

MACQUARIE UNIVERSITY

# **An investigation of the origin of ultrabasic granofels, Fiordland, New Zealand**

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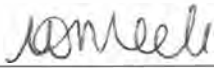
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### Statement of Originality

All the work submitted in this thesis is the original work of the author except where otherwise acknowledged. No part of this thesis has previously been submitted to any other university or institution.



**Signed** – Uvana Meek

9.10.15

**Dated**

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## **An investigation of the origin of ultrabasic granofels, Fiordland, New Zealand**

### **Abstract**

Ultrabasic granofels are a minor component of the Pembroke Granulite that is hosted within granulite facies gabbroic gneiss, in Fiordland, New Zealand. The ultrabasic granofels unit consists predominantly of pargasite  $\pm$  clinozoisite. The unit has a rare earth element (REE) pattern that shows a typical amphibole cumulate middle-REE hump, which suggests an igneous cumulate origin. However, field relationships, including irregular contact boundaries and significant mineralogical changes, suggest the ultrabasic granofels has replaced part of the gabbroic gneiss as a result of metasomatic melt-rock interaction. Significant REE-enrichment in the ultrabasic granofels relative to the host gabbroic gneiss, low dihedral angles in plagioclase grains and geochemical evidence of mass increase all point to an influx of external melt. A model is presented for a reaction product that imitates an igneous cumulate; involving fluxing of granodioritic melt through a high to moderate strain zone. The system and melt are at an appropriate P-T where it is pargasite + clinozoisite-saturated. This results in the dissolution of plagioclase and quartz and crystallisation of pargasite + clinozoisite. Progressive melt-rock interaction leads to the formation of plagioclase-quartz-poor and pargasite-clinozoisite-rich rock domains. This model presents a new way of forming basic to ultrabasic rocks in the deep crust.



# **An investigation of the origin of ultrabasic granofels, Fiordland, New Zealand**

## **1. Introduction and background geology**

Ultrabasic rocks are common at or around the crust-mantle boundary. Potentially, important information can be obtained about the nature and origin of ultrabasic rocks that may inform on fractionation processes of melts within the lower crust and upper mantle. High-grade metamorphic ultrabasic rocks can be derived from a range of protolith materials and deep Earth processes (Daczko et al., 2012). Three key groups of ultramafic and ultrabasic rocks include mantle, cumulate and metasomatised rocks. Many ultrabasic rocks are upper-mantle-derived, commonly arising from the exhumation of mantle material, such as peridotite, dunite or pyroxenite. Samples may be brought to the surface as xenoliths or through tectonic processes (Reverdatto et al., 2008). Ultramafic cumulate rocks are produced by crystal accumulation of mafic minerals during magmatic fractionation. Hornblende-rich cumulates, such as hornblendite or hornblende peridotite, are inferred to form from H<sub>2</sub>O-rich fractionated magmas, which are commonly thought to be present near the base of many island arc systems (Daczko et al., 2012). Alternatively, conceptually ultrabasic rocks could be produced by metasomatism as a result of fluid or melt infiltration and subsequent fluid/rock interaction. Fluid- or melt-rock interaction is common in rocks within mantle (Kelemen et al., 1995) and the crust (Taylor and McLennan, 1995, Brown and Rushmer, 2006) and is one of the key processes in developing ultramafic and ultrabasic rocks. Metasomatism is commonly strongly linked to fluid/melt pathways such as fractures, faults, shear zones or highly permeable lithologies. In the mantle, significant changes in mineralogy associated with melt flux are common and described in many instances (Kelemen et al., 1992, Kelemen et al., 1995). However, this is not the case for the lower crust even though it is clear that melt must transfer through the lower crust to form igneous bodies at higher crustal levels that still retrain a lower crustal or even mantle geochemical signature (Taylor and McLennan, 1995). As the lower crust is chemically less extreme compared to the mantle which is dominated by olivine, establishing the signal of melt flux in the geological record may be hindered. In addition, if melt/fluid rock interaction occurred, the chemical variability of lower crustal rocks, will make it difficult to determine the nature and chemical composition of the metasomatising agent.

The ultrabasic rocks which form a minor component of the Pembroke Granulite, Fiordland, New Zealand, have not been studied in any detail. Their geochemical characteristics and their relationship to the main gabbroic gneiss unit is still unknown.

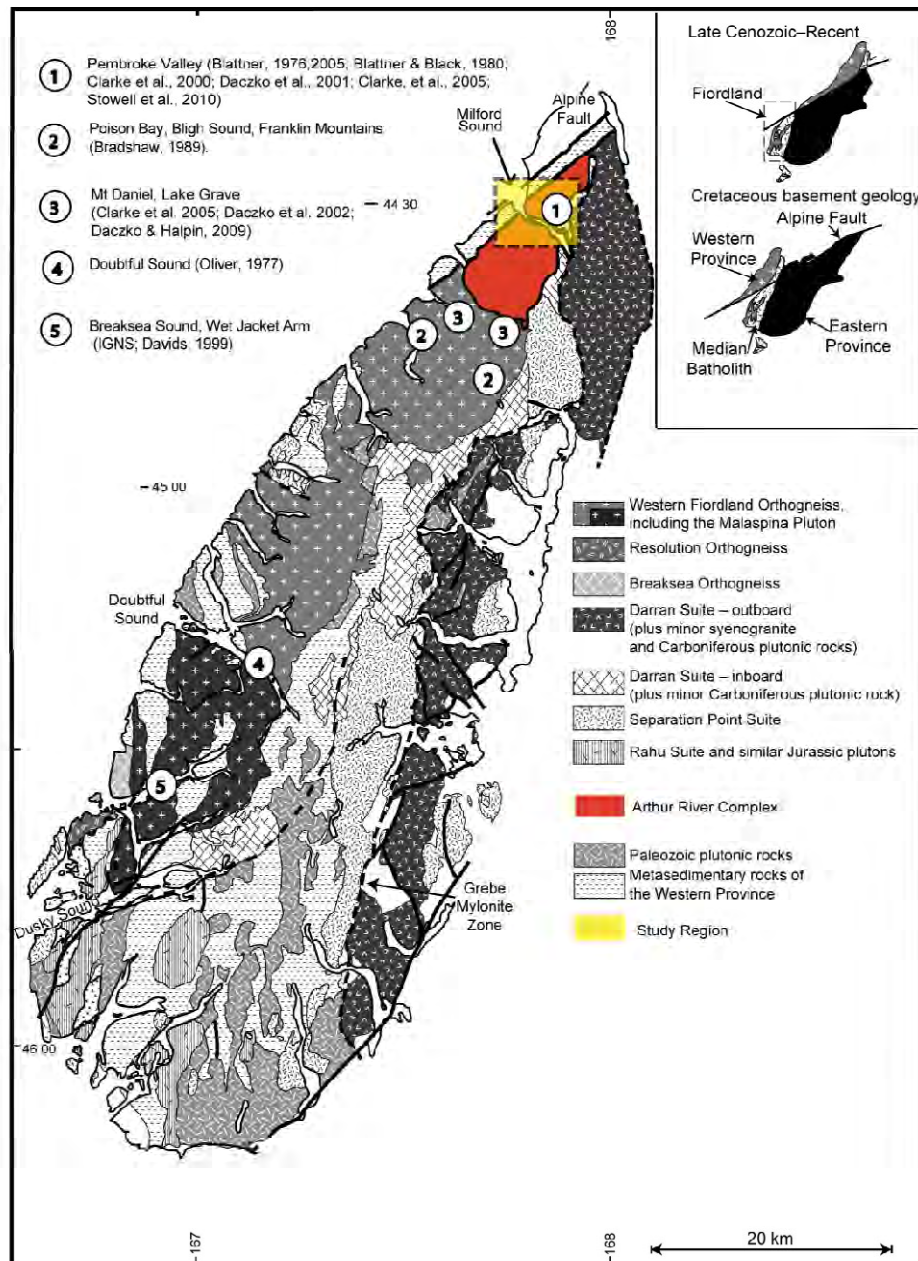
The aims of this thesis are therefore to:

