timing of high-pressure metamorphism (Nutman et al., 2008), based on the trace element chemistry and types of inclusions in zircon (garnet and clinopyroxene, no plagioclase). Similar U-Pb dates between 1892 ± 5 and 1870 ± 50 Ma have been collected from supposedly metamorphic zircon in supracrustal rocks across the Kuummiut Terrane (Kalsbeek et al., 1993; Nutman et al., 2008; Thrane et al., 2016; Nicoli et al., 2018). In the Isertoq Terrane, granulite-facies metamorphism (~7 kbar, ≥ 700 °C; Andersen et al., 1989; Nutman and Friend, 1989; Nutman et al., 2008) occurred between 1900 ± 30 and 1874 ± 9 Ma (zircon U-Pb; Kalsbeek et al., 1993; Nutman et al., 2008; Lebrun et al., 2018). The timing of collision between the North Atlantic and Rae cratons correlates with regional, medium- to lower-pressure amphibolite-facies metamorphism in the Kuummiut Terrane and is constrained at either 1870 to 1820 Ma (Nutman et al., 2008) or 1890 to 1840 Ma (Nicoli et al., 2018). A Sm-Nd garnet-clinopyroxene-whole rock isochron of 1817 ± 22 Ma for a basic dyke from the Kuummiut Terrane (Kalsbeek et al., 1993) has been interpreted to reflect isotopic equilibration during decompression and cooling (Nutman et al., 2008). Intrusion of post-tectonic granite, diorite, gabbro and pegmatite (1680 ± 9 to 1523 ± 12 Ma; Kalsbeek et al., 1993; Thrane et al., 2016) locally resulted in contact metamorphism in metasedimentary rocks (2.5 kbar and 580-650 °C; Nutman and Friend, 1989).

4.6 Petrology

Nine samples, including five metabasic dykes (samples 525224, 566216, 566218, 566240 and 566249), three basic supracrustal rocks (samples 524713, 524716 and 566277) and one garnet-kyanite schist (sample 566267) were chosen for U-Pb dating. The samples were collected from several localities in the Kuummiut Terrane during field trips to South-East Greenland in 2010 and 2014 (Fig. 4.1a; Dziggel and Müller, 2018). A brief petrographic description for each sample is given below, with mineral assemblages reported in Table 4.1. Overall, the garnet-kyanite schist shows a well-equilibrated amphibolite-facies mineral assemblage, whereas the basic dykes and supracrustal rocks contain complex mineral reaction textures typical of retrogressed eclogite (Müller et al., 2018).

Table 4.1

Mineral assemblages in the samples used for U-Pb dating (mineral abbreviations after Whitney and Evans, 2010).

Sample	HP-assemblage or relict HP-phases	Relict HT-phases	Retrograde assemblage	Dated grains
Retrogressed eclogite				
524713	Rt	Ilm	Di + Pl + Hbl + Grt _b + Ttn + Py	Ttn, Rt
524716	Grt _a + Qz	Ilm	Di + Pl + Hbl + Grt _b + Ttn + Bt	Ttn
525224	Rt + Qz	Opx + Ilm	Di + Pl + Hbl + Grt _b + Bt	Zrn, Rt
566216	Grt _a + Rt + Qz	Aug + Ilm	Di + Pl + Hbl + Grt _b + Ttn	Zrn, Ttn, Rt
566218	Grt _a + Rt + Qz	Aug + Opx + Ilm	Di + Pl + Hbl + Grt _b + Ttn	Zrn, Rt
566240	Rt	Ilm	Di + Pl + Hbl + Grt _b	Zrn, Rt
566277	Grt _a + Rt + Qz	Ilm	Di + Pl + Hbl + Grt _b + Ttn	Ttn, Rt
566249	Grt _a + Qz	Ilm	Di + Pl + Hbl + Grt _b + Bt + Ttn	Zrn
Garnet-kyanite schist				
566267			Grt + Bt + Ms + Ky + Qz + Pl + Rt + Ilm	Zrn, Mz

4.6.1 Retrogressed eclogite

Retrogressed eclogite consists of two domains, one dominated by clinopyroxene-plagioclase symplectites (clinopyroxene domain, Fig. 4.2a, b), and one dominated by garnet (garnet domain), in which garnet is surrounded by plagioclase ± hornblende ± pyroxene coronas (Fig. 4.2c-d; Müller et al., 2018). The clinopyroxene-plagioclase symplectite is interpreted to have formed after omphacite, with omphacite exsolving plagioclase during decompression (Mysen, 1972; Boland and Van Roermund, 1983; O'Brien, 1989; Joanny et al., 1991). In the samples analyzed, most clinopyroxene is diopside, except for a Na-rich sample, in which omphacite is preserved as inclusions in garnet, hornblende and the core of large clinopyroxene-grains (X_{Jd25-38}; Müller et al., 2018).

Sample 524713 is a relatively coarse-grained retrogressed eclogite. It is dominated by hornblende, garnet, plagioclase, diopside and ilmenite with minor amounts of rutile, titanite and

pyrite. Hornblende and plagioclase form symplectitic intergrowths with diopside and coronas around garnet. Several garnet grains are pseudomorphed by a symplectitic assemblage of hornblende, plagioclase and varying amounts of Fe- or Ti-phases (Fig. 4.2e; ilmenite, titanite, rutile, pyrite). Rutile, ilmenite and titanite often form intergrowths with each other, with titanite locally forming coronas around the former two (Fig. 4.2f). The Ti-phases occur as small grains in the clinopyroxene and garnet domains and as large grains surrounded by hornblende.

The medium- to coarse-grained retrogressed eclogite sample 524716 shows a predominance of hornblende, garnet and plagioclase. Diopside is minor and forms globular intergrowths with plagioclase and hornblende. Plagioclase and hornblende are also present in coronas around garnet together with biotite. Large garnet ($\geq 1000 \mu\text{m}$) preserves prograde Ca-rich cores (Grt_a ; $\text{Alm}_{50}\text{GrS}_{31}\text{Prp}_{15}\text{Sps}_5$), while garnet rims and smaller grains are Mg-rich (Grt_b ; $\text{Alm}_{43}\text{GrS}_{24}\text{Prp}_{30}\text{Sps}_2$; Müller et al., 2018). Ilmenite, titanite and quartz represent minor components and occur in both domains.

Sample 525224 is a foliated, medium-grained retrogressed eclogite. The foliation is defined by the preferred grain orientation of garnet and hornblende. Garnet, hornblende and plagioclase dominate the sample, followed by diopside and biotite. Quartz, ilmenite and orthopyroxene are minor components, whereas zircon and rutile are accessory components occurring in intergrowths with hornblende, biotite and plagioclase in the garnet domain (Fig. 4.2g). Diopside forms globular symplectites with quartz, plagioclase and hornblende (Fig. 4.2a). Garnet is rimmed by almost all major and minor minerals.

Hornblende, plagioclase, diopside and garnet make up the dominant assemblage of the fine- to medium-grained retrogressed eclogite sample 566216. Rutile, quartz, ilmenite, titanite and augite are minor components. Zircon is an accessory phase. Both, zircon and the Ti-phases mostly occur in the garnet domain. Plagioclase and hornblende form worm-like symplectites with and after diopside (Fig. 4.2b) and coronas around garnet. Diopside + hornblende coronas separate garnet and quartz (Fig. 4.2d). Augite occurs both in coronas around garnet and in the clinopyroxene domain. Rutile and ilmenite are surrounded by titanite. Large garnet grains preserve Ca-rich cores, while the majority of garnet is Mg-rich (Müller et al., 2018).

Sample 566218 is a fine-grained retrogressed eclogite that is primarily composed of plagioclase, hornblende, garnet and diopside. Minor phases are ilmenite, rutile, titanite, quartz, augite and orthopyroxene. Zircon is an accessory phase and occurs in both domains. Hornblende and plagioclase form worm-like intergrowths with and after diopside and coronas around garnet. Hornblende also forms coronas between garnet and quartz, together with diopside and orthopyroxene. Augite occurs in the symplectitic intergrowths in both domains. Titanite replaces ilmenite and rutile. The Ti-phases occur in both domains. Garnet preserves Ca-rich cores and Mg-rich rims (Müller et al., 2018).

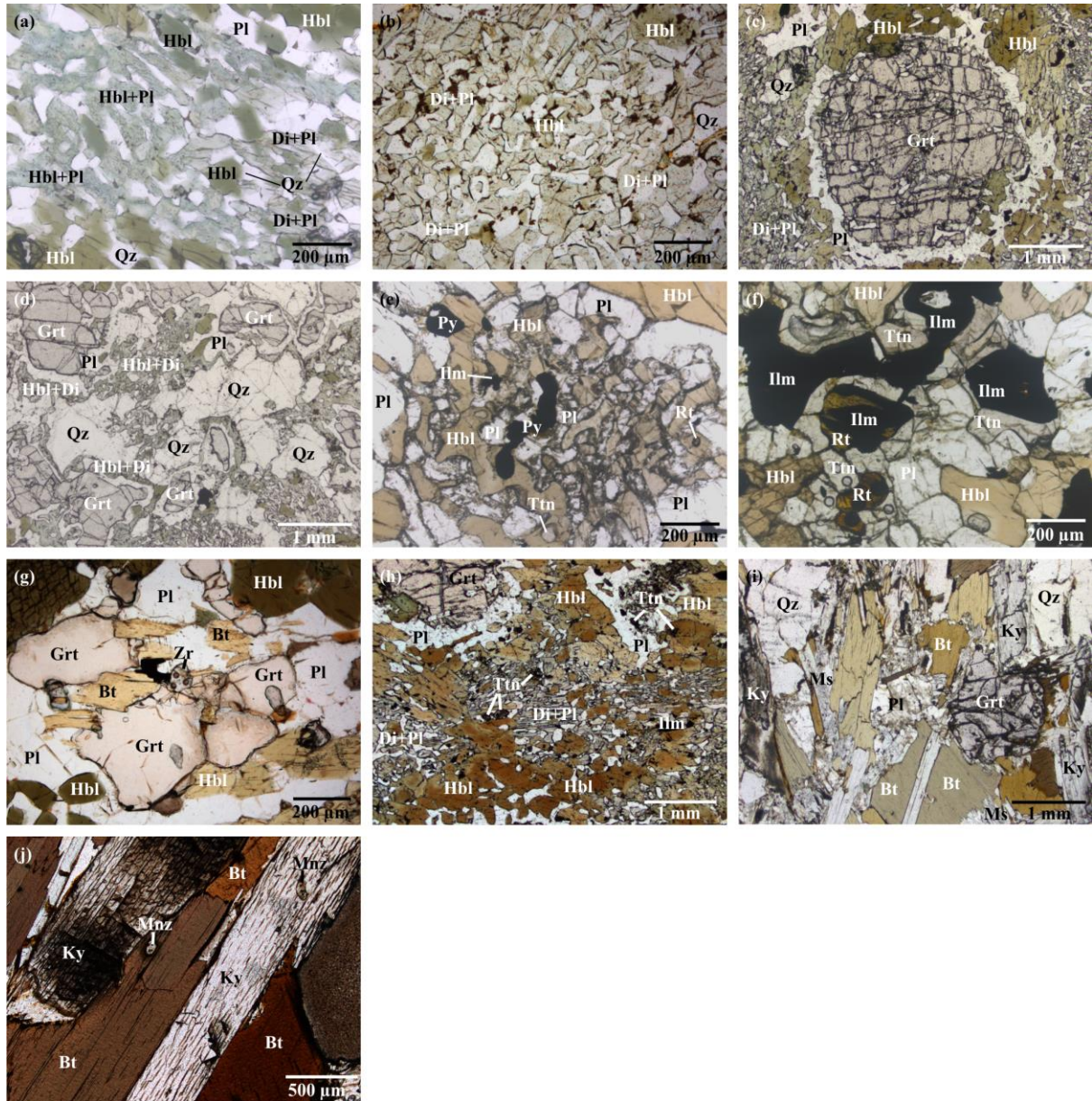


Fig. 4.2. Photomicrographs of the metabasic rock samples (a-h) and the garnet-kyanite schist (i, j), showing mineral assemblages and microstructures (a-f, i), and the association of accessory minerals with the dominant high-pressure amphibolite-facies mineral assemblage (g, h, j). (a) Globular intergrowths of diopside, plagioclase, hornblende and quartz (sample 525224). (b) Worm-like intergrowths of diopside, plagioclase and hornblende (sample 566216). (c) Garnet grain surrounded by plagioclase-hornblende coronas (sample 566277). (d) Thin, string-like diopside and hornblende coronas between garnet and quartz (sample 566216). (e) Hornblende, plagioclase and titanite in pseudomorphs after garnet (sample 524713). (f) Titanite coronas around rutile and ilmenite (sample 524713). (g) Zircon between the Mg-rich garnet rim and surrounding Pl-corona (sample 525224). Holes represent laser spots from U-Pb dating. (h) Titanite intergrown with the hornblende-plagioclase symplectites after diopside and garnet (sample 566277). (i) Mineral assemblage in the garnet-kyanite schist (sample 566267). (j) Monazite intergrown with biotite and kyanite (sample 566267).

The retrogressed eclogite sample 566240 is relatively fine-grained and shows a weak foliation defined by the preferred grain orientation of garnet and hornblende. Diopside and plagioclase form worm-like intergrowths with hornblende. Hornblende also forms coronas around garnet, whereas plagioclase coronas are less common. Rutile and ilmenite are minor components and mostly occur in the garnet domain or are surrounded by hornblende. Zircon occurs as an accessory phase in the garnet domain.

Sample 566249 is a fine-grained retrogressed eclogite, dominated by garnet, hornblende, plagioclase and diopside. Ilmenite, biotite, quartz and titanite occur as minor components. Zircon is an accessory phase. Plagioclase and hornblende form symplectitic intergrowths with and after diopside and coronas around garnet.

Sample 566277 is a retrogressed eclogite dominated by coarse-grained garnet and fine-grained diopside, hornblende and plagioclase. The latter three occur in worm-like intergrowths in the clinopyroxene domain, whereas hornblende and plagioclase also occur in coronas around garnet (Fig. 4.2c). Garnet in this sample, however, is partially pseudomorphed by plagioclase and hornblende. Minor components are rutile, ilmenite, titanite and quartz, occurring in both domains. Titanite forms coronas around rutile and ilmenite or is associated with plagioclase, hornblende and diopside (Fig. 4.2h).

4.6.2 Garnet-kyanite schist

The garnet-kyanite schist (sample 566267) is relatively coarse-grained and dominated by biotite, kyanite, quartz and garnet, with minor muscovite and plagioclase (Fig. 4.2i). Zircon, monazite, rutile, ilmenite and graphite are accessory phases. Monazite is usually associated with kyanite and biotite (Fig. 4.2j). The rock contains a well-developed foliation that is defined by the preferred grain orientation of biotite and kyanite.

4.7 Analytical Methods

4.7.1 U-Pb dating

U-Pb dating was carried out during two sessions, in-situ on polished thin sections (zircon, monazite, titanite and rutile) at Goethe-University Frankfurt (GUF) and in zircon grain mounts at the Department of Petrology and Economic Geology, Geological Survey of Denmark and Greenland (GEUS, Copenhagen).

Seven retrogressed eclogites (samples 524713, 524716, 525224, 566216, 566218, 566240 and 566277) and the garnet-kyanite schist (sample 566267) were used for in-situ dating, whereas zircon of very fine grain size (30-50 μm) from three retrogressed eclogites (samples 566218, 566240 and 566249) was separated and handpicked from <80 μm fractions. Standard magnetic and density

procedures were applied for separation. The small zircon grains were mounted in one-inch epoxy mounts, which subsequently were polished to expose a central cross-section for most of the grains.

Analytical procedures for LA-ICP-MS analysis are quite similar in the two laboratories (see Table 4.2 for detailed operating conditions and instrument settings) and follow the methods outlined in Gerdes and Zeh (2006, 2009), Frei and Gerdes (2009) and Kolb et al. (2012). Positioning of laser spots was aided by backscatter electron (BSE) imaging, carried out using a JEOL JXA-8900 R Electron Microprobe at the Institute of Applied Mineralogy and Economic Geology, RWTH Aachen University and a Philips XL40 SEM at GEUS, respectively. After image acquisition, the samples were thoroughly cleaned to remove carbon coating and other contamination, before being mounted in a two-volume ablation cell (S-155, Laurin Technic at GUF; TV2 10x10x2 cm, New Wave Research at GEUS). U-(Th)-Pb isotopes were analyzed using a Thermo-Fisher Scientific Element II, single-collector, doubly-focusing, magnetic sector field ICP-MS, coupled to a RESolution (Resonetics) 193 nm ArF Excimer laser system (CompexPro 102, Coherent) at GUF and a NWR 213 (New Wave Research) UV-laser system, equipped with a frequency quintupled Nd:YAG (neodymium-doped yttrium aluminum garnet) solid state laser at GEUS. The data were mostly acquired in low resolution, time-resolved, peak-jumping pulse counting mode, except for monazite, where ^{208}Pb , ^{232}Th and ^{238}U were measured in analog mode. Spot size varied depending on the microstructure and the U and Th content of the different minerals. Penetration depths are at 15-20 μm . Helium was used as carrier gas to flush the sample cell of the laser ablation system, mixed in the ablation funnel with argon gas and directed via Nylon 6 (at GUF) and Tygon® (at GEUS) tubing towards the mass spectrometer. At GUF, small amounts of nitrogen gas (6 ml/min) were added to the sample argon flow in order to suppress the formation of polyatomic species. Prior to analysis, the signal was tuned for maximum sensitivity via line scans across reference zircon GJ-1 (Jackson et al., 2004), while keeping oxide formation rate well below 0.5% ($^{254}\text{UO}/^{238}\text{U}$).

The raw isotope data were corrected offline in MS Excel© (Gerdes and Zeh, 2006; Frei and Gerdes, 2009) for significant contributions of common Pb, based on the interference- and background-corrected ^{204}Pb signal and a model Pb composition (Stacey and Kramers, 1975). At GUF, the ^{204}Pb content for each ratio was determined after the method described in Millonig et al. (2013). Within-run Pb/U-fractionation was corrected for each analysis using a linear regression through all ratios. Reference zircon GJ-1 was used for instrumental mass bias and drift correction in a standard bracketing mode. Apart from zircon, this correction with the GJ-1 standard was also applied for U-Pb isotope analysis of monazite, titanite and rutile, which are characterized by a different mineral matrix.

Accuracy and reproducibility of the method was checked by repeated analysis (n= 6-16) of reference zircon 91500 (Wiedenbeck et al., 1995), Plešovice zircon (Sláma et al., 2008), Namaqualand (GUF ID-TIMS age of 1027 ± 1 Ma; Wolfgang Dörr, personal communication, 2017) and Manangotry monazite (Horstwood et al., 2003), Bear Lake (Alleinikoff et al., 2007) and Namaqualand titanite (GUF ID-TIMS age of 1023 ± 3 Ma; Wolfgang Dörr, personal communication, 2017) and Tábor

Table 4.2

Operating conditions and instrument settings for BSE-imaging and U-Pb LA-ICP-MS analysis.

	GUF	GEUS
<u>BSE-imaging</u>		
Instrument	JEOL JXA-8900 R Electron Microprobe, Institute of Applied Mineralogy and Economic Geology, RWTH Aachen University	Philips XL40 SEM, GEUS
Acceleration voltage	15-20 kV	17 kV
Probe current	30 nA	~70 nA
Spot Size	1 µm	6.5 µm
<u>LA-ICP-MS</u>		
Instrument	Thermo-Fisher Scientific Element II	Thermo-Fisher Scientific Element II
Type Of MS	Magnetic sector field	Magnetic sector field
Resolution mode	Low	Low
Analysis mode	Time-resolved analysis	Time-resolved analysis
Scan Mode	E-Scan	E-Scan
Scanned isotopes	^{202}Hg , $^{204}(\text{Pb} + \text{Hg})$, ^{206}Pb , ^{207}Pb , ^{208}Pb , ^{232}Th , ^{238}U ; ^{235}U calculated	^{202}Hg , $^{204}(\text{Pb} + \text{Hg})$, ^{206}Pb , ^{207}Pb , ^{208}Pb , ^{232}Th , ^{238}U ; ^{235}U calculated
Detector mode	In Zrn, Ttn and Rt: pulse-counting; In Mz: analog for ^{208}Pb , ^{232}Th and ^{238}U , pulse-counting for all other isotopes	pulse-counting
<u>Laser settings</u>		
Type of laser	Resonetics, 193 nm, ArF Excimer (CompexPro 102, Coherent)	New Wave Research, 213 nm, Nd:YAG
Repetition rate	5.5 and 8 Hz	10 Hz
Laser fluence	<2 J/cm ²	10 J/cm ²
Cell type	S-155 two-volume cell (Laurin Technic)	TV2 two-volume cell (New Wave Research/ ESI)
Carrier gas	He (0.4 L/min)	He (0.85 L/min)
Sample Gas	Ar (0.75 L/min) + N ₂ (6 ml/min)	Ar (0.925 L/min)
Spot size	13-67 µm	25 µm
Penetration depth	~ 15-20 µm	~ 15-20 µm
Background collection	21 s	30 s
Specimen ablation	20 s	30 s
Washout delay	24 s (with 2 s pre-ablation)	35 s
<u>Data acquisition</u>		
Type of Sample	Polished thin section	Grain mount
Minerals analyzed	Zrn, Mz, Ttn, Rt	Zrn
Number of analyses	470	85
<u>Standardization</u>		
Primary Standard	GJ-1	GJ-1
Reference Standards	91500 zircon, Namaqualand and Manangotry monazite, Bear Lake and Namaqualand titanite, Tábor and Sand River Quartzite rutile	91500 and Plešovice zircon

(Janoušek and Gerdes, 2003) and Sand River Quartzite rutile (Schmitt and Zack, 2012). All standards were analyzed under the same conditions as the samples and reproduce recommended $^{207}\text{Pb}/^{206}\text{Pb}$ (for 91500 zircon) and $^{206}\text{Pb}/^{238}\text{U}$ ages (all others) of the standards at limits $\leq 2.6\%$ (Table 4.3).

Table 4.3

Mean $^{207}\text{Pb}/^{206}\text{Pb}$ (91500 zircon) and $^{206}\text{Pb}/^{238}\text{U}$ (all others) ages (in Ma) of the reference standards with 2 standard deviation uncertainties.

Standard	GUF Day 1	GUF Day 2	GEUS	Reference
91500 Zircon	1057.5 $\pm$ 26.6 (n=10)	1070.9 $\pm$ 18.1 (n=6)	1063 $\pm$ 28.1 (n=13)	Wiedenbeck et al., 1995; 1065.4 $\pm$ 0.3
Plešovice Zircon			328.33 $\pm$ 17.70 (n=15)	Sláma et al., 2008; 337.13 $\pm$ 0.37
Namaqualand Monazite		1029 $\pm$ 12 (n=9)		GUF ID-TIMS; 1027 $\pm$ 1
Manangotry Monazite		554.5 $\pm$ 20.9 (n=5)		Horstwood et al., 2003; 552.9 $\pm$ 10.1
Bear Lake Titanite	1053.4 $\pm$ 15.3 (n=16)	1054.2 $\pm$ 16.2 (n=6)		Alleinikoff et al., 2007; 1047.1 $\pm$ 0.4
Namaqualand Titanite	1018 $\pm$ 19.8 (n=9)			GUF ID-TIMS; 1023 $\pm$ 3
Tábor Rutile	337.6 $\pm$ 4.7 (n=11)	336.4 $\pm$ 4.4 (n=6)		Janoušek and Gerdes, 2003; 336.8 $\pm$ 0.8
Sand River Quartzite Rutile	2016 $\pm$ 51 (n=10)			Schmitt and Zack, 2012; 1998 $\pm$ 6

To account for a systematic offset of monazite, titanite and rutile analyses after correction with GJ-1 as primary standard, the U-Pb isotope ratios were corrected by offset factors of 0.95, 1.055-1.075, and 1.06, respectively. These factors are based on the deviations for Namaqualand monazite, Bear Lake titanite and Tábor rutile, respectively, used as secondary standards.

Zircon 91500 shows that results between the laboratories at GUF ($^{207}\text{Pb}/^{206}\text{Pb}$ age = 1062.5 $\pm$ 26.7 Ma, 2 std. dev., n = 16) and GEUS ($^{207}\text{Pb}/^{206}\text{Pb}$ age = 1063.0 $\pm$ 28.1 Ma, 2 std. dev., n = 13) are identical within limits of error. U-Pb concordia-diagrams were constructed using the MS Excel© Add-In Isoplot (Ludwig, 1998) in Version 4.15. Uncertainties reported in the text and figures are at 2 σ -level. Detailed LA-SF-ICP-MS U-Pb data can be found in the Appendix (Tables A6-A12).

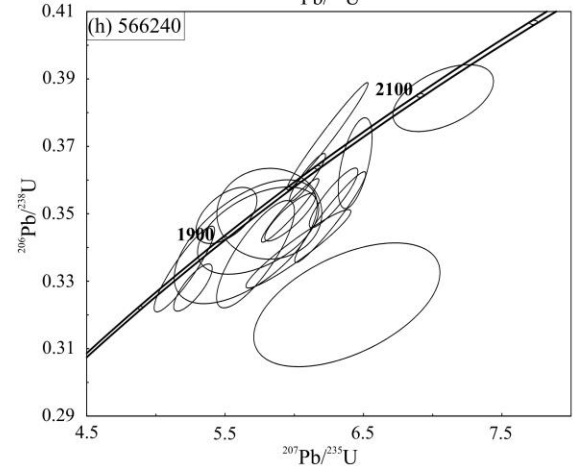
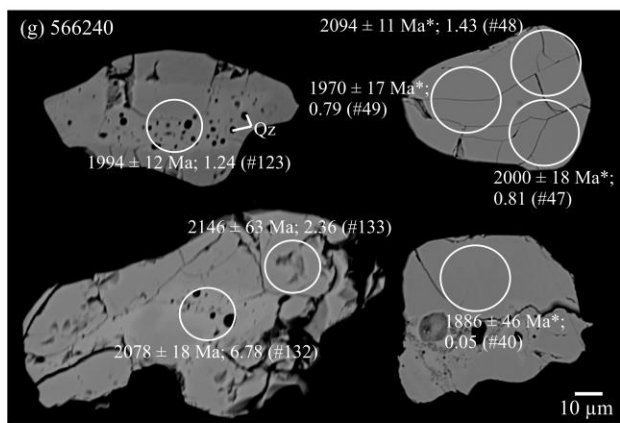
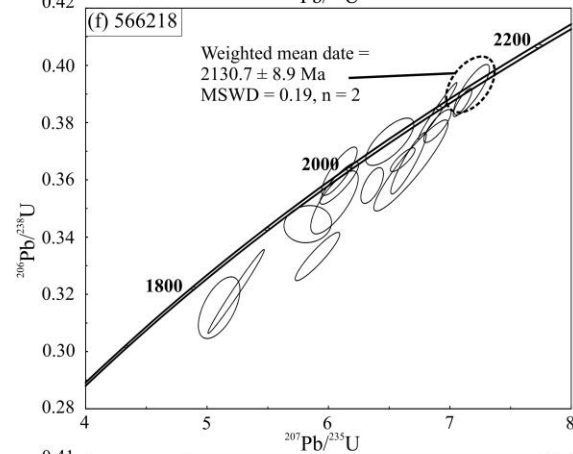
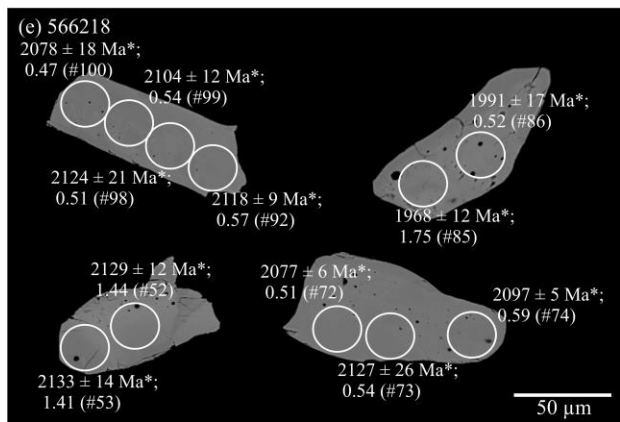
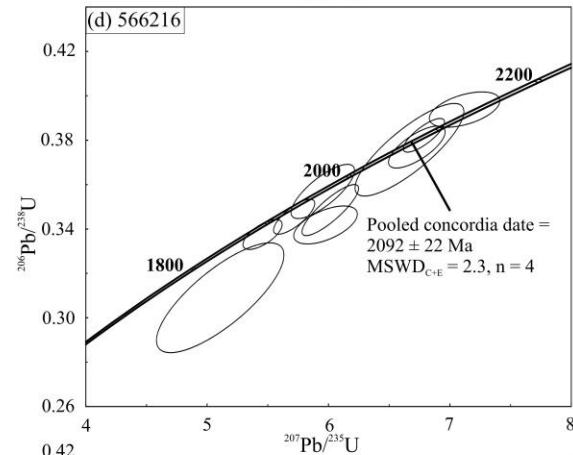
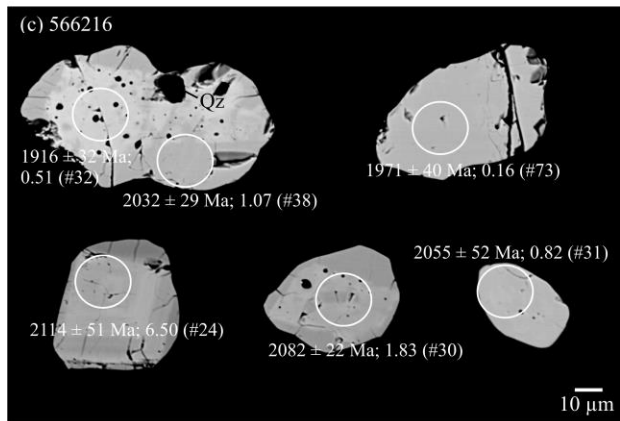
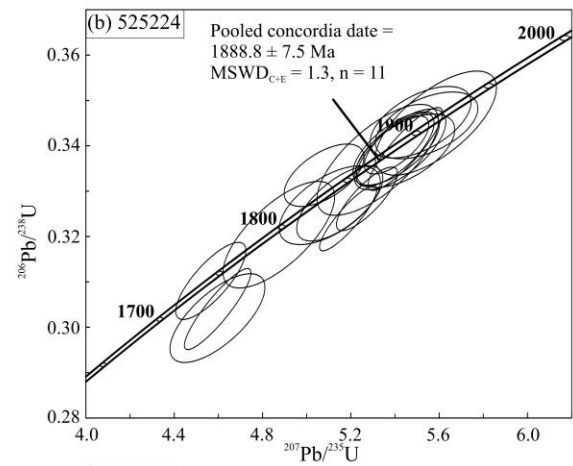
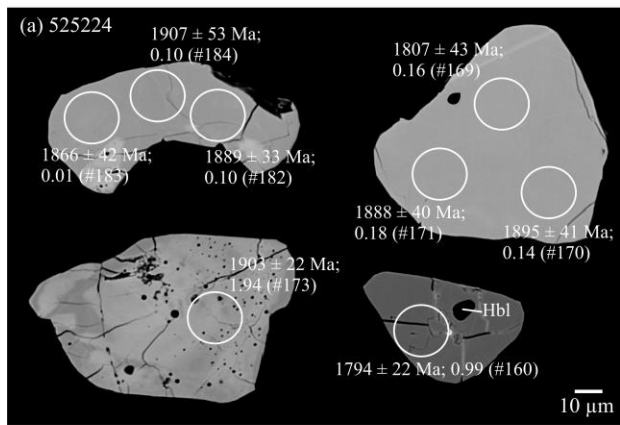
4.7.2 Trace element analysis of zircon

In order to provide information on the conditions of zircon growth in the samples analyzed, the trace element concentrations in zircon were analyzed in two retrogressed eclogites (525224, 566240) and the garnet-kyanite schist (566267). The trace element analyses were performed at GEUS, using the same ablation and ICP-MS system utilized for U-Pb dating. Sites for the additional laser ablation analyses were carefully selected using BSE-images and ablation pits produced during U-Pb dating. Analyzed elements include the REE's (Y, La, Ce, Pr, Nd, Sm, Eu, Gd, Tb, Dy, Ho, Er, Tm, Yb, Lu), Hf, Ta and Nb. The laser was run with a frequency of 5 Hz and an energy density of 10 J/cm³. Measurements lasted for about 105 s, with 30 s measured for the background and 40 s on the peak. Washout delays between measurements were 35 s. The spot size was set to 10 µm, due to the small size of the zircon grains. This spot size yielded detection limits in the order of 0.1 ppm. Zircon GJ-1 (Jackson et al., 2004) and NIST-612 glass (GeoRem online database and Jochum et al., 2011) were used as external standards for instrument tuning (e.g. mass offset, oxide production) and calibration of the measurements. Si²⁹ was used as the internal standard element. Quantitative trace element compositions were determined using the analysis software IOLITE (Paton et al., 2011). The trace element data are reported in the Appendix (Table A13).

4.8 U-Pb dating and trace element geochemistry of zircon

4.8.1 U-Pb dating

Zircon in sample 525224 occurs as inclusions in hornblende, plagioclase and ilmenite near garnet (Fig. 4.2g) or as inclusions in the garnet rim. Grains vary from 10x25 to 60x100 µm in size and have subhedral to anhedral, variably resorbed shapes (Fig. 4.3a). About half of the zircon grains appear nearly homogenous in BSE-image. The remaining, mainly smaller, grains show BSE-bright domains that are surrounded by BSE-dark domains (Fig. 4.3a), with the interface usually being irregular and corroded. Inclusions in zircon are usually of very small grain size (<5 µm) and concentrated in BSE-bright domains, while BSE-dark domains and nearly unzoned grains are typically free of inclusions. Only some inclusions in zircon could be identified as quartz and hornblende, while the majority is too small for analysis. Zircon is often intensely fractured, with zoned grains showing thinning of the fractures towards the BSE-bright domains. Compared to BSE-dark domains, these brighter domains are relatively enriched in U and Pb and usually show higher Th/U ratios, which overall vary from 0.01 to 2.36 (Table A6). The 19 spots analyzed yield a range in ²⁰⁷Pb/²³⁵U ratios between 4.57 and 5.59 (Table A6), with the majority of the data showing ²⁰⁷Pb/²³⁵U ratios larger than 5.27 and defining a pooled concordia date of 1889 ± 8 Ma (Fig. 4.3b; MSWD = 1.3, n = 11). About half of the zircons with these high isotope ratios have Th/U ratios <0.1, typical of metamorphic zircon. Predominantly magmatic Th/U ratios (>0.1) are observed for zircons with low



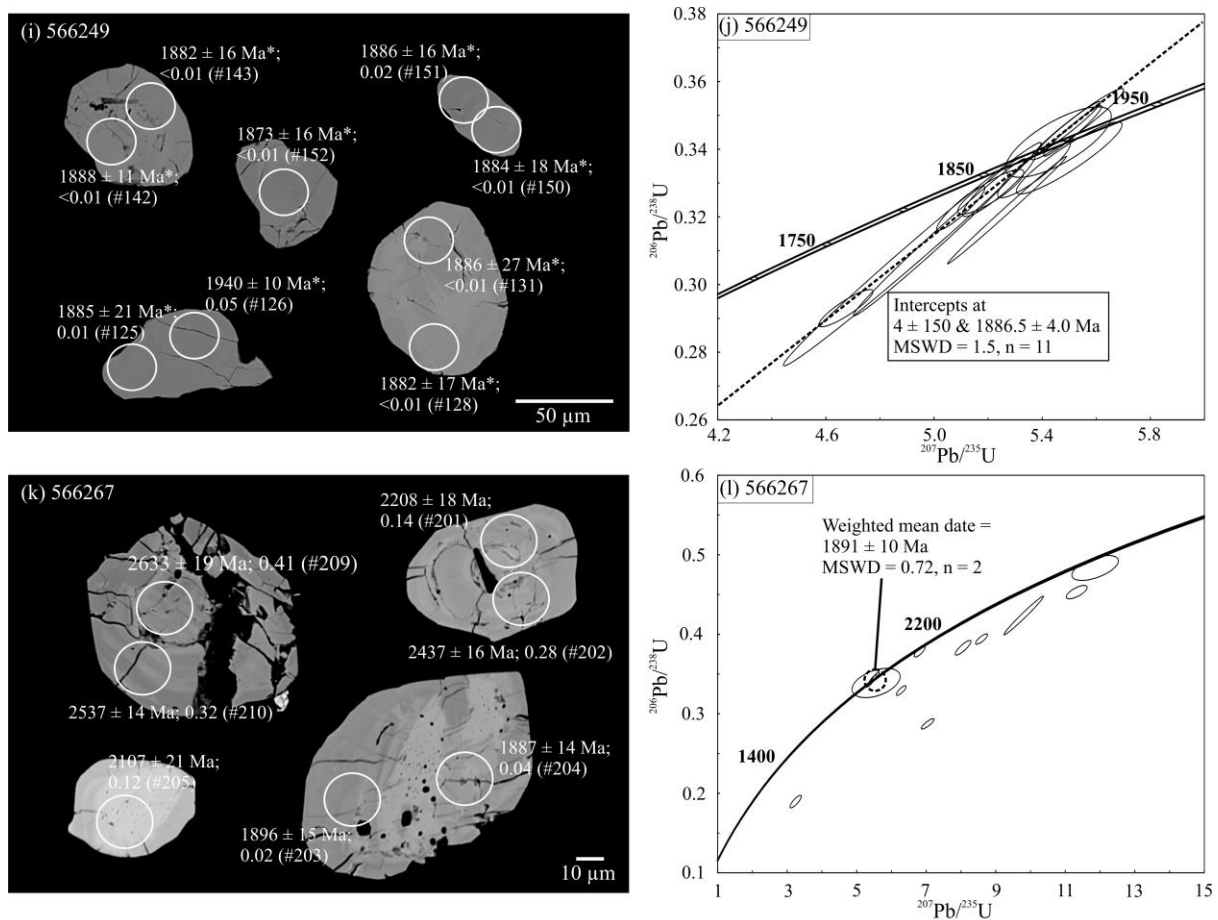


Fig. 4.3. BSE-images of representative zircon grains and U-Pb concordia diagrams from sample (a, b) 525224, (c, d) 566216, (e, f) 566218, (g, h) 566240, (i, j) 566249 and (k, l) 566267. White circles indicate the position of laser spots (spot size; 20 μm at GUF, 25 μm at GEUS). Annotated text indicates the $^{207}\text{Pb}/^{206}\text{Pb}$ date with 2σ -errors, followed by the Th/U ratio and the measurement number. An asterisk indicates a measurement at GEUS.

$^{207}\text{Pb}/^{235}\text{U}$ ratios, which yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 1807 ± 43 and $1746 \pm 32 \text{ Ma}$. BSE-bright, inclusion-rich domains and/or magmatic Th/U ratios for zircon in this sample can be taken as evidence for incomplete recrystallization of magmatic zircon during metamorphism.

Zircon in sample 566216 occurs as inclusions in the Mg-rich garnet-rim, or in hornblende-plagioclase coronas around garnet. The grains are subhedral to anhedral, resorbed and between 25×40 and $50 \times 90 \mu\text{m}$ in size (Fig. 4.3c). About half of the grains are unzoned, whereas the other half shows inclusion-rich, BSE-bright domains with faint fractures and corroded boundaries, surrounded by fractured and inclusion-poor BSE-dark domains. Most inclusions are too small to be analyzed and only quartz could be identified. Uranium and Pb contents are higher in the BSE-bright domains, whereas Th/U ratios range between 0.48 and 6.50 for all grains (Table A6). The highest Th/U ratio is recorded in a grain with faint oscillatory zoning (Fig. 4.3c). Ten spots analyzed plot close to the concordia with $^{207}\text{Pb}/^{235}\text{U}$ ratios ranging between 5.11 and 7.12 (Fig. 4.3d). The four spots with the highest isotope ratios yield a pooled concordia date of $2092 \pm 22 \text{ Ma}$ ($\text{MSWD} = 0.75$). This date is

obtained from unzoned grains and inclusion-rich or oscillatory zoned domains, and most likely approximates the timing of dyke emplacement into the Archean TTG gneiss. Analyses with lower $^{207}\text{Pb}/^{235}\text{U}$ ratios (<6.10) are slightly discordant, with the most concordant analyses yielding $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 2032 ± 32 and 1916 ± 32 Ma. This may reflect either partial Pb-loss or incomplete recrystallization of magmatic zircon during the metamorphic evolution.