- 1) characterise the unit in terms of geochemistry, mineralogy, microstructure and field relationships
- 2) establish its structural, metamorphic and/or igneous history
- 3) interpret the origin of the unit considering the two main scenarios (a) igneous origin by fractionation and/or accumulation and (b) metasomatic origin by melt/fluid fluxing and melt/rock interaction. In the latter case, this would imply that the unit represents a melt "conduit" in the lower crust.

## **Background geology**

The geology of New Zealand's South Island is characterized by three major provinces, which incorporate a number of tectonostratigraphic terranes. The main provinces are the Eastern Province, the Western Province and Median Batholith (Figure 1.1). These are mainly separated by fault contacts (Clarke et al. 2000). The Eastern and Western provinces are predominantly comprised of volcanic and sedimentary strata while the Median Batholith comprises magmatic arc rocks (Bradshaw, 1989, Clarke et al., 2000, Daczko et al., 2001a, Daczko and Halpin, 2009). The Eastern and Western provinces are the basement rocks intruded by the Median Batholith, and are divided on the basis of chronological and geochemical constraints. The Eastern province is mainly composed of Permian to Early Cretaceous turbiditic, volcanic and volcanoclastic sedimentary rocks. The Western Province is dominantly Lower Palaeozoic metasedimentary rocks and Palaeozoic-Mesozoic orthogneisses (Muir et al., 1998, Wandres et al., 1998).

The Median Batholith was emplaced and subsequently metamorphosed during arc-continent collision intruding the two basement provinces. It consists of Carboniferous to Early Cretaceous plutonic arc rocks representing a 250 Ma record of magmatism. The Middle Cretaceous Western Fiordland Orthogneiss (WFO) dominates the Median Batholith and consists of two-pyroxene-hornblende gneiss.



**Figure 1.1.** Regional geology of Fiordland highlighting the three main provinces of south New Zealand (inset). The main figure, modified from Allibone et al. (2009b) and Clarke et al. (2005) also shows the field of study, Pembroke Granulite (yellow) and the location of the Arthur River Complex (red). Additionally, surrounding geological suites are highlighted.

The WFO represents hundreds of sheet-like intrusions of ultramafic to dioritic compositions which have experienced varying metamorphic conditions and deformation. Emplacement ages at mid-crustal conditions range between 126-120 Ma in northern Fiordland and 120-115 Ma at greater depths in southern Fiordland (Allibone et al., 2009a). It is estimated that approximately ~10% of the WFO shows a transition to garnet granulite (Daczko and Halpin, 2009).

The Arthur River Complex (ARC) is part of the Median Batholith and is comprised of two main sections: a homogeneous, mafic body called the Milford Gneiss and the

compositionally banded Harrison Gneiss (Clarke et al., 2000, Hollis et al., 2003). These rocks are slightly older than the WFO, but show similar rock types. Crystallisation ages, determined from magmatic zircons, indicates ages between 136 and 129 Ma (Hollis et al., 2003) or may be as old as c. 355 Ma (Tulloch et al., 2009).

The field of study, the Pembroke Granulite, is a low strain body that sits within the Milford Gneiss of the ARC, and its composed predominantly of two-pyroxene hornblende gabbroic gneiss with intermittent pods of dioritic compositions and 'ultramafic blebs' (Clarke et al. 2000, Daczko et al 2001a). The emplacement age of the Pembroke Granulite has so far been difficult to constrain. Stowell et al. (2010) found the oldest zircons, between 163 - 150 Ma, to reflect either inheritance or intrusion ages.

The Pembroke Granulite is known to have experienced several deformation and metamorphic events ( $P \sim 8.0$  kbar and  $T = 750$  °C) prior to rapid cooling and exhumation.  $S_1$  is a pervasive, amphibolite facies gneissosity, characterised by strongly aligned hornblende grains and clinopyroxene + orthopyroxene + hornblende + quartz clusters (Clarke et al., 2000). This is the earliest deformation event ( $D_1$ ) recorded, the  $S_1$  gneissosity formed at  $134.2 \pm 2.9$  Ma (Stowell et al., 2010).  $D_2$  is a localised partial melting event, which shows leucosome material surrounding peritectic garnets. Brittle rectilinear fracturing transported the melt throughout the gabbroic gneiss, causing melt-rock interactions to form the garnet reactions zones surrounding the dykes (where the two pyroxene-hornblende rock was transformed into garnet granulite) (Daczko et al., 2001a). This reaction is caused by dehydration reactions occurring in isochemical conditions and is characterised by pseudomorphing of hornblende by garnet (Blattner, 1976, Bradshaw, 1989, Clarke et al., 2000, Daczko et al., 2001a, Smith et al., 2015).  $D_1$  and  $D_2$  fabrics are both deformed by two later deformation events,  $D_3$  and  $D_4$ . These two events both occurred at granulite facies conditions forming narrow (cm-m scale), steeply dipping mylonitic shear zones ( $D_3$ ) followed by larger (metre scale) shallowly dipping thrust faults (Daczko et al., 2001b).

This study aims to characterise the expression of an ultrabasic unit situated within the Pembroke Granulite. Previously, these have been termed 'ultramafic bleb' (Clarke et al., 2000), and were considered to be cumulate layers fractionating from a magmatic source. This unit is investigated using field relationships, petrography and geochemistry to determine the likely origin.

## 2. Field relationships

Fieldwork for this project was undertaken during an 8-day visit to the South Island of New Zealand, from the 17<sup>th</sup>-25<sup>th</sup> January 2015. Mapping and sample collection was undertaken between 20<sup>th</sup>-24<sup>th</sup> January 2015 within the Pembroke Valley, immediately north of Milford Sound, Fiordland, New Zealand. The field area was accessed from Milford Sound by helicopter. The field party comprised myself, two supervisors, Nathan Daczko and Sandra Piazzolo, and another Masters of Research student, Victoria Elliott.

Rocks in the Pembroke Valley are well exposed due to recent glaciation. However, the flat and smooth outcrop surfaces restrict ease of in-situ sampling. Therefore, following detailed field mapping and observations, additional loose samples were collected from the cobbles and boulders available in the creek bed lower in the valley.

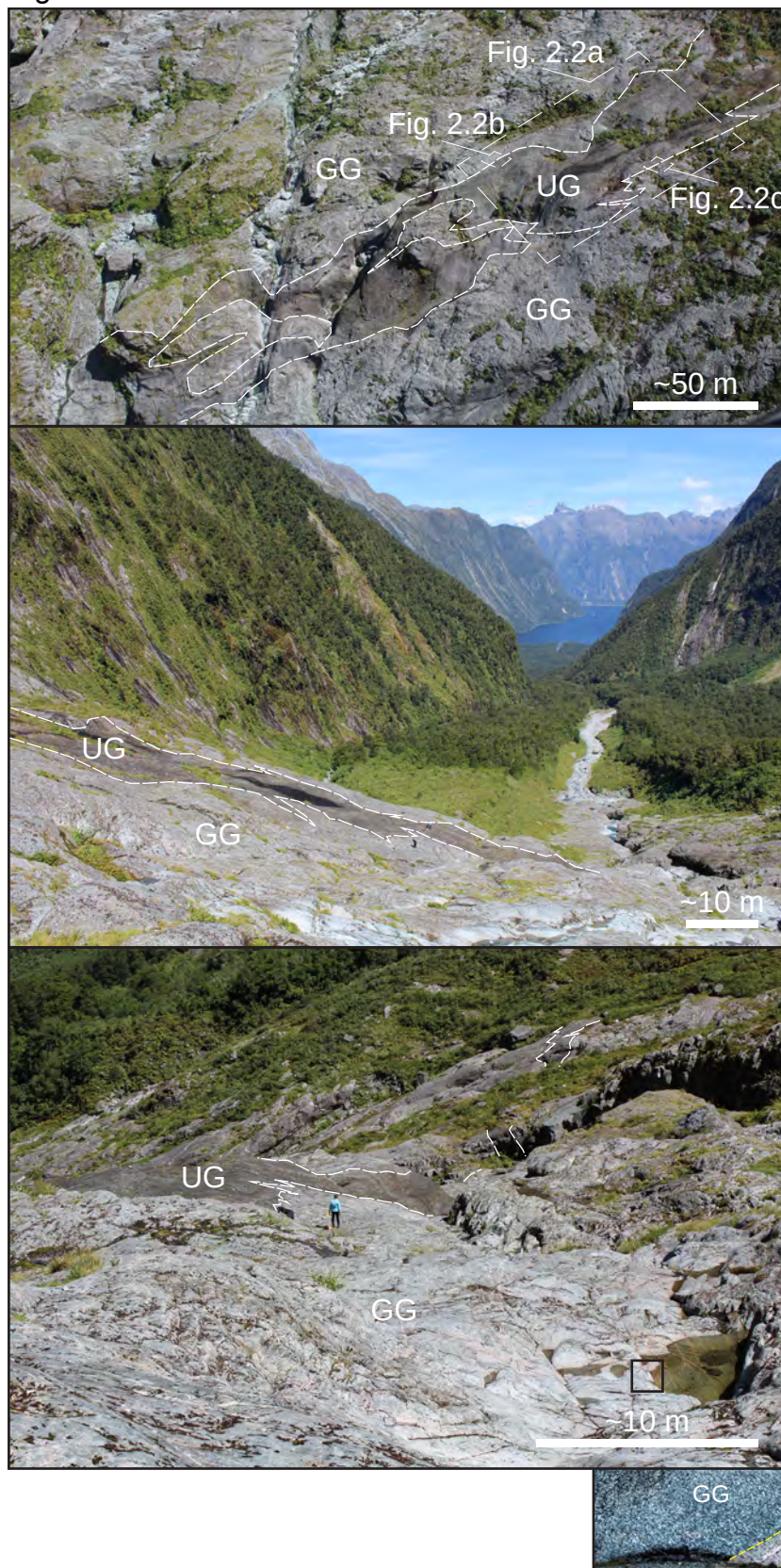
This study investigated rare amphibole-rich rocks within the Pembroke Valley. The outcrop examined in detail consists of three main rock types: (i) host gabbroic gneiss (GG), (ii) amphibole-rich rocks here called ultrabasic granofels and (iii) a narrow transition zone between the first two rock types. The outcrop pattern and extent of the ultrabasic granofels are shown in an aerial view, taken from the helicopter (Figure 2.1a) and in views from higher in the valley (Figure 2.1b,c). An overview map (60m x 35 m) was constructed to record the extent of the three rock types, their contact relationships and variability in mineralogy and strain within the units, along with sample locations (Figure 2.2a). Two small, detailed maps (~5m x ~8m) were constructed to document the detailed structural relationships between the three rock types (Figures 2.2b and 2.2c). Figure 2.3a shows that the ultrabasic granofels (UG) unit is not offset by a major shear zone in the valley ( $D_4$  thrust zone of Daczko et al. 2001a). Therefore, the emplacement or formation of the ultrabasic granofels unit is later than the previously recorded geological history of the Pembroke Granulite (Clarke et al., 2000, Daczko et al., 2001a, Daczko et al., 2001b, Schröter et al., 2004, Clarke et al., 2005, Smith et al., 2015).