Zircon in sample 566218 is rare and was mostly analyzed in grain mount. Two grains were analyzed in thin section, where they occur as inclusions in plagioclase and hornblende rimming garnet. Zircon grains in thin section and grain mount are subhedral to anhedral, up to 100×50 μm large and show variable degrees of resorption (Fig. 4.3e). Faint BSE-bright and inclusion-rich domains, in most cases surrounded by BSE-dark, fractured domains are common (Fig. 4.3e). Due to the fine and patchy nature of the zoning, measurement spots often span both domains and yield inconclusive data regarding the distribution of U, Th and Pb. The Th/U ratios vary from 0.01 to 1.75, with most being typical of magmatic zircon (Tables A6 and A10). Sixteen variably discordant analyses give a range of $^{207}\text{Pb}/^{235}\text{U}$ ratios between 5.10 and 7.18 (Fig. 4.3f). Two concordant analyses with the highest isotope ratios (>7), determined from a BSE-dark and BSE-bright domain in the same zircon, yield a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ date of 2131 ± 9 Ma (MSWD = 0.19). This date is interpreted to approximate the timing of dyke emplacement. Analyses with lower $^{207}\text{Pb}/^{235}\text{U}$ ratios show variable discordance, the most concordant analyses yielding $^{207}\text{Pb}/^{206}\text{Pb}$ dates of 2097 ± 5 to 1968 ± 12 Ma. A patchy zoning, inclusion-rich domains and large range in Th/U ratios for these analyses indicates that the variable discordance may either be related to partial Pb-loss or incomplete recrystallization of magmatic zircon during metamorphism.

Sample 566240 contains anhedral zircon grains (50×70 to 100×80 μm , Fig. 4.3g). Zircon occurs as inclusions in the garnet rims and in hornblende-plagioclase coronas around garnet. The grains show a high degree of fracturing; BSE-bright and inclusion-rich domains are frequently present. Quartz occurs as inclusions in zircon. BSE-bright domains contain more U and Pb than surrounding BSE-dark domains, which have higher Th/U ratios (Tables A6 and A10). Overall, Th/U ratios range from <0.01 to 6.78. Zircon from this sample yields a range in $^{207}\text{Pb}/^{235}\text{U}$ ratios from 5.21 to 7.08, with a few analyses showing large errors (Fig. 4.3h). The oldest concordant data point, from a BSE-dark domain, yields a $^{207}\text{Pb}/^{235}\text{U}$ ratio of 7.08 and a $^{207}\text{Pb}/^{206}\text{Pb}$ date of 2146 ± 63 Ma. We interpret this date to approximate the timing of dyke emplacement. Analyses with $^{207}\text{Pb}/^{235}\text{U}$ ratios <7 are variably discordant and have been obtained from zircon grains with a large range in Th/U ratios, and locally inclusion-rich domains. The most concordant of these analyses yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates of 2052 ± 46 to 1903 ± 23 Ma. The variable discordance most likely relates to either partial Pb-loss or incomplete recrystallization of magmatic zircon during metamorphism. A $^{207}\text{Pb}/^{206}\text{Pb}$ date of 1886 ± 46 Ma was obtained from a BSE-dark domain with a metamorphic Th/U ratio and represents zircon that fully recrystallized during metamorphism.

Zircon in sample 566249 was analyzed in grain mount, as zircon in thin section is too small to

be analyzed. In this sample, zircon occurs as inclusions in hornblende, plagioclase and ilmenite. Grain shapes vary from sub- to anhedral and grain size varies from 20x20 to 90x70 μm (Fig. 4.3i). Most grains show resorbed margins and faint patchy zoning. Inclusions are rare. The Th/U ratios are <0.62 for all grains, with most grains showing metamorphic Th/U ratios (Table A10). Eleven out of thirteen analyses define a discordia with an upper intercept date of 1887 ± 4 Ma (Fig. 4.3j; MSWD = 1.5). Due to the low Th/U ratios and patchy to unzoned nature of the grains, we interpret this date to reflect the age of zircon recrystallization during metamorphism.

In the garnet-kyanite schist (sample 566267), zircon predominantly occurs as inclusions in biotite but may also be observed in plagioclase and kyanite. Grain shapes range from ovoid to subrounded and angular (Fig. 4.3k), and grain size varies from 45x25 to 75x100 μm . Zircon is unzoned or shows patchy and oscillatory zoning. Mineral inclusions occur in BSE-bright domains, whereas surrounding darker domains are commonly fractured. Most inclusions are too small to be analyzed, except for a graphite inclusion. The Th/U ratios range from 0.02 to 0.91, with oscillatory zoned zircon having higher Th contents (Table A6). The majority of the 12 spots analyzed plot below the concordia, with $^{207}\text{Pb}/^{235}\text{U}$ ratios showing a large range between 3.25 and 11.86 (Fig. 4.3l). High $^{207}\text{Pb}/^{235}\text{U}$ ratios (>6.2) were obtained from oscillatory-zoned grains, from which the most concordant data points yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates at 2634 ± 63 Ma and 2107 ± 21 Ma. We interpret these grains to be detrital, yielding age information of their provenance area, with the apparently youngest grain at 2107 ± 21 Ma giving a maximum age for the deposition of the precursor of the metasedimentary rock. Low isotope ratios are derived from zircon grains with patchy zoning, where two analyses in one grain yield concordant U-Pb dates that are identical within error (Fig. 4.3k; 1896 ± 15 and 1887 ± 14 Ma, $^{207}\text{Pb}/^{206}\text{Pb}$). A weighted mean date of 1891 ± 10 Ma for these two analyses (MSWD = 0.72) is considered to reflect the age of zircon recrystallization during metamorphism, based on the patchy zoning and low Th/U ratio.

4.8.2 Trace element geochemistry

In retrogressed eclogite, the BSE-bright and inclusion-rich zircon domains show a distinct positive Ce anomaly, a negative Eu anomaly and HREE enrichment (Fig. 4.4a, b). These domains yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates around ~ 1903 Ma, with Th/U ratios between 0.72 and 1.94. Similar REE signatures were also obtained from zircon with oscillatory zoning and discordant $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 2437 ± 16 and 2208 ± 18 Ma in the garnet-kyanite schist (Fig. 4.4c). Based on the high Th/U ratios and oscillatory zoning patterns, we interpret these REE signatures as magmatic in origin.

The homogeneous zircon grains or domains in both retrogressed eclogite and garnet-kyanite schist show overall lower REE contents, with a generally weak to absent Eu anomaly and flat HREE patterns (Fig. 4.4a-c). These grains and domains yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates of 1895 ± 41 to 1888 ± 40 Ma (Fig. 4.4a), 1955 ± 108 Ma (Fig. 4.4b) and 1887 ± 14 Ma (Fig. 4.4c), and have low Th/U ratios between 0.02 and 0.34. The REE signatures are indicative of coeval growth of zircon and garnet in the

absence of plagioclase (Rubatto, 2002).

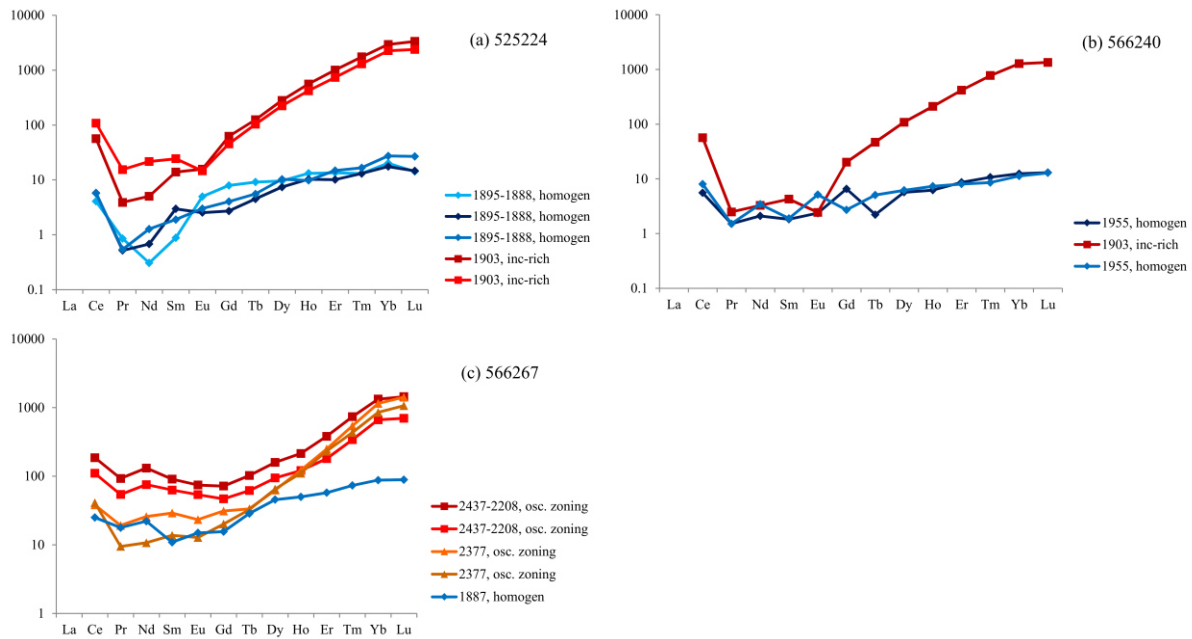


Fig. 4.4. Chondrite-normalized REE patterns of zircon from two retrogressed eclogite samples (*a*-525224, *b*-566240) and the garnet-kyanite schist (*c*-566267). Normalization values after McDonough and Sun (1995).

4.9 U-Pb dating of monazite

Monazite was analyzed in the garnet-kyanite schist (sample 566267). It occurs as inclusions in biotite, kyanite and plagioclase (Fig. 4.2j) and shows variable grain shapes from rounded to subrounded to angular and elongate (Fig. 4.5a). The grain boundaries show different degree of resorption. Maximum grain size is 200x85 μm . Monazite shows faint BSE-bright and inclusion-rich domains, surrounded by BSE-dark domains that are poor in inclusions. Inclusions are rounded to subrounded and usually <5 μm in size. Large inclusions are quartz and muscovite. Monazite appears relatively homogeneous or variably fractured. Uranium and Pb contents show large variations, with Th/U ratios ranging from 4.97 to 40.63 (Table A7), unrelated to the zones with different BSE-intensity. Despite this, U-Pb data are mostly concordant (Fig. 4.5b), with 59 out of 62 analyses defining a pooled concordia date of 1882 ± 3 Ma (MSWD = 0.97). Based on the presence of patchy zoning, this date relates to monazite recrystallization during metamorphism. Two concordant analyses measured in rims yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates of 1854 ± 15 Ma and 1837 ± 16 Ma, and are interpreted to reflect further monazite recrystallization during subsequent metamorphic stages.

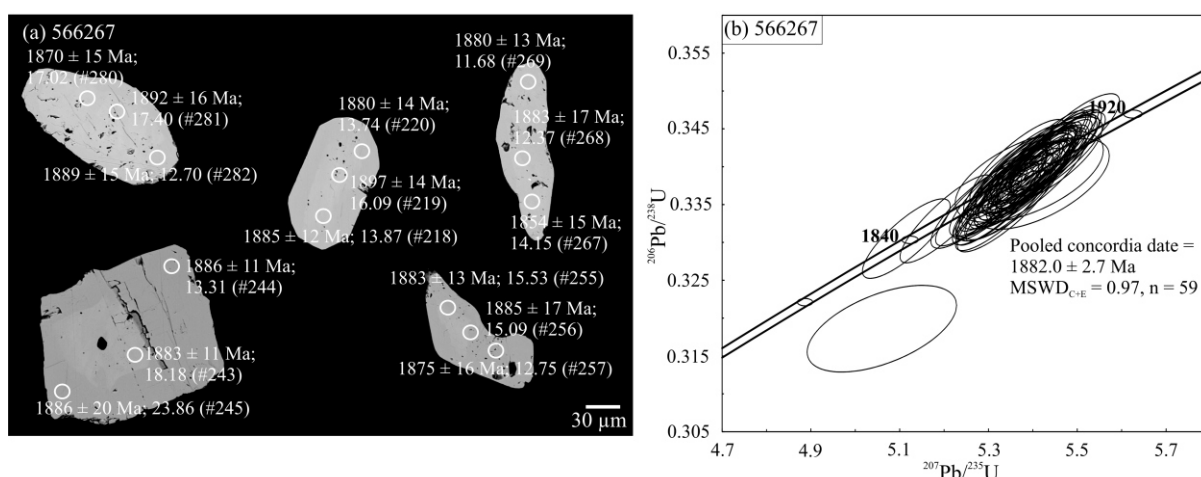


Fig. 4.5. BSE-images of representative monazite grains **(a)** and U-Pb concordia-diagram **(b)** from sample 566267. White circles indicate the position of laser spots (13 µm). Annotated text indicates the $^{207}\text{Pb}/^{206}\text{Pb}$ date with 2σ -errors, followed by the Th/U ratio and the measurement number.

4.10 U-Pb dating of titanite

Titanite in sample 524713 forms coronas around rutile and ilmenite, or occurs intergrown with ilmenite and as inclusions in hornblende and plagioclase (Figs. 4.2f and 4.6a). The grain shapes are irregular and anhedral. Depending on the degree of rutile and/or ilmenite-replacement, corona widths vary from <20 to more than 100 µm. Titanite in intergrowths and as inclusions shows a grain size from 50x120 to 100x300 µm. Coronas and single titanite grains are variably fractured and contain holes (Fig. 4.6a). U-Pb analysis and BSE-imaging indicate the preservation of multiple titanite generations or grains within individual coronas. Nevertheless, titanite has a relatively uniform isotopic composition, with Th/U ratios ranging between 0.14 and 1.20 and $^{207}\text{Pb}/^{235}\text{U}$ ratios between 4.83 and 5.52 (Table A8). Forty-six analyses give a range of concordant U-Pb data, from which 32 analyses with $^{207}\text{Pb}/^{235}\text{U}$ ratios >5.2 define a pooled concordia date of 1884 ± 4 Ma (MSWD = 1.02, Fig. 4.6b). Based on the presence of titanite coronas around the high-pressure phase rutile (Müller et al., 2018), this date is interpreted to reflect titanite growth during retrogression. Analyses with $^{207}\text{Pb}/^{235}\text{U}$ ratios <5.2 give a range of $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 1869 ± 28 and 1782 ± 27 Ma. They most likely reflect ongoing titanite growth during subsequent thermal stages.

In sample 524716, titanite is mostly present as relatively large, anhedral grains (140x90 to 430x450 µm; Fig. 4.6c) in hornblende coronas around garnet. Locally, titanite contains inclusions of rutile or is intergrown with ilmenite. Some titanite grains show different BSE-intensities, in conjunction with different U-Pb dates. The Th/U ratios vary from 0.78 to 2.49 (Table A8), with the highest values being recorded from titanite associated with hornblende. Thirty-six analyses give a range of $^{207}\text{Pb}/^{235}\text{U}$ ratios between 4.74 and 5.56 (Fig. 4.6d). The majority of the analyses with

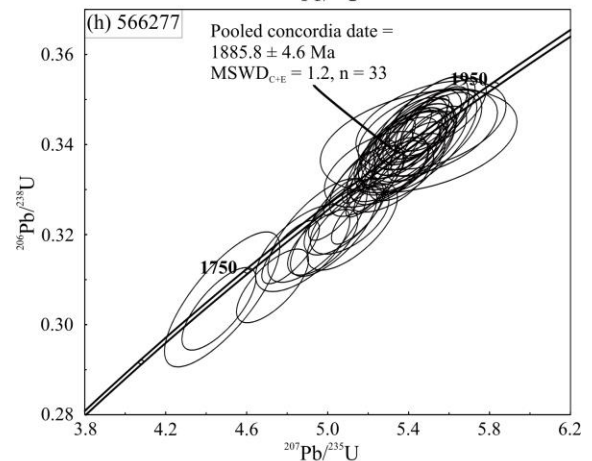
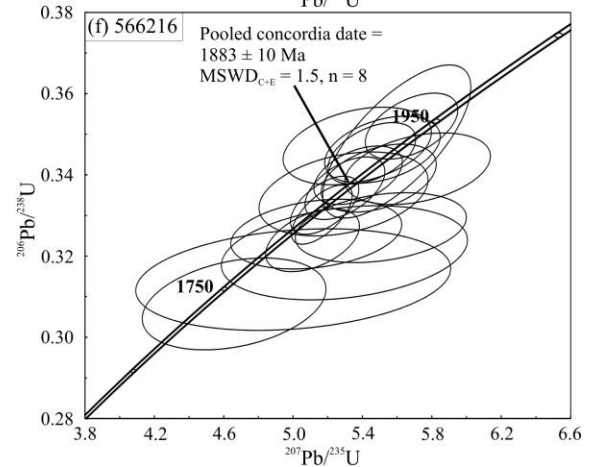
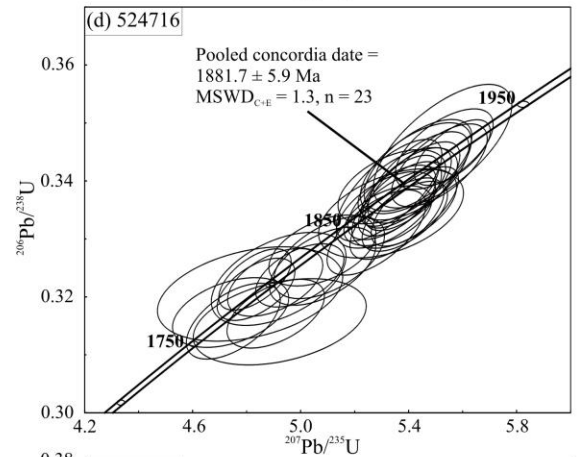
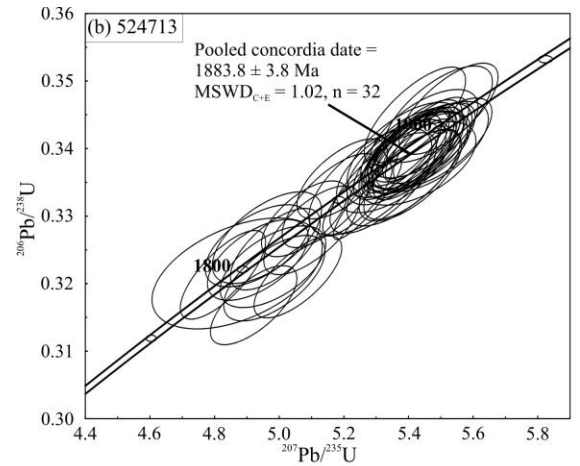
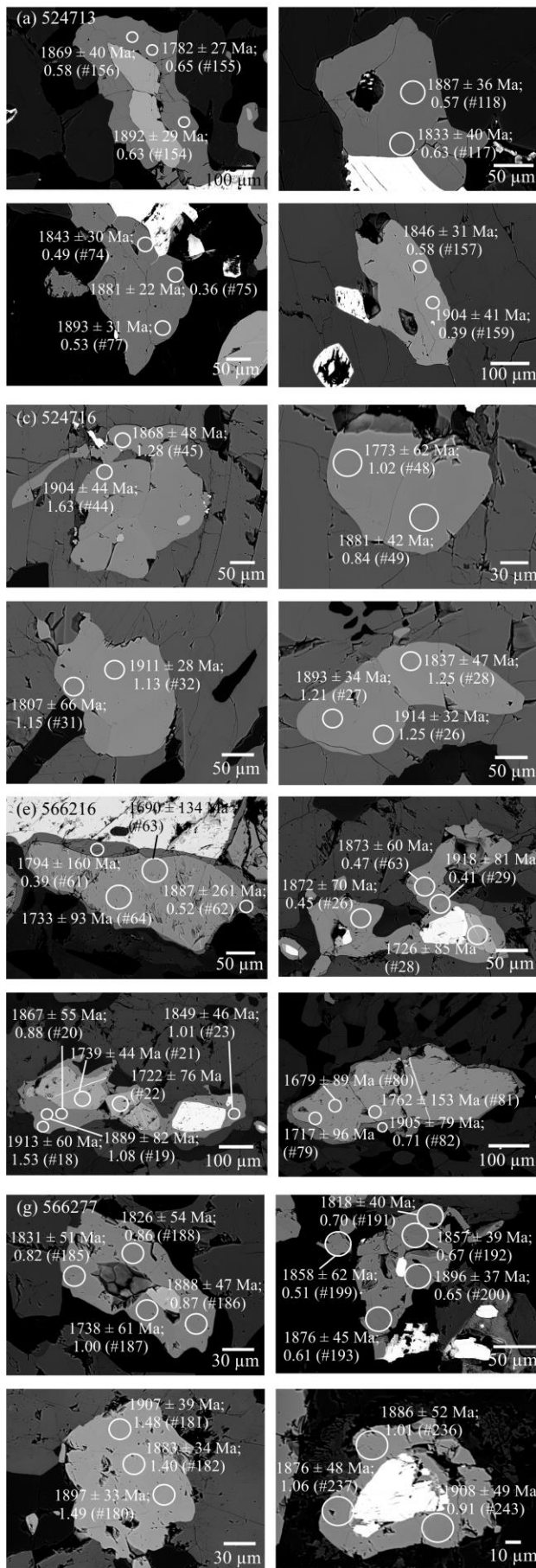


Fig. 4.6: BSE-images of representative titanite ($\pm$ rutile and ilmenite) grains and coronas and U-Pb concordia-diagrams for titanite from sample (a, b) 524713, (c, d) 524716, (e, f) 566216 and (g, h) 566277. White circles indicate the position of laser spots (Ttn – 20 and 30 μ m, Rt -30 and 43 μ m). Annotated text indicates the $^{207}\text{Pb}/^{206}\text{Pb}$ date with 2σ -errors, followed by the Th/U ratio and the measurement number.

$^{207}\text{Pb}/^{235}\text{U}$ ratios >5.3 define a pooled concordia date of 1882 ± 6 Ma (MSWD = 1.3, $n = 23$), which represents the age of titanite growth during retrograde metamorphism. Younger $^{207}\text{Pb}/^{206}\text{Pb}$ dates (1854 ± 50 to 1773 ± 62 Ma) were obtained for analyses with $^{207}\text{Pb}/^{235}\text{U}$ ratios <5.3 and can be assigned to titanite growth during subsequent thermal stages.

In sample 566216, titanite mostly occurs as coronas around and intergrown with rutile and ilmenite (Fig. 4.6e). Corona widths vary from <10 to $70 \mu\text{m}$ and rare single grains are up to $50 \times 50 \mu\text{m}$. Titanite shows a range in Th/U ratios from 0.12 to 2.02 and $^{207}\text{Pb}/^{235}\text{U}$ ratios from 4.66 to 5.72 (Table A8). The highest Th/U ratios were obtained from a large titanite grain near ilmenite. Eighteen analyses yield variably discordant data (Fig. 4.6f), of which the eight most concordant analyses with $^{207}\text{Pb}/^{235}\text{U}$ ratios between 5.18 and 5.52 define a pooled concordia date of 1883 ± 10 Ma (MSWD = 1.5). This date is mostly obtained from titanite in coronas around rutile and is therefore interpreted to reflect the age of titanite growth during retrograde metamorphism. Titanite in coronas around and intergrown with ilmenite yields lower isotope ratios, with the most concordant analyses having $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 1872 ± 70 and 1794 ± 160 Ma. These dates represent ongoing titanite growth during subsequent thermal stages.

Titanite in sample 566277 occurs as coronas around rutile and ilmenite and as anhedral inclusions in garnet, hornblende and plagioclase (Figs. 4.2h and 4.6g). Coronas vary between <10 and $30 \mu\text{m}$, whereas single grains are up to $110 \times 140 \mu\text{m}$ large. A faint BSE-zoning and coronas made up of multiple grains is observed. The Th/U and $^{207}\text{Pb}/^{235}\text{U}$ ratios show relatively large variation (0.60 to 3.57 and 4.46 to 5.60, respectively; Table A8), with the highest Th/U ratios obtained from a titanite inclusion in the Mg-rich garnet rim. Out of 50 analyses, 33 with $^{207}\text{Pb}/^{235}\text{U}$ ratios >5.2 yield a pooled concordia date of 1886 ± 5 Ma (MSWD = 1.2, Fig. 4.6h). In line with the replacement of rutile by titanite, we interpret this date as the age of titanite growth during retrograde metamorphism. Analyses with lower isotope ratios yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 1861 ± 43 and 1738 ± 61 Ma, reflecting further titanite growth during subsequent thermal stages.

4.11 U-Pb dating of rutile

Rutile in sample 524713 occurs as inclusions in plagioclase and hornblende or is intergrown with titanite and ilmenite (Figs. 4.6a and 4.7a). Grain shapes are anhedral to euhedral and grain size varies from 60×95 to $130 \times 280 \mu\text{m}$. Rutile locally contains ilmenite lamellae and fractures. Uranium

and Pb contents are very low (Table A9). Seven, variably discordant analyses show large errors and define a discordia with an upper intercept date of 1743 ± 96 Ma (Fig. 4.7b; MSWD = 0.49). Because rutile in the retrogressed eclogite formed during high-pressure metamorphism (Müller et al., 2018, and below), this intercept is interpreted to reflect the timing at which rutile cooled below its closure temperature for Pb diffusion. The discordant trend of the data, however, indicates that the U-Pb system in rutile did not remain closed, with the lower intercept of the discordia indicating recent Pb-loss (Fig. 4.7b).

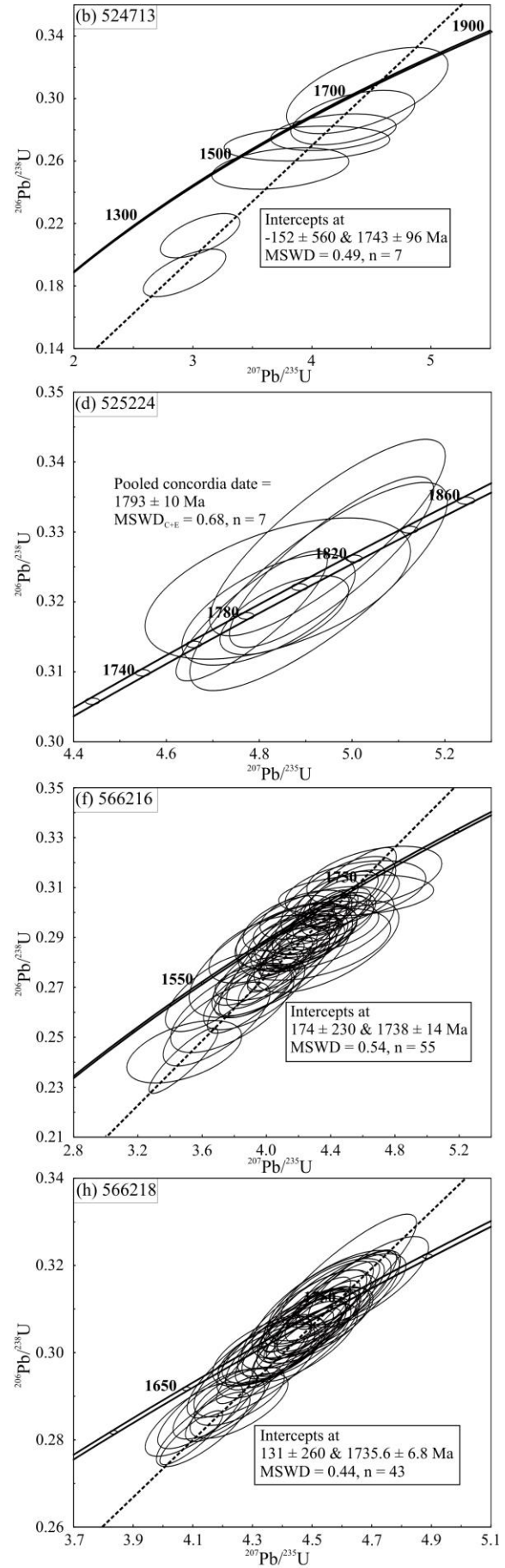
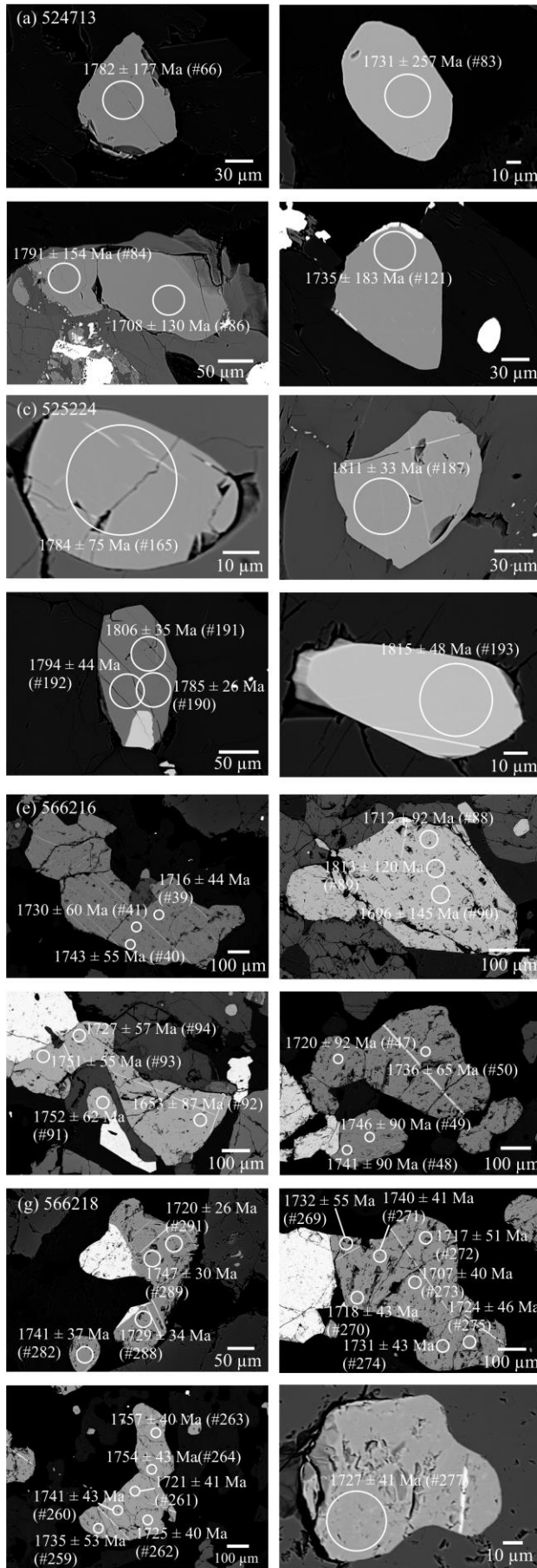
Sample 525224 contains only few rutile grains that occur as inclusions in garnet, hornblende and plagioclase. The rutile grains show ilmenite lamellae and fractures; one grain is intergrown with ilmenite (Fig. 4.7c). Grain shapes are subhedral to euhedral; grain size varies from 55×40 to 95×180 μm . Rutile in this sample has slightly higher U and Pb concentrations than in sample 524713 (Table A9). The seven spots analyzed define a pooled concordia date of 1793 ± 10 Ma (MSWD = 0.68, Fig. 4.7d), which is interpreted as a cooling age.

Rutile in sample 566216 is intergrown with ilmenite and titanite or occurs as single grains surrounded by amphibole and plagioclase (Figs. 4.6e and 4.7e). Grain shapes are mainly anhedral and grain size ranges from 50×50 to 530×480 μm . All rutile grains show ilmenite lamellae of varying width and are severely fractured. Uranium and Pb contents are low and show large variations (Table A9). Fifty-five concordant to variably discordant analyses define a discordia with an upper intercept date of 1738 ± 14 Ma (MSWD = 0.54, Fig. 4.7f). We interpret this date as a cooling age, whereas the lower intercept indicates recent Pb-loss. A concordant data point with low $^{207}\text{Pb}/^{235}\text{U}$ ratio (<4) gives a $^{207}\text{Pb}/^{206}\text{Pb}$ date of 1617 ± 91 Ma, which probably indicates minor resetting of the U-Pb system in rutile after cooling below the closure temperature for Pb diffusion.

Rutile in sample 566218 occurs as inclusions in plagioclase, garnet and hornblende and commonly shows ilmenite lamellae or is intergrown with ilmenite. Grain size varies from 130×100 to 370×450 μm . Most grains are anhedral, strongly fractured (Fig. 4.7g) and show high U contents (Table A9). The 43 spots analyzed define a discordia with an upper intercept date of 1736 ± 7 Ma (Fig. 4.7h; MSWD = 0.44). This date reflects the timing when rutile cooled below its closure temperature for Pb diffusion. The lower intercept indicates that the U-Pb system was disturbed during a recent Pb-loss event.

In sample 566240, rutile occurs as inclusions in hornblende and plagioclase and usually shows ilmenite lamellae (Fig. 4.7i). Anhedral grains are most common and grain size varies from 130×150 to 180×200 μm . Uranium and Pb contents are very similar to rutile in sample 566216 (Table A9). The 38 spots analyzed define a discordia with an upper intercept date of 1732 ± 11 Ma (Fig. 4.7j; MSWD = 0.55). This date is interpreted to reflect closure during cooling, while recent Pb-loss is indicated by the lower intercept of the discordia.

Rutile in sample 566277 occurs in intergrowths with ilmenite and titanite or as inclusions in amphibole, plagioclase and garnet (Fig. 4.6g and 4.7k). Only few grains show ilmenite lamellae.



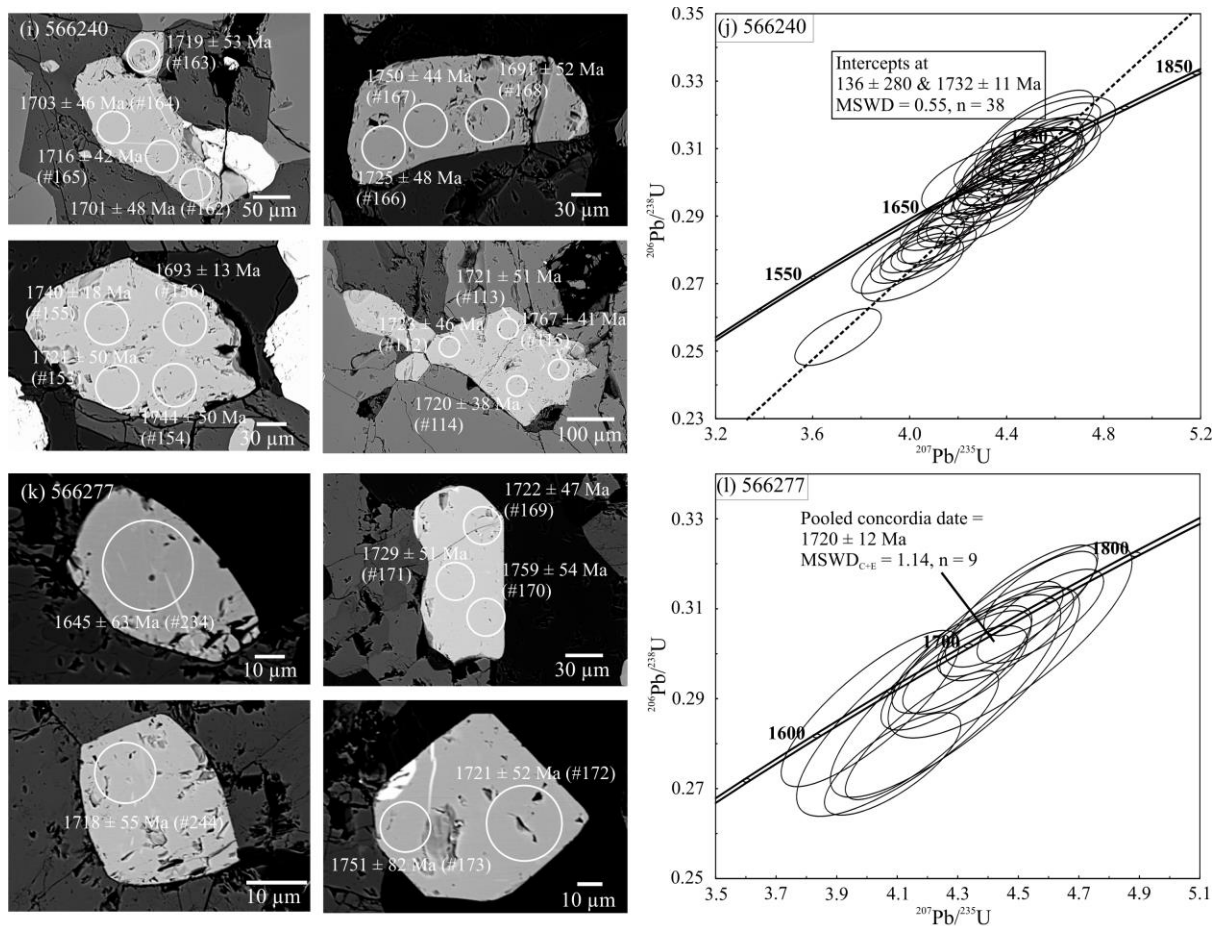


Fig. 4.7. BSE-images of representative rutile grains ($\pm$ ilmenite and titanite) and U-Pb concordia-diagrams from sample (a, b) 524713, (c, d) 525224, (e, f) 566216, (g, h) 566218, (i, j) 566240 and (k, l) 566277. White circles indicate the position of laser spots (20, 30 and 43 μ m). Annotated text indicates the $^{207}\text{Pb}/^{206}\text{Pb}$ date with 2σ -errors, followed by the measurement number.

Grain shapes are euhedral to anhedral, with grain sizes between 80x90 and 70x150 μ m (Fig. 4.7k). Uranium and Pb contents are low (Table A9). Fourteen analyses yield mostly concordant data with $^{207}\text{Pb}/^{235}\text{U}$ ratios between 4.01 and 4.57 (Fig. 4.7l). A pooled concordia date of 1720 ± 12 Ma (MSWD = 1.14, $n = 9$) can be calculated for the most concordant analyses with isotope ratios >4.29 . We interpret this date as a cooling age. A concordant rutile analysis with lower isotope ratios gives a $^{207}\text{Pb}/^{206}\text{Pb}$ date of 1645 ± 63 Ma, which is indicated to reflect minor resetting of the U-Pb system in rutile after cooling below the closure temperature for Pb diffusion.

4.12 Discussion

4.12.1 Interpretation of the results of U-(Th)-Pb analysis

In this study, the most concordant analyses of zircon, monazite, titanite and rutile yield a large range in $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 2634 ± 63 and 1617 ± 91 Ma (Tables 4.4 and A6-A10). In the following

and for clarity, the data are subdivided into four different groups:

- 1) detrital and magmatic zircon of sedimentary and magmatic precursor rocks (2634 – 2092 Ma)
- 2) eclogite- to high-pressure amphibolite-facies zircon, titanite and monazite (1891 – 1882 Ma)
- 3) medium- to low-pressure amphibolite-facies zircon, titanite, monazite and rutile (1872 – 1773 Ma)
- 4) late-stage rutile (1743 – 1617 Ma)

Table 4.4

Summary of $^{207}\text{Pb}/^{206}\text{Pb}$ dates (in Ma) from U-Pb analyses in this study. Intercepts of the discordia are shown in bold (upper, lower), pooled concordia dates are in cursive and further dates (ranges, single analysis) are unformatted. For further discussion see text.

Sample	Zircon	Monazite	Titanite	Rutile
524713			1884 ± 4 ; 1869-1782	1743 ± 96, -152 ± 560
524716			1882 ± 6 ; 1854-1773	
525224	1889 ± 8 ; 1807-1746			1793 ± 10
566216	2092 $\pm$ 22; 2032-1916		1883 ± 10 ; 1872-1794	1738 ± 14, 174 ± 230 , 1617 $\pm$ 91
566218	2131 $\pm$ 9, 2097-1968			1736 ± 7, 131 ± 260
566240	2146 $\pm$ 63, 2052-1903, 1886 $\pm$ 46			1732 ± 11, 136 ± 280
566249	1887 ± 4, 4 ± 150			
566277			1886 ± 5 ; 1861-1738	1720 ± 12 , 1645 $\pm$ 63
566267	2634 $\pm$ 63, 2107 $\pm$ 21, 1891 $\pm$ 10	1882 ± 3 , 1854 $\pm$ 15, 1837 $\pm$ 16		

4.12.1.1 Detrital and magmatic zircon of sedimentary and magmatic precursor rocks (2634 – 2092 Ma)

Concordant detrital zircon in the garnet-kyanite schist yields the oldest $^{207}\text{Pb}/^{206}\text{Pb}$ date obtained in this study (2634 $\pm$ 63 Ma) and provides a maximum age for the deposition of the precursor of the metasedimentary rocks at 2107 $\pm$ 21 Ma (Fig. 4.3k). These dates are generally consistent with the results of previous studies (Kalsbeek and Taylor, 1989; Kalsbeek et al., 1993; Nutman et al., 2008; Thrane et al., 2016), documenting the predominance of Archean detritus and final deposition of the sediments in the Paleoproterozoic. Nevertheless, the oldest concordant zircon in this sample has an Archean $^{207}\text{Pb}/^{206}\text{Pb}$ date of 2634 $\pm$ 63 Ma, which is somewhat younger than previously constrained intrusion ages for the TTG gneiss in the Kuummiut Terrane (T_{DM} and zircon U-Pb; 3076 $\pm$ 14 to 2707 $\pm$ 15 Ma; Kalsbeek et al., 1993; Nutman et al., 2008; Kokfelt et al., 2016a). The magmatic Th/U ratio of this zircon (0.6) indicates that this date may be related to late Archean magmatic activity (zircon U-Pb; 2681 $\pm$ 16 to 2646 $\pm$ 5 Ma; Kokfelt et al., 2016a), rather than regional granulite-facies metamorphism around 2720 $\pm$ 6 Ma (zircon U-Pb; Nutman et al., 2008).

In the basic dykes, the oldest zircon dates (ca. 2146 to 2092 Ma) were mostly obtained from

patchy, BSE-dark and unzoned domains (Fig. 4.3c, e, g). The type of zoning and subhedral to anhedral shape of these grains may be taken as evidence for a metamorphic origin (Hoskin and Black, 2000; Corfu et al., 2003b), but contrasts with the predominance of magmatic Th/U ratios. In addition, one zircon grain with a $^{207}\text{Pb}/^{206}\text{Pb}$ date of 2114 ± 51 Ma (sample 566216) shows faint oscillatory zoning (Fig. 4.3c). Based on the oscillatory zoning and magmatic Th/U ratios, we interpret these dates to approximate the age of dyke emplacement into the Archean TTG gneiss. The data give intrusion ages at 2146 ± 63 Ma (sample 566240), 2131 ± 9 Ma (sample 566218), and 2092 ± 22 Ma (sample 566216). Together with the published zircon U-Pb age of 2015 ± 15 Ma from a metabasic dyke in the northern part of the Kuummiut Terrane (Nutman et al., 2008), these data indicate that dyke emplacement occurred during multiple intrusive stages (Kolb, 2014).

In many cases, the U-Pb isotope system of zircon of this first of the four groups was disturbed as indicated by discordance. A range in $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 2097 ± 5 and 1903 ± 23 Ma was determined for the most concordant analyses, recorded in inclusion-rich, BSE-bright and surrounding BSE-dark domains, as well as patchy and unzoned domains, of predominantly anhedral to subhedral zircon in the garnet-kyanite schist and retrogressed eclogite. These grains are characterized by resorbed margins, show a range in magmatic Th/U ratios (0.1-1.94), and the ca. 2097 to 1903 Ma dates lie in between the estimated ages of dyke emplacement and metamorphism. The chondrite-normalized REE patterns of some inclusion-rich zircon domains and grains with oscillatory zoning show typical magmatic signatures, with a weak negative Eu anomaly and HREE enrichment (Fig. 4.4; Rubatto, 2002; Nutman et al., 2008; Rubatto, 2017). The grains and domains are thus interpreted to reflect magmatic zircon that either experienced partial Pb-loss or incomplete recrystallization during metamorphism (Gebauer et al., 1997; Hoskin and Black, 2000), while the REE patterns remained relatively undisturbed.

Inclusion-rich, BSE-bright domains often yield younger $^{207}\text{Pb}/^{206}\text{Pb}$ dates than the surrounding magmatic, BSE-dark domains (Fig. 4.3c, g) and are characterized by higher U contents (Table A6). In addition, these domains have resorbed and irregular boundaries, suggesting that they experienced resorption during dyke emplacement and zircon-melt interaction. The BSE-bright domains are thus interpreted to represent inherited Archean grains that were incorporated into the magma during ascent. The high U content in the inherited grains led to volume expansion due to differential metamictization and eventually caused fracturing in the surrounding magmatic domains (Corfu et al., 2003b). The fractures form a network of pathways along which fluids can preferentially infiltrate and recrystallize the metamict, inclusion-rich domains, while the BSE-dark domains preserve older dates (Fig. 4.3c, Corfu et al., 2003b).

4.12.1.2 Eclogite- to high-pressure amphibolite-facies zircon, titanite and monazite (1891 – 1882 Ma)

Monazite and titanite, as well as zircon in some samples, predominantly yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 1891 ± 10 and 1882 ± 3 Ma (Table 4.4). In zircon, these dates have been obtained from

both inclusion-rich, BSE-bright and fractured, BSE-dark domains from three retrogressed eclogites and the garnet-kyanite schist (Fig. 4.3a, g, i, k). Monazite in the garnet-kyanite schist has a well-defined date of 1882 ± 3 Ma (Fig. 4.5b), with only two analyses giving slightly younger dates. For titanite, dates of 1886 ± 5 to 1882 ± 6 Ma have mostly been obtained from coronas around rutile and ilmenite, whereas the single grains and titanite in intergrowths often comprise multiple titanite generations (Fig. 4.6a, c, e, g). The consistency and low error of the 1891 ± 10 to 1882 ± 3 Ma dates among minerals of varying closure temperatures, derived from samples of different composition (Table 4.1), and from different localities in the Kuummiut Terrane (Fig. 4.1a), are testimony to the regional nature and extent of the eclogite- to high-pressure amphibolite-facies metamorphic stages. These dates are also consistent with the results of previous geochronological studies on the timing of regional high-grade metamorphism (Bridgwater et al., 1978; Pedersen and Bridgwater, 1979; Kalsbeek et al., 1993; Nutman et al., 2008).

While mineral inclusions are ubiquitous in BSE-bright and inclusion-rich zircon domains, no high-pressure mineral inclusions (as observed by Nutman et al., 2008) could be identified in our samples. Nevertheless, the chondrite-normalized REE patterns of homogeneous zircon with low Th/U ratios (Fig. 4.4) show lower REE content than magmatic zircons, with a weak to absent Eu anomaly and depressed HREE abundances. Such trace element signatures are typical of coeval zircon and garnet growth at eclogite-facies conditions, where plagioclase is absent (Rubatto, 2002; Kohn and Kelly, 2017; Rubatto, 2017). This interpretation is consistent with the results of Nutman et al. (2008), who assigned an age of 1867 ± 28 Ma to the growth of zircon at eclogite-facies conditions, based on the trace element data and inclusion content of zircon. However, despite the presence of eclogite-facies REE signatures, the zircons in our samples mainly occur in hornblende-plagioclase coronas around garnet (Fig. 4.2g) or as inclusions in the Mg-rich garnet rim, and the metamorphic zircon dates are, within error, indistinguishable from those of amphibolite-facies titanite. This may be interpreted to indicate that the ages of the eclogite- and high-pressure amphibolite-facies stages overlap within error, i.e. that the exhumation of the Kuummiut Terrane was rapid. An alternative and equally reasonable interpretation is that the trace element signatures in zircon were not reset during retrogression. Such a lack in the readjustment of the trace element content in zircon has also been observed in amphibolite-facies zircon rims in UHP metamorphic rocks (Liu and Liou, 2011; Gilotti et al., 2014), and also characterizes the magmatic zircon grains in retrogressed eclogite. In addition, the textural relationships indicate that magmatic zircon was variably recrystallized during fluid-induced high-pressure amphibolite-facies retrogression, where garnet was replaced by hornblende-plagioclase symplectites (Müller et al., 2018). Recrystallization of zircon only locally reached completion in areas of presumed higher fluid activity (i.e. sample 525224 and 566249 based on the high biotite and hornblende content), whereas samples dominated by clinopyroxene-plagioclase symplectites (i.e. 566216, 566218, 566240) predominantly yield variably discordant U-Pb dates, caused by either partial Pb-loss or incomplete recrystallization of magmatic zircon during metamorphism (Table 4.4).

If this interpretation is correct, the decoupling of the U-Pb and trace element systematics in zircon also means that the 1867 ± 28 Ma zircon age of Nutman et al. (2008) is unlikely to reflect zircon growth at eclogite-facies conditions.

The mineral assemblage of biotite + kyanite + quartz + garnet + muscovite + plagioclase in the garnet-kyanite schist also records high-pressure amphibolite-facies conditions (Müller et al., 2018). This, in combination with the well-defined U-Pb monazite date of 1882 ± 3 Ma, common association of monazite with biotite and kyanite, and presence of inclusion-rich domains in monazite similar to zircon, indicates that an earlier monazite generation was completely recrystallized during retrogression (Ayers et al., 2002). Recrystallization of monazite most likely took place via coupled dissolution-reprecipitation in the presence of a fluid, as suggested by the patchy zoning and chemical heterogeneity (Teufel and Heinrich, 1997; Townsend et al., 2000; Sindern et al., 2012).

In contrast to zircon and monazite, titanite is usually free of inclusions but the PT-conditions of titanite formation can be derived from pseudosection modelling and microtextural analysis. Figure 4.8a shows the stability fields for titanite and rutile derived from seven retrogressed eclogite samples, combined with the PTt-path for the high-pressure rocks (Müller et al., 2018; this study). In the samples investigated, titanite stability is limited to PT-conditions below ~ 12.3 kbar and 810°C , except for temperatures $< 500^\circ\text{C}$, where the stability field of titanite extends towards higher pressures (Fig. 4.8a). The previously constrained PT-path (Müller et al., 2018) only crosses the titanite stability field at high-pressure amphibolite-facies conditions, and the titanite coronas around rutile (Fig. 4.2f) show that rutile is replaced by titanite during retrogression (Carswell et al., 1996; O'Brien, 1997; Dachs and Proyer, 2001; Lucassen et al., 2012). Furthermore, titanite is frequently associated with plagioclase and hornblende in the symplectites replacing clinopyroxene and garnet (Fig. 4.2h). Therefore, the 1886 ± 5 to 1882 ± 6 Ma titanite dates reflect the timing of titanite growth during fluid-induced, high-pressure amphibolite-facies retrogression.

4.12.1.3 Medium- to low-pressure amphibolite-facies zircon, titanite, monazite and rutile (1872 – 1773 Ma)

Two monazite rims in the garnet-kyanite schist yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates of 1854 ± 15 and 1837 ± 16 Ma (Fig. 4.5b). A range in U-Pb dates between ca. 1872 and 1773 Ma was obtained from composite titanite grains and coronas (Fig. 4.6a, c, g; Table 4.4). Zircon in sample 525224 and 566240 yields $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 1853 ± 53 and 1794 ± 22 Ma, from BSE-dark domains and unzoned grains with a magmatic Th/U ratio (0.16-1.27). Rutile in sample 525224 has been dated at 1793 ± 10 Ma.

Apart from rutile, these dates are interpreted to reflect ongoing, but volumetrically minor mineral growth and recrystallization after the high-pressure amphibolite-facies stage. The older dates between 1872 ± 70 and 1821 ± 31 Ma overlap with the proposed age for the collision of the Rae and North Atlantic cratons (1890 to 1820 Ma; Nutman et al., 2008; Kolb, 2014; Nicoli et al., 2018).

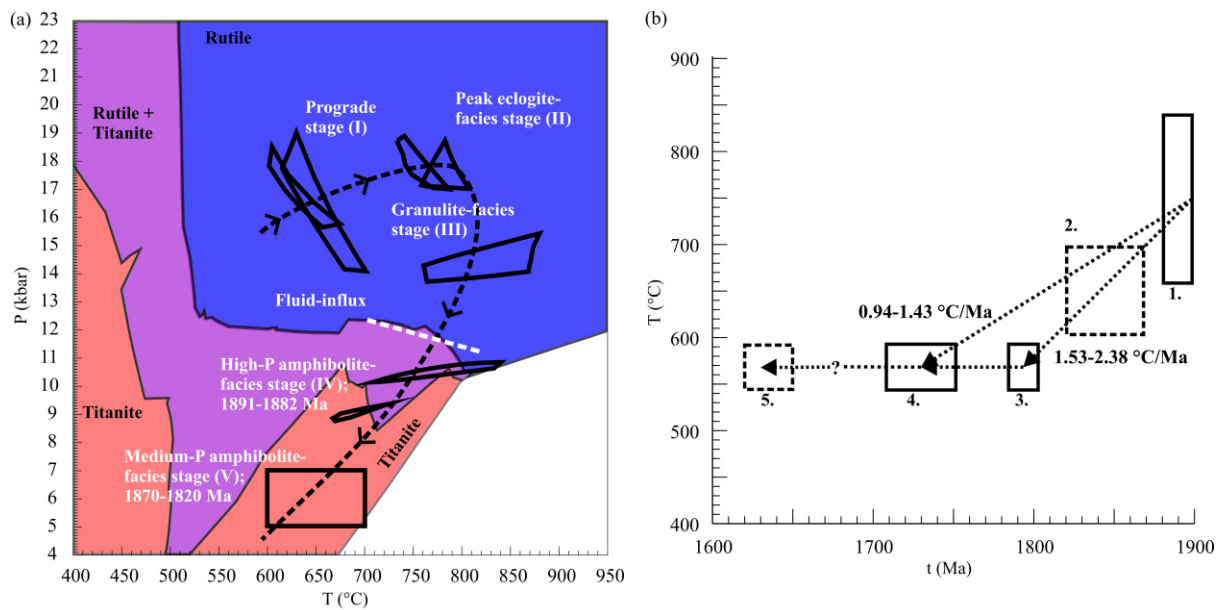


Fig. 4.8: Metamorphic (a) and thermal evolution (b) for the high-pressure rocks of the Kuummiut Terrane. (a) The PT-path is modified after Müller et al. (2018), with PT-data for the medium-pressure amphibolite-facies stage V from Nutman et al. (2008), Baden (2013) and Nicoli et al. (2018). Color-coded areas have been determined from several pseudosections of retrogressed eclogite samples and indicate the stability field of rutile (blue), titanite (red), and rutile + titanite (purple). Ages for the metamorphic stages are from this study. (b) The Tt-diagram depicts the thermal evolution from high-pressure amphibolite-facies metamorphism towards the waning stages of metamorphic and magmatic activity. PT- and age data are from this and earlier studies (see text). 1. High-pressure amphibolite-facies metamorphism between 1891 ± 10 and 1882 ± 3 Ma (660-840 °C), 2. Medium-pressure amphibolite-facies metamorphism at ca. 1870-1820 Ma (~600-700 °C), 3. Local cooling below the closure temperature for Pb diffusion in rutile, following a minor thermal event at 1793 ± 10 Ma (569 ± 24 °C). 4. Regional rutile cooling following a minor thermal event between 1738 ± 14 and 1720 ± 12 Ma (569 ± 24 °C), 5. Partial resetting of the U-Pb system with subsequent cooling around 1645 ± 63 and 1617 ± 91 Ma (569 ± 24 °C) due to post-tectonic intrusions.

PT-conditions for regional medium- to lower-pressure amphibolite-facies retrogression during this stage are constrained at ~5-7 kbar and 600-700 °C (Nutman et al., 2008; Baden, 2013; Nicoli et al., 2018).

An 1817 ± 22 Ma Sm-Nd garnet-clinopyroxene-whole rock isochron date for a metabasic dyke (Kalsbeek et al., 1993) has been interpreted to reflect isotopic re-equilibration during decompression and cooling (Nutman et al., 2008). Further evidence for cooling is given by rutile. In the samples investigated, rutile formed at eclogite- or high-pressure granulite-facies conditions (Fig. 4.8a), but mostly yields younger U-Pb dates than its retrograde replacement product. This discrepancy can be explained by a lower closure temperature for Pb diffusion in rutile. Based on the absence in correlation between rutile grain size and U-Pb date (Fig. 4.7g), we adopt a closure temperature of 569 ± 24 °C, representing a weighted mean from the seven most reliable closure temperature profiles of rutile in a study on rutile age zoning by Kooijman et al. (2010). Closure temperatures for Pb diffusion

in titanite, in contrast, are assumed to be around 800 °C, consistent with the results of Kohn (2017). The 1793 ± 10 Ma date for rutile in sample 525224 is therefore interpreted to indicate the age at which rutile cooled below its closure temperature. An absence in correlation between rutile grain size and U-Pb date in this sample may indicate a minor thermal stage around 1800 Ma, resetting the U-Pb system in rutile, although there is no petrological evidence to support this theory.

4.12.1.4 Late-stage rutile (1743 – 1617 Ma)

Most of the remaining rutile dates vary between 1743 ± 96 and 1720 ± 12 Ma (Fig. 4.7b, f, h, j, l). Similar dates were also obtained from a few titanite grains in sample 566277 (1743 ± 41 and 1738 ± 61 Ma; Fig. 4.6g) and a zircon grain in sample 525224 (1746 ± 32 Ma; Fig. 4.3b). Based on the phase relations established above, we interpret these rutile dates to also reflect cooling. Isotopic closure most likely occurred at similar temperature to that of rutile in sample 525224 (569 ± 24 °C; Kooijman et al., 2010), based on the absence of a correlation between grain size and U-Pb date for the younger rutile. Furthermore, the 1743 ± 96 to 1720 ± 12 Ma cooling ages are consistent with a U-Pb age for metamorphic zircon from a kyanite-bearing metasedimentary rock of the Helheim unit (1740 ± 40 Ma; Nutman et al., 2008). This may indicate that the U-Pb systematics in rutile were reset during a thermal stage at 1743 ± 96 to 1720 ± 12 Ma. The different apparent cooling ages for rutile imply that either locally, cooling in the Kuummiut Terrane may have proceeded faster or that some high-pressure rocks were not affected thoroughly by late thermal stages. A minor thermal overprint is suggested by the youngest $^{207}\text{Pb}/^{206}\text{Pb}$ dates for concordant rutile (1645 ± 63 and 1617 ± 91 Ma), which correlate with zircon U-Pb ages of major, post-tectonic intrusive events across the Kuummiut Terrane (1680 ± 9 to 1545 ± 15 Ma; Kalsbeek et al., 1993; Kokfelt et al., 2016b; Thrane et al., 2016).

4.12.2 Thermal evolution

The newly collected U-Pb data allow a more detailed insight into the metamorphic evolution of the Kuummiut Terrane. Chondrite-normalized REE patterns from metamorphic zircon indicate coeval zircon and garnet growth at eclogite-facies conditions (Fig. 4.4). The absolute age of the eclogite-facies stage is unknown, however, the clockwise, near-isothermal decompression PT-path implies that the Kuummiut Terrane initially experienced a rapid and tectonically-driven exhumation (Fig. 4.8a; Müller et al., 2018). The common occurrence of zircon in hornblende-plagioclase coronas around garnet and in the Mg-rich garnet rim indicates that zircon coexisted with the high-pressure amphibolite-facies assemblage, consistent with the titanite (and monazite) dates between 1886 ± 5 and 1882 ± 3 Ma. Following the high-pressure amphibolite-facies stage, the Kuummiut Terrane experienced a prolonged period of retrogression and cooling. This is recorded by the ca. 1872 to 1773 Ma dates predominantly obtained from titanite and the ca. 1746 to 1720 Ma dates predominantly recorded in rutile.

The cooling history following the tectonically-driven exhumation can be constrained using the

new geochronological data in combination with closure temperature estimates and published PT-estimates (Fig. 4.8b). PT-conditions for the high-pressure amphibolite-facies stage are estimated at 8.8-10.9 kbar and 660-840 °C (Müller et al., 2018). Dates between 1872 ± 70 and 1821 ± 31 Ma correlate with regional medium- to lower-pressure amphibolite-facies metamorphism (~5-7 kbar, 600-700 °C; Nutman et al., 2008; Baden, 2013, Nicoli et al., 2018), interpreted to reflect collision of the Rae and North Atlantic cratons (Nutman et al., 2008; Kolb, 2014; Nicoli et al., 2018). The rutile date of 1793 ± 10 Ma (sample 525224) may point to an at least local cooling below the closure temperature for Pb diffusion in rutile (569 ± 24 °C, Kooijman et al., 2010), corresponding to a cooling rate of 1.53-2.38 °C/Ma between 1891 ± 10 and 1793 ± 10 Ma (Fig. 4.8b). However, the majority of the rutile cooling ages range from 1738 ± 14 to 1720 ± 12 Ma (Fig. 4.7), with the youngest rutile dates at 1645 ± 63 and 1617 ± 91 Ma. The younger cooling ages suggest that the Kuummiut Terrane, as a whole, only cooled below 569 ± 24 °C between 1738 ± 14 and 1720 ± 12 Ma, corresponding to a cooling rate of 0.94-1.43 °C/Ma. This was followed by a prolonged period during which the Kuummiut Terrane presumably stayed at relatively high temperature and at the same crustal level (Fig. 4.8b). Rutile dates at 1793 ± 10 and 1738 ± 14 to 1720 ± 12 Ma, however, may also reflect a succession of minor thermal stages, at which the U-Pb system in rutile was reset due to heating before cooling below ~569 °C again. The youngest rutile dates at 1645 ± 63 and 1617 ± 91 Ma correlate with the emplacement of post-tectonic intrusions (<1680 Ma; Kalsbeek et al., 1993; Kokfelt et al., 2016b; Thrane et al., 2016), with the closure temperature for Pb diffusion in rutile (569 ± 24 °C; Kooijman et al., 2010) being consistent with PT-conditions of contact metamorphism (2.5 kbar and 580-650 °C) obtained from a metasedimentary rock in vicinity to a granite-gabbro complex (Nutman and Friend, 1989). The data show that the Kuummiut Terrane experienced relatively slow, erosion-controlled cooling with only minor thermal perturbations during the waning stages of metamorphic and magmatic activity. Such slow cooling rates following the main period of tectonic and metamorphic activity are typical in ancient orogens (> 1000 Ma; Dunlap, 2000).

In conclusion, our data show that U-Pb dating of accessory minerals with different origin and closure temperatures can provide important constraints on the thermal evolution of Paleoproterozoic orogens. However, the survival of inherited magmatic and eclogite-facies ages critically depends on the degree of retrogression, and problems remain in unequivocally correlating zircon and monazite dates to a specific metamorphic stage. Geochronological studies on retrogressed eclogite are further complicated by the low yield and small grain size of zircon from mafic lithologies, as most information of the pre-metamorphic and prograde evolution may be lost in minerals with lower closure temperature for Pb diffusion or the tendency to grow or recrystallize during retrogression.

4.12.3 Comparison with other studies on Paleoproterozoic eclogite

Based on previous geochronological data (Kalsbeek et al., 1993; Nutman et al., 2008), Kolb (2014) proposed that the basic dykes and other meta-volcanic rocks of the Kuummiut Terrane were

emplaced during Paleoproterozoic extension and basin formation. This may be supported by overlapping $^{207}\text{Pb}/^{206}\text{Pb}$ dates of detrital (2107 ± 21 Ma) and magmatic zircon (2146 ± 63 to 2092 ± 22 Ma) in the garnet-kyanite schist and retrogressed eclogite, respectively, indicating more or less contemporaneous dyke emplacement and sedimentation. A similar tectonic setting has been proposed for dykes in the Belomorian Belt (Imayama et al., 2017; Liu et al., 2017) and in the Nagssugtoqidian Orogen in West Greenland (Van Gool et al., 2002).

Where geochronological data are available, the timing of high-pressure metamorphism is quite similar between the different Paleoproterozoic orogens, with eclogite-facies ages of 1904 ± 0.3 Ma for the Snowbird tectonic zone (Baldwin et al., 2004), and 1868 ± 17 to 1865 ± 15 Ma, 1909 ± 11 to 1897 ± 10 Ma, and ~ 1890 Ma for the Salma-, Gridino- and Kuru-Vaara-type eclogites of the Belomorian mobile belt (Imayama et al., 2017; Yu et al., 2017; Liu et al., 2017). The data presented in this study overlap with the above ages. Similarities are also observed for the estimated PT-conditions and overall PT-evolution of the Paleoproterozoic eclogites. PT-estimates of 17-19 kbar, 740-810 °C (Nagssugtoqidian Orogen, Müller et al., 2018), 18-20 kbar, 920-1000 °C (Snowbird Tectonic Zone, Baldwin et al., 2004), 16-18 kbar, 750-770 °C (Salma-type eclogite, Belomorian Belt, Imayama et al., 2017), >18 kbar, 695-755 °C (Gridino-type eclogite, Belomorian Belt, Yu et al., 2017) and 18-20 kbar, 720-820 °C (Kuru-Vaara-type eclogite, Belomorian Belt, Liu et al., 2017) have been reported for the eclogite-facies stage, followed by an initially rapid, tectonically-controlled exhumation and slow, erosion-controlled cooling towards the end of metamorphic activity (Bibikova et al., 2001; Baldwin et al., 2004; Berman et al., 2007; Liu et al., 2017; Imayama et al., 2017; Yu et al., 2017). Assuming that the calculated pressures approximate the lithostatic load, these data correspond to geothermal gradients of 11 °C/km to 13 °C/km and maximum subduction depths of 67-75 km. These geothermal gradients are consistent with eclogite formation in warmer subduction zones in the Paleoproterozoic, due to higher mantle temperature (Bradley et al., 2011; Anderson et al., 2012; Yu et al., 2017; Brown and Johnson, 2018; Loose and Schenk, 2018).

Nevertheless, a generally warmer thermal environment in Paleoproterozoic orogens appears to be in conflict with two recently discovered UHP metamorphic complexes, most notably in the Nagssugtoqidian Orogen in West Greenland (Glassley et al., 2014) but also the Trans-Hudson Orogen in North America (Weller and St-Onge, 2017). The presence of these Paleoproterozoic UHP metamorphic complexes may reflect a local phenomenon rather than globally acting plate-tectonic processes (Brown and Johnson, 2018) or indicate a preservation problem of Paleoproterozoic UHP metamorphic rocks (Weller and St-Onge, 2017). Our data (Müller et al., 2018; this study) and those of Glassley et al. (2014) show that both high-pressure and UHP rocks can be preserved in a single Paleoproterozoic orogenic belt, suggesting that their formation may be linked in time and space.

4.13 Conclusions

- Detrital zircon in the garnet-kyanite schist yields Archean to Paleoproterozoic dates and confines the maximum deposition of the metasediment precursor at 2107 ± 21 Ma.
- Dyke emplacement occurred during multiple intrusive events at 2146 ± 63 , 2092 ± 22 and possibly 2015 ± 15 Ma (Nutman et al., 2008). Detrital and magmatic zircon dates overlap within error, possibly indicating near-contemporaneous dyke emplacement and sedimentation during Paleoproterozoic extension and basin formation.
- The metamorphic PT-evolution was associated with partial to complete recrystallization of zircon and monazite, titanite growth, and the resetting of the U-Pb system in rutile.
- Zircon, monazite and titanite yield U-Pb dates of 1891 ± 10 to 1882 ± 3 Ma, which are interpreted to reflect recrystallization and mineral growth at high-pressure amphibolite-facies conditions.
- The Kuummiut Terrane initially experienced relatively rapid, tectonically-driven exhumation, followed by slow, erosion-controlled cooling during the waning stage of metamorphic and magmatic activity.

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4.15 References

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5 Conclusion and outlooks

5.1 Summary

In this thesis, I investigated the tectonometamorphic evolution and geochronology of variably retrogressed high-pressure rocks from the Kuummiut Terrane in the Nagssugtoqidian Orogen in South-East Greenland. The overall goal was to quantify the PTt-evolution of the Kuummiut Terrane and to gain insight into the geodynamic processes and subduction depths in this Paleoproterozoic orogen. The main results presented in Chapters 2–4 are summarized as follows:

- The Kuummiut Terrane is dominated by highly-foliated Archean TTG gneiss and a variety of supracrustal rocks. Mineral assemblages in the exhumed rocks mostly reflect high-pressure amphibolite-facies conditions; well-equilibrated eclogite *sensu stricto* (containing garnet and omphacite) was not found.
- Relict eclogite-facies mineral assemblages and decompression textures are preserved in metabasic dykes intruding the Archean TTG gneiss, and in boudins and boudinaged layers of basic to ultrabasic supracrustal rock. The degree of retrogression and type of replacement assemblage is interpreted to be mainly dependent on the amount of fluid available during retrograde metamorphism.
- Texturally well-equilibrated eclogite-facies mineral assemblages were identified in garnet-pyroxenite; high-pressure amphibolite-facies mineral assemblages with only minor reaction textures were identified in garnet-amphibolite and garnet-kyanite schist.
- In the metabasic dykes and supracrustal rocks, the retrograde replacement reactions ceased prior to completion, resulting in the preservation of complex mineral reaction textures typical of retrogressed eclogite, and the formation of two mineralogically and chemically distinct domains. A clinopyroxene domain consists of fine-grained clinopyroxene-plagioclase symplectite after omphacite. Omphacite is only preserved in a Na-rich retrogressed eclogite sample. In a garnet domain, coarse-grained garnet is surrounded and/or variably pseudomorphed by plagioclase + hornblende $\pm$ clinopyroxene $\pm$ orthopyroxene.
- Through a combination of mineral textural analysis, conventional geothermobarometry and pseudosection modelling, a clockwise PT-evolution and a near-isothermal decompression path have been obtained. Large garnet grains in retrogressed eclogite often have a Ca-rich core, interpreted as prograde in origin, and pseudosection modelling indicates PT-conditions of 14-19 kbar and 600-750 °C. Ultrabasic garnet-pyroxenite constrains the eclogite-facies PT-conditions to 17-19 kbar and 740-810 °C. The Na-rich retrogressed eclogite records near-isothermal decompression to high-pressure granulite-facies conditions (13.8-15.4 kbar, 760-

880 °C). Subsequent decompression and cooling to high-pressure amphibolite-facies conditions (8.2-12.2 kbar, 680-750 °C) are recorded in garnet-amphibolite and garnet-kyanite schist. The near-isothermal decompression path implies that exhumation was initially rapid and tectonically-controlled.

- LA-SF-ICP-MS U-Pb dating of zircon, monazite, titanite and rutile gives a large range in $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 2634 ± 63 and 1617 ± 91 Ma.
- Archean to Paleoproterozoic dates were determined for detrital zircon in the garnet-kyanite schist, with the youngest detrital grain confining the maximum deposition of the metasediment precursor to 2107 ± 21 Ma.
- Magmatic zircon in retrogressed eclogite indicates that dyke emplacement occurred during multiple intrusive events (at 2146 ± 63 , 2092 ± 22 and possibly also 2015 ± 15 Ma; Nutman et al., 2008). Furthermore, overlapping dates for detrital and magmatic zircon possibly indicate near-contemporaneous dyke emplacement and sedimentation during Paleoproterozoic rifting and basin formation.
- The metamorphic evolution of the Kuummiut Terrane was associated with partial to complete recrystallization of zircon and monazite, titanite growth and the resetting of the U-Pb system in rutile.
- The majority of the zircon, monazite and titanite dates are between 1891 ± 10 and 1882 ± 3 Ma. Although, the REE patterns for metamorphic zircon imply growth at eclogite-facies conditions, the REE data for magmatic zircon and the microstructural setting of zircon, monazite and titanite imply that the above range in dates most likely reflects mineral growth and recrystallization at high-pressure amphibolite-facies conditions.
- Further retrogression occurred in relation to the final stages of compressional deformation in the Kuummiut Terrane (Nutman et al., 2008; Baden, 2013; Kolb, 2014; Nicoli et al., 2018), as reflected by titanite and monazite dates between 1872 ± 70 and 1821 ± 31 Ma.
- Rutile dates at 1793 ± 10 Ma, 1738 ± 14 to 1720 ± 12 Ma, 1645 ± 63 Ma and 1617 ± 91 Ma, record relatively slow, erosion-controlled cooling during the waning stages of metamorphic and magmatic activity.

In summary, the data presented in this study provide detailed insight into the tectonometamorphic evolution of the Kuummiut Terrane. With the new data, some previous observations were (re-) confirmed or better constrained, including multiple stages of dyke emplacement (Chapter 4; Kolb, 2014) and near-contemporaneous dyke emplacement and sedimentation (Chapter 4; Kolb, 2014). Other results are new, including the first report of prograde, eclogite-facies conditions, and overall PT-path, as well as the temperature-time evolution during retrogression.

5.2 Research needs

Despite a long history of research in the Nagssugtoqidian Orogen in South-East Greenland (Andrews et al., 1973; Bridgwater et al., 1973; Wright et al., 1973; Bridgwater, 1976; Bridgwater et al., 1976; Bridgwater et al., 1978; Bridgwater and Myers, 1979; Pedersen and Bridgwater, 1979; Myers, 1987; Andersen et al., 1989; Kalsbeek, 1989; Vannucci et al., 1989; Bridgwater et al., 1990; Mengel et al., 1990; Messiga et al., 1990; Kalsbeek et al., 1993; Nutman et al., 2008; Kolb, 2014, Kolb et al., 2016; Nicoli et al., 2018, Lebrun et al., 2018; this thesis), some questions remain:

- 1) What is the timing of eclogite-facies metamorphism?
- 2) Can the PTt-conditions of prograde metamorphism be constrained more precisely?
- 3) How and along which structures was the Kuummiut Terrane exhumed?

5.2.1 What is the timing of eclogite-facies metamorphism?

As discussed in Chapter 4, the 1891 ± 10 to 1882 ± 3 Ma zircon, monazite and titanite dates obtained in this study and the 1867 ± 28 Ma zircon U-Pb date from Nutman et al. (2008) most likely reflect mineral growth and recrystallization at high-pressure amphibolite-facies conditions. PT-data indicates an initially rapid exhumation, with the timing of eclogite- and high-pressure amphibolite-facies metamorphism probably being within error of each other. The exact age of eclogite-facies metamorphism, however, remains unknown.

Constraining the timing of eclogite-facies metamorphism is of paramount importance for the determination of subduction and exhumation rates, as well as in confining the age range for prograde metamorphism and retrograde high-pressure granulite-facies metamorphism. However, the presence of domainal re-equilibration volumes, limited fluid-availability, and scarcity or minuteness of datable material, makes it very difficult to collect reliable U-Pb data in retrogressed eclogite. One possibility to establish the timing of eclogite-facies metamorphism may be a higher sample throughput, although this would require some kind of pre-selection step (e.g. one thin section per sample) to determine which samples may yield the most datable material. This method may potentially yield larger and more zircon (and monazite) grains with larger inclusions, or zircon grains that occur as inclusion in eclogite-facies minerals (Ca-rich garnet, omphacite, rutile). A method to potentially correlate the U-Pb dates of zircon to the metamorphic stages on a PT-path is Ti-in-zircon thermometry (Ferry and Watson, 2007). This method, however, is severely compromised in the Kuummiut Terrane due to the near-isothermal decompression path, with roughly similar temperatures for the eclogite-, granulite- and amphibolite-facies metamorphic stage

An alternative to analyzing retrogressed eclogite may be the analysis of texturally well-equilibrated samples, such as garnet-pyroxenite, from which the eclogite-facies PT-conditions were

constrained. This ultrabasic rock, however, is even more devoid of U-bearing accessory phases than retrogressed eclogite. Bulk-rock or multigrain-dating methods (i.e. Ar-Ar, Sm-Nd, Lu-Hf, Rb-Sr, Pb-Pb) may only give information about the prograde and/or retrograde metamorphic evolution, due to their relatively low closure temperatures (Harrison, 1981; Mezger et al., 1992; Christensen et al., 1994; Scherer et al., 2000). Based on the scarcity of U-bearing accessory phases in mafic rocks (Timmermann et al., 2004), it may also be worth analyzing felsic lithologies, which in general should yield a larger amount of zircon. Low-strain domains, in particular, may preserve relatively well-equilibrated high-pressure mineral assemblages in various lithologies and provide information on the eclogite-facies stage. Based on the dominance of TTG gneiss in the Kuummiut Terrane, such domains, however are probably very hard to identify.

5.2.2 Can the PTt-conditions of prograde metamorphism be constrained more precisely?

Ca-rich cores (Grt_a) of large garnet grains in retrogressed eclogite are the only evidence for the PT-conditions during prograde metamorphism (Chapter 3). Combined mineral chemical data for Grt_a and intersecting X_{Grs} and X_{Fe} isopleths yield a relatively large range in PT-conditions from 14-19 kbar and 600-750 °C. The PT-data may be more precisely constrained by conducting pseudosection modelling in additional retrogressed eclogite samples containing Ca-rich garnet. Furthermore, the inclusion assemblage within Ca-rich garnet may be used to determine pressures and temperatures via conventional geothermobarometric methods. In the available QEMSCAN images, quartz, titanite, epidote and chlorite were identified as the major inclusions in Grt_a, with minor plagioclase, rutile and apatite. These mineral phases, however, probably do not encompass all inclusions within Grt_a, as very small grains tend to be overlooked by QEMSCAN-analysis. Nonetheless, all of the above minerals except for titanite (apatite not included in the modelling) can be present at the PT-conditions determined for the prograde stage. At these conditions, rutile replaces titanite. As such, the titanite grains within Ca-rich garnet may possibly represent inherited grains that remained metastable during metamorphism. Zr-in-sphene geothermobarometry may be used to determine the origin of the titanite inclusions (Hayden et al., 2008). From the remaining phases, temperatures may be estimated via Ti-in-quartz (Wark and Watson, 2006), garnet-chlorite (Grambling, 1990), garnet-epidote (Perchuk, 1991) and Zr-in-rutile thermometry (Ferry and Watson, 2007).

In contrast to the PT-data, the timing of prograde metamorphism is unknown. Prograde ages may be used to further confine the timing of dyke emplacement and eclogite-facies metamorphism, allowing the estimation of subduction rates. During future fieldwork, it may be worth sampling mafic lithologies with relatively large garnet grains, as these grains potentially may preserve Ca-rich cores with inclusions that are large enough to be analyzed (i.e. zircon). From the known inclusion assemblage, rutile, titanite and apatite are suitable for U-Pb dating. Rutile and apatite, however, are likely to only give cooling ages due to their relatively low closure temperatures for Pb diffusion (400-700 °C; Mezger et al., 1989; Cherniak et al., 1991; Krogstad and Walker, 1994; Cherniak, 2000; Vry

and Baker, 2006; Kooijman et al., 2010; Warren et al., 2011). Closure temperatures for titanite, in contrast, were recently reported to be around 800 °C (Kohn, 2017). Depending on its origin, it may either give the age of titanite formation prior to the actual prograde stage or an age of titanite formation or recrystallization during prograde metamorphism.

5.2.3 How and along which structures was the Kuummiut Terrane exhumed?

During fieldwork, a complex structural geology was identified throughout the Kuummiut Terrane (Chapter 2), hindering regional correlations of the different structural fabrics and deformation events. Furthermore, high-strain fabrics are typically observed in TTG gneiss and garnet-kyanite or graphite-schist, whereas retrogressed eclogite and garnet-pyroxenite are mostly at low strain. Consequently, the eclogite-facies mineral assemblages in retrogressed eclogite and garnet-pyroxenite formed prior to deformation in surrounding country rock. Due to these factors, the nature of the early exhumation in the Kuummiut Terrane was constrained using a combination of PT- and age data (Chapter 3 and 4). Furthermore, the exact nature of the tectonic event(s) terminating subduction and initiating exhumation, as well as the structures along which the high-pressure rocks were exhumed, are unknown. To resolve these issues, more detailed structural work is necessary, in particular from the bounding shear zones and country gneiss.