### Gabbroic Gneiss

The host gabbroic gneiss is well exposed throughout much of the valley and is the main rock type. It consists of two-pyroxene-pargasite mafic clusters that define the gneissic foliation ( $S_1$ ) within a plagioclase-rich rock (Daczko et al. 2001a.). This unit contains narrow (< 20mm wide) felsic (plagioclase-rich  $\pm$  garnet-quartz-K-feldspar) dykes



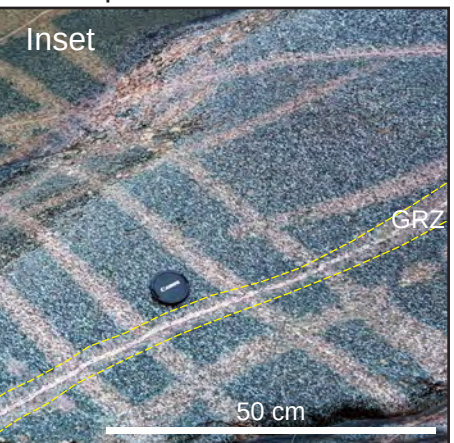
Figure 2.1



(a) A helicopter aerial view of the ultrabasic granofels (UG; boundaries highlighted by dashed lines) and gabbroic gneiss (GG) situated within the field area, Pembroke Valley, Fiordland, New Zealand. Note the irregular shape of the boundaries and apophyses.

(b) View of the same ultrabasic granofels (UG) as in (a) from higher in the Pembroke Valley. Harrison Cove, Milford Sound in the background. People for scale.

(c) The ultrabasic granofels (UG) extends across the valley and cuts the major D4 thrust zone (Daczko et al., 2001). People for scale. Inset: Gabbroic gneiss (GG) with garnet reaction zones (GRZ; one outlined with yellow dashed lines) in a rectilinear pattern. The GRZ involves dehydration of hornblende to form garnet, clinopyroxene and quartz, changing the grey-green GG into a pink band.



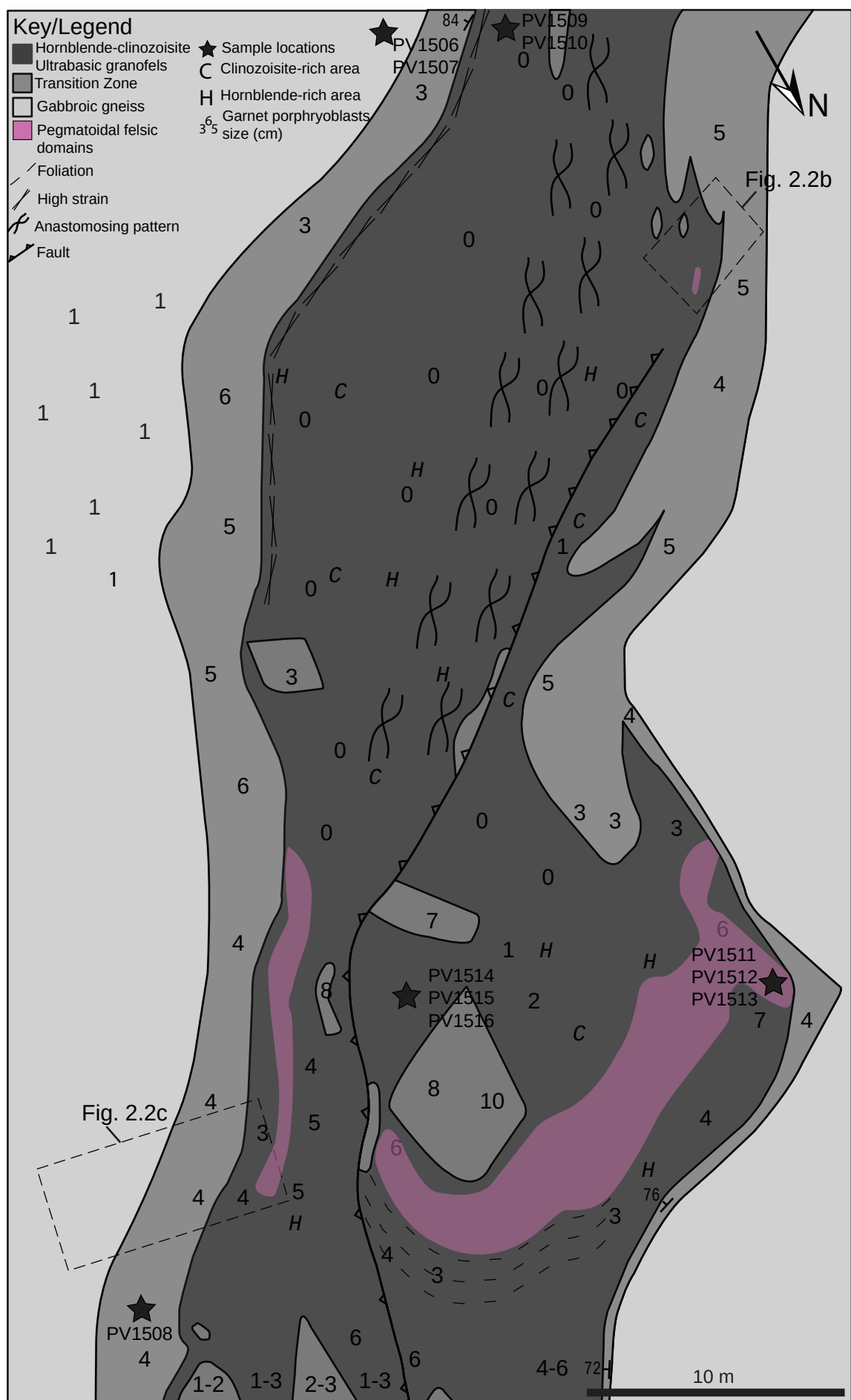
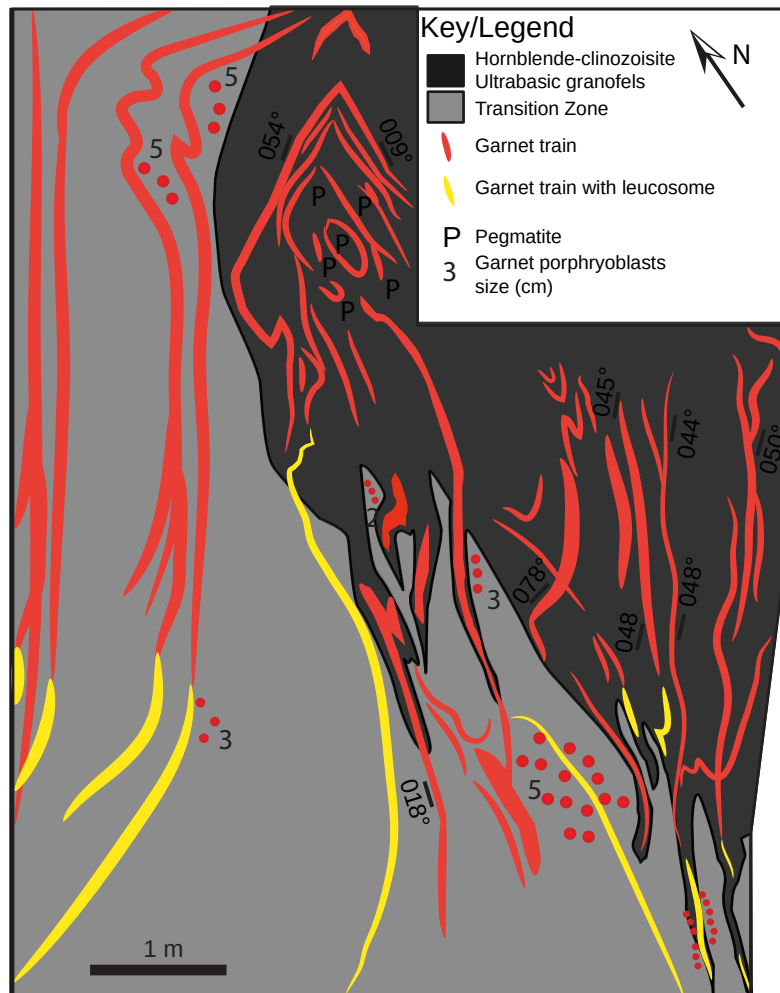