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Appendix 1 – List of localities (from aFieldwork¹)

Locality ID	Locality Name	Date	Latitude	Longitude
Camp 1 (15.07-17.07.2014)				
1	Camp 1	16.07.2014	66.4143692	-38.1832291
2	Diorite gneiss	16.07.2014	66.4154169	-38.1829386
3	Grt ² -amphibolite	16.07.2014	66.4159667	-38.1820439
4	TTG gneiss	16.07.2014	66.4164828	-38.1805751
5	Grt-Bt rock	16.07.2014	66.4176359	-38.1797003
6	Grt-Bt-Pl-Qz gneiss	16.07.2014	66.4182904	-38.1765059
7	Ol-Hbl-Pl-Opx schist	16.07.2014	66.419087	-38.1780403
8	Grt-Pl-Qz-Hbl gneiss	16.07.2014	66.4197229	-38.1762198
9	Folded gneiss	16.07.2014	66.4207007	-38.1763812
10	Sheared amphibolite	16.07.2014	66.4213525	-38.1797777
11	Grt-amphibolite	16.07.2014	66.4207449	-38.1809773
12	TTG gneiss 2	17.07.2014	66.4197963	-38.1822434
13	Grt-amphibolite	17.07.2014	66.4306376	-38.1833079
14	Grt-amphibolite	17.07.2014	66.431567	-38.1828318
15	Grt-amphibolite/gneiss contact	17.07.2014	66.4321467	-38.1870393
16	Pyroxenite	17.07.2014	66.4319722	-38.1923289
17	Grt-rich layer in amphibolite	17.07.2014	66.4295463	-38.1977401
Camp 2 (18.07-22.07.2014)				
18	Camp 2	18.07.2014	65.9731367	-37.9927371
19	Basic boudin next to camp	18.07.2014	65.9733745	-37.9942623
20	Garnet and kyanite	18.07.2014	65.9730386	-37.9980445
21	Garnet with bright stripes	18.07.2014	65.9733414	-37.9984331
22	Eclogite gravel	18.07.2014	65.9734869	-37.9957144
23	Trondhjemite gneiss	19.07.2014	65.9724519	-37.9996722
24	Grt-amphibolite with relict eclogitic assemblages	19.07.2014	65.9738418	-38.0047792
25	Grt-amphibolite	19.07.2014	65.9735006	-38.0093765
26	Weathered amphibolite with eclogitic relicts	19.07.2014	65.9734668	-38.0102148
27	TTG gneiss	19.07.2014	65.9719638	-37.990133
28	Weathered amphibolite	19.07.2014	65.9725571	-37.9889499
29	Green orthoamphibole schist	19.07.2014	65.9725559	-37.9876542
30	Grt-Am rich schist	19.07.2014	65.9725044	-37.9853406
31	Green lens in TTG gneiss	20.07.2014	65.9720104	-38.0075451
32	Grt-amphibolite with eclogitic relicts	20.07.2014	65.9694852	-38.0075268
33	Retrogressed eclogite	20.07.2014	65.9698572	-38.0073142
34	Banded amphibolite	20.07.2014	65.9628507	-38.0104586
Reco 1 (21.07.2014)				
35	Gneiss at reco	21.07.2014	65.7846899	-37.8707219
36	Darker gneiss with two lineations	21.07.2014	65.7865528	-37.8697936
37	Anorthosite, leucogabbro	21.07.2014	65.7863942	-37.7870473
38	Gneiss	21.07.2014	65.7860352	-37.78898
Camp 2 (continued)				
39	Retrogressed eclogite	21.07.2014	65.9812112	-37.9964542
41	Gneiss	21.07.2014	65.9783968	-38.0032856
42	Rusty gneiss	21.07.2014	65.9744279	-37.9983094
43	Gneiss southeast	22.07.2014	65.9686137	-37.986601
44	Gneiss further south with fold axis	22.07.2014	65.9671351	-37.9859184

Locality ID	Locality Name	Date	Latitude	Longitude
45	Ultramafite	22.07.2014	65.9631661	-37.9832736
46	Intrusive plutonic body	22.07.2014	65.962582	-37.9827871
47	Garnet 2	22.07.2014	65.9600077	-37.98125
48	Garnet 3	22.07.2014	65.9573893	-37.981391
49	Gneiss above garnet	22.07.2014	65.955281	-37.9894707
50	Fold axis in gneiss	22.07.2014	65.964689	-37.9950668
51	Garnet 4	22.07.2014	65.9680474	-37.9979684
52	Garnet 5	22.07.2014	65.9720899	-37.9948328
	Reco 2 (23.07.2014)			
53	Reco 2 stop 1	23.07.2014	66.1428578	-37.9921478
54	Reco 2 stop 2	23.07.2014	66.1092147	-37.9703311
55	Reco 2 stop 3 graphite locality	23.07.2014	66.0879704	-38.0452815
	Camp 3 (23.07-26.07.2014)			
56	Camp 3	23.07.2014	65.8299301	-36.820587
57	Rusty gneiss east of lake	24.07.2014	65.8338188	-36.8210995
58	Gneiss	24.07.2014	65.8347846	-36.8243015
59	Gneiss above second lake	24.07.2014	65.8384297	-36.8452227
60	Amphibolite boudin	24.07.2014	65.8387032	-36.8466636
61	Gneiss with amphibolite boudin	24.07.2014	65.8390373	-36.848823
62	Retrogressed eclogite	24.07.2014	65.8386818	-36.8483786
63	Amphibolite with many veins	24.07.2014	65.8409785	-36.853023
64	Gneiss next to boudins	24.07.2014	65.8418087	-36.856116
65	Retrogressed eclogite	24.07.2014	65.8408408	-36.8564342
66	Boulder with amphibolite schist containing eclogitic boudins	24.07.2014	65.8395705	-36.8569557
67	Gneiss north of camp	24.07.2014	65.8364129	-36.8367181
68	Lens in gneiss	25.07.2014	65.8251824	-36.8224606
69	Gneiss and amphibolite	25.07.2014	65.8234792	-36.8298863
70	Gneiss and amphibolite 2	25.07.2014	65.8201519	-36.8312214
71	Gneiss on southeastern ridge of Blokken	25.07.2014	65.8146143	-36.8357427
72	Grt-amphibolite next to Ky-Grt schist	25.07.2014	65.8155549	-36.8393393
73	Gneiss next to river	25.07.2014	65.8159743	-36.8585064
74	Gneiss next to whirlpool	25.07.2014	65.8122639	-36.8498237
75	Gneiss with two foliations	25.07.2014	65.8127294	-36.8484661
76	Amphibolite east of camp	26.07.2014	65.8297171	-36.811809
77	Grt-amphibolite further south	26.07.2014	65.8298889	-36.8087335
78	Folded pegmatite next to gneiss and mafic rock	26.07.2014	65.8301101	-36.8057039
79	Basic dyke	26.07.2014	65.8291137	-36.8014487
80	Gravel sample and sample depot	26.07.2014	65.8284472	-36.7979007
81	Amphibolite southeast of sample depot	26.07.2014	65.827711	-36.7934204
82	Retrogressed eclogite	26.07.2014	65.8277115	-36.7925239
83	Amphibolite southeast of camp	26.07.2014	65.8299382	-36.8066318
	Camp 4 (27.07-29.07.2014)			
84	Camp 4	27.07.2014	65.9269585	-37.2392755
85	Rusty amphibolite	27.07.2014	65.9316094	-37.2251759
86	Metasediment boulders	27.07.2014	65.9335338	-37.2103312
87	Rusty rock at river	27.07.2014	65.9269476	-37.2237561
88	Orthogneiss	28.07.2014	65.9390583	-37.186814
89	Thrust	28.07.2014	65.9535348	-37.1877429

Locality ID	Locality Name	Date	Latitude	Longitude
90	Metasediment	28.07.2014	65.9533648	-37.1907216
91	Gneiss	28.07.2014	65.9527994	-37.1909666
92	Metasediment 2	28.07.2014	65.9524238	-37.1920559
93	Folded gneiss	28.07.2014	65.9521198	-37.1924814
94	Grt-amphibolite	28.07.2014	65.951022	-37.196586
95	Eclogitic gravel below dyke	28.07.2014	65.9474612	-37.1974457
96	Basic dyke in gneiss	29.07.2014	65.9383653	-37.3181214
97	Gneiss and diorite	29.07.2014	65.9393586	-37.3202546
98	Gneiss with two foliations	29.07.2014	65.9399869	-37.3179609
99	Retrogressed eclogite	29.07.2014	65.9402906	-37.3175326
100	Gneiss	29.07.2014	65.9422329	-37.322893
101	Folded gneiss	29.07.2014	65.9445591	-37.3228576
102	Gneiss with cleavages	29.07.2014	65.9462256	-37.3219136
103	Gneiss near rivers	29.07.2014	65.950238	-37.3047406
	Reco 3 (30.07.2014)			
105	Reco 3 stop 1 Grt-amphibolite	30.07.2014	66.1650454	-37.4959128
106	Retrogressed eclogite	30.07.2014	65.9799691	-38.1081672
107	Reco 3 stop 2	30.07.2014	65.9807571	-38.1085326
	Camp 5 (30.07-01.08.2014)			
104	Camp 5/joint camp with team 6	30.07.2014	65.9981805	-38.0684711
108	Gneiss near water spot	31.07.2014	66.1644091	-37.4734786
109	Rusty gneiss	31.07.2014	66.1638346	-37.4397718
110	Amphibolite	31.07.2014	66.1631799	-37.4325998
111	Orthogneiss and amphibolite	31.07.2014	66.1602743	-37.415424
112	Rusty gneiss on way back	31.07.2014	66.1636684	-37.4329752
113	Rusty gneiss 3	31.07.2014	66.1635318	-37.4348591
114	Gneiss with weathered out mineral	31.07.2014	66.1648091	-37.4547365
115	Carbonate	31.07.2014	66.1646451	-37.4624584
116	Gneiss north of camp	01.08.2014	66.168042	-37.488465
117	Gneiss further north	01.08.2014	66.1724615	-37.486114
118	Gneiss	01.08.2014	66.1678865	-37.4794225
119	Gneiss 4	01.08.2014	66.1690005	-37.4674542
120	Break spot	01.08.2014	66.171153	-37.4520787
121	Gneiss with two foliations	01.08.2014	66.1706219	-37.4523836
122	Gneiss near lunch spot	01.08.2014	66.1685971	-37.455742
123	Gneiss with S-C structures	01.08.2014	66.1665723	-37.4586377
124	Carbonate bounded by two gneiss units and pegmatite	01.08.2014	66.1664312	-37.4585168
125	Carbonate with weathered out top	01.08.2014	66.1651396	-37.4578064
126	Rust zone	01.08.2014	66.1648544	-37.4532315
	Camp 6 (02.08-05.08.2014)			
127	Camp 6	02.08.2014	66.2050494	-37.140727
128	Ultramafite	02.08.2014	66.2137728	-37.1430236
129	Dyke	02.08.2014	66.2162702	-37.1484911
130	Between two glaciers before the metasediment	02.08.2014	66.2195582	-37.1563205
131	Grt-amphibolite	02.08.2014	66.2193483	-37.1659689
132	Double folded gneiss	02.08.2014	66.2171659	-37.1634799
133	Garnet pyroxene rock	03.08.2014	66.2214444	-37.1584455
134	Lunch stop day 2	03.08.2014	66.217112	-37.1396243

Locality ID	Locality Name	Date	Latitude	Longitude
135	Rusty gneiss	03.08.2014	66.2171433	-37.1335244
136	Gneiss	03.08.2014	66.2176171	-37.1280966
137	Amphibolite	03.08.2014	66.2193786	-37.1242063
138	Orthogneiss	03.08.2014	66.2187147	-37.1238404
139	Ultramafite east of camp	04.08.2014	66.2066656	-37.1299225
140	Felsic lens between ultramafite	04.08.2014	66.2067835	-37.1285035
141	Grt gneiss	04.08.2014	66.2067287	-37.1265549
142	Grt-amphibolite	04.08.2014	66.2078166	-37.1264735
143	Grt-amphibolite with pegmatite containing blue mineral	04.08.2014	66.2088878	-37.1202985
144	On top of glacier rim	04.08.2014	66.2092236	-37.1162027
145	Folded gneiss	04.08.2014	66.2095215	-37.1172124
146	At rim of fjord next to granulite	05.08.2014	66.3056872	-37.0867153
	Camp 7 (05.08-07.08.2014)			
147	Camp 7	05.08.2014	65.6659843	-37.1008245
148	Garnet gneiss	06.08.2014	65.6742014	-37.1051587
149	Sheared gneiss	06.08.2014	65.6756164	-37.1094523
150	Dunite	06.08.2014	65.6764765	-37.1078638
151	Contact between gneiss and ultramafic pod	06.08.2014	65.6802787	-37.1084682
152	Gneiss with garnet	06.08.2014	65.6830581	-37.1061793
153	Gneiss at last lake before fjord	06.08.2014	65.6887744	-37.1006139
154	Gneiss on other side of lake	06.08.2014	65.6859316	-37.1019297
155	Grt-rich gneiss	06.08.2014	65.684064	-37.1002263
156	Gneiss near waterfall	06.08.2014	65.682667	-37.0976758
157	Ultramafic rock near fjord	07.08.2014	65.6761789	-37.1120971
158	Gneiss in high valley	07.08.2014	65.6922155	-37.1023848
159	Lunch stop at lake	07.08.2014	65.6925565	-37.1259017
160	Gneiss near western end of valley	07.08.2014	65.6918768	-37.1443086
161	Gneiss 2	07.08.2014	65.6920331	-37.1432878
162	Dunite	07.08.2014	65.6916845	-37.1389492

¹ Official sample collector: Annika Dziggel (ADZ), ² Mineral abbreviations after Whitney and Evans (2010, Appendix 3).

Appendix 2 – List of samples

Sample	Collector	Rock Type	GPS		Section No.	Chemical analysis			PT		U-Pb dating
			Latitude	Longitude		EPMA	QS	XRF Major	TC	Conv.	
Collected 1986:											
318349	ANU	Eclogite			29011	x	x	x	x	x	
318350	ANU	Eclogite			29010	x	x	x	x	x	
Collected 2010:											
524713	BMS	Grt ¹ -amphibolite	65.9483	-38.1747	28807	x		x	x	x	x
524715	BMS	Eclogite	65.9977	-38.0692	28808	x		x	x	x	
524716	BMS	Eclogite	65.9977	-38.0692	28809	x		x	x	x	x
524731	BMS	Amphibolite (medium-grained)	65.0810	-40.0992	28810	x		x	x	x	
524733	BMS	Amphibolite (fine-grained)	65.0810	-40.0992							
524746	BMS	Mafic dyke/Dolerite	65.9703	-37.7249							
524769	BMS	Grt-amphibolite	66.3884	-37.4274	28811			x			
524771	BMS	Amphibolite	66.3884	-37.4274	28812			x			
524790	PKA	Amphibolite	65.6672	-39.0915	28813						
524798	BMS	Amphibolite	65.6417	-38.4327	28814			x			
525105	JKOL	Metabasite	65.9019	-37.5527		x		x			
525115	JKOL	Amphibolite	65.6457	-37.4946	28877	x		x	x	x	
525117	JKOL	Gneiss	66.1228	-38.0143							
525118	JKOL	Gneiss	66.1216	-38.0141							
525124	JKOL	Amphibolite	65.8913	-37.1180	28878						
525125	JKOL	Amphibolite	65.8913	-37.1180	28879			x			
525126	JKOL	Amphibolite	65.8913	-37.1180	28880						
525129	JKOL	Amphibolite	65.8913	-37.1180	28881			x			

Sample	Collector	Rock Type	GPS		Section No.	Chemical analysis			PT		U-Pb dating
			Latitude	Longitude		EPMA	QS	XRF Major	TC	Conv.	
525158	JKOL	Amphibolite	65.6940	-38.6598	28815			x			
525164	JKOL	Grt-amphibolite	65.6807	-38.5371	28816			x			
525218	KT	Grt-bearing gneiss	65.7900	-37.2691	28867						
525224	KT	retrogressed eclogite	65.8491	-37.8189	28868	x	x	x	x	x	x
525225a	KT	retrogressed eclogite	65.8340	-37.8274	28869	x		x	x	x	
525225b	KT	retrogressed eclogite	65.8340	-37.8274	28870	x		x	x	x	
525231	KT	Orthogneiss, Grt-bearing	65.8914	-37.1186	28871						
525232	KT	Amphibolite	65.9059	-37.1268	28872	x		x	x	x	
525239	KT	Grt-bearing felsic schist	65.8841	-37.0136	28873						
525250	KT	Eclogite	65.8197	-36.8027	28874	x	x	x		x	
525251	KT	Eclogite	65.7886	-36.8025	28875	x		x	x	x	
525253	KT	Grt-Qz-Bt-Am schist	65.7912	-36.7960	28876						
Collected 2014:											
566201	ADZ	Grt-amphibolite	66.4160	-38.1820	29218	x		x	x	x	
566202	ADZ	Garnetite schist	66.4176	-38.1797							
566203	ADZ	Grt-Hbl-Pl-Qz gneiss	66.4197	-38.1762							
566204	ADZ	Sheared amphibolite	66.4207	-38.1810	566204			x			
566205	ADZ	Cpx-rich layer	66.4306	-38.1833	566205			x			
566206	ADZ	Grt-amphibolite	66.4306	-38.1833	29219	x		x	x	x	
566207	ADZ	Grt-amphibolite	66.4306	-38.1833	29220			x			
566208	ADZ	Grt-amphibolite	66.4316	-38.1828							
566209	ADZ	Pyroxenite	66.4320	-38.1923							
566210	ADZ	Layered amphibolite	66.4295	-38.1977	29221			x			
566211	ADZ	Garbenschiefer ultramafite	65.9733	-37.9984							
566212	ADZ	Kyanite	65.9730	-37.9980							
566213	ADZ	Garnet	65.9730	-37.9980							
566214	ADZ	Grt-bearing gneiss	65.9730	-37.9980							

Sample	Collector	Rock Type	GPS		Section No.	Chemical analysis			PT		U-Pb dating
			Latitude	Longitude		EPMA	QS	XRF Major	TC	Conv.	
566215	ADZ	Grt-bearing gneiss	65.9730	-37.9980							
566216	ADZ	Eclogitic gravel	65.9735	-37.9957	566216	x	x	x	x	x	x
566217	ADZ	Eclogitic gravel	65.9735	-37.9957	566217			x			
566218	ADZ	Eclogitic gravel	65.9735	-37.9957	566218	x		x	x	x	x
566219	ADZ	Pegmatite next to gravel	65.9735	-37.9957							
566220	ADZ	Basic boudin with eclogitic patches	65.9734	-37.9943	29222			x			
566221	ADZ	Basic boudin with eclogitic patches	65.9734	-37.9943							
566222	ADZ	Amphibolite with retrogressed eclogite	65.9738	-38.0048	29223			x			
566223	ADZ	Pink amphibolite	65.9738	-38.0048	566223	x		x	x	x	
566224	ADZ	Grt-amphibolite	65.9735	-38.0094	566224			x	x		
566225	ADZ	Weathered Grt-amphibolite	65.9735	-38.0102	29224			x			
566226	ADZ	Grt-amphibolite with green mineral	65.9735	-38.0102							
566227	ADZ	Weathered Grt-amphibolite with pyroxene	65.9726	-37.9889							
566228	ADZ	Weathered Grt-amphibolite	65.9726	-37.9889	29225	x		x	x	x	
566229	ADZ	Garnet with inclusions	65.9726	-37.9889							
566230	ADZ	Green orthoamphibole-rich schist	65.9726	-37.9877							
566231	ADZ	Red-green Grt- and Am-rich schist	65.9725	-37.9853				x			
566232	ADZ	Red-green Grt- and Am-rich schist	65.9725	-37.9853							
566233	ADZ	Red-green Grt- and Am-rich schist	65.9725	-37.9853							
566234	ADZ	Pegmatite with blue mineral	65.9730	-37.9980							
566235	ADZ	Green boudin surrounded by gneiss	65.9720	-38.0075							
566236	ADZ	Amphibolite with retrogressed eclogite	65.9695	-38.0075							
566237	ADZ	Retrogressed eclogite	65.9699	-38.0073	566237			x			
566238	ADZ	Retrogressed eclogite	65.9699	-38.0073	566238	x		x		x	
566239	ADZ	Retrogressed eclogite	65.9699	-38.0073							
566240	ADZ	Zebra-striped retrogressed eclogite	65.9812	-37.9965	566240	x		x			x
566241	ADZ	Diorite	65.9626	-37.9828							

Sample	Collector	Rock Type	GPS		Section No.	Chemical analysis			PT		
			Latitude	Longitude		EPMA	QS	XRF Major	TC	Conv.	U-Pb dating
566242	ADZ	Rusty colored Grt-gneiss	65.9600	-37.9813							
566243	ADZ	Ultramafic rock	65.9600	-37.9813							
566244	ADZ	Grt-gneiss with sillimanite	65.9680	-37.9980							
566245	ADZ	Grt-gneiss	65.9721	-37.9948							
566246	ADZ	Grt-gneiss	65.9721	-37.9948							
566247	ADZ	Grt-gneiss	65.9730	-37.9980							
566248	ADZ	Retrogressed eclogite, fine-grained	65.8387	-36.8484	566248			x			
566249	ADZ	Retrogressed eclogite, unsheared	65.8387	-36.8484	566249	x		x	x	x	x
566250	ADZ	Retrogressed eclogite	65.8387	-36.8484	566250			x			
566251	ADZ	Retrogressed eclogite	65.8408	-36.8564	566251			x			
566252	ADZ	Eclogitic boudin	65.8396	-36.8570	29226			x	x		
566253	ADZ	Eclogitic boudin	65.8396	-36.8570							
566254	ADZ	Ky-Grt schist	65.8156	-36.8393							
566255	ADZ	Grt-amphibolite	65.8156	-36.8393							
566256	ADZ	Grt-amphibolite	65.8297	-36.8118							
566257	ADZ	Grt-amphibolite	65.8299	-36.8087	566257			x			
566258	ADZ	Grt-amphibolite	65.8299	-36.8087	566258			x			
566259	ADZ	Retrogressed eclogite	65.8291	-36.8014	566259			x			
566260	ADZ	Retrogressed eclogite	65.8291	-36.8014	566260			x			
566261	ADZ	Eclogitic gravel	65.8284	-36.7979							
566262	ADZ	Retrogressed eclogite	65.8277	-36.7934	566262			x			
566263	ADZ	Retrogressed eclogite	65.8277	-36.7925	566263	x		x	x	x	
566264	ADZ	Amphibolite lense	65.8299	-36.8066	566264			x	x		
566265	ADZ	Grt-Ky-Bt-Qz schist	65.9335	-37.2103							
566266	ADZ	River outcrop with rusty gneiss	65.9269	-37.2238							
566267	ADZ	Grt-Ky Gneiss	65.9534	-37.1907	29227	x		x	x	x	x
566268	ADZ	Grt-Ky Gneiss	65.9524	-37.1921							

Sample	Collector	Rock Type	GPS		Section No.	Chemical analysis			PT		
			Latitude	Longitude		EPMA	QS	XRF Major	TC	Conv.	U-Pb dating
566269	ADZ	Eclogitic gravel	65.9475	-37.1974	566269	x		x		x	
566270	ADZ	Eclogitic gravel	65.9475	-37.1974	566270			x			
566271	ADZ	Eclogitic gravel	65.9475	-37.1974	566271	x		x	x	x	
566272	ADZ	Diorite	65.9394	-37.3203							
566273	ADZ	Grt-pyroxenite	65.9403	-37.3175	29172	x		x	x	x	
566274	ADZ	Granite	65.9462	-37.3219							
566275	ADZ	Retrogressed eclogite	65.9800	-38.1082	566275	x		x		x	
566276	ADZ	Retrogressed eclogite	65.9800	-38.1082	566276			x			
566277	ADZ	Retrogressed eclogite	65.9800	-38.1082	566277	x	x	x	x	x	x
566278	ADZ	Retrogressed eclogite	65.9800	-38.1082	566278			x			
566279	ADZ	Eclogite	65.9808	-38.1085	29228	x		x	x	x	
566280	ADZ	Eclogite	65.9808	-38.1085	29229			x			
566281	ADZ	Magnetite-rich rock	65.9808	-38.1085	29230			x			
566282	ADZ	Magnetite-rich rock	65.9808	-38.1085							
566283	ADZ	Eclogite	65.9808	-38.1085	29231			x			
566284	ADZ	Siderite-rich rock	66.1664	-37.4585							
566285	ADZ	Rust zone	66.1664	-37.4585							
566286	ADZ	Laminated quartz	66.1664	-37.4585							
566287	ADZ	Green carbonate	66.1664	-37.4585							
566288	ADZ	Rust zone	66.1649	-37.4532							
566289	ADZ	Sillimanite schist	66.1635	-37.4349							
566290	ADZ	Grt-amphibolite	66.2193	-37.1660							
566291	ADZ	Grt-amphibolite with pyroxene	66.2193	-37.1660							
566292	ADZ	Pyroxene-Grt rock	66.2193	-37.1660	566292			x			
566293	ADZ	Grt-amphibolite with pyroxene	66.2214	-37.1584	566293			x			
566294	ADZ	Rusty gneiss	66.2176	-37.1281							
566295	ADZ	Rusty gneiss	66.2176	-37.1281							

Sample	Collector	Rock Type	GPS		Section No.	Chemical analysis			PT		
			Latitude	Longitude		EPMA	QS	XRF Major	TC	Conv.	U-Pb dating
566296	ADZ	Amphibolite sequence within metasediments	66.2194	-37.1242							
566297	ADZ	Grt-gneiss	66.2067	-37.1266							
566298	ADZ	Mylonitic amphibolite	66.2078	-37.1265							
566299	ADZ	Pegmatite	66.2089	-37.1203							
566301	ADZ	Mineral sample, Carbonate boudin	66.2089	-37.1203							
566302	ADZ	Granulite	66.3057	-37.0867	29232			x			
566303	ADZ	Granulite	66.3057	-37.0867							
566304	ADZ	Granulite	66.3057	-37.0867							
566305	ADZ	Garnet gneiss	65.6742	-37.1052							
566306	ADZ	Amphibolite at contact to dunite	65.6765	-37.1079							
566307	ADZ	Ultramafite	65.6803	-37.1085							
566308	ADZ	Gabbroic lens in gneiss	65.6888	-37.1006							
566309	ADZ	Ultramafic body in gneiss	65.6888	-37.1006							
566310	ADZ	Grt-Sil gneiss	65.6841	-37.1002							
566311	ADZ	Fine-grained Grt-Sil gneiss	65.6920	-37.1433							
566312	ADZ	Grt-Sil gneiss	65.6920	-37.1433							
566313	ADZ	Dunite	65.6917	-37.1389							

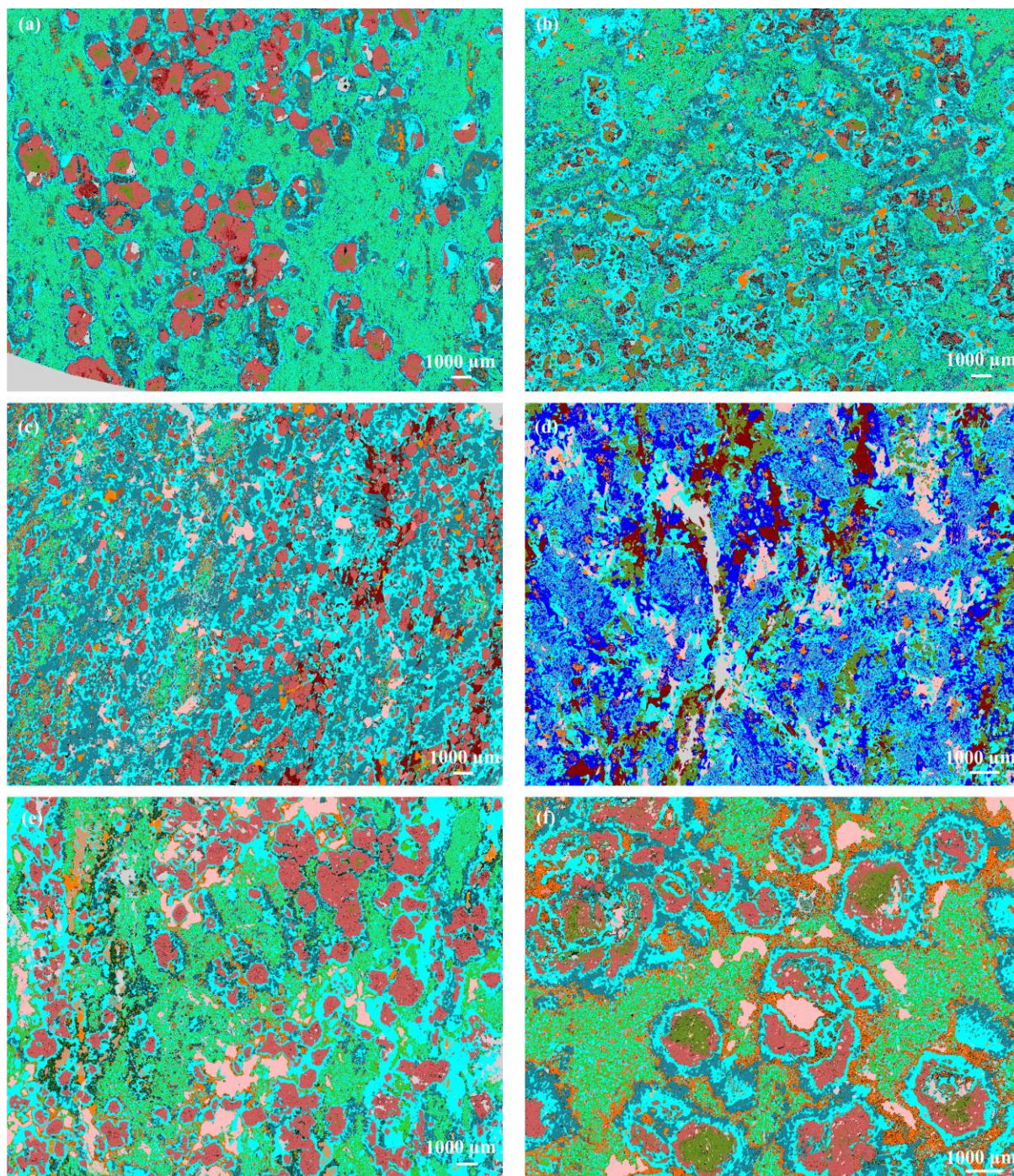
¹ Mineral abbreviations after Whitney and Evans (2010, Appendix 3). GPS = GPS-position of sample, Section No. = Thin section number, EPMA = Electron Probe Micro Analysis, QS = QEMSCAN, XRF = X-Ray Fluorescence, PT = Pressure-Temperature calculations, TC = THERMOCALC, Conv. = Conventional Geothermobarometry, ANU = Allen Nutman, BMS = Bo Møller Stensgaard, PKA = Per Kalvig, JKOL = Jochen Kolb, KT = Kristine Thrane, ADZ = Annika Dziggel.

Appendix 3 – List of mineral abbreviations (after Whitney and Evans, 2010)

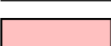
<u>Abbreviation</u>	<u>Mineral</u>	<u>Abbreviation</u>	<u>Mineral</u>	<u>Abbreviation</u>	<u>Mineral</u>
Cpx	Clinopyroxene	Am	Amphibole	Grt	Garnet
Opx	Orthopyroxene	Act	Actinolite	Alm	Almandine
Di	Diopside	Hbl	Hornblende	Grs	Grossular
Aug	Augite	Ath	Anthophyllite	Prp	Pyrope
Omp	Omphacite	Prg	Pargasite	Sps	Spessartine
Wo	Wollastonite	Ts	Tschermakite		
Jd	Jadeite	Hst	Hastingsite		
En	Enstatite	Ged	Gedrite		
Fs	Ferrosilite				

<u>Abbreviation</u>	<u>Mineral</u>	<u>Abbreviation</u>	<u>Mineral</u>	<u>Abbreviation</u>	<u>Mineral</u>
Pl	Plagioclase	Ilm	Ilmenite	Chl	Chlorite
An	Anorthite	Rt	Rutile	Bt	Biotite
Ab	Albite	Ttn	Titanite	Ms	Muscovite
Or	Orthoclase	Hem	Hematite	Zrn	Zircon
Qz	Quartz	Mag	Magnetite	Mz	Monazite
Ky	Kyanite	Fo	Forsterite	Liq	Melt
Sil	Sillimanite	Crd	Cordierite		
		Ep	Epidote		
		St	Staurolite		

Appendix 4 – QEMSCAN



QEMSCAN-images of the retrogressed eclogite samples 318349 (a), 318350 (b), 525224 (c), 525250 (d), 566216 (e) and 566277 (f).

	Garnet _b (high Mg)		Ilmenite
	Magnesio-Hornblende (525250)		Rutile
	Garnet _a (high Ca, others)		Titanite
	Diopside		Magnetite
	Augite-Omphacite (318349)		Biotite
	Ferrohornblende (others)		Background
	Orthopyroxene		Epidote
	Other pyroxene		Chlorite
	Magnesio-Hornblende		Apatite
	Hornblende high Si		Zircon
	Plagioclase		
	Quartz		

Legend for QEMSCAN-images.

Appendix 5 – Bulk-rock geochemistry

Bulk-rock major element geochemistry, used for pseudosection modelling (Chapter 3.7) was conducted on 67 samples (Appendix 2).

For each sample, a piece of about 20 g was cut from the hand specimen, avoiding leucosomes and quartz veins. The rock pieces were dried before being ground to a fine powder by crushing and milling in an agate mortar. Excess water from cutting was removed by drying the resulting powders at room temperature for 24h.

For bulk-rock major element analysis, 2-2.5 g of the powders are placed in a ceramic crucible, weighed and then dried overnight at ca. 105 °C. Subsequently, the dried sample is cooled till room temperature in a desiccator, weighed and then heated in a muffle furnace to 1000 °C for 2h. After cooling below 200 °C in the furnace, the heated sample is removed, cooled till room temperature and weighed again to determine the loss on ignition (LOI). The LOI is calculated from the difference in weight between the dried and heated sample powder (in g) and the difference in weight between the dried and heated sample powder divided by the amount of sample weighed in (in %). After weighing, 0.5 g of the heated powder were homogenized with 5 g of Dilithiumtetra-Lithiummetaborate (Fluxana), acting as a fluxing agent, in an agate mortar by hand. The homogenized mixture was placed in a platinum crucible and melted at 1150 °C to form a borate glass fusion pellet. Surface tension in the fusion pellets was achieved by inserting an ammonium iodide pill into the oven at the start of the smelting. The finished fusion pellets were analyzed using a SPECTRO X-LAB 2000 energy-dispersive spectrometer, after the method described in Kramm (2017).

Bulk-rock major element composition

sample	rock type	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Cr ₂ O ₃	SO ₃	LOI
318349	Eclogite	45.22	0.74	14.32	15.63	0.10	7.64	10.19	4.86	<0.20	<0.10	<0.20	0.23	-0.11
318350	Eclogite	45.91	2.77	11.19	17.04	0.24	6.32	9.25	5.10	<0.20	0.51	<0.20	0.23	-0.37
524713	Grt-amphibolite	44.43	1.53	13.20	16.48	0.24	8.94	10.72	2.16	0.38	<0.10	<0.20	0.24	1.00
524715	Eclogite	39.45	2.55	11.60	32.04	0.49	7.40	5.21	0.82	<0.20	<0.10	<0.20	0.26	-1.61
524716	Eclogite	48.25	1.22	15.27	16.12	0.23	4.96	9.95	1.80	1.00	<0.10	<0.20	0.27	0.30
524731	Amphibolite (medium-grained)	52.75	1.08	13.25	13.17	0.20	5.55	9.74	2.76	1.07	0.12	<0.20	<0.20	0.68
524769	Grt-amphibolite	50.71	0.95	14.29	13.73	0.19	6.31	9.83	2.49	0.39	<0.10	<0.20	0.26	0.50
524771	Amphibolite	50.97	1.13	6.87	32.28	1.11	3.05	2.34	0.41	<0.20	<0.10	<0.20	0.29	-1.53
524798	Amphibolite	55.74	0.76	13.73	9.98	0.15	5.67	9.35	2.84	0.84	0.11	<0.20	0.23	0.36
525105	Metabasite	50.44	1.16	12.33	15.81	0.25	5.81	11.57	2.12	0.38	<0.10	<0.20	<0.20	0.09
525115	Amphibolite	48.09	1.89	8.89	24.15	0.18	13.93	0.78	0.62	1.12	<0.10	<0.20	0.33	1.34
525125	Amphibolite	47.02	4.16	12.05	17.40	0.32	5.67	8.80	2.65	0.43	0.36	<0.20	0.26	0.92
525129	Amphibolite	44.33	1.64	13.44	18.37	0.53	4.67	11.37	2.10	0.80	0.16	<0.20	0.61	0.78
525143	Metabasite	51.30	1.21	12.41	14.70	0.24	5.59	9.83	2.49	0.61	0.13	<0.20	0.25	-0.06
525158	Amphibolite	48.71	1.33	13.23	15.92	0.28	5.59	9.94	2.14	0.75	0.14	<0.20	<0.20	0.01
525164	Grt-amphibolite	50.68	1.52	12.31	12.98	0.18	8.25	8.89	2.43	0.43	0.11	<0.20	0.23	0.85
525224	retrogressed eclogite	48.03	1.47	15.85	13.80	0.18	6.18	8.87	2.75	1.00	0.25	<0.20	0.23	0.00
525225a	retrogressed eclogite	46.37	0.94	13.70	15.78	0.34	7.42	10.73	2.37	0.72	<0.10	<0.20	0.25	0.58
525225b	retrogressed eclogite	48.61	0.98	13.76	12.75	0.16	7.13	11.62	2.50	0.66	<0.10	<0.20	0.23	0.47
525232	Amphibolite	50.85	0.91	12.36	13.26	0.21	7.06	11.83	2.29	0.61	<0.10	<0.20	<0.20	0.44
525250	Eclogite	50.42	2.38	12.29	16.79	0.22	4.23	8.24	2.88	1.25	0.25	<0.20	0.23	-0.36
525251	Eclogite	49.06	2.60	11.44	17.87	0.20	4.80	8.47	2.51	0.82	0.28	<0.20	0.23	-0.56
566201	Grt-amphibolite	42.17	1.08	13.90	16.28	0.26	10.17	11.81	2.76	0.55	0.27	<0.20	<0.20	0.44
566204	Sheared amphibolite	45.27	1.20	11.30	11.07	0.20	7.35	19.31	2.35	<0.20	<0.10	<0.20	<0.20	-0.08
566205	Cpx-rich layer	47.47	1.10	8.95	11.76	0.22	7.98	17.44	1.58	0.79	0.15	<0.20	0.27	1.70
566206	Grt-amphibolite	51.34	0.69	16.11	11.28	0.25	7.89	11.04	1.00	0.60	<0.10	<0.20	<0.20	0.25
566207	Grt-amphibolite	50.72	1.08	14.68	10.75	0.13	7.22	10.26	3.08	0.63	0.13	<0.20	<0.20	0.46

sample	rock type	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Cr ₂ O ₃	SO ₃	LOI
566210	Layered amphibolite	51.68	1.70	13.14	19.32	0.26	3.66	8.75	0.67	<0.20	<0.10	<0.20	<0.20	-1.32
566216	Eclogitic gravel	50.21	1.37	15.50	10.63	0.31	7.48	10.95	3.34	0.24	0.12	<0.20	<0.20	-0.08
566217	Eclogitic gravel	49.48	0.78	16.06	9.52	0.20	7.18	11.76	3.17	0.21	0.13	<0.20	<0.20	0.22
566218	Eclogitic gravel	46.03	0.48	21.28	7.85	0.09	6.78	12.70	2.83	<0.20	0.20	<0.20	<0.20	0.17
566220	Basic boudin with eclogitic patches	51.22	0.67	10.78	16.23	0.22	10.74	6.26	1.60	0.71	0.25	<0.20	<0.20	0.72
566222	Amphibolite with retrogressed eclogite	49.24	0.49	14.76	11.35	0.20	8.81	12.07	1.81	<0.20	<0.10	<0.20	<0.20	0.52
566223	Pink amphibolite	52.61	0.57	15.17	13.18	0.23	8.05	7.64	1.86	0.48	<0.10	<0.20	<0.20	0.04
566224	Grt-amphibolite	48.66	1.09	10.47	11.37	0.19	9.92	14.41	2.01	0.27	0.14	<0.20	<0.20	0.38
566225	Weathered Grt-amphibolite	45.86	1.05	8.08	13.02	0.24	18.62	11.44	1.07	<0.20	<0.10	0.27	<0.20	2.65
566228	Weathered Grt-amphibolite	50.62	1.71	8.30	11.94	0.14	14.88	10.90	0.63	0.29	<0.10	<0.20	<0.20	0.44
566231	Red-green Grt- and Am-rich schists	46.15	0.10	13.89	13.27	0.22	15.48	7.03	2.21	<0.20	<0.10	0.30	<0.20	0.65
566237	retrogressed eclogite	42.25	0.36	13.99	12.51	0.22	8.21	15.59	1.65	0.85	<0.10	<0.20	<0.20	2.69
566238	retrogressed eclogite	46.32	0.38	13.36	12.40	0.15	9.55	13.87	2.71	<0.20	<0.10	<0.20	<0.20	0.34
566240	zebra striped retrogressed eclogite	47.45	1.25	15.63	13.40	0.21	8.10	10.50	2.70	0.24	0.11	<0.20	<0.20	-0.21
566248	retrogressed eclogite, fine-grained	49.77	0.84	12.84	15.18	0.23	7.24	10.33	3.11	<0.20	<0.10	<0.20	<0.20	-0.09
566249	retrogressed eclogite, unsheared	47.87	3.43	11.47	18.39	0.27	4.55	9.43	3.28	0.71	0.21	<0.20	<0.20	-0.61
566250	retrogressed eclogite	48.69	4.05	10.95	18.67	0.28	3.97	9.19	2.85	0.33	0.45	<0.20	<0.20	-0.84
566251	retrogressed eclogite	48.74	3.06	11.74	16.67	0.20	4.84	10.03	3.54	0.56	0.27	<0.20	<0.20	-0.46
566252	Eclogitic boudin	44.08	1.00	14.93	16.99	0.51	4.58	16.47	0.60	<0.20	<0.10	<0.20	<0.20	1.38
566257	Grt-amphibolite	49.85	2.46	11.89	16.76	0.26	4.67	8.83	3.56	1.11	0.18	<0.20	<0.20	-0.44
566258	Grt-amphibolite	50.19	2.49	12.22	16.75	0.24	4.48	8.64	3.39	0.99	0.16	<0.20	<0.20	-0.27
566259	retrogressed eclogite	48.18	2.44	14.23	16.68	0.40	3.78	9.02	4.24	0.87	0.23	<0.20	<0.20	-0.50
566260	retrogressed eclogite	49.90	2.27	13.70	13.92	0.21	4.02	8.34	4.74	1.32	0.14	<0.20	<0.20	-0.21
566262	retrogressed eclogite	50.04	2.63	11.52	17.72	0.25	4.24	8.42	3.15	0.88	0.23	<0.20	<0.20	-0.54
566263	retrogressed eclogite	57.41	1.75	11.75	16.51	0.40	1.77	8.27	2.10	0.59	0.65	<0.20	<0.20	-1.07
566264	Amphibolite lense	50.97	2.56	11.71	17.07	0.30	3.78	8.12	3.50	0.64	0.25	<0.20	<0.20	-0.20
566267	Grt-Ky Gneiss	55.20	0.74	26.15	9.80	0.21	1.88	0.25	0.37	2.79	<0.10	<0.20	<0.20	0.85

sample	rock type	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Cr ₂ O ₃	SO ₃	LOI
566269	Eclogitic gravel	47.86	0.92	13.12	14.83	0.37	6.81	13.53	3.13	<0.20	<0.10	<0.20	<0.20	-0.61
566270	Eclogitic gravel	51.16	1.09	12.76	14.32	0.23	6.67	11.06	0.95	0.33	<0.10	<0.20	<0.20	0.12
566271	Eclogitic gravel	50.05	1.20	12.68	15.74	0.23	6.22	10.42	3.23	0.24	<0.10	<0.20	<0.20	-0.28
566273	Grt-pyroxenite	50.92	0.13	2.34	9.38	0.27	17.33	15.75	0.81	<0.20	<0.10	0.40	<0.20	-0.07
566275	retrogressed eclogite	45.40	0.93	12.40	14.81	0.25	7.87	12.67	0.72	0.35	0.14	<0.20	<0.20	0.22
566276	retrogressed eclogite	50.02	0.89	13.38	13.31	0.22	7.53	10.95	3.04	<0.20	<0.10	<0.20	<0.20	-0.28
566277	retrogressed eclogite	50.34	0.87	12.80	13.24	0.20	7.54	11.07	3.00	<0.20	<0.10	<0.20	<0.20	-0.30
566278	retrogressed eclogite	48.78	0.50	13.17	11.42	0.17	9.41	12.45	3.17	0.24	<0.10	<0.20	<0.20	0.17
566279	Eclogite	53.59	0.21	1.15	8.18	0.22	18.42	17.14	0.74	<0.20	<0.10	0.39	<0.20	0.60
566280	Eclogite	45.48	1.75	11.13	13.05	0.47	11.72	12.27	1.69	0.30	0.29	<0.20	<0.20	0.33
566281	Magnetite-rich rock	51.64	0.19	0.84	9.82	0.19	19.89	13.27	0.30	<0.20	<0.10	0.42	<0.20	0.55
566283	Eclogite	38.43	1.74	12.67	31.32	0.42	8.08	4.82	0.76	<0.20	<0.10	<0.20	<0.20	-0.73
566292	Pyroxene-Grt rock	46.39	0.32	12.44	11.05	0.20	7.50	18.89	1.10	<0.20	<0.10	0.27	<0.20	1.80
566293	Grt-Amphibolite with pyroxene	46.23	0.44	12.20	15.83	0.25	5.97	15.90	2.15	<0.20	<0.10	<0.20	<0.20	0.16
566302	Granulite	49.52	1.54	15.16	13.41	0.20	4.39	9.39	3.81	0.90	0.35	<0.20	<0.20	0.17

Oxides in wt%, Loss on ignition (LOI) in %.