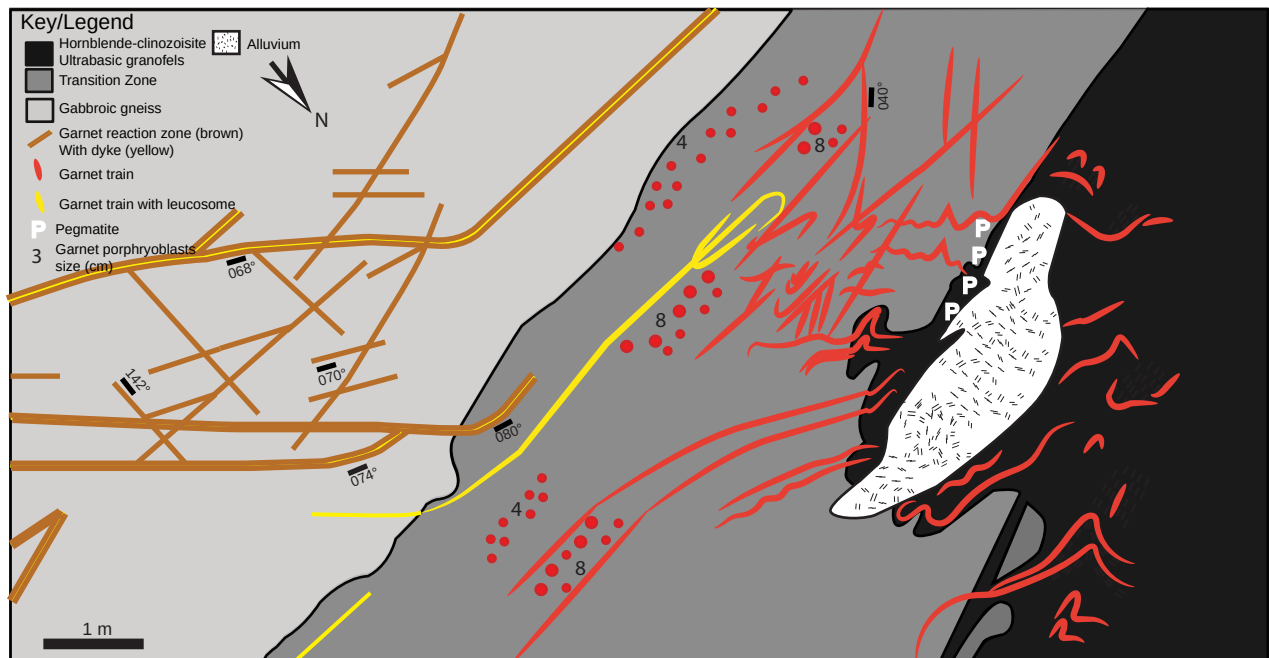
Figure 2.2a. Overall geological map of the study area including the gabbroic gneiss, transition zone and ultrabasic granofels. Small map areas are indicated.



2.2b



2.2c



Figures 2.2b. Field map of close-up relationships of the northern side of the ultrabasic granofels (area marked in Figure 2.2a). GRZ within the transition zone often change from garnet + leucosome into garnet trains  $\pm$  leucosome or pargasite. Large garnet grains are common to the transition zone. GRZ extend into the ultrabasic granofels and become intensely deformed, dismembered and complexly folded (e. g. sheath fold). Areas of high pegmatite material (marked by P) intermittent. 2.2c. Field map of close-up relationships of the southern side of the ultrabasic granofels (area marked in Figure 2.2a). Gabbroic gneiss GRZ with or without felsic dyke in three orientations. GRZ leading into transition zone are bent due to strain. and begin to complexly get deformed with increasing proximity to the ultrabasic granofels unit. Intense folding of GRZ (now garnet trains in the ultrabasic granofels) and pegmatite association within the UG unit.

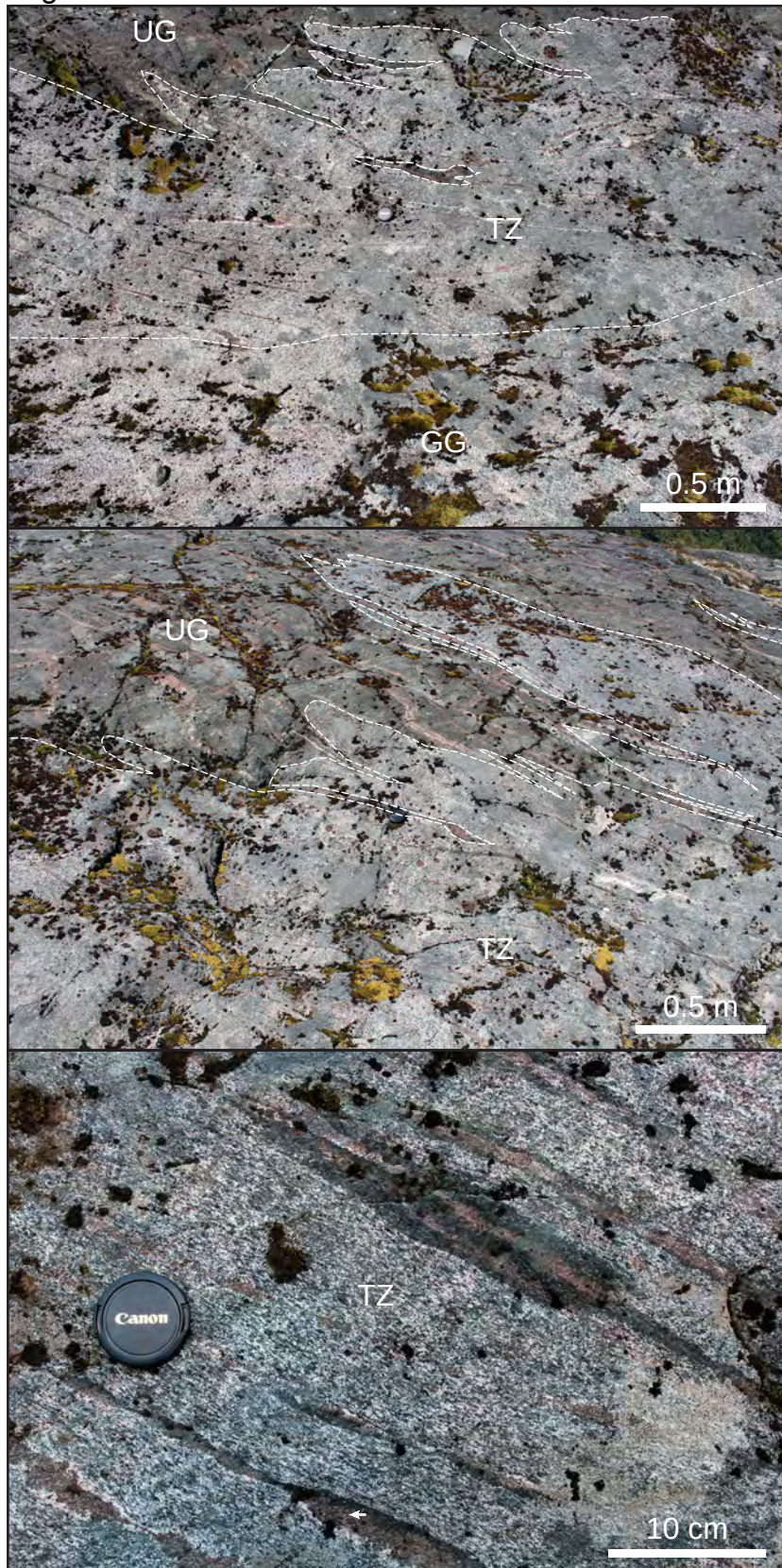


that cut the gneissic fabric in three orientations, following a rectilinear pattern. On either side of these dykes, 1-8cm wide reaction zones developed, changing the gabbroic gneiss into garnet granulite, and hence these domains of the gabbroic gneiss unit have been termed garnet reaction zones, (GRZ; Figure 2.3c inset). (Blattner 1976, Daczko et al. 2001a, Daczko and Halpin 2009). It is common for the garnet reaction zones to exist without an associated dyke structure (Figure 2.1c; see inset), inferring the dyke has collapsed and closed (Daczko & Halpin, 2009; Smith et al., 2015). The garnet reaction zones with or without dykes are exceptional “marker beds” during subsequent strain (Daczko et al., 2001b; Smith et al., 2015) and rock modification.

### **Transition Zone**

A thin (metre scale; ranging 2-7 m wide) transition zone (TZ) separates the gabbroic gneiss and ultrabasic granofels units (Figures 2.3a). This unit is termed the transition zone as the key structural features of the gabbroic gneiss (S1 gneissosity and garnet reaction zones) are progressively modified and may be traced from the gabbroic gneiss through the transition zone and into the ultrabasic granofels (Figure 2.2b,c). The northern and southern boundaries between the gabbroic gneiss and the transition zones are regular (Figure 2.3a), whereas the boundary between the transition zone and ultrabasic granofels is highly irregular in the north (Figure 2.3a,b) and more regular in the south (Figure 2.3a). Rounded pods of transition zone are also observed within the ultrabasic granofels unit (Figure 2.5a; 2.2a,c). Each garnet reaction zone is modified along strike from the gabbroic gneiss into a garnet train surrounded by leucosome (lower left Figure 2.3c; lower part of inset Figure 2.4b, Figure 2.5b) to a garnet train surrounded by pargasite amphibole (upper right Figure 2.3c; upper part of inset Figure 2.4b). The garnet trains surrounded by leucosome or pargasite amphibole are typically 5–20 mm wide. In many cases, garnet trains can alternate between being leucosome or hornblende-enclosed. As a result, the presence of garnet + pargasite amphibole or garnet + leucosome can change multiple times along a single garnet train within the transition zone. The transition zone is also characterised by large, randomly distributed garnet grains (<100 mm across) surrounded by thin leucosome (<10 mm across; Figure 2.5a, b). The mineralogy of the transition zone largely resembles the host gabbroic gneiss, being dominated by plagioclase (Figure 2.3a). However, pyroxene grains in the gabbroic gneiss are pseudomorphed by pargasite amphibole in the transition zone. Garnet reaction zones with or without dykes show an increased mode of garnet in the transition zone.