Appendix 6 – Zircon U-Pb data (GUF)

Date	Sample	Grain	Meas. No	²⁰⁷ Pb ^a (cps)	U ^b (ppm)	Pb ^b (ppm)	Th ^b U	²⁰⁶ Pb ^c (%)	²⁰⁶ Pb ^d ²³⁸ U	±2s ^e (%)	²⁰⁷ Pb ^d ²³⁵ U	±2s ^e (%)	²⁰⁷ Pb ^d ²⁰⁶ Pb	±2s ^e (%)	rho ^f	²⁰⁶ Pb ²³⁸ U	±2s (Ma)	²⁰⁷ Pb ²³⁵ U	±2s (Ma)	²⁰⁷ Pb ²⁰⁶ Pb	±2s (Ma)	conc. ^g (%)
26.01.16	525224	Zr2	A160	13951	136	42.8	0.985	bdl	0.304	2.4	4.597	2.7	0.1097	1.2	0.90	1711	36	1749	23	1794	22	95
26.01.16	525224	Zr4	A169	2401	31	10.5	0.162	0.28	0.3335	1.7	5.079	2.9	0.1105	2.4	0.58	1855	27	1833	25	1807	43	103
26.01.16	525224	Zr4	A170	1732	21	7.3	0.144	bdl	0.3392	2.1	5.423	3.1	0.1116	2.3	0.69	1883	35	1888	27	1895	41	99
26.01.16	525224	Zr4	A171	1411	18	6.3	0.184	bdl	0.3394	1.9	5.405	3.0	0.1155	2.2	0.66	1884	32	1886	26	1888	40	100
26.01.16	525224	Zr5	A174	3222	22	7.4	0.121	2.75	0.318	2.9	4.717	3.8	0.1076	2.5	0.75	1780	45	1770	32	1759	46	101
26.01.16	525224	Zr5	A175	2972	30	10.7	0.299	0.45	0.3377	2.3	5.427	3.0	0.1166	2	0.75	1875	37	1889	26	1905	36	98
26.01.16	525224	Zr6	A172	13480	128	40.1	2.355	1.74	0.3055	1.9	4.468	5.2	0.1061	4.9	0.36	1718	28	1725	44	1734	90	99
26.01.16	525224	Zr6	A173	28096	366	128.8	1.942	1.77	0.3393	2.2	5.447	2.5	0.1165	1.2	0.87	1883	36	1892	22	1903	22	99
26.01.16	525224	Zr9	A176	1931	23	8.1	0.083	1.19	0.345	2.7	5.585	4.0	0.1174	3	0.68	1911	45	1914	35	1918	53	100
26.01.16	525224	Zr11	A179	11910	122	41.8	1.450	1.52	0.3305	1.4	5.24	2.1	0.115	1.5	0.68	1841	23	1859	18	1880	28	98
26.01.16	525224	Zr12	A167	1550	18	6.4	0.010	0.38	0.3389	2.0	5.447	3.4	0.1166	2.7	0.58	1881	32	1892	29	1905	49	99
26.01.16	525224	Zr12	A168	1506	18	6.5	0.026	0.42	0.3452	1.9	5.516	3.4	0.1159	2.8	0.55	1912	31	1903	30	1894	51	101
26.01.16	525224	Zr13	A166	4120	32	10.7	1.270	2.75	0.3204	3.0	4.874	4.2	0.1104	3	0.71	1792	47	1798	36	1805	54	99
26.01.16	525224	Zr15	A162	2350	24	7.6	0.328	0.33	0.3102	2.2	4.566	2.8	0.1068	1.8	0.79	1742	34	1743	24	1746	32	100
26.01.16	525224	Zr16	A182	2199	25	8.8	0.096	bdl	0.3391	1.9	5.401	2.7	0.1156	1.8	0.73	1882	32	1885	23	1889	33	100
26.01.16	525224	Zr16	A183	1990	22	7.6	0.013	0.57	0.3273	2.1	5.15	3.1	0.1141	2.3	0.67	1826	33	1844	27	1866	42	98
26.01.16	525224	Zr16	A184	1895	22	7.7	0.103	0.15	0.3445	1.7	5.543	3.4	0.1167	2.9	0.51	1908	29	1907	30	1907	53	100
26.01.16	525224	Zr18	A180	1459	17	5.9	0.033	bdl	0.3269	2.2	5.105	3.7	0.1133	3	0.59	1823	35	1837	32	1853	53	98
26.01.16	525224	Zr18	A181	3595	32	9.8	0.277	0.75	0.302	2.6	4.596	3.8	0.1104	2.8	0.69	1701	40	1749	32	1806	50	94
26.01.16	525224	Zr19	A177	13800	179	60.9	1.248	bdl	0.3283	1.7	5.273	2.1	0.1165	1.2	0.81	1830	28	1865	18	1903	22	96
26.01.16	525224	Zr19	A178	34869	386	129.9	1.413	0.01	0.324	1.8	5.188	2.0	0.1162	0.95	0.88	1809	29	1851	18	1898	17	95
26.01.16	525224	Zr21	A185	5885	63	22.3	0.188	bdl	0.3402	1.5	5.407	2.2	0.1153	1.6	0.70	1888	25	1886	19	1885	28	100
26.01.16	525224	Zr22	A188	2082	21	7.5	0.070	1.25	0.3393	2.0	5.46	4.4	0.1167	4	0.45	1883	32	1894	39	1907	71	99
26.01.16	525224	Zr23	A189	1654	19	6.7	0.088	0.68	0.3369	3.0	5.322	4.2	0.1146	3	0.70	1872	48	1872	37	1874	54	100
25.01.16	566216	Zr1	A31	1925	47	17	0.82	bdl	0.3418	2.0	5.976	3.6	0.1268	2.9	0.57	1895	34	1972	32	2055	52	92

Date	Sample	Grain	Meas. No	²⁰⁷ Pb ^a (cps)	U ^b (ppm)	Pb ^b (ppm)	Th ^b U	²⁰⁶ Pbc ^c (%)	²⁰⁶ Pb ^d ²³⁸ U	±2s ^e (%)	²⁰⁷ Pb ^d ²³⁵ U	±2s ^e (%)	²⁰⁷ Pb ^d ²⁰⁶ Pb	±2s ^e (%)	rho ^f	²⁰⁶ Pb ²³⁸ U	±2s (Ma)	²⁰⁷ Pb ²³⁵ U	±2s (Ma)	²⁰⁷ Pb ²⁰⁶ Pb	±2s (Ma)	conc. ^g (%)
25.01.16	566216	Zr3	A73	475	12	4.6	0.162	1.12	0.3569	2.8	5.953	3.6	0.121	2.3	0.78	1968	48	1969	32	1971	40	100
25.01.16	566216	Zr5	A25	72	1.8	0.6	1.21	bdl	0.309	6.5	5.108	8.4	0.12	5.3	0.78	1736	100	1838	74	1956	94	89
25.01.16	566216	Zr6	A24	25325	414	171	6.50	2.0	0.3938	1.6	7.121	3.3	0.1312	2.9	0.49	2140	30	2127	30	2114	51	101
25.01.16	566216	Zr7	A52	5477	126	49.9	0.772	0.38	0.3769	2.0	6.728	2.8	0.1295	2	0.72	2062	36	2076	25	2091	34	99
25.01.16	566216	Zr8	A51	8631	200	72.0	0.843	0.58	0.3455	1.9	5.717	2.4	0.1201	1.6	0.77	1913	31	1934	21	1957	28	98
25.01.16	566216	Zr9	A30	21155	427	171	1.83	0.60	0.3822	1.6	6.786	2.0	0.1288	1.3	0.79	2087	29	2084	18	2082	22	100
25.01.16	566216	Zr10	A17	243	6	2.3	0.48	bdl	0.3757	4.5	6.667	5.5	0.1287	3.1	0.83	2056	80	2068	50	2081	55	99
25.01.16	566218	Zr1	A250	10977	375	148.5	1.65	0.74	0.3776	2.0	6.729	2.3	0.1293	1.2	0.85	2065	35	2076	21	2088	22	99
25.01.16	566218	Zr3	A248	4901	146	57.0	1.17	bdl	0.3733	1.8	6.505	2.5	0.1264	1.7	0.73	2045	32	2047	22	2049	30	100
25.01.16	566240	Zr1	A138	312	9.5	3.4	0.052	bdl	0.345	3.1	5.739	6.2	0.1207	5.4	0.49	1911	51	1937	55	1966	97	97
25.01.16	566240	Zr2	A141	300	9	3.0	0.34	bdl	0.3417	4.4	5.649	7.5	0.1199	6	0.59	1895	72	1924	67	1955	108	97
25.01.16	566240	Zr2	A142	319	9	3.3	0.91	0.24	0.3442	3.4	5.842	4.8	0.1231	3.4	0.70	1907	56	1953	43	2002	61	95
25.01.16	566240	Zr2	A143	10727	362	123.1	0.72	0.47	0.3281	1.7	5.268	2.2	0.1165	1.3	0.81	1829	28	1864	19	1903	23	96
25.01.16	566240	Zr3	A132	15078	347	129.0	6.78	0.05	0.3547	2.0	6.284	2.3	0.1285	1	0.89	1957	35	2016	20	2078	18	94
25.01.16	566240	Zr3	A133	481	15	6.0	2.36	bdl	0.3843	2.1	7.076	4.2	0.1336	3.6	0.50	2096	38	2121	38	2146	63	98
25.01.16	566240	Zr4	A134	112	2.3	0.9	0.004	1.56	0.3858	6.2	6.739	6.9	0.1267	3	0.90	2103	113	2078	63	2053	54	102
25.01.16	566240	Zr4	A135	82	2.3	0.8	0.002	0.78	0.3532	6.9	5.948	8.1	0.1222	4.2	0.85	1950	118	1968	73	1988	75	98
25.01.16	566240	Zr6	A123	16144	425	160	1.24	bdl	0.3608	1.6	6.096	1.7	0.1226	0.68	0.92	1986	27	1990	15	1994	12	100
25.01.16	566240	Zr7	A122	558	17	5.8	0.49	2.27	0.3336	3.1	5.206	3.4	0.1132	1.5	0.91	1856	50	1854	30	1852	26	100
26.01.16	566267	Mz19	A248	8990	106	37.2	0.906	6.82	0.3387	4.4	5.561	10.2	0.1191	9.2	0.44	1880	73	1910	92	1943	164	97
26.01.16	566267	Mz19	A249	11806	119	63.0	0.599	1.75	0.4834	2.6	11.86	4.6	0.1779	3.8	0.57	2542	56	2593	44	2634	63	97
26.01.16	566267	Zr1	A209	87833	482	150.9	0.414	0.41	0.2873	1.8	7.044	2.1	0.1779	1.1	0.85	1628	26	2117	19	2633	19	62
26.01.16	566267	Zr1	A210	58081	279	128.0	0.318	0.86	0.4236	4.7	9.805	4.8	0.1679	0.81	0.99	2277	91	2417	45	2537	14	90
26.01.16	566267	Zr4	A200	66825	682	134.9	0.176	8.33	0.1895	3.4	3.245	4.1	0.1242	2.3	0.83	1119	35	1468	32	2018	40	55
26.01.16	566267	Zr5	A201	39788	296	102.8	0.135	1.22	0.3292	1.5	6.283	1.8	0.1385	1.1	0.82	1834	24	2016	16	2208	18	83

Date	Sample	Grain	Meas. No	²⁰⁷ Pb ^a (cps)	U ^b (ppm)	Pb ^b (ppm)	Th ^b U	²⁰⁶ Pb ^c (%)	²⁰⁶ Pb ^d ²³⁸ U	±2s ^e (%)	²⁰⁷ Pb ^d ²³⁵ U	±2s ^e (%)	²⁰⁷ Pb ^d ²⁰⁶ Pb	±2s ^e (%)	rho ^f	²⁰⁶ Pb ²³⁸ U	±2s (Ma)	²⁰⁷ Pb ²³⁵ U	±2s (Ma)	²⁰⁷ Pb ²⁰⁶ Pb	±2s (Ma)	conc. ^g (%)
26.01.16	566267	Zr8	A207	90417	643	263.4	0.230	0.10	0.383	2.0	8.062	2.3	0.1527	1.2	0.85	2090	35	2238	21	2377	20	88
26.01.16	566267	Zr9	A205	81389	654	259.9	0.117	1.30	0.3783	1.5	6.813	1.9	0.1307	1.2	0.78	2068	26	2087	17	2107	21	98
26.01.16	566267	Zr10	A203	46414	348	124.8	0.020	0.45	0.3455	1.6	5.525	1.8	0.116	0.83	0.89	1913	27	1904	16	1896	15	101
26.01.16	566267	Zr10	A204	41274	401	142.1	0.039	0.53	0.3417	1.9	5.438	2.0	0.1155	0.8	0.92	1895	31	1891	17	1887	14	100

^a Within run background-corrected mean ²⁰⁷Pb signal in cps (counts per second).

^b U and Pb content and Th/U ratio were calculated relative to GJ-1 reference zircon.

^c percentage of the common Pb on the ²⁰⁶Pb. bdl = below detection limit.

^d corrected for background, within-run Pb/U fractionation (in case of ²⁰⁶Pb/²³⁸U) and common Pb using Stacy and Kramers (1975) model Pb composition and subsequently normalized to GJ-1 (ID-TIMS value/measured value); ²⁰⁷Pb/²³⁵U calculated using ²⁰⁷Pb/²⁰⁶Pb/(²³⁸U/²⁰⁶Pb*1/137.88)

^e ²⁰⁶Pb/²³⁸U error is the quadratic additions of the within run precision (2 SE) and the external reproducibility (2 SD) of the reference zircon. ²⁰⁷Pb/²⁰⁶Pb error propagation (²⁰⁷Pb signal dependent) following Gerdes & Zeh (2009). ²⁰⁷Pb/²³⁵U error is the quadratic addition of the ²⁰⁷Pb/²⁰⁶Pb and ²⁰⁶Pb/²³⁸U uncertainty.

^f rho is the ²⁰⁶Pb/²³⁸U/²⁰⁷Pb/²³⁵U error correlation coefficient.

^g degree of concordance = ²⁰⁶Pb/²³⁸U age / ²⁰⁷Pb/²⁰⁶Pb age x 100.

Appendix 7 – Monazite U-Pb data (GUF)

Date	Sample	Grain	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pbc ^c	²⁰⁶ Pbd ^d	±2s ^e	²⁰⁷ Pbd ^d	±2s ^e	²⁰⁷ Pbd ^d	±2s ^e	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g	²⁰⁸ Pb	±2s	²⁰⁸ Pb	±2s
			No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)	²³² Th	(%)	²³² Th	(Ma)
26.01.2016	566267	Mz1	A288	69211	1685	575.9	19.027	0.42	0.3305	1.3	5.115	1.6	0.1123	0.9	0.82	1841	21	1839	13	1837	16	100	0.0929	1.7	1796	30
26.01.2016	566267	Mz1	A289	110718	2500	881.7	15.676	0.29	0.3405	1.4	5.367	1.7	0.1144	0.93	0.83	1889	23	1880	14	1870	17	101	0.0973	1.5	1877	27
26.01.2016	566267	Mz2	A280	107786	2513	872.4	17.017	0.12	0.3352	1.4	5.285	1.6	0.1144	0.83	0.86	1864	22	1866	14	1870	15	100	0.0967	1.7	1865	30
26.01.2016	566267	Mz2	A281	107325	2254	782.0	17.401	0.36	0.3346	1.4	5.34	1.7	0.1158	0.9	0.85	1861	23	1875	15	1892	16	98	0.0965	1.7	1863	30
26.01.2016	566267	Mz2	A282	98146	2270	787.6	12.695	bdl	0.3347	1.4	5.333	1.6	0.1156	0.84	0.86	1861	23	1874	14	1889	15	98	0.0963	1.6	1859	28
26.01.2016	566267	Mz3	A277	87238	1981	696.3	15.182	bdl	0.3389	1.2	5.405	1.3	0.1157	0.6	0.89	1882	19	1886	11	1891	11	100	0.0981	1.6	1892	29
26.01.2016	566267	Mz3	A278	111041	2395	836.2	13.175	0.12	0.3369	1.3	5.333	1.5	0.1148	0.76	0.87	1872	21	1874	13	1877	14	100	0.0964	1.6	1861	29
26.01.2016	566267	Mz3	A279	101746	2424	850.5	12.890	0.09	0.3388	1.3	5.342	1.5	0.1144	0.8	0.84	1881	21	1876	13	1870	15	101	0.0968	1.6	1867	28
26.01.2016	566267	Mz4	A215	48780	1243	410.5	20.357	1.39	0.3186	1.5	5.06	2.7	0.1152	2.3	0.54	1783	23	1829	23	1883	41	95	0.0948	1.9	1830	33
26.01.2016	566267	Mz4	A216	100299	2732	948.1	13.788	0.11	0.3348	1.1	5.333	1.3	0.1156	0.61	0.88	1862	18	1874	11	1889	11	99	0.0967	1.4	1865	25
26.01.2016	566267	Mz4	A217	88292	2369	835.7	15.753	0.61	0.3401	1.2	5.431	1.3	0.1158	0.6	0.90	1887	20	1890	12	1893	11	100	0.0975	1.4	1881	25
26.01.2016	566267	Mz5	A275	94789	2214	777.7	14.672	bdl	0.339	1.5	5.388	1.7	0.1153	0.9	0.85	1882	24	1883	15	1885	16	100	0.0980	1.6	1890	29
26.01.2016	566267	Mz5	A276	121033	2710	952.7	10.329	0.27	0.3392	1.3	5.389	1.6	0.1153	0.93	0.80	1883	20	1883	13	1884	17	100	0.0980	1.5	1890	27
26.01.2016	566267	Mz6	A270	81225	1941	679.3	15.754	bdl	0.3376	1.4	5.369	1.7	0.1154	0.98	0.82	1875	23	1880	15	1885	18	99	0.0966	1.7	1864	31
26.01.2016	566267	Mz6	A271	103811	2504	876.4	13.611	bdl	0.3377	1.4	5.375	1.6	0.1155	0.93	0.83	1876	22	1881	14	1887	17	99	0.0974	1.5	1879	27
26.01.2016	566267	Mz9	A264	32602	738	261.4	38.564	0.16	0.3415	1.3	5.436	1.6	0.1155	1	0.77	1894	21	1890	14	1887	19	100	0.0975	1.5	1881	26
26.01.2016	566267	Mz9	A265	83095	1990	695.8	14.342	0.17	0.3375	1.4	5.351	1.7	0.115	0.96	0.82	1874	23	1877	15	1880	17	100	0.0980	1.6	1889	29
26.01.2016	566267	Mz9	A266	110091	2399	831.4	14.487	0.52	0.3347	1.3	5.274	1.6	0.1143	0.88	0.83	1861	22	1865	14	1869	16	100	0.0975	1.8	1880	33
26.01.2016	566267	Mz10	A261	39717	1010	354.4	28.335	0.23	0.3385	1.2	5.387	1.5	0.1154	0.93	0.80	1880	20	1883	13	1887	17	100	0.0971	1.6	1872	29
26.01.2016	566267	Mz10	A262	31151	748	263.7	40.629	bdl	0.34	1.3	5.42	1.6	0.1156	0.88	0.83	1887	22	1888	14	1890	16	100	0.0977	1.8	1885	32
26.01.2016	566267	Mz10	A263	39611	909	320.2	32.176	0.86	0.3401	1.3	5.376	1.6	0.1147	0.87	0.84	1887	22	1881	14	1875	16	101	0.0974	1.6	1879	28
26.01.2016	566267	Mz11	A272	157887	3707	1307.9	7.451	0.02	0.3403	1.4	5.426	1.6	0.1157	0.76	0.88	1888	23	1889	14	1890	14	100	0.0973	1.6	1877	29
26.01.2016	566267	Mz11	A273	105672	2456	856.4	9.593	0.29	0.3365	1.5	5.327	1.6	0.1148	0.75	0.89	1870	24	1873	14	1877	14	100	0.0970	1.6	1872	29
26.01.2016	566267	Mz11	A274	75960	1720	602.9	15.057	0.34	0.3382	1.3	5.38	1.5	0.1154	0.85	0.83	1878	21	1882	13	1886	15	100	0.0977	1.8	1883	33
26.01.2016	566267	Mz12	A267	94273	2348	809.4	14.153	0.10	0.3331	1.4	5.206	1.6	0.1134	0.83	0.86	1853	23	1854	14	1854	15	100	0.0900	1.7	1742	28
26.01.2016	566267	Mz13	A258	98457	2528	887.7	11.778	0.08	0.339	1.3	5.347	1.4	0.1144	0.65	0.89	1882	21	1876	12	1871	12	101	0.0979	1.5	1889	27

Date	Sample	Grain	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pbc ^c	²⁰⁶ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g	²⁰⁸ Pb	±2s	²⁰⁸ Pb	±2s
			No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)	²³² Th	(%)	²³² Th	(Ma)
26.01.2016	566267	Mz13	A259	70220	1536	538.5	20.370	bdl	0.3384	1.3	5.347	1.5	0.1146	0.74	0.86	1879	21	1876	13	1874	13	100	0.0979	1.7	1887	31
26.01.2016	566267	Mz13	A260	58046	1549	538.1	18.060	0.70	0.3352	1.4	5.328	1.5	0.1153	0.74	0.88	1864	22	1873	13	1885	13	99	0.0980	1.6	1889	28
26.01.2016	566267	Mz14	A243	115674	2798	984.2	18.177	0.06	0.3394	1.2	5.389	1.4	0.1152	0.62	0.89	1884	20	1883	12	1883	11	100	0.0976	1.6	1882	29
26.01.2016	566267	Mz14	A244	88982	2316	814.6	13.313	0.19	0.3394	1.4	5.397	1.5	0.1154	0.63	0.91	1884	23	1884	13	1886	11	100	0.0971	1.6	1872	29
26.01.2016	566267	Mz14	A245	57347	1225	426.5	23.857	0.33	0.336	1.4	5.345	1.8	0.1154	1.1	0.78	1867	23	1876	16	1886	20	99	0.0976	1.6	1882	29
26.01.2016	566267	Mz15	A235	32366	830	291.5	38.028	bdl	0.3387	1.5	5.392	2.5	0.1155	2.1	0.58	1880	24	1884	22	1887	37	100	0.0980	1.8	1890	33
26.01.2016	566267	Mz15	A236	100519	2576	903.9	13.962	0.26	0.3384	1.3	5.391	1.4	0.1156	0.6	0.91	1879	21	1883	12	1889	11	99	0.0980	1.5	1889	28
26.01.2016	566267	Mz15	A237	104607	2680	951.0	15.070	0.10	0.3424	1.4	5.447	1.5	0.1154	0.7	0.89	1898	22	1892	13	1887	13	101	0.0978	1.8	1887	33
26.01.2016	566267	Mz16	A255	89385	2082	733.8	15.528	bdl	0.3401	1.2	5.401	1.4	0.1152	0.74	0.85	1887	20	1885	12	1883	13	100	0.0978	1.6	1886	29
26.01.2016	566267	Mz16	A256	97576	2269	798.6	15.094	0.11	0.3396	1.4	5.397	1.7	0.1153	0.93	0.83	1885	23	1884	14	1885	17	100	0.0982	1.6	1893	29
26.01.2016	566267	Mz16	A257	98915	2330	812.2	12.745	bdl	0.3366	1.4	5.321	1.7	0.1147	0.89	0.85	1870	23	1872	14	1875	16	100	0.0983	1.5	1894	27
26.01.2016	566267	Mz17	A253	121871	2868	1001.3	13.278	bdl	0.3368	1.2	5.358	1.4	0.1154	0.71	0.86	1871	19	1878	12	1886	13	99	0.0978	1.5	1887	28
26.01.2016	566267	Mz17	A254	103100	2497	876.1	12.490	bdl	0.3384	1.3	5.399	1.5	0.1157	0.7	0.88	1879	21	1885	13	1891	13	99	0.0983	1.8	1895	32
26.01.2016	566267	Mz18	A250	32314	779	272.7	37.989	bdl	0.3377	1.4	5.419	2.3	0.1164	1.8	0.63	1875	24	1888	20	1902	32	99	0.0975	1.5	1881	27
26.01.2016	566267	Mz18	A251	128880	3052	1077.7	10.645	0.07	0.3407	1.2	5.417	1.4	0.1153	0.62	0.90	1890	20	1887	12	1885	11	100	0.0966	1.6	1864	28
26.01.2016	566267	Mz18	A252	121466	3006	1045.2	12.504	0.09	0.3355	1.3	5.32	1.4	0.1151	0.69	0.88	1865	21	1872	12	1881	12	99	0.0978	1.7	1885	30
26.01.2016	566267	Mz20	A232	110438	2825	985.3	12.130	0.77	0.3364	1.5	5.36	1.7	0.1156	0.9	0.85	1869	24	1879	15	1889	16	99	0.0978	1.8	1886	32
26.01.2016	566267	Mz20	A233	98792	2590	917.4	13.711	0.26	0.3416	1.3	5.456	1.6	0.1159	0.89	0.83	1895	22	1894	14	1893	16	100	0.0973	2.0	1877	36
26.01.2016	566267	Mz20	A234	90067	2173	765.5	16.703	0.12	0.3399	1.3	5.414	1.5	0.1156	0.67	0.89	1886	21	1887	13	1889	12	100	0.0975	1.6	1881	28
26.01.2016	566267	Mz21	A224	103111	2582	916.0	16.026	bdl	0.3422	1.3	5.452	1.5	0.1156	0.86	0.83	1897	21	1893	13	1889	15	100	0.0968	1.5	1867	27
26.01.2016	566267	Mz21	A225	109466	2762	979.0	17.698	0.05	0.3419	1.3	5.453	1.5	0.1157	0.75	0.87	1896	22	1893	13	1891	14	100	0.0971	1.7	1872	31
26.01.2016	566267	Mz21	A226	105206	2590	925.2	17.021	bdl	0.3445	1.2	5.494	1.5	0.1157	0.88	0.81	1908	20	1900	13	1891	16	101	0.0982	1.5	1893	27
26.01.2016	566267	Mz22	A218	105838	2850	1004.5	13.874	0.13	0.3401	1.3	5.405	1.4	0.1153	0.65	0.89	1887	21	1886	12	1885	12	100	0.0974	1.6	1879	29
26.01.2016	566267	Mz22	A219	122184	3035	1065.5	16.088	0.08	0.3385	1.3	5.417	1.6	0.1161	0.78	0.86	1880	22	1887	13	1897	14	99	0.0975	1.8	1880	32
26.01.2016	566267	Mz22	A220	102602	2663	931.4	13.737	0.07	0.3375	1.3	5.349	1.5	0.115	0.8	0.85	1875	21	1877	13	1880	14	100	0.0972	1.5	1875	27
26.01.2016	566267	Mz23	A230	113508	2911	1031.0	10.721	0.01	0.3416	1.3	5.448	1.5	0.1157	0.85	0.83	1895	21	1892	13	1891	15	100	0.0980	1.4	1890	26
26.01.2016	566267	Mz23	A231	107526	2759	969.5	10.973	0.30	0.339	1.3	5.396	1.4	0.1155	0.6	0.90	1882	21	1884	12	1887	11	100	0.0968	1.5	1868	26
26.01.2016	566267	Mz24	A223	99473	2558	904.7	11.438	0.24	0.3412	1.5	5.413	1.7	0.1151	0.78	0.89	1893	25	1887	15	1881	14	101	0.0981	1.8	1891	33

Date	Sample	Grain	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pb ^c	²⁰⁶ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g	²⁰⁸ Pb	±2s	²⁰⁸ Pb	±2s
			No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)	²³² Th	(%)	²³² Th	(Ma)
26.01.2016	566267	Mz25	A227	84976	2144	760.6	11.798	0.08	0.3422	1.2	5.439	1.4	0.1153	0.69	0.86	1897	19	1891	12	1885	12	101	0.0979	1.5	1888	26
26.01.2016	566267	Mz25	A228	91114	2263	795.8	10.502	0.39	0.3394	1.3	5.384	1.7	0.1151	1.1	0.79	1884	22	1882	15	1881	19	100	0.0978	1.5	1886	26
26.01.2016	566267	Mz25	A229	78453	2047	723.6	12.905	0.02	0.3409	1.4	5.437	1.8	0.1157	1.2	0.76	1891	22	1891	16	1891	21	100	0.0967	1.6	1866	29
26.01.2016	566267	Mz26	A246	139521	3346	1185.4	4.965	0.12	0.3419	1.2	5.429	1.4	0.1152	0.55	0.92	1896	21	1889	12	1883	10	101	0.0970	1.7	1872	31
26.01.2016	566267	Mz26	A247	93563	1924	682.9	9.588	2.06	0.3423	1.5	5.449	1.9	0.1155	1.3	0.75	1898	24	1893	17	1888	23	101	0.0974	1.9	1878	34

^a Within run background-corrected mean ²⁰⁷Pb signal in cps (counts per second).

^b U and Pb content and Th/U ratio were calculated relative to GJ-1 reference zircon.

^c percentage of the common Pb on the ²⁰⁶Pb. bdl = below detection limit.

^d corrected for background, within-run Pb/U fractionation (in case of ²⁰⁶Pb/²³⁸U) and common Pb using Stacy and Kramers (1975) model Pb composition and subsequently normalized to GJ-1 (ID-TIMS value/measured value); ²⁰⁷Pb/²³⁵U calculated using ²⁰⁷Pb/²⁰⁶Pb/(²³⁸U/²⁰⁶Pb*1/137.88)

^e ²⁰⁶Pb/²³⁸U error is the quadratic additions of the within run precision (2 SE) and the external reproducibility (2 SD) of the reference zircon. ²⁰⁷Pb/²⁰⁶Pb error propagation (²⁰⁷Pb signal dependent) following Gerdes & Zeh (2009). ²⁰⁷Pb/²³⁵U error is the quadratic addition of the ²⁰⁷Pb/²⁰⁶Pb and ²⁰⁶Pb/²³⁸U uncertainty.

^f rho is the ²⁰⁶Pb/²³⁸U/²⁰⁷Pb/²³⁵U error correlation coefficient.

^g degree of concordance = ²⁰⁶Pb/²³⁸U age / ²⁰⁷Pb/²⁰⁶Pb age x 100.

Appendix 8 – Titanite U-Pb data (GUF)

Date	Sample	Grain	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pb ^c	²⁰⁶ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g
			No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)
26.01.16	524713	Ttn1	A61	43467	384	132.3	1.196	bd1	0.3325	1.5	5.273	1.6	0.1150	0.7	0.90	1851	24	1864	14	1881	13	98
26.01.16	524713	Ttn1	A62	28524	246	87.0	1.023	bd1	0.3407	1.3	5.427	1.5	0.1155	0.8	0.84	1890	21	1889	13	1888	14	100
26.01.16	524713	Ttn2	A64	8721	53	17.5	0.502	11.68	0.3215	2.1	4.870	4.4	0.1099	3.9	0.47	1797	33	1797	38	1798	71	100
26.01.16	524713	Ttn2	A65	6698	50	17.5	0.529	2.33	0.3355	1.4	5.195	2.2	0.1123	1.7	0.65	1865	23	1852	19	1838	30	101
26.01.16	524713	Ttn3	A98	4713	29	10.0	0.387	2.14	0.3349	1.8	5.322	2.5	0.1153	1.8	0.72	1862	29	1872	22	1885	32	99
26.01.16	524713	Ttn3	A99	4961	31	11.1	0.498	1.37	0.3396	1.4	5.402	2.1	0.1154	1.5	0.68	1885	23	1885	18	1886	28	100
26.01.16	524713	Ttn3	A100	4612	29	10.1	0.483	1.77	0.3403	1.6	5.473	2.6	0.1167	2.0	0.62	1888	26	1896	23	1906	37	99
26.01.16	524713	Ttn3	A101	4753	30	10.7	0.429	1.53	0.3397	1.4	5.433	2.4	0.1160	1.9	0.58	1885	23	1890	21	1896	35	99
26.01.16	524713	Ttn4	A96	7778	54	19.1	0.328	0.82	0.3382	1.4	5.362	2.0	0.1150	1.5	0.69	1878	23	1879	17	1880	26	100
26.01.16	524713	Ttn4	A97	5304	34	12.0	0.464	1.70	0.3400	1.7	5.410	2.4	0.1154	1.6	0.73	1887	28	1886	20	1887	29	100
26.01.16	524713	Ttn5	A90	8291	51	17.9	0.547	1.92	0.3380	1.5	5.363	2.3	0.1151	1.8	0.65	1877	25	1879	20	1882	32	100
26.01.16	524713	Ttn5	A91	8993	63	20.8	0.430	1.78	0.3210	1.5	5.058	2.2	0.1143	1.6	0.70	1794	24	1829	19	1869	28	96
26.01.16	524713	Ttn5	A92	5365	37	12.6	0.554	1.22	0.3323	1.7	5.158	2.4	0.1126	1.6	0.72	1849	28	1846	20	1842	29	100
26.01.16	524713	Ttn5	A93	8050	54	19.2	0.387	1.25	0.3394	1.5	5.419	2.0	0.1158	1.3	0.74	1884	24	1888	17	1893	24	100
26.01.16	524713	Ttn5	A94	6041	40	14.0	0.557	1.15	0.3410	1.6	5.506	2.3	0.1171	1.6	0.69	1892	25	1901	20	1913	30	99
26.01.16	524713	Ttn5	A95	12258	86	29.0	0.315	1.46	0.3239	1.3	5.039	1.9	0.1129	1.3	0.72	1809	21	1826	16	1846	23	98
26.01.16	524713	Ttn6	A107	2057	11	3.9	0.413	10.30	0.3424	1.7	5.495	3.7	0.1164	3.3	0.46	1898	28	1900	32	1902	60	100
26.01.16	524713	Ttn6	A108	4606	29	10.1	0.581	2.04	0.3398	1.7	5.385	2.7	0.1150	2.0	0.64	1886	28	1882	23	1879	37	100
26.01.16	524713	Ttn7	A109	5219	34	11.3	0.439	1.54	0.3208	1.6	4.985	2.6	0.1128	2.0	0.63	1793	25	1817	22	1844	36	97
26.01.16	524713	Ttn7	A110	4139	25	8.8	0.590	1.65	0.3451	1.7	5.491	2.5	0.1155	1.9	0.67	1911	28	1899	22	1887	33	101
26.01.16	524713	Ttn7	A111	5184	32	11.5	0.556	1.18	0.3459	1.6	5.524	2.3	0.1159	1.6	0.70	1915	26	1904	20	1893	29	101
26.01.16	524713	Ttn7	A112	5618	34	11.9	0.532	0.96	0.3362	1.6	5.411	2.3	0.1168	1.6	0.72	1868	26	1887	19	1907	28	98
26.01.16	524713	Ttn8	A113	5522	33	11.2	0.676	2.23	0.3268	1.7	4.962	2.8	0.1102	2.2	0.61	1823	27	1813	24	1802	40	101
26.01.16	524713	Ttn8	A114	6791	45	15.2	0.557	1.01	0.3296	1.4	5.078	2.1	0.1118	1.6	0.65	1836	22	1832	18	1829	29	100
26.01.16	524713	Ttn8	A115	6700	41	14.6	0.423	1.44	0.3408	1.4	5.463	2.1	0.1163	1.6	0.67	1891	23	1895	19	1900	29	100
26.01.16	524713	Ttn9	A118	4460	26	9.2	0.566	1.53	0.3422	1.6	5.447	2.5	0.1155	2.0	0.62	1897	26	1892	22	1887	36	101

Date	Sample	Grain	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pb ^c	²⁰⁶ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g
			No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)
26.01.16	524713	Ttn9	A119	6003	40	13.3	0.403	3.90	0.3249	1.3	4.972	2.6	0.1110	2.2	0.49	1814	20	1815	22	1816	41	100
26.01.16	524713	Ttn10	A78	3283	16	5.7	0.713	6.34	0.3397	1.8	5.439	4.1	0.1162	3.7	0.43	1885	29	1891	36	1898	67	99
26.01.16	524713	Ttn10	A79	6864	50	17.6	0.729	0.70	0.3384	1.4	5.419	1.9	0.1162	1.3	0.75	1879	24	1888	17	1898	23	99
26.01.16	524713	Ttn10	A80	6935	50	17.5	0.680	1.17	0.3406	1.3	5.416	1.9	0.1153	1.4	0.68	1890	22	1887	17	1885	26	100
26.01.16	524713	Ttn10	A81	5812	42	15.0	0.771	1.12	0.3448	1.7	5.420	2.4	0.1141	1.6	0.73	1910	29	1888	20	1865	29	102
26.01.16	524713	Ttn10	A82	7131	52	18.1	0.862	0.70	0.3360	1.6	5.406	2.4	0.1167	1.7	0.68	1868	27	1886	21	1907	31	98
26.01.16	524713	Ttn11	A74	6008	45	14.6	0.485	2.26	0.3168	1.5	4.921	2.2	0.1127	1.6	0.68	1774	23	1806	19	1843	30	96
26.01.16	524713	Ttn11	A75	9679	74	26.1	0.359	0.83	0.3392	1.4	5.380	1.8	0.1151	1.2	0.75	1883	23	1882	16	1881	22	100
26.01.16	524713	Ttn11	A77	5719	40	14.2	0.525	1.28	0.3395	1.5	5.420	2.3	0.1158	1.7	0.65	1884	24	1888	20	1893	31	100
26.01.16	524713	Ttn12	A133	5480	37	11.8	0.141	3.67	0.3108	1.6	4.697	2.8	0.1096	2.3	0.57	1745	25	1767	24	1793	42	97
26.01.16	524713	Ttn12	A134	5711	35	11.8	0.567	1.23	0.3256	1.5	4.933	2.7	0.1099	2.2	0.57	1817	24	1808	23	1798	40	101
26.01.16	524713	Ttn12	A135	4999	30	10.4	0.631	1.03	0.3335	1.7	5.321	2.4	0.1157	1.8	0.68	1855	27	1872	21	1891	32	98
26.01.16	524713	Ttn12	A136	5188	29	10.1	0.728	1.94	0.3374	1.7	5.354	2.7	0.1151	2.1	0.62	1874	27	1877	23	1882	38	100
26.01.16	524713	Ttn13	A157	4280	23	7.8	0.577	1.73	0.3279	1.6	5.101	2.4	0.1128	1.7	0.69	1828	26	1836	20	1846	31	99
26.01.16	524713	Ttn13	A158	4347	25	8.1	0.522	1.52	0.3165	1.5	4.819	3.2	0.1105	2.8	0.48	1772	23	1788	27	1807	50	98
26.01.16	524713	Ttn13	A159	4354	23	7.9	0.393	1.63	0.3378	1.5	5.428	2.7	0.1166	2.3	0.55	1876	24	1889	24	1904	41	99
26.01.16	524713	Rt6	A132	4775	29	9.6	0.627	2.28	0.3235	1.5	4.934	2.6	0.1107	2.1	0.58	1807	24	1808	22	1810	38	100
26.01.16	524713	Rt8	A125	4615	26	9.1	0.486	1.75	0.3369	2.0	5.396	2.8	0.1162	2.0	0.72	1872	33	1884	25	1898	35	99
26.01.16	524713	Rt8	A126	5153	31	10.9	0.621	1.19	0.3400	1.6	5.438	2.4	0.1160	1.8	0.67	1887	26	1891	21	1896	32	100
26.01.16	524713	Rt8	A127	4745	28	9.8	0.520	1.45	0.3401	1.8	5.430	2.4	0.1158	1.5	0.76	1887	30	1890	20	1893	27	100
26.01.16	524713	Rt9	A140	11150	72	24.9	0.605	0.22	0.3326	1.3	5.242	1.7	0.1143	1.0	0.79	1851	21	1859	14	1870	18	99
26.01.16	524713	Rt9	A141	6683	43	14.3	0.429	1.09	0.3251	1.5	4.988	2.3	0.1113	1.7	0.67	1814	24	1817	19	1821	31	100
26.01.16	524713	Rt9	A142	6032	36	12.0	0.552	1.60	0.3193	1.8	4.882	2.4	0.1109	1.6	0.74	1786	27	1799	20	1815	29	98
26.01.16	524713	Rt9	A143	6290	38	13.4	0.438	0.92	0.3401	1.4	5.389	2.1	0.1150	1.6	0.66	1887	22	1883	18	1879	28	100
26.01.16	524713	Rt10	A153	5021	28	9.4	0.611	1.69	0.3278	1.5	5.106	2.3	0.1130	1.8	0.63	1828	23	1837	20	1848	33	99
26.01.16	524713	Rt10	A154	4888	26	9.2	0.625	1.06	0.3401	1.5	5.429	2.1	0.1158	1.6	0.67	1887	24	1890	19	1892	29	100
26.01.16	524713	Rt10	A155	5421	31	10.3	0.647	1.42	0.3213	1.7	4.826	2.2	0.1090	1.5	0.75	1796	26	1790	19	1782	27	101
26.01.16	524713	Rt10	A156	4759	26	9.2	0.580	1.32	0.3351	1.9	5.282	2.9	0.1143	2.2	0.66	1863	31	1866	25	1869	40	100

Date	Sample	Grain	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pb ^c	²⁰⁶ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g
			No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)
26.01.16	524716	Ttn1	A06	7627	78	26.2	2.226	1.69	0.3253	1.7	5.082	3.2	0.1133	2.7	0.53	1816	27	1833	27	1854	48	98
26.01.16	524716	Ttn1	A07	7722	80	27.0	2.348	1.43	0.3280	1.5	5.081	3.2	0.1124	2.8	0.48	1828	24	1833	27	1839	51	99
26.01.16	524716	Ttn1	A08	7434	68	23.5	2.156	1.97	0.3363	1.6	5.338	2.5	0.1152	2.0	0.62	1869	25	1875	22	1882	35	99
26.01.16	524716	Ttn2	A09	4804	43	15.3	2.489	2.85	0.3455	1.6	5.532	2.7	0.1162	2.1	0.61	1913	27	1906	23	1898	38	101
26.01.16	524716	Ttn3	A42	5703	45	15.4	1.255	2.93	0.3284	1.5	5.059	2.6	0.1118	2.1	0.59	1831	24	1829	22	1828	37	100
26.01.16	524716	Ttn3	A43	4540	32	10.4	1.345	4.74	0.3173	1.8	4.679	2.9	0.1070	2.3	0.62	1776	28	1764	25	1749	42	102
26.01.16	524716	Ttn3	A44	3086	17	5.8	1.625	6.71	0.3352	1.7	5.384	2.9	0.1165	2.4	0.56	1863	27	1882	26	1904	44	98
26.01.16	524716	Ttn3	A45	6407	49	17.4	1.284	2.40	0.3390	1.5	5.338	3.1	0.1142	2.6	0.51	1882	25	1875	26	1868	48	101
26.01.16	524716	Ttn4	A46	3908	24	8.5	1.153	4.93	0.3427	1.7	5.514	2.6	0.1168	2.0	0.65	1899	28	1903	23	1907	36	100
26.01.16	524716	Ttn4	A47	5429	37	13.0	1.287	3.46	0.3339	1.5	5.386	2.4	0.1170	1.9	0.61	1857	24	1883	21	1911	34	97
26.01.16	524716	Ttn5	A48	4036	30	9.8	1.019	3.73	0.3173	1.5	4.741	3.7	0.1084	3.4	0.42	1777	24	1775	32	1773	62	100
26.01.16	524716	Ttn5	A49	3822	27	9.5	0.841	3.67	0.3427	1.9	5.435	3.0	0.1151	2.3	0.63	1899	31	1890	26	1881	42	101
26.01.16	524716	Ttn6	A50	4147	29	10.3	0.967	3.23	0.3406	1.9	5.436	2.8	0.1158	2.1	0.67	1889	31	1891	24	1892	38	100
26.01.16	524716	Ttn6	A51	3382	20	6.5	1.141	5.80	0.3185	1.5	4.914	4.1	0.1119	3.8	0.36	1782	23	1805	35	1831	69	97
26.01.16	524716	Ttn6	A52	4865	35	12.3	0.938	2.34	0.3350	1.6	5.316	2.8	0.1151	2.3	0.57	1863	26	1871	24	1882	42	99
26.01.16	524716	Ttn6	A53	3365	23	7.7	0.784	4.21	0.3311	1.8	5.281	2.8	0.1157	2.2	0.64	1844	29	1866	24	1891	39	97
26.01.16	524716	Ttn6	A54	3887	26	9.1	0.887	3.38	0.3335	1.7	5.357	2.7	0.1165	2.0	0.65	1855	28	1878	23	1904	36	97
26.01.16	524716	Ttn6	A55	4824	32	11.2	1.201	3.90	0.3395	1.5	5.397	2.8	0.1153	2.3	0.54	1884	25	1884	24	1885	42	100
26.01.16	524716	Ttn7	A56	3045	18	6.1	1.227	5.63	0.3344	1.9	5.335	4.1	0.1157	3.7	0.46	1860	30	1874	36	1891	66	98
26.01.16	524716	Ttn7	A57	4642	33	11.6	1.320	2.85	0.3388	1.7	5.394	2.5	0.1155	1.8	0.67	1881	27	1884	22	1888	33	100
26.01.16	524716	Ttn8	A31	3456	27	9.0	1.151	3.97	0.3231	2.0	4.919	4.1	0.1104	3.6	0.49	1805	32	1805	36	1807	66	100
26.01.16	524716	Ttn8	A32	5279	42	15.1	1.130	2.15	0.3449	1.3	5.562	2.1	0.1170	1.6	0.65	1910	22	1910	18	1911	28	100
26.01.16	524716	Ttn8	A38	4527	36	12.1	1.059	3.11	0.3299	1.9	4.992	2.9	0.1098	2.3	0.64	1838	30	1818	25	1796	41	102
26.01.16	524716	Ttn9	A29	5854	50	17.3	1.241	1.85	0.3369	1.6	5.368	2.3	0.1156	1.8	0.66	1872	25	1880	20	1889	32	99
26.01.16	524716	Ttn9	A30	8141	76	25.2	1.180	1.98	0.3230	1.3	4.831	2.3	0.1085	1.8	0.59	1805	21	1790	19	1774	33	102
26.01.16	524716	Ttn10	A26	4477	37	12.9	1.206	2.61	0.3356	1.7	5.423	2.4	0.1172	1.8	0.68	1865	27	1888	21	1914	32	97
26.01.16	524716	Ttn10	A27	3791	30	10.5	1.245	3.42	0.3391	1.8	5.414	2.7	0.1158	1.9	0.69	1882	30	1887	23	1893	34	99
26.01.16	524716	Ttn10	A28	4608	39	12.9	1.250	2.72	0.3171	1.5	4.908	3.0	0.1123	2.6	0.51	1776	24	1804	26	1837	47	97

Date	Sample	Grain	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pbc ^c	²⁰⁶ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g
			No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)
26.01.16	524716	Ttn13	A22	5989	53	17.8	1.168	2.41	0.3237	1.4	4.997	2.7	0.1120	2.3	0.51	1808	22	1819	23	1832	42	99
26.01.16	524716	Ttn13	A23	4999	42	14.6	0.947	2.73	0.3383	1.5	5.367	2.2	0.1151	1.6	0.67	1879	24	1880	19	1881	30	100
26.01.16	524716	Ttn13	A24	4620	38	13.5	1.113	3.23	0.3384	2.1	5.328	3.0	0.1142	2.1	0.71	1879	35	1873	26	1868	38	101
26.01.16	524716	Ttn13	A25	1880	8	3.0	0.973	10.39	0.3460	2.5	5.526	3.8	0.1159	2.8	0.66	1915	42	1905	33	1894	51	101
26.01.16	524716	Ttn14	A39	5861	35	12.1	1.468	6.22	0.3340	1.3	5.323	2.5	0.1156	2.2	0.51	1858	20	1872	22	1889	39	98
26.01.16	524716	Ttn14	A40	2818	16	5.2	0.936	7.67	0.3157	1.9	4.923	5.3	0.1131	5.0	0.36	1769	30	1806	46	1851	90	96
26.01.16	524716	Ttn14	A41	5365	39	13.7	1.125	3.31	0.3422	1.9	5.427	3.1	0.1151	2.5	0.60	1897	31	1889	27	1881	44	101
26.01.16	524716	Ttn15	A14	4161	32	10.3	0.910	4.74	0.3158	1.7	4.751	3.0	0.1091	2.5	0.57	1769	26	1776	25	1785	45	99
26.01.16	524716	Ttn15	A15	7303	69	22.9	1.256	2.63	0.3213	1.4	4.835	2.4	0.1092	2.0	0.56	1796	21	1791	21	1786	37	101
26.01.16	524716	Ttn15	A16	4313	17	5.7	1.345	16.29	0.3205	2.1	4.813	5.8	0.1090	5.4	0.36	1792	33	1787	50	1782	99	101
26.01.16	524716	Ttn15	A17	4933	42	13.9	1.063	3.38	0.3225	1.5	4.906	2.9	0.1104	2.5	0.50	1802	23	1803	25	1805	46	100
26.01.16	524716	Ttn15	A18	7158	57	19.4	1.365	3.56	0.3268	1.8	5.106	3.3	0.1134	2.8	0.54	1823	28	1837	28	1854	50	98
25.01.16	566216	Rt8	A60	5385	19	6.5	0.715	24	0.335	2.5	5.291	7.5	0.1145	7.1	0.33	1864	41	1867	66	1872	128	100
25.01.16	566216	Rt9	A61	2828	21	6.8	0.390	24	0.308	3.0	4.66	9.3	0.1097	8.8	0.32	1732	46	1760	81	1794	160	97
25.01.16	566216	Rt9	A62	3910	19	6.0	0.518	34	0.314	3.3	5.001	14.8	0.1155	14	0.22	1761	50	1819	134	1887	261	93
25.01.16	566216	Rt12	A82	4353	14	5.2	0.75	52	0.348	4.4	5.598	6.2	0.1166	4.4	0.71	1926	74	1916	55	1905	79	101
25.01.16	566216	Rt13	A86	5136	12	4.1	0.25	35.94	0.321	3.0	5.296	10.9	0.1198	11	0.27	1794	46	1868	98	1953	188	92
25.01.16	566216	Rt13	A87	6600	23	7.8	0.37	23.10	0.327	2.1	5.506	7.3	0.1221	7	0.29	1824	34	1901	65	1988	125	92
25.01.16	566216	Rt14	A75	1225	7	2.5	0.12	23	0.334	3.9	5.204	5.8	0.1129	4.3	0.67	1859	63	1853	50	1847	77	101
25.01.16	566216	Rt14	A76	2758	31	10.5	0.64	16	0.325	2.2	5.134	7.8	0.1144	7.5	0.27	1816	34	1842	69	1871	136	97
25.01.16	566216	Rt22	A18	19929	96	35	1.53	14	0.352	1.9	5.683	3.8	0.1171	3.3	0.49	1944	32	1929	34	1913	60	102
25.01.16	566216	Rt22	A19	13978	44	16	1.08	24	0.346	1.9	5.515	4.9	0.1156	4.5	0.39	1916	32	1903	43	1889	82	101
25.01.16	566216	Rt22	A20	15966	92	32	0.88	12	0.337	1.9	5.299	3.6	0.1142	3	0.53	1871	30	1869	31	1867	55	100
25.01.16	566216	Rt22	A23	19575	136	47	1.01	9	0.333	1.8	5.182	3.1	0.1131	2.6	0.57	1851	28	1850	27	1849	46	100
25.01.16	566216	Rt23	A26	12026	46	16	0.50	20	0.324	2.0	5.112	4.3	0.1145	3.9	0.45	1809	31	1838	37	1872	70	97
25.01.16	566216	Rt23	A27	19887	100	36	0.98	13	0.345	1.8	5.454	3.8	0.1146	3.3	0.47	1912	30	1893	33	1873	60	102
25.01.16	566216	Rt23	A29	18243	75	26	1.00	22	0.339	2.0	5.494	4.9	0.1175	4.5	0.41	1883	33	1900	43	1918	81	98

Date	Sample	Grain	Meas. No	²⁰⁷ Pb ^a (cps)	U ^b (ppm)	Pb ^b (ppm)	Th ^b U	²⁰⁶ Pb ^c (%)	²⁰⁶ Pb ^d ²³⁸ U	±2s ^e (%)	²⁰⁷ Pb ^d ²³⁵ U	±2s ^e (%)	²⁰⁷ Pb ^d ²⁰⁶ Pb	±2s ^e (%)	rho ^f	²⁰⁶ Pb ²³⁸ U	±2s (Ma)	²⁰⁷ Pb ²³⁵ U	±2s (Ma)	²⁰⁷ Pb ²⁰⁶ Pb	±2s (Ma)	conc. ^g (%)
25.01.16	566216	Rt24	A98	16555	41	14.5	0.821	34	0.342	2.1	5.718	6.0	0.1215	5.6	0.35	1894	35	1934	53	1978	100	96
25.01.16	566216	Rt24	A99	16115	66	23.9	1.302	40	0.347	2.3	5.37	6.5	0.1123	6.1	0.35	1921	38	1880	57	1836	111	105
25.01.16	566216	Rt24	A100	25695	270	98.0	2.017	16	0.349	1.8	5.74	4.1	0.1194	3.6	0.45	1929	31	1937	36	1947	65	99
25.01.16	566216	Rt24	A101	14106	82	28.8	0.227	13	0.34	1.7	5.375	3.9	0.1147	3.5	0.45	1886	28	1881	34	1876	63	101
25.01.16	566216	Ttnex1	A111	4618	26	8.9	0.322	34	0.328	3.2	5.694	11.1	0.1261	11	0.29	1827	51	1930	101	2044	188	89
25.01.16	566216	Rt10	A65	8369	84	28.7	0.096	4.91	0.33	1.8	5.166	2.8	0.1135	2.1	0.65	1839	29	1847	24	1856	38	99
25.01.16	566277	Ttn2	A185	883	22	7.4	0.82	3.13	0.321	2.9	4.95	4.1	0.1119	2.8	0.72	1794	46	1811	35	1831	51	98
25.01.16	566277	Ttn2	A186	767	22	7.4	0.87	2.31	0.322	2.4	5.122	3.6	0.1155	2.6	0.68	1798	38	1840	31	1888	47	95
25.01.16	566277	Ttn2	A187	551	20	6.3	1.00	4.14	0.306	4.0	4.479	5.2	0.1064	3.3	0.77	1719	61	1727	44	1738	61	99
25.01.16	566277	Ttn2	A188	728	21	7.0	0.86	1.2	0.317	2.5	4.877	3.9	0.1116	3	0.64	1775	39	1798	33	1826	54	97
25.01.16	566277	Ttn3	A177	6568	208	71.1	1.68	bd1	0.331	1.8	5.181	2.3	0.1136	1.5	0.76	1842	29	1849	20	1858	27	99
25.01.16	566277	Ttn3	A178	14168	212	72.7	1.54	0.02	0.331	1.7	5.257	2.0	0.1151	1	0.85	1845	27	1862	17	1881	19	98
25.01.16	566277	Ttn3	A179	14168	215	76.7	1.68	bd1	0.344	1.7	5.467	2.0	0.1153	1.1	0.84	1905	28	1895	17	1885	20	101
25.01.16	566277	Ttn4	A174	12475	198	69.5	1.60	bd1	0.34	2.0	5.364	2.3	0.1146	1.1	0.87	1885	33	1879	20	1873	20	101
25.01.16	566277	Ttn4	A175	8239	250	81.6	1.39	0.27	0.316	1.8	4.817	2.4	0.1106	1.6	0.75	1771	28	1788	20	1809	29	98
25.01.16	566277	Ttn4	A176	6698	229	73.4	1.75	0.87	0.311	1.9	4.613	3.0	0.1075	2.3	0.65	1747	30	1752	25	1758	42	99
25.01.16	566277	Ttn5	A180	2850	41	14.7	1.49	0.84	0.346	1.7	5.531	2.5	0.1161	1.8	0.67	1914	28	1905	22	1897	33	101
25.01.16	566277	Ttn5	A181	2664	39	13.4	1.48	0.68	0.335	1.8	5.391	2.8	0.1167	2.2	0.64	1863	30	1883	25	1907	39	98
25.01.16	566277	Ttn5	A182	3012	44	15.5	1.40	1.03	0.343	1.7	5.444	2.6	0.1152	1.9	0.68	1901	29	1892	22	1883	34	101
25.01.16	566277	Ttn7	A204	6685	170	60.1	2.36	1.09	0.34	1.7	5.399	2.5	0.1151	1.8	0.69	1889	29	1885	22	1881	33	100
25.01.16	566277	Ttn7	A205	4428	121	41.8	3.57	2.1	0.333	1.9	5.31	3.3	0.1157	2.7	0.59	1853	31	1871	29	1891	48	98
25.01.16	566277	Ttn8	A201	3161	47	16.5	1.36	0.57	0.337	2.0	5.401	2.8	0.1163	2	0.70	1872	32	1885	24	1900	36	99
25.01.16	566277	Ttn8	A202	3116	47	16.3	1.54	0.94	0.334	1.8	5.306	3.0	0.1152	2.4	0.61	1858	30	1870	26	1884	43	99
25.01.16	566277	Ttn8	A203	3066	44	15.6	1.41	1.33	0.341	1.9	5.5	2.8	0.1171	2.1	0.67	1890	31	1901	24	1912	37	99
25.01.16	566277	Ttn9	A191	3512	50	15.9	1.26	0.49	0.309	2.2	4.728	3.1	0.1111	2.2	0.70	1734	33	1772	26	1818	40	95
25.01.16	566277	Ttn9	A192	3137	45	15.5	1.01	0.92	0.332	1.9	5.193	2.9	0.1135	2.2	0.67	1847	31	1852	25	1857	39	100
25.01.16	566277	Ttn9	A193	2785	42	14.1	1.31	0.27	0.323	1.9	5.109	3.1	0.1147	2.5	0.61	1805	30	1838	27	1876	45	96

Date	Sample	Grain	Meas. No	²⁰⁷ Pb ^a (cps)	U ^b (ppm)	Pb ^b (ppm)	Th ^b U	²⁰⁶ Pb ^c (%)	²⁰⁶ Pb ^d ²³⁸ U	±2s ^e (%)	²⁰⁷ Pb ^d ²³⁵ U	±2s ^e (%)	²⁰⁷ Pb ^d ²⁰⁶ Pb	±2s ^e (%)	rho ^f	²⁰⁶ Pb ²³⁸ U	±2s (Ma)	²⁰⁷ Pb ²³⁵ U	±2s (Ma)	²⁰⁷ Pb ²⁰⁶ Pb	±2s (Ma)	conc. ^g (%)
25.01.16	566277	Ttn9	A199	3101	43	14.9	1.47	2.2	0.335	2.1	5.24	4.0	0.1136	3.4	0.51	1860	33	1859	35	1858	62	100
25.01.16	566277	Ttn9	A200	2885	41	14.8	1.33	0.61	0.345	1.7	5.514	2.7	0.116	2	0.65	1910	29	1903	23	1896	37	101
25.01.16	566277	Ttn10	A206	1263	15	5.3	1.15	3.9	0.34	2.1	5.373	3.4	0.1148	2.7	0.62	1884	34	1881	29	1877	48	100
25.01.16	566277	Ttn10	A207	1188	13	4.7	1.07	4.0	0.339	2.1	5.37	3.3	0.1151	2.6	0.63	1880	34	1880	29	1881	46	100
25.01.16	566277	Ttn10	A208	1233	14	5.1	0.80	3.2	0.342	2.5	5.484	4.1	0.1163	3.3	0.60	1896	41	1898	36	1900	60	100
25.01.16	566277	Ttn10	A209	1211	14	5.1	1.26	3.7	0.341	2.7	5.45	7.3	0.1161	6.8	0.38	1890	45	1893	65	1897	121	100
25.01.16	566277	Ttn10	A210	997	12	4.1	0.90	4.0	0.343	2.3	5.495	5.1	0.1163	4.6	0.45	1899	38	1900	45	1901	82	100
25.01.16	566277	Ttn10	A211	1208	14	5.1	1.17	3.3	0.344	2.3	5.511	5.5	0.1161	5	0.43	1907	39	1902	48	1897	89	101
25.01.16	566277	Ttn11	A212	2865	38	13.4	0.68	0.23	0.338	2.2	5.379	3.8	0.1153	3.1	0.58	1879	36	1882	33	1885	55	100
25.01.16	566277	Ttn11	A213	3368	50	18.1	1.01	0.15	0.347	2.0	5.498	2.9	0.1151	2.1	0.67	1918	33	1900	25	1881	39	102
25.01.16	566277	Ttn11	A214	4900	77	27.1	1.03	bdl	0.342	1.8	5.428	3.0	0.1153	2.4	0.59	1895	29	1889	26	1884	44	101
25.01.16	566277	Ttn12	A215	2727	40	14.4	0.65	1.19	0.344	1.9	5.449	3.2	0.1149	2.5	0.61	1906	32	1893	28	1878	46	102
25.01.16	566277	Ttn12	A216	1119	33	11.0	0.66	1.7	0.321	2.5	5.03	3.5	0.1138	2.4	0.73	1793	40	1824	30	1861	43	96
25.01.16	566277	Ttn12	A217	1491	45	14.9	0.60	0.9	0.324	2.1	5.08	3.2	0.1138	2.3	0.67	1809	34	1833	27	1861	42	97
25.01.16	566277	Ttn13	A218	15309	268	91.9	1.76	bdl	0.331	1.8	5.106	2.0	0.1118	0.98	0.87	1845	28	1837	17	1829	18	101
25.01.16	566277	Ttn13	A219	6978	250	84.2	1.73	bdl	0.326	1.7	5.043	2.2	0.1124	1.4	0.78	1817	27	1827	19	1838	25	99
25.01.16	566277	Ttn13	A220	7720	264	88.7	1.75	bdl	0.325	1.7	5.154	2.1	0.1152	1.3	0.79	1812	27	1845	18	1883	24	96
25.01.16	566277	Ttn14	A223	5232	86	30.9	0.89	bdl	0.347	1.7	5.601	2.7	0.117	2	0.64	1922	28	1916	23	1910	37	101
25.01.16	566277	Ttn14	A224	4917	81	28.5	0.98	0.10	0.34	1.9	5.466	2.5	0.1168	1.7	0.74	1885	31	1895	22	1907	30	99
25.01.16	566277	Ttn15	A227	2779	38	13.5	1.34	3.8	0.341	2.2	5.379	3.2	0.1143	2.4	0.67	1894	36	1882	28	1869	43	101
25.01.16	566277	Ttn15	A228	1512	51	15.8	1.02	1.5	0.304	2.5	4.463	3.3	0.1067	2.3	0.74	1709	37	1724	28	1743	41	98
25.01.16	566277	Ttn15	A229	1771	54	19.5	1.11	1.4	0.346	2.2	5.507	3.3	0.1155	2.5	0.66	1915	36	1902	29	1887	45	102
25.01.16	566277	Ttn17	A245	2769	47	15.4	0.68	bdl	0.318	1.9	4.976	2.8	0.1134	2.1	0.66	1782	29	1815	24	1855	39	96
25.01.16	566277	Ttn17	A246	3605	57	20.2	0.67	0.56	0.34	1.9	5.382	3.0	0.1147	2.3	0.63	1888	31	1882	26	1876	42	101
25.01.16	566277	Ttn17	A247	3894	63	22.2	0.69	0.78	0.343	1.7	5.436	2.8	0.1151	2.2	0.62	1900	29	1891	24	1881	40	101
25.01.16	566277	Ttn18	A236	647	17	5.9	1.01	4.4	0.339	2.4	5.389	3.8	0.1154	2.9	0.65	1881	40	1883	33	1886	52	100
25.01.16	566277	Ttn18	A237	832	23	8.1	1.06	2.1	0.337	2.3	5.326	3.5	0.1148	2.7	0.65	1871	37	1873	30	1876	48	100
25.01.16	566277	Ttn18	A243	794	21	7.5	0.91	3.28	0.341	2.7	5.481	3.8	0.1168	2.7	0.69	1889	44	1898	33	1908	49	99

Date	Sample	Grain	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pb ^c	²⁰⁶ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g
			No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)
25.01.16	566277	Ttn19	A231	11962	208	71.9	1.63	bdl	0.334	1.7	5.203	2.0	0.113	1.1	0.83	1857	27	1853	18	1849	21	100
25.01.16	566277	Ttn19	A232	11143	186	63.9	1.76	0.57	0.332	1.9	5.232	2.2	0.1145	1.2	0.85	1846	30	1858	19	1872	21	99
25.01.16	566277	Ttn19	A233	13274	276	96.8	1.87	0.22	0.338	1.8	5.3	2.1	0.1137	1.1	0.85	1878	29	1869	18	1859	20	101

^a Within run background-corrected mean ²⁰⁷Pb signal in cps (counts per second).

^b U and Pb content and Th/U ratio were calculated relative to GJ-1 reference zircon.

^c percentage of the common Pb on the ²⁰⁶Pb. bdl = below detection limit.

^d corrected for background, within-run Pb/U fractionation (in case of ²⁰⁶Pb/²³⁸U) and common Pb using Stacy and Kramers (1975) model Pb composition and subsequently normalized to GJ-1 (ID-TIMS value/measured value); ²⁰⁷Pb/²³⁵U calculated using ²⁰⁷Pb/²⁰⁶Pb/(²³⁸U/²⁰⁶Pb*1/137.88)

^e ²⁰⁶Pb/²³⁸U error is the quadratic additions of the within run precision (2 SE) and the external reproducibility (2 SD) of the reference zircon. ²⁰⁷Pb/²⁰⁶Pb error propagation (²⁰⁷Pb signal dependent) following Gerdes & Zeh (2009). ²⁰⁷Pb/²³⁵U error is the quadratic addition of the ²⁰⁷Pb/²⁰⁶Pb and ²⁰⁶Pb/²³⁸U uncertainty.

^f rho is the ²⁰⁶Pb/²³⁸U/²⁰⁷Pb/²³⁵U error correlation coefficient.

^g degree of concordance = ²⁰⁶Pb/²³⁸U age / ²⁰⁷Pb/²⁰⁶Pb age x 100.

Appendix 9 – Rutile U-Pb data (GUF)

Date	Sample	Grain	Meas. No	²⁰⁷ Pb ^a (cps)	U ^b (ppm)	Pb ^b (ppm)	Th ^b U	²⁰⁶ Pb ^c (%)	²⁰⁶ Pb ^d ²³⁸ U	±2s ^e (%)	²⁰⁷ Pb ^d ²³⁵ U	±2s ^e (%)	²⁰⁷ Pb ^d ²⁰⁶ Pb	±2s ^e (%)	rho ^f	²⁰⁶ Pb ²³⁸ U	±2s (Ma)	²⁰⁷ Pb ²³⁵ U	±2s (Ma)	²⁰⁷ Pb ²⁰⁶ Pb	±2s (Ma)	conc. ^g (%)
26.01.16	524713	Rt2	A66	181	1.3	0.4	bdl	bdl	0.2784	3.4	4.181	10.3	0.1089	9.7	0.33	1583	48	1670	88	1782	177	89
26.01.16	524713	Rt3	A83	137	1.3	0.4	bdl	bdl	0.2712	3.4	3.959	14.4	0.1059	14.0	0.23	1547	46	1626	124	1731	257	89
26.01.16	524713	Rt5	A121	290	0.9	0.3	bdl	22.28	0.3051	7.4	4.465	12.4	0.1062	10.0	0.60	1716	112	1725	109	1735	183	99
26.01.16	524713	Rt7	A84	99	1.4	0.4	0.015	bdl	0.2878	4.9	4.344	9.8	0.1095	8.5	0.50	1631	72	1702	84	1791	154	91
26.01.16	524713	Rt7	A86	156	1.3	0.3	0.011	bdl	0.2122	5.5	3.060	8.9	0.1046	7.0	0.61	1240	62	1423	71	1708	130	73
26.01.16	524713	Ttn5	A89	314	1.6	0.3	0.013	bdl	0.1883	6.6	2.930	9.7	0.1129	7.1	0.68	1112	68	1390	76	1847	128	60
26.01.16	524713	Ttn11	A73	58	1.3	0.3	bdl	bdl	0.2552	4.3	3.735	12.6	0.1062	12.0	0.34	1465	56	1579	106	1735	217	84
26.01.16	525224	Rt1	A165	2586	11.0	3.7	0.005	6.09	0.3219	2.6	4.840	4.8	0.1091	4.1	0.53	1799	40	1792	41	1784	75	101
26.01.16	525224	Rt2	A186	2378	12.4	4.2	0.003	0.30	0.3292	3.5	4.947	4.2	0.1090	2.3	0.83	1834	56	1810	36	1783	42	103
26.01.16	525224	Rt3	A187	4993	15.8	5.2	bdl	bdl	0.3181	1.4	4.854	2.3	0.1107	1.8	0.62	1781	23	1794	20	1811	33	98
26.01.16	525224	Rt4	A190	4384	11.0	3.6	bdl	bdl	0.3180	2.4	4.784	2.8	0.1091	1.4	0.86	1780	37	1782	24	1785	26	100
26.01.16	525224	Rt4	A191	4144	10.6	3.6	bdl	0.08	0.3257	3.1	4.956	3.6	0.1104	1.9	0.85	1818	49	1812	31	1806	35	101
26.01.16	525224	Rt4	A192	4190	10.4	3.4	0.001	bdl	0.3189	2.0	4.820	3.2	0.1097	2.4	0.64	1784	32	1788	27	1794	44	99
26.01.16	525224	Rt5	A193	2237	11.0	3.7	bdl	bdl	0.3222	3.8	4.928	4.6	0.1109	2.6	0.82	1801	60	1807	40	1815	48	99
25.01.16	566216	Rt1	A39	1203	7.2	1.8	bdl	bdl	0.2406	4.4	3.484	5.0	0.1051	2.4	0.88	1390	56	1524	41	1716	44	81
25.01.16	566216	Rt1	A40	1520	9.0	2.6	bdl	bdl	0.2775	2.6	4.079	4.0	0.1067	3.0	0.65	1579	36	1650	33	1743	55	91
25.01.16	566216	Rt1	A41	1015	5.8	1.8	bdl	bdl	0.3068	3.3	4.480	4.7	0.1059	3.3	0.71	1725	51	1727	40	1730	60	100
25.01.16	566216	Rt2	A42	515	6.5	1.9	0.024	bdl	0.2907	4.6	4.232	5.5	0.1056	3.0	0.83	1645	66	1680	46	1725	55	95
25.01.16	566216	Rt2	A43	478	6.0	1.8	0.067	0.11	0.2879	3.3	4.205	4.7	0.1060	3.4	0.70	1631	47	1675	39	1731	62	94
25.01.16	566216	Rt2	A44	566	6.6	1.9	0.060	0.25	0.2871	3.8	4.200	4.6	0.1061	2.6	0.82	1627	55	1674	39	1734	49	94
25.01.16	566216	Rt3	A45	783	4.6	1.4	0.002	bdl	0.2920	2.8	4.284	5.4	0.1064	4.6	0.53	1652	41	1690	45	1739	83	95
25.01.16	566216	Rt3	A46	734	4.2	1.2	bdl	0.65	0.2679	3.1	3.918	5.0	0.1061	4.0	0.61	1530	42	1617	41	1733	73	88
25.01.16	566216	Rt5	A47	286	3.3	1.0	bdl	0.94	0.3064	4.3	4.448	6.6	0.1053	5.0	0.65	1723	65	1721	56	1720	92	100
25.01.16	566216	Rt5	A49	245	3.5	1.1	bdl	bdl	0.2959	3.0	4.359	5.8	0.1069	4.9	0.53	1671	45	1705	49	1746	90	96

Date	Sample	Grain	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pb ^c	²⁰⁶ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g
			No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)
25.01.16	566216	Rt5	A50	222	2.6	0.7	bdl	bdl	0.2684	5.1	3.932	6.2	0.1063	3.6	0.82	1533	70	1620	52	1736	65	88
25.01.16	566216	Rt7	A55	295	3.9	1.1	bdl	0.46	0.2679	4.8	3.679	6.8	0.0996	4.9	0.70	1530	65	1567	56	1617	91	95
25.01.16	566216	Rt7	A56	263	3.2	0.9	bdl	bdl	0.2765	5.3	4.028	7.4	0.1057	5.3	0.71	1573	74	1640	62	1726	96	91
25.01.16	566216	Rt7	A57	366	2.9	0.9	0.049	3.94	0.2934	4.7	4.304	9.3	0.1064	8.0	0.51	1658	70	1694	80	1739	148	95
25.01.16	566216	Rt8	A58	180	1.6	0.5	0.010	bdl	0.3144	3.1	4.672	7.4	0.1078	6.6	0.43	1762	49	1762	63	1763	122	100
25.01.16	566216	Rt8	A59	335	1.9	0.6	bdl	1.57	0.2904	3.7	4.185	7.4	0.1046	6.4	0.50	1644	54	1671	63	1706	119	96
25.01.16	566216	Rt9	A63	417	2.4	0.6	0.016	2.06	0.2442	4.2	3.489	8.4	0.1036	7.2	0.50	1409	53	1525	68	1690	134	83
25.01.16	566216	Rt9	A64	498	3.2	1.0	bdl	0.22	0.2905	3.4	4.247	6.1	0.1061	5.1	0.56	1644	50	1683	51	1733	93	95
25.01.16	566216	Rt11	A66	4150	28.2	8.7	bdl	bdl	0.2997	1.7	4.419	2.7	0.1070	2.0	0.65	1690	26	1716	22	1748	37	97
25.01.16	566216	Rt11	A72	4078	30.3	9.4	bdl	0.04	0.3010	1.8	4.425	2.7	0.1067	2.0	0.68	1696	27	1717	23	1743	36	97
25.01.16	566216	Rt12	A79	1754	8.2	2.4	0.117	15	0.2896	3.1	4.197	6.1	0.1051	5.2	0.51	1640	44	1673	51	1717	96	96
25.01.16	566216	Rt12	A80	600	8.5	2.6	bdl	0.06	0.2991	2.7	4.247	5.5	0.1030	4.8	0.49	1687	40	1683	46	1679	89	100
25.01.16	566216	Rt12	A81	467	2.8	0.9	0.006	10	0.3069	2.2	4.559	8.6	0.1078	8.4	0.25	1725	33	1742	75	1762	153	98
25.01.16	566216	Rt13	A84	356	4.3	1.4	bdl	0.05	0.3090	3.1	4.555	7.8	0.1069	7.2	0.40	1736	48	1741	68	1748	132	99
25.01.16	566216	Rt13	A85	502	2.7	0.8	0.025	13.08	0.2782	6.0	4.078	7.5	0.1063	4.5	0.80	1582	84	1650	63	1738	83	91
25.01.16	566216	Rt14	A77	413	1.3	0.3	0.007	9	0.2615	5.5	3.788	8.3	0.1051	6.2	0.66	1497	74	1590	69	1716	114	87
25.01.16	566216	Rt14	A78	508	3.5	1.1	bdl	0.71	0.2939	2.9	4.309	7.4	0.1064	6.8	0.40	1661	43	1695	63	1738	125	96
25.01.16	566216	Rt14	A74	371	2.9	0.9	bdl	bdl	0.3056	3.1	4.504	5.9	0.1069	5.0	0.52	1719	46	1732	50	1748	92	98
25.01.16	566216	Rt17	A88	270	2.0	0.5	bdl	0.73	0.2553	5.4	3.690	7.4	0.1049	5.0	0.73	1466	71	1569	61	1712	92	86
25.01.16	566216	Rt17	A89	224	1.2	0.4	bdl	2.07	0.2863	5.2	4.374	8.4	0.1108	6.6	0.62	1623	76	1707	72	1813	120	90
25.01.16	566216	Rt17	A90	174	1.1	0.3	bdl	bdl	0.2810	6.7	4.028	10.3	0.1040	7.8	0.65	1597	95	1640	88	1696	145	94
25.01.16	566216	Rt18	A91	903	6.0	1.8	bdl	0.12	0.2852	2.6	4.214	4.3	0.1072	3.4	0.60	1618	37	1677	36	1752	62	92
25.01.16	566216	Rt18	A92	805	6.0	1.8	bdl	bdl	0.2957	3.1	4.140	5.6	0.1016	4.7	0.54	1670	45	1662	47	1653	87	101
25.01.16	566216	Rt18	A93	1237	7.9	2.4	0.123	1.21	0.3003	2.6	4.434	4.0	0.1071	3.0	0.65	1693	39	1719	33	1751	55	97
25.01.16	566216	Rt18	A94	1159	8.5	2.7	0.006	0.16	0.3076	2.4	4.482	3.9	0.1057	3.1	0.61	1729	36	1728	33	1727	57	100
25.01.16	566216	Rt19	A07	3510	10.8	3.3	0.008	bdl	0.3019	1.9	4.381	2.7	0.1053	1.9	0.72	1700	29	1709	22	1719	35	99
25.01.16	566216	Rt19	A08	2670	8.6	2.6	bdl	bdl	0.2903	2.0	4.236	3.0	0.1059	2.3	0.65	1643	28	1681	25	1730	42	95
25.01.16	566216	Rt19	A09	3763	10.0	3.1	bdl	1.09	0.2998	1.9	4.450	3.2	0.1077	2.5	0.61	1691	29	1722	27	1760	46	96

Date	Sample	Grain	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pb ^c	²⁰⁶ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g
			No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)
25.01.16	566216	Rt20	A10	1934	12.7	3.6	0.004	bdl	0.2759	2.3	4.040	3.2	0.1062	2.2	0.72	1571	32	1642	26	1736	41	90
25.01.16	566216	Rt20	A11	2752	13.6	3.8	0.006	3.0	0.2722	2.2	3.933	5.3	0.1048	4.8	0.42	1552	31	1621	43	1711	88	91
25.01.16	566216	Rt20	A12	1689	11.0	3.4	bdl	0.02	0.3040	2.5	4.386	3.5	0.1047	2.5	0.71	1711	37	1710	29	1708	45	100
25.01.16	566216	Rt20	A13	1856	11.9	3.6	bdl	bdl	0.2953	2.2	4.313	3.5	0.1060	2.8	0.63	1668	33	1696	30	1731	51	96
25.01.16	566216	Rt21	A14	3273	9.7	2.8	bdl	bdl	0.2836	1.8	4.177	2.7	0.1069	2.0	0.67	1609	26	1669	22	1746	37	92
25.01.16	566216	Rt21	A15	1415	4.0	1.2	bdl	0.09	0.2908	2.2	4.252	4.1	0.1061	3.5	0.53	1645	32	1684	34	1733	64	95
25.01.16	566216	Rt21	A16	3658	10.7	3.1	bdl	0.07	0.2818	2.0	4.098	2.7	0.1055	1.9	0.72	1601	28	1654	23	1723	35	93
25.01.16	566216	Rt22	A21	1242	6.8	1.9	0.001	bdl	0.2765	4.0	4.056	4.6	0.1064	2.4	0.86	1574	56	1646	39	1739	44	90
25.01.16	566216	Rt22	A22	2735	8.8	2.6	0.002	11	0.2854	2.5	4.148	4.8	0.1054	4.1	0.52	1619	36	1664	40	1722	76	94
25.01.16	566216	Rt23	A28	8158	4.9	1.5	0.123	69	0.2938	9.0	4.280	10.1	0.1057	4.6	0.89	1661	132	1690	87	1726	85	96
25.01.16	566216	Rt24	A107	266	3.1	0.9	bdl	0.41	0.2872	4.1	4.299	9.0	0.1086	8.0	0.45	1628	59	1693	77	1776	147	92
25.01.16	566216	Rt24	A108	805	10.0	2.9	bdl	0.64	0.2853	3.9	3.983	7.0	0.1013	5.8	0.56	1618	56	1631	58	1648	108	98
25.01.16	566216	Rt25	A95	1006	14.6	4.5	bdl	bdl	0.2983	2.7	4.422	4.8	0.1075	4.0	0.56	1683	40	1716	40	1758	73	96
25.01.16	566216	Rt25	A96	1142	14.9	4.4	bdl	bdl	0.2868	2.7	4.263	4.3	0.1078	3.4	0.62	1626	39	1686	36	1763	62	92
25.01.16	566216	Rt25	A97	1744	20.5	5.7	bdl	1.16	0.2723	2.9	3.881	4.4	0.1034	3.4	0.64	1552	39	1610	36	1686	63	92
25.01.16	566216	Rtex1	A109	1415	14.2	4.2	bdl	3.52	0.2901	3.2	4.260	4.9	0.1065	3.7	0.66	1642	46	1686	41	1741	67	94
25.01.16	566216	Rtex1	A110	1040	13.1	4.1	bdl	1.15	0.3006	2.5	4.400	4.8	0.1062	4.1	0.52	1694	37	1712	41	1735	75	98
25.01.16	566218	Rt1	A302	15409	140.6	44.8	bdl	3.0	0.3102	2.3	4.511	3.1	0.1055	2.1	0.74	1742	35	1733	26	1723	39	101
25.01.16	566218	Rt1	A303	16630	128.2	39.9	bdl	3.8	0.3024	2.1	4.451	2.7	0.1068	1.7	0.79	1703	32	1722	23	1745	30	98
25.01.16	566218	Rt2	A300	7654	81.0	25.5	bdl	bdl	0.3065	2.3	4.484	2.7	0.1061	1.3	0.87	1724	35	1728	22	1734	24	99
25.01.16	566218	Rt2	A301	9051	90.6	26.3	bdl	0.0	0.2824	2.5	4.135	3.1	0.1063	1.9	0.80	1603	35	1661	26	1736	34	92
25.01.16	566218	Rt3	A296	8556	85.4	27.5	0.003	0.69	0.3131	2.0	4.586	2.5	0.1063	1.5	0.81	1756	31	1747	21	1737	27	101
25.01.16	566218	Rt3	A298	11477	96.1	28.3	bdl	2.21	0.2866	2.0	4.208	2.6	0.1065	1.6	0.77	1624	29	1675	21	1741	30	93
25.01.16	566218	Rt4	A299	7299	74.5	21.5	bdl	bdl	0.2805	1.8	4.095	2.1	0.1059	1.1	0.86	1594	26	1653	17	1730	20	92
25.01.16	566218	Rt5	A294	9713	99.4	30.8	bdl	bdl	0.3007	2.4	4.430	2.6	0.1069	1.1	0.90	1695	35	1718	22	1747	21	97
25.01.16	566218	Rt5	A295	8702	96.2	30.4	bdl	0.08	0.3074	2.4	4.504	2.9	0.1063	1.5	0.85	1728	37	1732	24	1737	28	99
25.01.16	566218	Rt6	A292	12009	93.8	31.1	0.083	16.2	0.3209	1.8	4.954	4.0	0.1120	3.6	0.46	1794	29	1812	34	1832	64	98