Figure 2.3



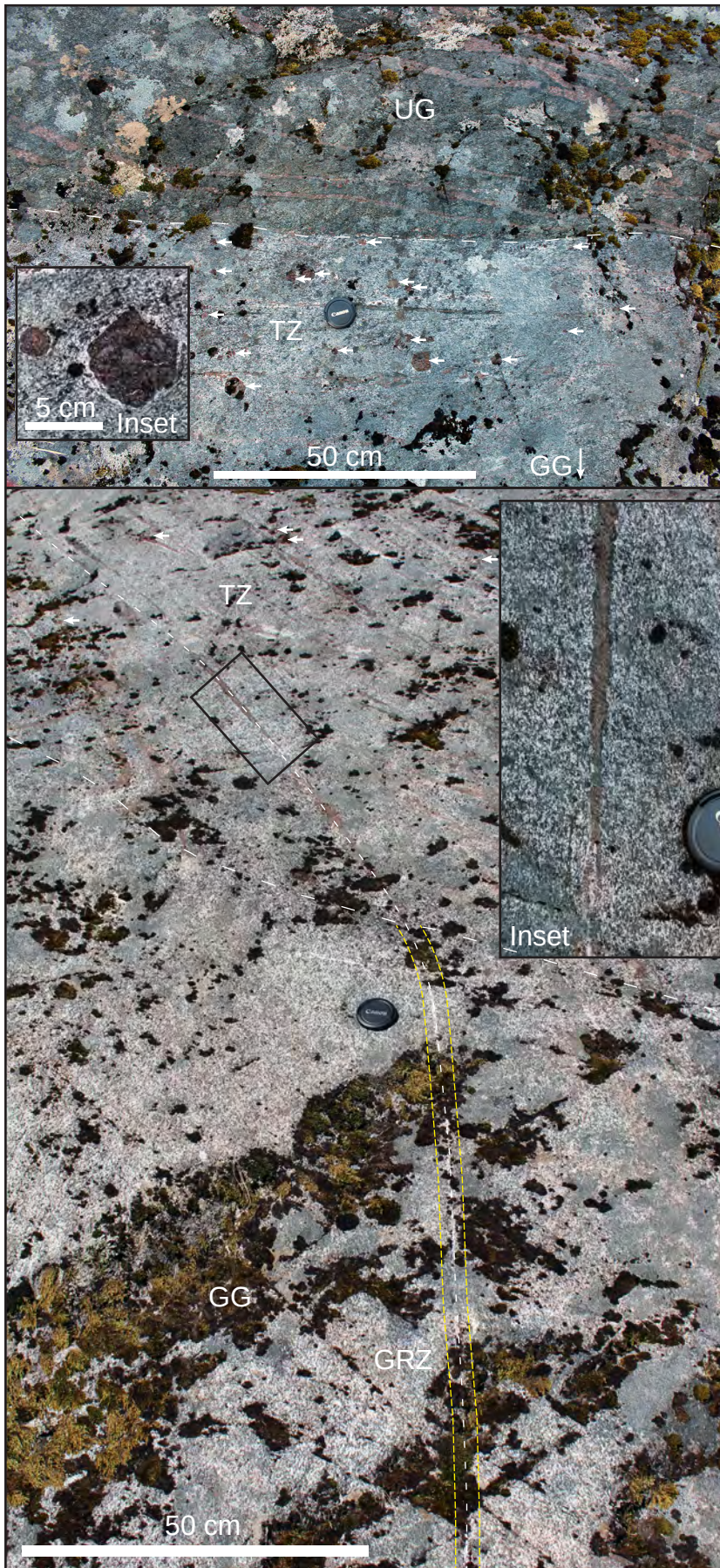
(a) Transition zone (TZ) between ultrabasic granofels (UG) and gabbroic gneiss (GG) with garnet reaction zones (GRZ). Note the irregular contact between UG and TZ and smooth contact between TZ and GG.

(b) Close up of the contact between the ultrabasic granofels (UG) and transition zone (TZ). Note the apophyses of UG are elongate along modified GRZ in the TZ.

(c) Variations in modified garnet reaction zones in the transition zone (TZ): (i) garnet trains surrounded by leucosome (lower left), (ii) garnet trains surrounded by hornblende (upper right). Note one garnet train has leucosome on the lower left and hornblende on the upper right (arrow).



Figure 2.4

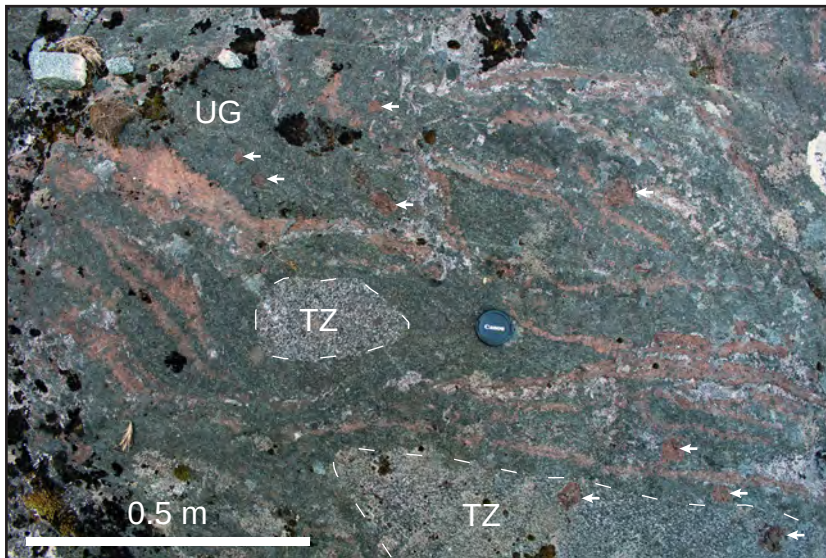


(a) Contact (dashed line) between the ultrabasic granofels (UG) and a narrow (1-4 m wide) transition zone (TZ) that is between the UG and gabbroic gneiss (GG; off the base of the photograph). Garnet (~2-7 cm; white arrows) surrounded by leucosome is characteristic of the TZ and is shown in inset.

(b) Contact (white dashed line) between the transition zone (TZ) and gabbroic gneiss (GG) with garnet reaction zones (GRZ - one outlined with yellow dashed lines). GRZ are modified and bent into the TZ. GRZ can be traced along strike where they change into garnet trains that may or may not be surrounded by leucosome and then hornblende (compare the two insets).



Figure 2.5



(a) Transition zone (TZ) rocks may be enclosed by the ultra-basic granofels (UG) unit with garnetite bands. Large (2-5 cm) garnet grains are present within both the TZ and UG.



(b) Contact (dashed line) between gabbroic gneiss (GG) with garnet reaction zones (GRZ) and transition zone (TZ). Large (2-7 cm) garnet grains (white arrows) are surrounded by leucosome. Modified GRZ within the transition zone comprise garnet trains surrounded by leucosome. Inset shows detail of euhedral to anhedral garnet grain shapes with the garnet trains. Note the ratio of garnet to leucosome varies.