Date	Sample	Grain	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pb ^c	²⁰⁶ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g
			No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)
25.01.16	566218	Rt6	A293	2274	24.6	7.7	bdl	bdl	0.3028	2.7	4.428	3.4	0.1061	2.1	0.80	1705	41	1718	29	1734	38	98
25.01.16	566218	Rt7	A280	3270	66.1	21.0	bdl	bdl	0.3087	2.4	4.548	3.0	0.1069	1.7	0.81	1734	37	1740	25	1747	32	99
25.01.16	566218	Rt7	A281	3808	59.5	19.1	0.012	5.0	0.3124	2.7	4.584	3.5	0.1064	2.3	0.76	1753	42	1746	30	1739	42	101
25.01.16	566218	Rt8	A282	4149	76.8	24.1	bdl	0.13	0.3052	2.3	4.483	3.1	0.1066	2.0	0.76	1717	35	1728	26	1741	37	99
25.01.16	566218	Rt8	A288	4124	79.1	25.0	bdl	0.13	0.3067	2.1	4.475	2.8	0.1058	1.8	0.75	1725	32	1726	23	1729	34	100
25.01.16	566218	Rt8	A289	4897	97.2	30.7	bdl	0.21	0.3066	2.1	4.518	2.6	0.1069	1.6	0.79	1724	31	1734	22	1747	30	99
25.01.16	566218	Rt8	A290	5961	85.0	26.5	0.001	1.8	0.3034	2.8	4.429	3.8	0.1059	2.5	0.76	1708	43	1718	32	1730	45	99
25.01.16	566218	Rt8	A291	5219	100.3	31.8	bdl	0.14	0.3087	2.3	4.482	2.7	0.1053	1.4	0.85	1734	35	1728	23	1720	26	101
25.01.16	566218	Rt9	A277	2774	45.4	14.1	0.002	1.6	0.3031	3.1	4.418	3.9	0.1057	2.2	0.82	1707	47	1716	32	1727	41	99
25.01.16	566218	Rt10	A278	3663	72.8	21.5	bdl	0.29	0.2870	2.6	4.149	3.3	0.1049	2.0	0.78	1627	37	1664	27	1712	38	95
25.01.16	566218	Rt11	A279	3218	68.6	22.0	bdl	0.16	0.3115	2.5	4.530	3.2	0.1055	2.0	0.78	1748	38	1737	27	1724	36	101
25.01.16	566218	Rt13	A276	1563	15.2	4.9	0.007	0.42	0.3128	2.8	4.603	3.4	0.1068	2.0	0.80	1754	43	1750	29	1745	37	101
25.01.16	566218	Rt14	A269	3161	14.2	4.5	0.010	8.7	0.3110	3.3	4.546	4.4	0.1060	3.0	0.73	1746	50	1739	38	1732	55	101
25.01.16	566218	Rt14	A270	1961	20.2	6.2	bdl	bdl	0.2997	2.0	4.347	3.1	0.1052	2.4	0.64	1690	29	1702	26	1718	43	98
25.01.16	566218	Rt14	A271	1554	16.1	4.8	bdl	bdl	0.2900	1.9	4.256	2.9	0.1065	2.2	0.65	1642	27	1685	24	1740	41	94
25.01.16	566218	Rt14	A272	1553	12.3	3.8	bdl	2.0	0.3005	2.2	4.355	3.5	0.1052	2.8	0.63	1694	33	1704	30	1717	51	99
25.01.16	566218	Rt14	A273	1387	14.5	4.2	bdl	bdl	0.2847	2.4	4.104	3.3	0.1046	2.2	0.75	1615	35	1655	27	1707	40	95
25.01.16	566218	Rt14	A274	1817	17.6	5.5	0.001	0.39	0.3035	2.4	4.433	3.4	0.1060	2.4	0.71	1709	36	1719	28	1731	43	99
25.01.16	566218	Rt14	A275	1906	19.6	6.0	0.003	0.21	0.3006	2.1	4.373	3.3	0.1055	2.5	0.63	1694	31	1707	27	1724	46	98
25.01.16	566218	Rt15	A265	1638	14.4	4.6	bdl	0.3	0.3137	2.1	4.569	3.1	0.1057	2.3	0.66	1759	32	1744	26	1726	43	102
25.01.16	566218	Rt15	A266	1193	11.4	3.7	bdl	0.3	0.3126	2.6	4.558	3.7	0.1058	2.7	0.70	1753	40	1742	31	1728	49	101
25.01.16	566218	Rt15	A268	1402	12.6	4.0	bdl	0.5	0.3072	2.0	4.539	3.0	0.1072	2.2	0.69	1727	31	1738	25	1752	40	99
25.01.16	566218	Rt16	A259	2082	11.2	3.5	0.056	5.9	0.3020	3.4	4.420	4.4	0.1062	2.9	0.76	1701	50	1716	37	1735	53	98
25.01.16	566218	Rt16	A260	1422	13.3	4.1	bdl	0.1	0.3029	2.4	4.448	3.4	0.1065	2.3	0.72	1705	36	1721	28	1741	43	98
25.01.16	566218	Rt16	A261	1673	14.6	4.7	0.002	0.3	0.3160	4.1	4.590	4.6	0.1054	2.2	0.88	1770	63	1747	39	1721	41	103
25.01.16	566218	Rt16	A262	1475	12.9	3.8	bdl	bdl	0.2883	2.3	4.197	3.2	0.1056	2.2	0.72	1633	33	1673	26	1725	40	95
25.01.16	566218	Rt16	A263	1718	16.0	5.0	0.054	bdl	0.3022	2.4	4.476	3.2	0.1075	2.2	0.74	1702	36	1726	27	1757	40	97
25.01.16	566218	Rt16	A264	1517	14.3	4.2	bdl	bdl	0.2878	2.1	4.255	3.2	0.1073	2.3	0.67	1630	31	1685	26	1754	43	93

Date	Sample	Grain	Meas. No	²⁰⁷ Pb ^a (cps)	U ^b (ppm)	Pb ^b (ppm)	Th ^b U	²⁰⁶ Pb ^c (%)	²⁰⁶ Pb ^d ²³⁸ U	±2s ^e (%)	²⁰⁷ Pb ^d ²³⁵ U	±2s ^e (%)	²⁰⁷ Pb ^d ²⁰⁶ Pb	±2s ^e (%)	rho ^f	²⁰⁶ Pb ²³⁸ U	±2s (Ma)	²⁰⁷ Pb ²³⁵ U	±2s (Ma)	²⁰⁷ Pb ²⁰⁶ Pb	±2s (Ma)	conc. ^g (%)
25.01.16	566218	Rt17	A257	1132	21.1	6.6	bdl	bdl	0.3063	2.3	4.506	3.4	0.1067	2.5	0.68	1723	36	1732	29	1744	46	99
25.01.16	566218	Rt17	A258	1181	19.5	5.9	bdl	0.2	0.2967	3.2	4.290	4.0	0.1049	2.3	0.81	1675	48	1691	33	1712	43	98
25.01.16	566218	Rt19	A252	3062	46.1	14.7	bdl	1.7	0.3100	2.9	4.576	3.5	0.1071	2.0	0.83	1741	45	1745	30	1750	36	99
25.01.16	566218	Rt19	A253	2377	42.3	13.7	bdl	0.01	0.3145	2.1	4.635	3.2	0.1069	2.4	0.67	1763	33	1756	27	1748	43	101
25.01.16	566218	Rt19	A254	3051	36.6	11.5	0.015	4.1	0.3063	3.4	4.431	4.2	0.1050	2.4	0.81	1722	51	1718	35	1713	45	101
25.01.16	566218	Rt19	A255	2662	43.1	14.0	bdl	bdl	0.3162	2.7	4.689	3.5	0.1076	2.3	0.77	1771	42	1765	30	1759	41	101
25.01.16	566218	Rt19	A256	2617	42.9	13.6	0.002	0.9	0.3076	3.0	4.480	4.0	0.1057	2.6	0.76	1729	46	1727	34	1726	48	100
25.01.16	566218	Rt20	A249	3041	50.4	16.2	bdl	bdl	0.3119	3.0	4.582	3.6	0.1066	1.9	0.85	1750	47	1746	30	1742	35	100
25.01.16	566240	Rt1	A136	845	8.6	2.8	0.009	3.27	0.3151	2.2	4.603	3.3	0.1060	2.5	0.67	1766	34	1750	28	1731	45	102
25.01.16	566240	Rt1	A137	672	10.7	3.3	0.003	0.28	0.3056	2.7	4.471	4.6	0.1061	3.8	0.58	1719	41	1726	39	1734	69	99
25.01.16	566240	Rt2	A144	1173	9.9	3.1	0.001	bdl	0.3092	2.5	4.573	3.5	0.1073	2.4	0.73	1737	39	1744	29	1754	43	99
25.01.16	566240	Rt3	A162	1230	7.9	2.3	0.013	3.5	0.2877	2.2	4.133	3.4	0.1042	2.6	0.65	1630	32	1661	28	1701	48	96
25.01.16	566240	Rt3	A163	1047	8.0	2.5	0.013	0.64	0.2995	2.8	4.346	4.0	0.1053	2.9	0.70	1689	42	1702	34	1719	53	98
25.01.16	566240	Rt3	A164	1055	8.7	2.5	0.056	bdl	0.2817	2.5	4.053	3.5	0.1044	2.5	0.70	1600	35	1645	29	1703	46	94
25.01.16	566240	Rt3	A165	993	8.1	2.3	0.004	0.26	0.2835	2.2	4.107	3.1	0.1051	2.3	0.69	1609	31	1656	26	1716	42	94
25.01.16	566240	Rt4	A166	1463	12.3	3.5	bdl	bdl	0.2779	2.1	4.044	3.4	0.1056	2.6	0.62	1581	29	1643	28	1725	48	92
25.01.16	566240	Rt4	A167	1434	11.9	3.6	bdl	bdl	0.2959	2.1	4.368	3.2	0.1071	2.4	0.66	1671	31	1706	27	1750	44	95
25.01.16	566240	Rt4	A168	1697	11.1	3.2	bdl	2.6	0.2768	2.9	3.955	4.0	0.1037	2.8	0.72	1575	40	1625	33	1691	52	93
25.01.16	566240	Rt5	A161	1652	8.7	2.5	0.045	4.6	0.2743	2.9	4.012	4.3	0.1061	3.2	0.66	1563	40	1637	36	1734	59	90
25.01.16	566240	Rt6	A157	1104	8.4	2.6	bdl	0.73	0.3070	2.4	4.475	4.0	0.1057	3.2	0.61	1726	37	1726	34	1727	58	100
25.01.16	566240	Rt6	A158	1200	8.6	2.7	0.005	1.0	0.3006	2.2	4.423	3.6	0.1067	2.9	0.60	1694	33	1717	31	1744	53	97
25.01.16	566240	Rt6	A159	1436	9.6	2.5	0.007	2.6	0.2542	2.7	3.705	3.9	0.1057	2.8	0.69	1460	36	1572	32	1727	52	85
25.01.16	566240	Rt7	A153	1235	9.6	2.9	bdl	0.60	0.2902	2.2	4.214	3.5	0.1054	2.7	0.63	1643	32	1677	29	1721	50	95
25.01.16	566240	Rt7	A154	1263	9.2	2.9	bdl	1.31	0.3109	2.4	4.573	3.7	0.1067	2.7	0.67	1745	37	1744	31	1744	50	100
25.01.16	566240	Rt7	A155	955	7.7	2.3	bdl	bdl	0.2893	2.5	4.246	2.6	0.1065	1.0	0.93	1638	36	1683	22	1740	18	94
25.01.16	566240	Rt7	A156	2601	9.2	2.8	0.014	12.7	0.3008	2.8	4.304	4.8	0.1038	3.9	0.58	1695	41	1694	41	1693	73	100
25.01.16	566240	Rt8	A139	696	10.0	3.0	bdl	0.20	0.2913	2.7	4.294	4.3	0.1069	3.3	0.62	1648	39	1692	36	1747	61	94

Date	Sample	Grain	Meas. No	²⁰⁷ Pb ^a (cps)	U ^b (ppm)	Pb ^b (ppm)	Th ^b U	²⁰⁶ Pb ^c (%)	²⁰⁶ Pb ^d ²³⁸ U	±2s ^e (%)	²⁰⁷ Pb ^d ²³⁵ U	±2s ^e (%)	²⁰⁷ Pb ^d ²⁰⁶ Pb	±2s ^e (%)	rho ^f	²⁰⁶ Pb ²³⁸ U	±2s (Ma)	²⁰⁷ Pb ²³⁵ U	±2s (Ma)	²⁰⁷ Pb ²⁰⁶ Pb	±2s (Ma)	conc. ^g (%)
25.01.16	566240	Rt8	A140	694	10.0	3.2	bdl	bdl	0.3074	3.0	4.533	4.6	0.1070	3.5	0.65	1728	46	1737	39	1749	64	99
25.01.16	566240	Rt9	A125	649	8.7	2.7	bdl	bdl	0.2963	2.7	4.357	4.2	0.1067	3.3	0.63	1673	40	1704	35	1743	60	96
25.01.16	566240	Rt9	A126	1190	9.2	2.9	bdl	bdl	0.3079	2.2	4.520	3.5	0.1065	2.7	0.64	1730	34	1735	30	1740	50	99
25.01.16	566240	Rt9	A127	1178	9.4	3.0	bdl	0.24	0.3077	2.0	4.506	3.7	0.1062	3.1	0.55	1730	31	1732	31	1736	56	100
25.01.16	566240	Rt10	A128	1804	14.2	4.6	bdl	bdl	0.3113	2.1	4.525	3.4	0.1055	2.8	0.60	1747	32	1736	29	1722	51	101
25.01.16	566240	Rt10	A129	1765	13.2	3.8	0.052	0.76	0.2806	2.1	4.099	3.5	0.1060	2.8	0.59	1594	29	1654	29	1731	52	92
25.01.16	566240	Rt10	A130	1886	13.7	4.0	0.043	0.67	0.2840	2.4	4.151	3.6	0.1060	2.7	0.66	1612	34	1664	30	1732	50	93
25.01.16	566240	Rt10	A131	1410	10.6	3.3	bdl	0.48	0.2994	2.4	4.368	3.5	0.1058	2.6	0.67	1688	35	1706	30	1729	48	98
25.01.16	566240	Rt11	A121	1562	8.6	2.8	bdl	1.94	0.3155	2.6	4.593	4.6	0.1056	3.8	0.56	1768	40	1748	39	1725	69	102
25.01.16	566240	Rt12	A112	860	6.4	1.9	0.006	0.31	0.2909	2.9	4.229	3.8	0.1055	2.5	0.75	1646	42	1680	32	1723	46	96
25.01.16	566240	Rt12	A113	923	6.8	2.1	0.007	0.58	0.3035	2.8	4.410	3.9	0.1054	2.7	0.71	1709	42	1714	33	1721	51	99
25.01.16	566240	Rt12	A114	941	6.6	2.0	0.030	0.96	0.3008	3.0	4.367	3.6	0.1053	2.0	0.83	1695	45	1706	31	1720	38	99
25.01.16	566240	Rt12	A115	1202	7.1	2.2	0.536	2.25	0.3020	2.3	4.499	3.2	0.1081	2.2	0.72	1701	35	1731	27	1767	41	96
25.01.16	566240	Rt13	A116	1271	7.0	2.2	0.049	3.83	0.3069	2.5	4.387	3.8	0.1037	2.9	0.66	1725	38	1710	32	1691	53	102
25.01.16	566240	Rt13	A117	950	6.9	2.2	0.001	0.21	0.3144	3.4	4.514	4.4	0.1042	2.9	0.75	1762	52	1734	38	1700	54	104
25.01.16	566240	Rt13	A118	1023	6.9	2.1	bdl	1.23	0.3025	2.5	4.397	3.8	0.1055	2.9	0.65	1704	37	1712	32	1722	53	99
25.01.16	566240	Rt13	A119	1122	6.9	2.1	0.004	2.42	0.3002	2.6	4.389	3.9	0.1061	2.8	0.68	1692	39	1710	32	1733	52	98
25.01.16	566240	Ex1Rt	A145	1196	9.8	2.8	0.004	0.13	0.2764	2.6	4.016	3.9	0.1054	2.9	0.66	1573	36	1638	32	1722	53	91
25.01.16	566240	Ex1Rt	A151	1047	8.4	2.6	0.002	0.01	0.2981	2.7	4.377	2.9	0.1065	1.1	0.92	1682	40	1708	24	1740	21	97
25.01.16	566240	Ex1Rt	A152	1006	8.7	2.6	bdl	0.24	0.2941	3.0	4.323	3.9	0.1066	2.5	0.77	1662	44	1698	33	1743	46	95
26.01.16	566267	Rt1	A199	1333	4.1	1.2	0.001	bdl	0.2799	2.8	4.115	4.3	0.1067	3.3	0.64	1591	39	1657	36	1743	61	91
25.01.16	566277	Rt2	A234	252	4.5	1.3	0.008	bdl	0.2878	4.9	4.011	6.0	0.1011	3.4	0.82	1631	71	1636	50	1645	63	99
25.01.16	566277	Rt3	A235	629	10.8	3.1	bdl	bdl	0.2819	4.2	4.182	4.9	0.1076	2.7	0.84	1601	59	1670	41	1759	49	91
25.01.16	566277	Rt4	A244	477	8.6	2.6	0.006	bdl	0.2963	3.7	4.295	4.7	0.1052	3.0	0.78	1673	54	1692	40	1718	55	97
25.01.16	566277	Rt6	A230	543	9.2	2.9	bdl	bdl	0.3045	2.8	4.462	4.1	0.1063	2.9	0.69	1713	43	1724	34	1737	53	99
25.01.16	566277	Rt7	A225	997	9.1	2.9	0.001	bdl	0.3083	2.8	4.573	3.7	0.1076	2.5	0.74	1732	42	1744	32	1759	46	98

Date	Sample	Grain	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pb ^c	²⁰⁶ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g
			No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)
25.01.16	566277	Rt7	A226	1144	9.4	3.0	bdl	1.0	0.3102	2.6	4.523	3.6	0.1058	2.5	0.72	1742	39	1735	30	1728	46	101
25.01.16	566277	Rt8	A221	620	9.8	3.0	0.227	bdl	0.2987	3.3	4.314	4.4	0.1048	2.9	0.75	1685	49	1696	37	1710	54	99
25.01.16	566277	Rt8	A222	1073	18.3	5.3	0.037	0.48	0.2837	2.4	4.007	3.6	0.1025	2.7	0.67	1610	35	1636	30	1669	49	96
25.01.16	566277	Rt9	A172	1000	16.3	5.2	bdl	0.32	0.3106	3.6	4.511	4.6	0.1054	2.8	0.78	1743	55	1733	39	1721	52	101
25.01.16	566277	Rt9	A173	150	4.2	1.3	0.020	bdl	0.3030	5.9	4.473	7.4	0.1071	4.5	0.80	1706	88	1726	63	1751	82	97
25.01.16	566277	Rt10	A169	743	10.5	3.3	bdl	0.21	0.3055	2.8	4.440	3.8	0.1054	2.6	0.74	1719	43	1720	32	1722	47	100
25.01.16	566277	Rt10	A170	635	9.9	2.8	bdl	bdl	0.2755	3.4	4.085	4.5	0.1076	3.0	0.75	1569	47	1651	37	1759	54	89
25.01.16	566277	Rt10	A171	1048	16.1	4.9	bdl	0.17	0.2964	2.7	4.324	3.9	0.1058	2.8	0.70	1674	40	1698	33	1729	51	97
25.01.16	566277	Rt11	A183	409	6.8	1.9	0.021	1.5	0.2692	5.8	3.801	6.5	0.1025	3.0	0.89	1537	79	1593	54	1669	56	92
25.01.16	566277	Rt11	A184	926	13.6	4.4	bdl	bdl	0.3134	2.4	4.573	3.3	0.1059	2.3	0.73	1757	38	1744	28	1729	42	102
25.01.16	566277	Rt13	A189	130	4.3	1.3	bdl	bdl	0.2914	6.5	4.291	7.8	0.1069	4.3	0.83	1648	96	1692	67	1746	79	94
25.01.16	566277	Rt13	A190	97	4.0	1.2	0.004	bdl	0.2885	7.0	4.204	8.5	0.1057	4.8	0.82	1634	102	1675	72	1727	88	95

^a Within run background-corrected mean ²⁰⁷Pb signal in cps (counts per second).

^b U and Pb content and Th/U ratio were calculated relative to GJ-1 reference zircon. bdl = below detection limit.

^c percentage of the common Pb on the ²⁰⁶Pb. bdl = below detection limit.

^d corrected for background, within-run Pb/U fractionation (in case of ²⁰⁶Pb/²³⁸U) and common Pb using Stacy and Kramers (1975) model Pb composition and subsequently normalized to GJ-1 (ID-TIMS value/measured value); ²⁰⁷Pb/²³⁵U calculated using ²⁰⁷Pb/²⁰⁶Pb/(²³⁸U/²⁰⁶Pb*1/137.88)

^e ²⁰⁶Pb/²³⁸U error is the quadratic additions of the within run precision (2 SE) and the external reproducibility (2 SD) of the reference zircon. ²⁰⁷Pb/²⁰⁶Pb error propagation (²⁰⁷Pb signal dependent) following

Gerdes & Zeh (2009). ²⁰⁷Pb/²³⁵U error is the quadratic addition of the ²⁰⁷Pb/²⁰⁶Pb and ²⁰⁶Pb/²³⁸U uncertainty.

^f rho is the ²⁰⁶Pb/²³⁸U/²⁰⁷Pb/²³⁵U error correlation coefficient.

^g degree of concordance = ²⁰⁶Pb/²³⁸U age / ²⁰⁷Pb/²⁰⁶Pb age x 100.

Appendix 10 – Zircon U-Pb data (GEUS)

Date	Sample	Grain	Meas. No	U ^a (ppm)	Pb ^a (ppm)	Th ^a U	²⁰⁶ Pb ^b ²³⁸ U	±2s ^c (absolute)	²⁰⁷ Pb ^b ²³⁵ U	±2s ^c (absolute)	²⁰⁷ Pb ^b ²⁰⁶ Pb	±2s ^c (absolute)	rho ^d	²⁰⁶ Pb ²³⁸ U	±2s (Ma)	²⁰⁷ Pb ²³⁵ U	±2s (Ma)	²⁰⁷ Pb ²⁰⁶ Pb	±2s (Ma)	FinalDisc. (%)
20.12.16	566218	Zr1	92	229	1120	0.568	0.3788	0.0047	6.898	0.092	0.132	0.00067	0.83	2071	22	2098	12	2118	9	2.4
20.12.16	566218	Zr1	98	190	801	0.512	0.3658	0.0088	6.66	0.12	0.132	0.0016	0.92	2009	42	2068	16	2124	21	5.3
20.12.16	566218	Zr1	99	225	1020	0.536	0.367	0.0034	6.612	0.083	0.13	0.00087	0.85	2015	16	2061	11	2104	12	4.21
20.12.16	566218	Zr1	100	218	844	0.466	0.3332	0.0069	5.91	0.15	0.129	0.0013	0.92	1854	33	1963	22	2078	18	10.8
20.12.16	566218	Zr3	85	650	1376	1.754	0.363	0.0069	6.09	0.12	0.121	0.00083	0.83	1996	33	1988	18	1968	12	-1
20.12.16	566218	Zr3	86	460	1440	0.524	0.3588	0.006	6.09	0.13	0.122	0.0012	0.87	1976	29	1988	19	1991	17	0.6
20.12.16	566218	Zr3	87	671	1450	0.362	0.3252	0.005	5.39	0.12	0.12	0.001	0.91	1815	24	1883	19	1952	15	6.8
20.12.16	566218	Zr3	87	350	770	0.278	0.3326	0.0099	5.49	0.2	0.12	0.0015	0.98	1850	48	1899	32	1950	22	5.1
20.12.16	566218	Zr4	72	170	680	0.513	0.358	0.0051	6.36	0.076	0.128	0.00046	0.65	1973	24	2027	10	2077	6.3	5
20.12.16	566218	Zr4	73	168	705	0.535	0.365	0.013	6.68	0.25	0.132	0.0019	0.93	2005	62	2069	32	2127	26	5.7
20.12.16	566218	Zr4	74	202	979	0.588	0.3841	0.0081	6.9	0.13	0.13	0.00035	0.98	2095	38	2099	17	2097	4.7	0.1
20.12.16	566218	Zr6	77	279	1500	0.614	0.3521	0.009	6.05	0.16	0.124	0.0022	0.77	1944	43	1982	23	2018	31	3.6
20.12.16	566218	Zr6	78	306	1640	0.469	0.3445	0.0053	5.83	0.16	0.122	0.0019	0.09	1908	25	1950	24	1990	28	4.1
20.12.16	566218	Zr6	79	306	650	0.342	0.321	0.012	5.24	0.19	0.118	0.00049	0.99	1793	61	1858	32	1928	7.5	7
20.12.16	566218	Zr7	52	635	7700	1.443	0.387	0.004	7.105	0.063	0.132	0.00088	0.85	2109	18	2125	7.9	2129	12	1.2
20.12.16	566218	Zr7	53	826	9500	1.412	0.3915	0.0072	7.1800	0.1200	0.1326	0.0011	0.84	2130	33	2133	15	2133	14	0.8
20.12.16	566218	Zr8	90	19.7	2.87	0.007	0.3153	0.0088	5.1	0.14	0.116	0.0032	0.58	1766	43	1835	24	1890	48	6
20.12.16	566240	Zr1	36	16.1	50	0.154	0.4100	0.0570	7	1	0.125	0.002	0.99	2200	260	2130	140	2024	28	-10
20.12.16	566240	Zr1	38	80.2	293	0.427	0.3490	0.0058	5.97	0.15	0.125	0.0015	0.86	1930	28	1971	23	2023	21	4.6
20.12.16	566240	Zr2	39	15	93	0.137	0.4870	0.0710	8	1	0.119	0.0034	0.98	2610	330	2200	110	1935	51	-35
20.12.16	566240	Zr2	40	10.69	5.9	0.047	0.3495	0.0068	5.51	0.18	0.116	0.0029	0.53	1932	32	1907	26	1886	46	-3
20.12.16	566240	Zr3	34	4.46	1.05	0.008	0.3380	0.0130	5.72	0.23	0.123	0.003	0.81	1874	61	1933	34	1992	43	5.8
20.12.16	566240	Zr4	47	1168	7830	0.813	0.3510	0.0077	5.97	0.17	0.123	0.0013	0.96	1939	37	1971	25	2000	18	2.5
20.12.16	566240	Zr4	48	1067	12020	1.431	0.3490	0.0110	6.26	0.21	0.13	0.00083	0.96	1930	53	2012	30	2094	11	7.8
20.12.16	566240	Zr4	49	681	4010	0.787	0.3730	0.0130	6.24	0.24	0.121	0.0012	0.98	2044	62	2009	34	1970	17	-3.8

Date	Sample	Grain	Meas. No	U ^a (ppm)	Pb ^a (ppm)	Th ^a U	²⁰⁶ Pb ^b ²³⁸ U	±2s ^c (absolute)	²⁰⁷ Pb ^b ²³⁵ U	±2s ^c (absolute)	²⁰⁷ Pb ^b ²⁰⁶ Pb	±2s ^c (absolute)	rho ^d	²⁰⁶ Pb ²³⁸ U	±2s (Ma)	²⁰⁷ Pb ²³⁵ U	±2s (Ma)	²⁰⁷ Pb ²⁰⁶ Pb	±2s (Ma)	FinalDisc. (%)
20.12.16	566240	Zr7	25	7.19	11	0.073	0.3230	0.0150	6.38	0.55	0.151	0.012	0.52	1804	72	2043	82	2330	140	22.8
20.12.16	566240	Zr7	26	4.9	1.2	0.027	0.3500	0.0110	5.82	0.31	0.123	0.0057	0.01	1933	52	1946	47	1996	81	3.3
20.12.16	566240	Zr7	33	35	198	0.336	0.5040	0.0750	8.4	1.3	0.121	0.0036	0.98	2610	320	2260	140	1971	53	-33
20.12.16	566240	Zr8	51	36.7	248	0.667	0.3650	0.0110	6.44	0.1	0.127	0.0033	0.57	2004	54	2037	14	2052	46	2.1
20.12.16	566249	Zr1	181	1860	2010	0.226	0.323	0.016	5.1	0.28	0.114	0.0015	0.96	1805	76	1835	47	1871	24	3.5
20.12.16	566249	Zr1	182	1600	3400	0.114	0.408	0.057	6.6	0.99	0.117	0.0016	1	2200	260	2100	160	1908	24	-15
20.12.16	566249	Zr1	183	1690	1340	0.141	0.325	0.021	5.17	0.36	0.114	0.0014	1	1810	100	1843	57	1870	21	3.6
20.12.16	566249	Zr2	150	817	31.4	0.007	0.3274	0.0054	5.21	0.1	0.115	0.0011	0.86	1826	26	1854	17	1884	18	3.1
20.12.16	566249	Zr2	151	373	33.1	0.015	0.293	0.0045	4.674	0.082	0.115	0.001	0.89	1657	23	1762	15	1886	16	12
20.12.16	566249	Zr3	125	257	12.2	0.005	0.3341	0.0069	5.37	0.11	0.115	0.0013	0.89	1858	33	1880	18	1885	21	1.4
20.12.16	566249	Zr3	126	200	159	0.049	0.322	0.013	5.27	0.18	0.119	0.00067	1	1798	62	1863	30	1940	10	5
20.12.16	566249	Zr3	127	1300	4600	0.513	0.3246	0.0056	5.426	0.057	0.121	0.0014	0.97	1812	27	1889	9.1	1974	20	9.6
20.12.16	566249	Zr4	103	463	86	0.011	0.3474	0.0086	5.53	0.14	0.115	0.00068	0.96	1922	41	1904	22	1880	11	-2
20.12.16	566249	Zr4	104	132	101	0.104	0.3374	0.0087	5.5	0.16	0.118	0.0014	0.92	1874	42	1900	24	1926	21	2.7
20.12.16	566249	Zr5	101	340	260	0.122	0.312	0.029	4.95	0.44	0.115	0.0017	0.99	1740	140	1802	75	1872	26	7.1
20.12.16	566249	Zr5	102	425	18.1	0.011	0.3421	0.0085	5.46	0.16	0.116	0.0023	0.74	1897	41	1895	25	1889	36	-0.5
20.12.16	566249	Zr6	176	338	43.7	0.031	0.2895	0.0076	4.594	0.093	0.116	0.0013	0.91	1639	38	1748	17	1896	21	13.5
20.12.16	566249	Zr7	177	4010	20300	0.621	0.3603	0.0059	5.72	0.069	0.115	0.0015	0.66	1984	28	1934	10	1881	24	-5.5
20.12.16	566249	Zr8	105	290	25	0.013	0.41	0.12	6.6	1.8	0.116	0.0015	1	2200	530	2020	210	1891	23	-17
20.12.16	566249	Zr8	124	400	29	0.010	0.47	0.1	7.5	1.6	0.115	0.00071	1	2440	430	2210	220	1884	11	-29
20.12.16	566249	Zr9	128	570	13.2	0.004	0.3295	0.0077	5.24	0.11	0.115	0.0011	0.92	1836	37	1859	18	1882	17	2.4
20.12.16	566249	Zr9	129	404	8.6	0.003	0.4008	0.0089	6.37	0.11	0.115	0.0019	0.73	2172	41	2028	15	1881	31	-15.6
20.12.16	566249	Zr9	130	950	27.3	0.003	0.36	0.01	5.71	0.16	0.115	0.00069	0.95	1981	48	1932	24	1885	11	-5.5
20.12.16	566249	Zr9	131	520	30	0.011	0.314	0.019	5.02	0.26	0.115	0.0017	0.99	1761	92	1821	43	1886	27	6.8
20.12.16	566249	Zr10	137	496	8.99	0.002	0.366	0.013	5.82	0.22	0.115	0.00099	0.98	2012	62	1948	33	1879	15	-7.1
20.12.16	566249	Zr10	138	480	24.7	0.009	0.375	0.037	6.01	0.65	0.116	0.0016	1	2050	170	2010	120	1888	25	-8.4
20.12.16	566249	Zr10	139	370	10	0.007	0.54	0.21	8.5	3.3	0.115	0.0014	1	2640	760	2180	280	1878	21	-40

Date	Sample	Grain	Meas. No	U ^a (ppm)	Pb ^a (ppm)	Th ^a U	²⁰⁶ Pb/ ²³⁸ U ^b	±2s ^c (absolute)	²⁰⁷ Pb/ ²³⁵ U ^b	±2s ^c (absolute)	²⁰⁷ Pb/ ²⁰⁶ Pb ^b	±2s ^c (absolute)	rho ^d	²⁰⁶ Pb/ ²³⁸ U	±2s (Ma)	²⁰⁷ Pb/ ²³⁵ U	±2s (Ma)	²⁰⁷ Pb/ ²⁰⁶ Pb	±2s (Ma)	FinalDisc. (%)
20.12.16	566249	Zr11	143	421	26.5	0.009	0.3465	0.0062	5.499	0.096	0.115	0.001	0.97	1918	30	1900	15	1882	16	-2.1
20.12.16	566249	Zr11	144	680	25	0.009	0.309	0.027	4.92	0.39	0.115	0.0015	0.99	1740	140	1805	67	1884	24	7.9
20.12.16	566249	Zr12	152	386	25.4	0.009	0.3225	0.0053	5.099	0.073	0.115	0.001	0.92	1802	26	1836	12	1873	16	3.8
20.12.16	566249	Zr12	153	490	17.2	0.006	0.3432	0.0089	5.42	0.13	0.115	0.00085	0.95	1901	42	1887	20	1876	13	-1.5
20.12.16	566249	Zr12	154	442	15	0.004	0.3493	0.008	5.55	0.13	0.115	0.0007	0.96	1931	38	1908	20	1878	11	-3.2
20.12.16	566249	Zr13	155	2770	4900	0.270	0.2928	0.0036	4.64	0.12	0.115	0.0022	0.81	1655	18	1757	22	1878	35	11.8
20.12.16	566249	Zr13	156	820	450	0.125	0.288	0.028	4.59	0.39	0.116	0.002	0.99	1630	140	1744	72	1888	31	13.4
20.12.16	566249	Zr13	157	1560	3600	0.233	0.381	0.042	6.02	0.69	0.115	0.0011	1	2080	200	1970	100	1873	16	-11

^a U and Pb content and Th/U ratio were calculated relative to GJ-1 reference zircon.

^b corrected for background, within-run Pb/U fractionation (in case of ²⁰⁶Pb/²³⁸U) and common Pb using Stacy and Kramers (1975) model Pb composition and subsequently normalized to GJ-1 (ID-TIMS value/measured value); ²⁰⁷Pb/²³⁵U calculated using ²⁰⁷Pb/²⁰⁶Pb/(²³⁸U/²⁰⁶Pb*1/137.88).

^c Quadratic addition of within-run errors (2 SD) and daily reproducibility of GJ-1 (2 SD).

^d rho is the ²⁰⁶Pb/²³⁸U/²⁰⁷Pb/²³⁵U error correlation coefficient.

Appendix 11 – U-Pb data for the reference standards (GUF)

Date	Standard	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pb ^c	²⁰⁶ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g	²⁰⁸ Pb	±2s	²⁰⁸ Pb	±2s
		No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)	²³² Th	(%)	²³² Th	(Ma)
25.01.16	Zrn 91500	A05	33444	426	79	0.45	0.00	0.1786	1.5	1.832	1.8	0.074	0.9	0.87	1060	15	1057	12	1052	17	101				
25.01.16	Zrn 91500	A37	32099	429	78	0.44	0.00	0.1775	1.5	1.838	1.8	0.075	0.9	0.87	1054	15	1059	12	1071	17	98				
25.01.16	Zrn 91500	A106	33458	508	93	0.45	0.00	0.1779	1.5	1.830	1.7	0.075	0.8	0.88	1056	15	1056	12	1058	17	100				
25.01.16	Zrn 91500	A150	32273	540	100	0.45	bdl	0.1796	1.5	1.850	1.8	0.075	1.0	0.84	1065	15	1064	12	1061	20	100				
25.01.16	Zrn 91500	A242	31904	674	130	0.45	bdl	0.1806	1.5	1.882	1.7	0.076	0.8	0.89	1070	15	1075	12	1084	16	99				
25.01.16	Zrn 91500	A287	29228	716	130	0.44	0.08	0.1791	1.5	1.837	1.8	0.074	1.0	0.83	1062	15	1059	12	1052	20	101				
25.01.16	Zrn 91500	A332	33527	993	180	0.46	0.06	0.1778	1.5	1.817	1.8	0.074	1.0	0.84	1055	15	1052	12	1045	20	101				
25.01.16	Zrn 91500	A387	31703	1202	220	0.46	0.09	0.1790	1.5	1.842	1.8	0.075	0.9	0.85	1062	15	1061	12	1059	19	100				
25.01.16	Zrn 91500	A498	26638	2776	510	0.43	0.01	0.1804	1.5	1.835	1.8	0.074	0.9	0.85	1069	15	1058	12	1035	19	103				
25.01.16	Zrn 91500	A553	32393	18696	3500	0.46	0.03	0.1792	1.5	1.845	1.7	0.075	0.9	0.87	1063	15	1062	12	1060	17	100				
26.01.16	Zrn 91500	A03	14077	201	37	0.45	0.08	0.1802	1.1	1.868	1.6	0.075	1.1	0.70	1068	11	1070	11	1073	23	100				
26.01.16	Zrn 91500	A35	13246	159	30	0.46	0.09	0.1803	1.1	1.870	1.8	0.075	1.4	0.62	1069	11	1070	12	1074	28	100				
26.01.16	Zrn 91500	A69	12681	133	25	0.46	0.40	0.1785	1.1	1.832	1.8	0.074	1.4	0.62	1059	11	1057	12	1053	28	101				
26.01.16	Zrn 91500	A104	12252	113	21	0.46	0.02	0.1789	1.1	1.856	1.7	0.075	1.3	0.64	1061	10	1066	11	1076	26	99				
26.01.16	Zrn 91500	A148	11640	92	17	0.46	0.17	0.1802	1.1	1.864	1.8	0.075	1.4	0.61	1068	11	1069	12	1070	28	100				
26.01.16	Zrn 91500	A196	11009	75	14	0.47	0.02	0.1803	1.1	1.874	1.8	0.075	1.5	0.61	1069	11	1072	12	1079	29	99				
26.01.16	Mz Mana.	A213	5398	209	300	76	0.31	0.0883	1.8	0.720	2.7	0.059	2.0	0.67	545	10	551	12	573	43	95	0.0276	1.2	551	7
26.01.16	Mz Mana.	A214	3443	129	260	109	0.00	0.0882	2.3	0.720	3.1	0.059	2.1	0.74	545	12	551	13	574	45	95	0.0277	1.2	553	6
26.01.16	Mz Mana.	A242	4938	176	260	77	0.00	0.0894	2.0	0.719	3.2	0.058	2.5	0.62	552	11	550	14	541	55	102	0.0275	1.4	549	8
26.01.16	Mz Mana.	A287	3181	99	200	107	0.46	0.0908	2.4	0.730	3.6	0.058	2.7	0.67	560	13	557	16	542	59	103	0.0273	1.2	544	7
26.01.16	Mz Mana.	A292	4868	150	230	80	0.46	0.0924	1.8	0.748	2.6	0.059	1.8	0.70	569	10	567	11	557	40	102	0.0278	1.3	554	7
26.01.16	Mz Namaqua.	A211	31235	502	1300	71.11	0.09	0.1713	1.1	1.733	1.4	0.073	0.8	0.82	1019	11	1021	9	1025	16	99	0.0521	1.2	1027	12
26.01.16	Mz Namaqua.	A212	30786	500	1300	71.68	0.00	0.1716	1.2	1.734	1.4	0.073	0.8	0.82	1021	11	1021	9	1022	16	100	0.0524	1.2	1032	12

Date	Standard	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pb ^c	²⁰⁶ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	²⁰⁷ Pb ^d	±2s ^e	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g	²⁰⁸ Pb	±2s	²⁰⁸ Pb	±2s
		No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)	²³² Th	(%)	²³² Th	(Ma)
26.01.16	Mz Namaqua.	A240	31020	458	1200	71	0.13	0.1735	1.2	1.763	1.5	0.074	0.8	0.82	1032	12	1032	10	1033	17	100	0.0519	1.3	1023	13
26.01.16	Mz Namaqua.	A241	30875	458	1200	71.59	0.00	0.1733	1.2	1.755	1.5	0.074	0.9	0.79	1030	11	1029	10	1027	18	100	0.0525	1.2	1034	12
26.01.16	Mz Namaqua.	A285	29542	403	1100	71.30	0.00	0.1733	1.2	1.758	1.5	0.074	0.9	0.78	1030	11	1030	10	1029	19	100	0.0519	1.2	1023	12
26.01.16	Mz Namaqua.	A286	29840	402	1100	72.11	0.00	0.1727	1.1	1.741	1.5	0.073	0.9	0.77	1027	11	1024	10	1017	19	101	0.0519	1.2	1022	12
26.01.16	Mz Namaqua.	A290	29052	385	1000	71.60	0.02	0.1735	1.2	1.750	3.0	0.073	2.8	0.38	1031	11	1027	20	1019	57	101	0.0525	1.1	1035	11
26.01.16	Mz Namaqua.	A291	29117	388	1000	72.20	0.20	0.1749	1.1	1.793	1.5	0.074	1.0	0.76	1039	11	1043	10	1051	19	99	0.0521	1.3	1026	13
26.01.16	Mz Namaqua.	A294	11973	149	410	71.77	0.00	0.1728	1.2	1.761	1.8	0.074	1.4	0.65	1027	11	1031	12	1039	28	99	0.0526	1.4	1036	14
25.01.16	Ttn Bear Lake	A04	36309	1008	280	1.84	2.77	0.17450	1.8	1.781	3.6	0.074	3.1	0.50	1037	17	1039	23	1042	62	100				
25.01.16	Ttn Bear Lake	A36	17437	496	140	2.74	2.79	0.17720	1.8	1.802	3.4	0.0738	2.9	0.52	1051	17	1046	22	1035	58	102				
25.01.16	Ttn Bear Lake	A70	16684	486	140	2.77	3.24	0.17720	1.7	1.782	3.2	0.0729	2.7	0.54	1052	17	1039	21	1012	54	104				
25.01.16	Ttn Bear Lake	A105	15515	479	140	2.79	2.79	0.17890	1.7	1.857	3.1	0.0753	2.6	0.56	1061	17	1066	21	1077	52	99				
25.01.16	Ttn Bear Lake	A149	14612	462	130	2.85	2.93	0.17970	1.7	1.840	3.2	0.0743	2.7	0.52	1065	16	1060	21	1049	55	102				
25.01.16	Ttn Bear Lake	A197	13151	472	130	2.86	2.65	0.17910	1.9	1.822	3.0	0.0738	2.3	0.63	1062	19	1053	20	1036	47	103				
25.01.16	Ttn Bear Lake	A241	14940	464	140	2.90	5.43	0.17840	1.8	1.808	3.2	0.0735	2.6	0.56	1058	17	1048	21	1027	53	103				
25.01.16	Ttn Bear Lake	A285	12474	552	160	2.83	2.62	0.17640	1.8	1.837	3.1	0.0756	2.5	0.58	1047	17	1059	20	1083	50	97				
25.01.16	Ttn Bear Lake	A313	12192	589	105	2.92	2.7	0.1781	1.8	1.830	3.4	0.0745	2.9	0.53	1057	18	1056	23	1056	58	100				
25.01.16	Ttn Bear Lake	A314	11599	586	103	2.88	2.6	0.1757	1.7	1.808	2.7	0.0747	2.2	0.61	1043	16	1048	18	1060	43	98				
25.01.16	Ttn Bear Lake	A315	11463	539	96	2.86	2.9	0.1783	1.7	1.831	3.6	0.0745	3.2	0.45	1058	16	1057	24	1055	65	100				
25.01.16	Ttn Bear Lake	A316	13046	556	99	2.88	4.6	0.1774	1.7	1.837	4.1	0.0751	3.7	0.42	1053	17	1059	27	1072	75	98				
25.01.16	Ttn Bear Lake	A317	12643	538	96	2.89	3.6	0.1788	1.6	1.826	4.2	0.0741	3.9	0.39	1060	16	1055	28	1044	79	102				
25.01.16	Ttn Bear Lake	A318	4987	519	92	2.80	2.60	0.17734	2.0	1.8	6	0.0733	5.9	0.32	1052	19	1043	41	1023	119	103				
25.01.16	Ttn Bear Lake	A319	5059	485	87	2.85	2.53	0.17835	1.9	1.9	6	0.0769	5.2	0.34	1058	18	1077	38	1118	104	95				
25.01.16	Ttn Bear Lake	A320	6404	503	90	2.86	5.34	0.17814	2.1	1.8	5	0.0729	4.5	0.42	1057	20	1042	33	1012	91	104				

Date	Standard	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pb ^c	²⁰⁶ Pb ^d	±2s ^c	²⁰⁷ Pb ^d	±2s ^c	²⁰⁷ Pb ^d	±2s ^c	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g	²⁰⁸ Pb	±2s	²⁰⁸ Pb	±2s
		No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)	²³² Th	(%)	²³² Th	(Ma)
	Ttn																								
26.01.16	Bear Lake Ttn	A04	12906	336	95	2.77	2.70	0.17600	1.2	1.794	2.0	0.0739	1.6	0.61	1045	12	1043	13	1040	32	101				
26.01.16	Bear Lake Ttn	A36	12067	267	76	2.83	2.77	0.17680	1.2	1.793	2.1	0.0736	1.7	0.58	1049	12	1043	14	1030	35	102				
26.01.16	Bear Lake Ttn	A70	12005	222	64	2.83	2.59	0.18020	1.3	1.853	2.3	0.0746	1.9	0.54	1068	12	1064	15	1057	39	101				
26.01.16	Bear Lake Ttn	A105	11298	186	53	2.86	2.34	0.17750	1.3	1.821	2.0	0.0744	1.5	0.66	1053	13	1053	13	1052	31	100				
26.01.16	Bear Lake Ttn	A149	10932	154	44	2.87	2.74	0.17720	1.2	1.821	2.0	0.0746	1.6	0.62	1051	12	1053	13	1056	31	100				
26.01.16	Bear Lake Ttn	A197	10413	129	37	2.89	2.39	0.17830	1.4	1.839	2.8	0.0748	2.4	0.51	1058	14	1059	18	1063	48	99				
25.01.16	Ttn Namaqua.	A304	14480	715	120	5.01	2.1	0.1687	1.8	1.677	3.7	0.0722	3.2	0.49	1005	17	1000	24	990	66	101				
25.01.16	Ttn Namaqua.	A305	15266	780	134	3.82	1.3	0.1716	1.8	1.722	2.9	0.0728	2.2	0.64	1021	17	1017	19	1009	45	101				
25.01.16	Ttn Namaqua.	A306	15354	779	133	3.97	1.5	0.1713	1.8	1.733	4.3	0.0734	3.9	0.42	1019	17	1021	28	1024	79	100				
25.01.16	Ttn Namaqua.	A307	14875	766	130	4.17	1.6	0.17	1.8	1.698	2.9	0.0725	2.3	0.60	1012	17	1008	19	999	48	101				
25.01.16	Ttn Namaqua.	A308	15085	809	137	3.87	2.0	0.1701	1.7	1.685	3.0	0.0719	2.5	0.57	1013	16	1003	19	982	51	103				
25.01.16	Ttn Namaqua.	A309	15506	800	135	3.77	1.8	0.1689	1.8	1.685	3.1	0.0724	2.6	0.58	1006	17	1003	20	997	52	101				
25.01.16	Ttn Namaqua.	A310	6621	761	131	5.59	1.6	0.1718	2.0	1.698	4.0	0.0717	3.5	0.49	1022	19	1008	26	978	71	104				
25.01.16	Ttn Namaqua.	A311	7366	767	134	5.55	2.1	0.1744	1.9	1.71	4.1	0.0711	3.6	0.47	1036	18	1012	26	961	73	108				
25.01.16	Ttn Namaqua.	A312	10195	763	131	5.53	5.6	0.1722	2.0	1.708	3.9	0.072	3.4	0.50	1024	19	1012	25	985	69	104				
25.01.16	Rt Tábor	A05	28652	177	13	0.00	15.25	0.0538	2.9	0.396	3.7	0.0534	2.4	0.77	338	9	339	11	345	54	98				
25.01.16	Rt Tábor	A37	24434	79	12	0.00	36.12	0.0538	2.7	0.395	3.3	0.0533	2.0	0.81	338	9	338	10	339	44	100				
25.01.16	Rt Tábor	A71	7262	225	11	0.00	0.24	0.0538	2.6	0.394	2.9	0.0531	1.3	0.90	338	9	337	8	333	29	101				
25.01.16	Rt Tábor	A106	3842	123	6.2	0.00	0.86	0.0532	2.5	0.390	3.2	0.0532	2.0	0.77	334	8	335	9	338	46	99				
25.01.16	Rt Tábor	A150	34354	122	14	0.00	36.05	0.0541	2.4	0.402	3.5	0.0539	2.6	0.69	339	8	343	10	366	58	93				
25.01.16	Rt Tábor	A198	16936	288	17	0.00	5.27	0.0543	2.8	0.401	3.7	0.0535	2.4	0.75	341	9	342	11	349	55	98				
25.01.16	Rt Tábor	A242	15888	85	9	0.00	28.25	0.0534	2.0	0.392	2.6	0.0533	1.7	0.76	335	7	336	7	340	38	99				
26.01.16	Rt Tábor	A05	2717	139	6.9	0.00	0.53	0.0535	2.6	0.393	3.9	0.0533	2.9	0.68	336	9	336	11	341	65	98				

Date	Standard	Meas.	²⁰⁷ Pb ^a	U ^b	Pb ^b	Th ^b	²⁰⁶ Pb ^c	²⁰⁶ Pb ^d	±2s ^c	²⁰⁷ Pb ^d	±2s ^c	²⁰⁷ Pb ^d	±2s ^c	rho ^f	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	conc. ^g	²⁰⁸ Pb	±2s	²⁰⁸ Pb	±2s
		No	(cps)	(ppm)	(ppm)	U	(%)	²³⁸ U	(%)	²³⁵ U	(%)	²⁰⁶ Pb	(%)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)	²³² Th	(%)	²³² Th	(Ma)
26.01.16	Rt Tábor	A37	2413	111	5.5	0.00	0.59	0.0537	1.9	0.397	3.3	0.0537	2.7	0.57	337	6	340	10	359	61	94				
26.01.16	Rt Tábor	A71	3901	137	7	0.00	1.00	0.0539	1.7	0.390	2.8	0.0525	2.3	0.59	338	6	335	8	309	52	109				
26.01.16	Rt Tábor	A106	3785	117	5.8	0.00	0.68	0.0531	2.1	0.393	3.2	0.0537	2.3	0.67	333	7	337	9	358	53	93				
26.01.16	Rt Tábor	A150	3306	91	4.6	0.00	0.47	0.0540	1.9	0.403	3.1	0.0541	2.5	0.62	339	6	344	9	374	55	91				
26.01.16	Rt Tábor	A198	2133	61	3	0.00	0.55	0.0533	2.3	0.385	4.0	0.0524	3.2	0.59	335	8	331	11	303	73	110				
25.01.16	Rt SRQ	A321	57955	971	363	0.00	0.1	0.3585	2.2	6.066	2.5	0.1227	1.2	0.87	1975	37	1985	22	1996	22	99				
25.01.16	Rt SRQ	A322	59451	1016	391	0.00	0.0	0.3689	2.3	6.315	2.5	0.1242	1.0	0.92	2024	40	2020	22	2017	17	100				
25.01.16	Rt SRQ	A323	58804	978	369	0.00	0.0	0.3617	2.3	6.121	2.7	0.1228	1.4	0.85	1990	39	1993	24	1997	25	100				
25.01.16	Rt SRQ	A324	55420	886	340	0.00	0.0	0.3674	2.2	6.296	2.5	0.1243	1.2	0.87	2017	38	2018	22	2019	22	100				
25.01.16	Rt SRQ	A325	28883	1225	483	0.00	0.0	0.3776	3.0	6.512	3.5	0.1251	1.9	0.85	2065	53	2048	32	2030	33	102				
25.01.16	Rt SRQ	A326	28875	1103	424	0.00	0.0	0.3674	2.4	6.313	3.1	0.1246	1.9	0.78	2017	42	2020	28	2024	34	100				
25.01.16	Rt SRQ	A327	22596	773	300	0.00	0.1	0.3718	2.1	6.318	2.3	0.1233	0.9	0.93	2038	37	2021	20	2004	15	102				
25.01.16	Rt SRQ	A328	25103	889	342	0.00	0.3	0.3691	2.5	6.234	2.7	0.1225	0.9	0.94	2025	44	2009	23	1993	16	102				
25.01.16	Rt SRQ	A329	992	14	5	0.02	0.8	0.3649	2.2	6.215	2.8	0.1236	1.7	0.78	2006	38	2007	25	2008	31	100				
25.01.16	Rt SRQ	A330	47630	833	316	0.00	0.2	0.3636	1.9	6.142	2.0	0.1226	0.7	0.93	1999	32	1996	18	1994	13	100				

^a Within run background-corrected mean ²⁰⁷Pb signal in cps (counts per second).

^b U and Pb content and Th/U ratio were calculated relative to GJ-1 reference zircon.

^c percentage of the common Pb on the ²⁰⁶Pb. bdl = below detection limit.

^d corrected for background, within-run Pb/U fractionation (in case of ²⁰⁶Pb/²³⁸U) and common Pb using Stacy and Kramers (1975) model Pb composition and subsequently normalized to GJ-1 (ID-TIMS value/measured value); ²⁰⁷Pb/²³⁵U calculated using ²⁰⁷Pb/²⁰⁶Pb/(²³⁸U/²⁰⁶Pb*1/137.88)

^e ²⁰⁶Pb/²³⁸U error is the quadratic additions of the within run precision (2 SE) and the external reproducibility (2 SD) of the reference zircon. ²⁰⁷Pb/²⁰⁶Pb error propagation (²⁰⁷Pb signal dependent) following Gerdes & Zeh (2009). ²⁰⁷Pb/²³⁵U error is the quadratic addition of the ²⁰⁷Pb/²⁰⁶Pb and ²⁰⁶Pb/²³⁸U uncertainty.

^f rho is the ²⁰⁶Pb/²³⁸U/²⁰⁷Pb/²³⁵U error correlation coefficient.

^g degree of concordance = ²⁰⁶Pb/²³⁸U age / ²⁰⁷Pb/²⁰⁶Pb age x 100.

Appendix 12 – U-Pb data for the reference standards (GEUS)

Date	Sample	Meas.	U ^a	Pb ^a	Th ^a	²⁰⁶ Pb ^b	±2s ^c	²⁰⁷ Pb ^b	±2s ^c	²⁰⁷ Pb ^b	±2s ^c	rho ^d	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	FinalDisc.
		No	(ppm)	(ppm)	U	²³⁸ U	(absolute)	²³⁵ U	(absolute)	²⁰⁶ Pb	(absolute)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)
20.12.16	Z_ Plešovice	016	1243	90.6	0.0524	0.0542	0.00037	0.3986	0.0042	0.05361	0.00058	0.6034	340.2	2.2	340.6	3	354	24	3.2
20.12.16	Z_ Plešovice	029	1020	57.9	0.0459	0.05423	0.00012	0.406	0.011	0.0544	0.0012	0.8125	340.46	0.75	345.9	8.2	385	51	10
20.12.16	Z_ Plešovice	042	916	60.5	0.0470	0.05396	0.00041	0.4023	0.006	0.05387	0.00059	0.5236	338.8	2.5	343.3	4.3	365	25	7.6
20.12.16	Z_ Plešovice	055	1250	113	0.0578	0.05364	0.00026	0.4073	0.0036	0.05464	0.00082	0.0915	336.8	1.6	346.9	2.6	396	34	18.4
20.12.16	Z_ Plešovice	068	1204	93.5	0.0555	0.05367	0.00043	0.3977	0.0052	0.05346	0.00088	0.1026	337	2.7	339.9	3.8	347	37	6.4
20.12.16	Z_ Plešovice	081	1317	105.7	0.0595	0.05305	0.00037	0.3914	0.0036	0.05355	0.00039	0.666	333.2	2.2	335.4	2.7	352	16	5.8
20.12.16	Z_ Plešovice	094	1397	89.4	0.0491	0.05277	0.00029	0.3866	0.0055	0.05262	0.00056	0.2328	331.5	1.8	331.9	4	312	25	-1.8
20.12.16	Z_ Plešovice	107	1318	89.9	0.0511	0.05098	0.00042	0.3823	0.0051	0.05431	0.00064	0.7205	320.5	2.6	328.7	3.8	384	26	15.8
20.12.16	Z_ Plešovice	120	1516	99.6	0.0504	0.05149	0.00034	0.3758	0.0059	0.05274	0.00046	0.618	323.6	2.1	323.9	4.3	326	25	-0.9
20.12.16	Z_ Plešovice	133	1540	102.3	0.0501	0.05156	0.00048	0.3782	0.0041	0.05325	0.00054	0.1336	324.1	2.9	325.7	3	339	23	2.5
20.12.16	Z_ Plešovice	146	1500	91.4	0.0495	0.05132	0.0005	0.3807	0.0034	0.05372	0.00047	0.5309	322.6	3.1	327.5	2.5	359	20	8.9
20.12.16	Z_ Plešovice	159	1321	87.8	0.0480	0.04998	0.00049	0.3681	0.0048	0.05307	0.00035	0.4451	314.4	3	318.2	3.6	332	15	7.8
20.12.16	Z_ Plešovice	172	1512	99	0.0510	0.05088	0.0005	0.3736	0.0043	0.05348	0.00069	0.4332	319.9	3.1	322.3	3.2	348	29	6.3
20.12.16	Z_ Plešovice	184	1426	112.2	0.0599	0.0514	0.00058	0.3752	0.0051	0.05275	0.00039	0.8834	323.1	3.5	323.5	3.8	318	17	-1.9
20.12.16	Z_ Plešovice	185	1800	110.7	0.0491	0.0507	0.00051	0.3711	0.0044	0.05279	0.00074	0.4005	318.8	3.1	320.5	3.2	319	32	-3
20.12.16	Z_91500	015	128.3	112.4	0.1946	0.1761	0.0017	1.813	0.024	0.07436	0.00088	0.5498	1045.7	9.4	1050	8.5	1051	24	-0.2
20.12.16	Z_91500	028	140.5	120.3	0.1919	0.1762	0.0021	1.833	0.029	0.07493	0.00096	0.4779	1046	11	1057	10	1066	26	3.1
20.12.16	Z_91500	041	144	119.4	0.1936	0.1772	0.002	1.857	0.035	0.0758	0.0012	0.5091	1052	11	1066	12	1088	33	3.8
20.12.16	Z_91500	054	141.5	113.3	0.1963	0.1744	0.0022	1.792	0.024	0.07513	0.00086	0.6676	1036	12	1042.6	8.6	1072	23	2.3
20.12.16	Z_91500	067	138.8	111.1	0.1943	0.1689	0.0019	1.745	0.022	0.07505	0.00078	0.6084	1006	10	1025.2	8.3	1069	21	6.1
20.12.16	Z_91500	080	106.6	78.1	0.1724	0.1743	0.0019	1.788	0.025	0.0746	0.00086	0.5546	1035	10	1041.1	9	1057	23	1.8
20.12.16	Z_91500	093	102.9	74.4	0.1700	0.173	0.0019	1.793	0.031	0.0748	0.0014	0.1144	1029	10	1042	11	1060	37	2.1
20.12.16	Z_91500	106	138	111.7	0.1920	0.1719	0.0018	1.776	0.023	0.0749	0.00096	0.5309	1023	10	1036.5	8.4	1065	26	3.4
20.12.16	Z_91500	119	132.5	106.6	0.1969	0.1708	0.0014	1.74	0.03	0.0735	0.0015	-0.05	1016.7	7.5	1023	11	1027	41	1.5

Date	Sample	Meas.	U ^a	Pb ^a	Th ^a	²⁰⁶ Pb ^b	±2s ^c	²⁰⁷ Pb ^b	±2s ^c	²⁰⁷ Pb ^b	±2s ^c	rho ^d	²⁰⁶ Pb	±2s	²⁰⁷ Pb	±2s	²⁰⁷ Pb	±2s	FinalDisc.
		No	(ppm)	(ppm)	U	²³⁸ U	(absolute)	²³⁵ U	(absolute)	²⁰⁶ Pb	(absolute)		²³⁸ U	(Ma)	²³⁵ U	(Ma)	²⁰⁶ Pb	(Ma)	(%)
20.12.16	Z_91500	132	146.3	115	0.1943	0.1699	0.0016	1.771	0.037	0.0752	0.0015	0.3651	1011.6	8.8	1035	14	1072	40	4.6
20.12.16	Z_91500	145	149.1	120.2	0.1960	0.1703	0.0017	1.755	0.035	0.075	0.0013	0.3179	1013.5	9.5	1029	13	1068	36	4.8
20.12.16	Z_91500	158	139.1	102.6	0.1878	0.1649	0.0023	1.718	0.032	0.0747	0.0014	0.3751	984	13	1015	12	1059	37	6.9
20.12.16	Z_91500	171	114.2	77.7	0.1745	0.1702	0.0014	1.752	0.033	0.0746	0.0016	0.0548	1013.4	7.8	1027	12	1065	40	4.3

^a U and Pb content and Th/U ratio were calculated relative to GJ-1 reference zircon.

^b corrected for background, within-run Pb/U fractionation (in case of ²⁰⁶Pb/²³⁸U) and common Pb using Stacy and Kramers (1975) model Pb composition and subsequently normalized to GJ-1 (ID-TIMS value/measured value); ²⁰⁷Pb/²³⁵U calculated using ²⁰⁷Pb/²⁰⁶Pb/(²³⁸U/²⁰⁶Pb*1/137.88).

^c Quadratic addition of within-run errors (2 SD) and daily reproducibility of GJ-1 (2 SD).

^d rho is the ²⁰⁶Pb/²³⁸U/²⁰⁷Pb/²³⁵U error correlation coefficient.

Appendix 13 – Zircon trace element geochemistry (samples)

Samples	Grain	Age (Ma)	Meas. No.	Y	Nb	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	Hf	Ta
566240	Zr7	1852 ± 26	7	5	0.60	0.008	0.30	0.00	0.01	0.20	0.00	0.40	0.09	0.48	0.14	0.44	0.14	0.81	0.15	10770	0.12
566240	Zr7	1852 ± 26	8	7	0.81	0.004	0.21	0.02	0.11	0.08	0.00	0.03	0.03	0.60	0.20	0.46	0.13	1.06	0.20	10300	-0.04
566240	Zr2	1955 ± 108?	9	13	0.53	0.004	3.40	0.14	0.96	0.27	0.13	1.30	0.08	1.40	0.34	1.38	0.26	2.00	0.32	12020	0.01
566240	Zr2	1903 ± 23	10	382	14.40	0.003	34.60	0.23	1.49	0.63	0.14	4.00	1.69	26.60	11.50	66.70	19.10	205.00	33.10	11100	18.20
566240	Zr2	1955 ± 108	11	12	0.69	0.008	4.90	0.14	1.57	0.28	0.29	0.54	0.18	1.52	0.40	1.29	0.21	1.81	0.32	11470	0.14
566240	Zr2 (mount)	1935-1886	63	1140	3.78	0.028	220.00	9.10	25.00	7.80	2.90	20.20	6.90	101.00	35.30	168.00	39.10	397.00	61.30	10020	6.16
566240	Zr2 (mount)	1935-1886	64	392	3.38	0.270	2480.00	109.00	370.00	68.00	33.80	52.00	6.80	46.00	13.80	67.00	16.70	179.00	30.30	9000	3.00
525224	Zr4	1895-1888	12	20	1.12	0.007	2.50	0.08	0.14	0.13	0.28	1.58	0.33	2.36	0.72	2.16	0.32	3.24	0.35	11500	0.22
525224	Zr4	1895-1888	13	14	0.72	0.009	3.53	0.05	0.31	0.44	0.14	0.54	0.16	1.82	0.56	1.62	0.32	2.82	0.36	10710	0.15
525224	Zr4	1895-1888	14	19	0.89	0.003	3.50	0.05	0.58	0.28	0.17	0.80	0.20	2.53	0.54	2.37	0.41	4.40	0.66	10550	0.21
525224	Zr6	1903 ± 22	21	884	3.65	0.001	34.60	0.36	2.30	2.06	0.88	12.40	4.51	69.00	30.50	161.00	43.20	474.00	82.30	10070	3.28
525224	Zr6	1903 ± 22	22	680	3.47	0.013	66.10	1.43	9.90	3.60	0.82	9.00	3.73	55.00	23.00	118.00	32.20	364.00	58.80	11200	2.59
525224	Zr12	1894 ± 51	23	8	0.65	0.004	0.19	0.10	0.40	0.15	0.05	1.00	-0.01	0.98	0.19	0.65	0.18	1.59	0.33	13500	-0.01
525224	Zr16	1889 ± 33	24	167	1.40	0.007	59.00	1.81	11.00	2.20	0.92	4.40	0.99	14.00	5.70	28.00	7.10	80.00	11.70	11900	2.00
525224	Zr18	1853 ± 53	25	36	0.99	0.008	2.35	-0.01	0.23	0.43	0.03	0.72	0.21	2.55	0.93	4.30	1.32	12.90	2.40	10470	0.48
525224	Zr18	1806 ± 50	26	25	0.48	0.006	13.40	0.22	0.91	0.38	0.16	1.17	0.20	3.09	0.88	3.08	0.66	5.80	0.72	7100	0.15
566249	Zr1	1871 ± 24	40	31	0.90	0.009	24.50	1.41	9.20	1.26	1.00	2.60	9.20	1.26	1.00	2.60	0.76	8.20	1.53	10800	0.42
566249	Zr3	1974 ± 20	42	13	0.27	0.004	0.40	0.11	0.34	0.22	0.18	1.89	0.34	0.22	0.18	1.89	0.40	4.01	0.67	10370	0.16
566249	Zr4	1926-1880	49	48	0.44	0.007	25.00	1.00	5.90	1.22	0.80	3.10	5.90	1.22	0.80	3.10	1.26	10.50	1.39	12450	0.24
566249	Zr9	1882 ± 17	50	44	0.59	0.006	12.00	0.60	4.00	1.50	0.95	7.40	4.00	1.50	0.95	7.40	0.66	4.90	0.63	11230	0.41
566249	Zr9	1881 ± 31	52	29	0.20	0.005	5.80	0.48	1.40	1.05	0.39	3.20	1.40	1.05	0.39	3.20	0.93	9.04	1.49	9620	0.83
566249	Zr12	1873 ± 16	55	71	2.91	0.024	279.00	17.90	113.00	29.80	14.90	27.50	113.00	29.80	14.90	27.50	0.56	3.68	0.37	11080	0.41
566249	Zr12	1878 ± 11	56	47	1.23	0.012	44.00	3.30	20.00	6.90	3.70	10.10	20.00	6.90	3.70	10.10	0.78	4.90	0.81	13800	0.46
566267	Zr5	2437-2208	27	427	4.30	0.014	114.00	8.60	60.00	13.40	4.19	14.30	3.69	39.10	11.70	61.00	18.20	214.00	35.50	11050	7.50

Samples	Grain	Age (Ma)	Analysis	Y	Nb	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	Hf	Ta
566267	Zr5	2437-2208	28	237	2.10	0.008	67.60	5.02	34.40	9.30	3.03	9.30	2.23	23.20	6.60	28.80	8.40	107.00	17.20	11850	3.00
566267	Zr8	2377 ± 20	35	217	6.00	0.008	23.40	1.78	11.80	4.30	1.31	6.20	1.21	15.40	6.70	40.00	13.40	185.00	34.90	9800	9.90
566267	Zr8	2377 ± 20	36	213	4.40	0.005	25.10	0.88	4.90	2.03	0.72	4.00	1.18	16.00	6.04	37.00	10.70	137.00	26.20	12500	6.20
566267	Zr10	1896 ± 15	37	235	6.20	0.005	58.90	4.76	26.00	5.50	3.55	7.50	1.81	21.70	7.10	32.40	9.36	93.40	18.10	12200	3.00
566267	Zr10	1887 ± 14	38	93	0.85	0.007	15.30	1.65	10.10	1.61	0.84	3.10	1.03	11.20	2.73	9.20	1.81	14.10	2.19	13300	0.40
566267	Zr10	1896 ± 15	39	67	0.77	0.007	25.60	2.90	15.00	3.40	1.25	3.50	0.79	9.00	1.96	5.90	1.24	9.90	1.45	9960	0.02

Concentrations (in ppm) are calculated relative to the titanium concentration in the GJ-1 reference zircon. Values in red are below the detection limit.

Appendix 13 – Zircon trace element geochemistry (standards)

Standards	Analysis	Y	Nb	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	Hf	Ta
NIST612	01	17	50.70	0.024	44.30	10.90	20.00	12.30	11.60	11.30	13.00	14.50	15.80	15.30	18.90	23.00	16.10	36.3	67.00
NIST612	02	15	48.00	0.026	44.80	11.12	16.20	15.10	13.60	13.30	14.76	17.00	16.80	17.60	20.20	23.90	18.00	43.2	69.90
NIST612	03	17	53.60	0.024	42.90	11.28	17.00	11.60	11.31	11.80	13.18	13.80	15.07	16.80	19.40	20.30	17.30	40.5	67.40
NIST612	19	14	45.90	0.023	38.90	11.90	13.90	10.90	12.00	11.70	11.80	14.00	15.40	15.20	17.90	20.00	17.20	37.7	68.10
NIST612	20	14	44.70	0.025	37.10	12.29	16.00	12.30	12.90	16.00	14.60	15.50	15.90	17.00	18.40	21.30	17.60	42.1	76.50
NIST612	33	16	47.30	0.029	42.10	16.00	16.20	12.50	12.50	11.70	13.80	16.00	16.03	16.90	19.40	20.50	17.30	39.9	77.10
NIST612	34	15	46.00	0.023	40.00	13.26	14.50	11.70	10.86	12.10	12.50	13.80	15.00	15.30	18.20	20.40	16.60	38.3	74.30
NIST612	47	15	46.70	0.026	41.90	17.70	13.40	13.90	12.00	12.20	13.80	15.10	15.30	16.00	18.20	21.20	17.90	37.4	86.90
NIST612	48	15	53.20	0.029	44.30	18.60	14.90	14.10	12.80	11.40	13.90	16.80	16.40	17.90	19.40	19.10	17.10	41.4	83.60
NIST612	61	16	50.90	0.026	40.00	19.10	11.80	12.10	13.00	9.40	13.20	15.00	16.60	18.30	19.40	24.60	16.40	41.2	91.00
NIST612	62	15	48.40	0.024	39.50	20.90	13.20	13.50	12.80	11.30	13.50	14.50	15.10	16.10	18.50	20.10	17.50	37.6	95.00
NIST612	70	16	52.20	0.028	41.10	23.60	11.70	12.60	11.14	12.10	13.11	15.50	16.70	16.80	18.93	21.30	17.20	37.5	96.90
NIST612	71	15	47.90	0.027	42.30	23.80	12.00	14.20	12.60	12.10	14.00	15.30	17.30	17.10	18.80	22.40	17.90	38.5	105
NIST612	72	16	55.30	0.027	39.70	23.90	12.20	12.90	12.10	11.50	13.70	15.40	16.50	15.90	17.30	20.30	16.90	39.0	93.90
<i>Average Preferred value (GeoRem)</i>		15.46	49.34	0.026	41.35	16.74	14.50	12.84	12.23	11.99	13.49	15.16	15.99	16.59	18.78	21.31	17.21	39.3	82.33
		38	40.00	35.800	38.70	37.20	35.90	38.10	35.00	36.70	36.00	36.00	38.00	38.00	38.00	39.20	36.90	35.0	40.00
NIST614	15	0	1.03	0.001	1.10	0.33	-0.30	-0.09	0.33	1.00	0.16	0.60	0.40	0.40	0.50	0.65	0.41	1.4	1.67
NIST614	29	0	1.16	0.000	-0.10	0.19	-0.10	-0.11	0.37	-0.65	0.19	0.25	0.40	0.65	0.48	0.39	0.42	0.6	1.57
NIST614	43	0	1.51	0.000	0.60	0.40	0.05	0.21	0.35	1.20	0.29	0.29	0.22	0.27	0.33	0.57	0.36	1.6	2.15
NIST614	57	1	0.68	0.000	1.33	0.36	-0.30	0.08	0.22	-0.30	0.32	0.42	0.21	0.38	0.27	0.81	0.29	1.0	1.44
NIST614	65	0	0.86	0.000	1.00	0.24	0.44	0.54	0.20	-0.60	0.38	0.12	0.26	0.43	0.37	0.35	0.23	0.3	1.71
<i>Average Preferred value (GeoRem)</i>		0	1.05	0.000	0.79	0.30	-0.04	0.13	0.29	0.13	0.27	0.34	0.30	0.43	0.39	0.55	0.34	1.0	1.71
		1	0.82	0.720	0.81	0.77	0.75	0.75	0.77	0.76	0.74	0.75	0.75	0.74	0.73	0.78	0.73	0.7	0.81

Standards	Analysis	Y	Nb	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	Hf	Ta
GJ1	04	246	1.69	0.005	15.80	0.02	0.34	1.55	1.08	5.90	1.84	19.50	7.20	29.20	7.37	64.70	11.70	6530	0.52
GJ1	05	224	1.77	0.006	13.70	0.03	0.75	1.30	0.94	6.11	1.68	19.90	6.28	28.70	6.46	69.70	11.23	6300	0.46
GJ1	06	228	1.56	0.007	14.80	0.02	0.71	1.61	1.04	6.70	1.98	19.70	6.60	30.50	6.73	74.30	11.60	6610	0.52
GJ1	17	217	1.35	0.005	14.50	0.03	0.59	1.55	0.96	5.80	1.93	20.90	6.77	29.00	6.96	70.80	11.30	6290	0.44
GJ1	18	217	1.75	0.007	12.50	0.02	0.83	1.37	0.86	6.23	1.65	20.30	6.25	28.30	7.03	66.10	11.30	6280	0.44
GJ1	31	234	1.38	0.006	14.90	0.03	0.47	1.70	0.90	6.44	1.63	20.30	6.06	29.10	6.70	68.70	10.70	5900	0.38
GJ1	32	232	1.83	0.005	15.80	0.05	0.23	1.83	0.84	7.00	1.86	20.70	6.58	29.50	6.79	66.70	10.83	6290	0.53
GJ1	45	227	1.18	0.006	14.20	0.03	0.56	1.26	0.95	7.10	1.77	20.70	6.53	29.50	6.60	69.90	11.70	6500	0.66
GJ1	46	228	1.70	0.006	13.40	0.03	0.62	1.45	0.91	5.90	1.75	20.00	6.05	27.00	6.25	62.30	10.60	5950	0.32
GJ1	59	221	1.70	0.005	14.00	-0.03	0.39	1.52	0.94	6.20	1.88	18.10	6.40	28.90	7.03	68.60	11.40	6240	0.47
GJ1	60	226	1.50	0.006	14.80	0.09	0.54	1.13	0.87	6.00	1.65	20.00	6.28	28.80	6.54	68.10	11.60	6230	0.42
GJ1	67	235	1.79	0.007	15.00	-0.10	0.55	1.41	0.95	6.51	1.84	21.40	6.63	29.70	7.24	70.80	12.30	6790	0.53
GJ1	68	236	1.59	0.007	14.50	0.03	0.84	1.57	1.03	6.80	1.88	20.60	6.99	29.20	6.87	72.40	11.30	6700	0.98
GJ1	69	227	1.82	0.006	13.80	0.08	0.43	1.75	0.94	5.90	1.86	19.00	6.47	28.70	6.52	67.30	11.08	6140	0.58
<i>Average Preferred value (Jackson et al., 2004)</i>		228	1.62	0.006	14.41	0.02	0.56	1.50	0.94	6.33	1.80	20.08	6.51	29.01	6.79	68.60	11.33	6339	0.52
		228	1.60	0.006	14.30	0.03	0.62	1.52	0.95	6.30	1.81	20.00	6.50	29.00	6.80	68.70	11.30	6341	0.45
Plešovice	16	242	5.50	0.005	2.26	0.04	1.29	1.21	0.29	5.20	2.00	25.10	7.03	27.50	5.82	46.90	5.34	10400	3.76
Plešovice	30	249	5.64	0.007	2.53	0.09	1.13	1.65	0.46	5.70	2.07	26.10	7.60	28.70	5.94	49.40	5.48	10280	4.28
Plešovice	44	241	5.71	0.006	2.65	0.09	1.19	1.37	0.36	5.80	2.10	26.00	7.49	27.00	6.02	46.90	5.00	9980	4.97
Plešovice	58	228	6.05	0.007	2.42	0.06	1.08	1.61	0.31	5.26	1.80	26.00	7.60	29.90	5.86	47.90	5.67	10300	5.39
Plešovice	66	239	6.20	0.005	3.10	0.02	1.19	1.48	0.41	4.80	2.02	26.10	7.67	28.90	5.17	45.10	5.22	9900	5.23
<i>Average Previous experiments (25 um)</i>		240	5.82	0.01	2.59	0.06	1.18	1.46	0.37	5.35	2.00	25.86	7.48	28.40	5.76	47.24	5.34	10172	4.73
				0.54-2.11	1.34-10.8	0.07-2.14	0.81-13.5	1.25-8.23	0.294-2.41	4.83-24.7	1.97-7.77	25.7-79.1	7.17-19.69	27.8-66.1	4.91-11.35	36.4-78.2	4.65-10.01	9180-11180	

Concentrations (in ppm) are calculated relative to the titanium concentration in the GJ-1 reference zircon. Values in red are below the detection limit.

Curriculum Vitae

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Born on 16th August 1987 in Münster

Education

2013 – present

Doctoral student, RWTH Aachen University

Thesis topic: Metamorphic evolution of relict eclogite-facies rocks in the Paleoproterozoic Nagssugtoqidian Orogen, South-East Greenland.

Supervisor: PD. Dr. A. Dziggel, PD. Dr. S. Sindern

2010 – 2013

Master of Science, Geosciences, WWU Münster

Thesis topic: Detailed study on microstructures and element mobility during fluid-mediated pseudomorphous gabbro-to-eclogite transformation.

Supervisor: Dr. T. John, Prof. Dr. A. Putnis

2007 – 2010

Bachelor of Science, Geosciences, WWU Münster

Thesis topic: Trace element mobility during fluid-mediated pseudomorphous gabbro-to-eclogite transformation.

Supervisor: Dr. T. John, Prof. Dr. A. Putnis

2004 – 2007

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Declaration of originality

I hereby declare in lieu of oath that the entirety of this dissertation is the result of my own work and includes nothing, which is not the outcome of the work done in collaboration, except where specifically indicated in the text. I have used no other than the cited sources. It has not been previously submitted, in part or whole, to any university or institution for any degree, diploma or other qualification.

Eidesstattliche Erklärung

Ich versichere hiermit an Eides statt, dass ich die vorliegende Arbeit selbstständig und ohne unzulässige fremde Hilfe verfasst habe. Ich habe keine anderen als die angegebenen Quellen und Hilfsmittel benutzt. Die Arbeit hat in gleicher oder ähnlicher Form noch keiner Prüfungsbehörde vorgelegen.

Ich versichere hiermit, die Grundsätze zur Sicherung guter wissenschaftlicher Praxis der Rheinisch-Westfälischen Technischen Hochschule Aachen zur Kenntnis genommen und eingehalten zu haben.

Signed/Unterschrift:

Sascha Müller

Heinsberg, 20.03.2019