## Ultrabasic Granofels

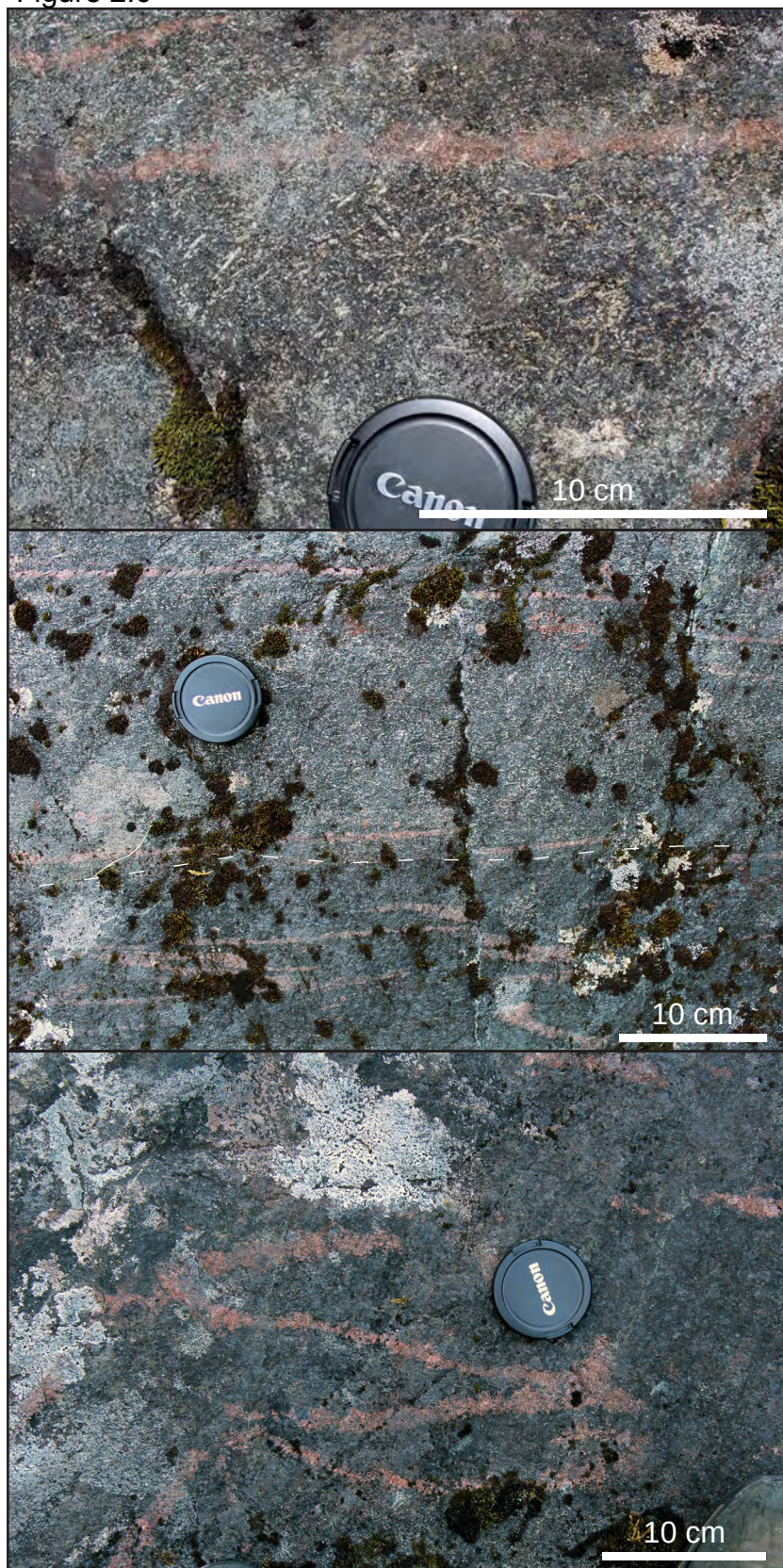
As noted above, the field observations presented here show that the ultrabasic granofels unit cuts a major thrust shear zone in the valley, indicating the unit is not a primary compositional band within the protolith gabbroic rocks of the Pembroke Granulite (Figure 2.1c). The ultrabasic granofels (UG) unit contains four rock types: (i) clinozoisite-rich hornblendite, (ii) hornblendite, (iii) minor garnetite, and (iv) minor pegmatoidal domains. The clinozoisite-rich hornblendite rock comprises 60–70 vol. % pargasite and 30–40 vol. % clinozoisite (Figure 2.5a). Tabular prismatic clinozoisite grains (<20 mm long) are pale green-brown in colour (Figure 2.6a,b). The clinozoisite is commonly randomly orientated, with the exception of areas of higher strain where they are aligned due to local deformation (Figure 2.2b, c). The hornblendite is predominantly pargasite and is homogenous (Figure 2.6b,c). The spatial relationships of these two main rock types in the ultrabasic granofels, defined by the presence or absence of clinozoisite, are sporadic and show no structural relationship to each other or surrounding garnetite bands.

Modified garnet reaction zones in the transition zone may be traced into the ultrabasic granofels unit (Figure 2.2b,c; 2.3a,b), where they become garnetite (Figures 2.6a,b,c and 2.7 a,b,c). Garnetite layers are commonly dismembered (Figure 2.7b), but may be traced as continuous features for up to 3 m, (Figure 2.2b,c). There are no apparent differences between garnetite bands hosted within clinozoisite-rich and clinozoisite-absent areas of the ultrabasic granofels unit. Garnetite layers may preserve the rectilinear patterns of garnet reaction zones in the gabbroic gneiss unit, reflecting lower strain areas of the ultrabasic granofels unit (Figure 2.7a). More commonly, intermediate to high strain areas of the ultrabasic granofels are characterised by variably folded and dismembered garnetite bands (Figure 2.7b) and in areas of considerable strain, sheath folds are preserved (Figure 2.7c).

Pegmatoidal domains are a minor component of the ultrabasic granofels unit (Figures 2.2a; 2.8a). These domains are commonly associated with coarse-grained, prismatic pargasite crystals (2-10 cm long), coarse-grained garnet grains up to 10 cm across) and tabular plagioclase feldspar grains (~2-7cm long) (Figure 2.7a,b,c). Interstitial pegmatoidal, or plagioclase-rich material is also found throughout the unit, commonly surrounding garnet trains (Figure 2.2b,c and Figure 2.7c). Coarse-grained garnets are common elsewhere throughout the ultrabasic granofels (1-10 cm) and grain size is variable across the unit (Figure 2.2a).



Figure 2.6



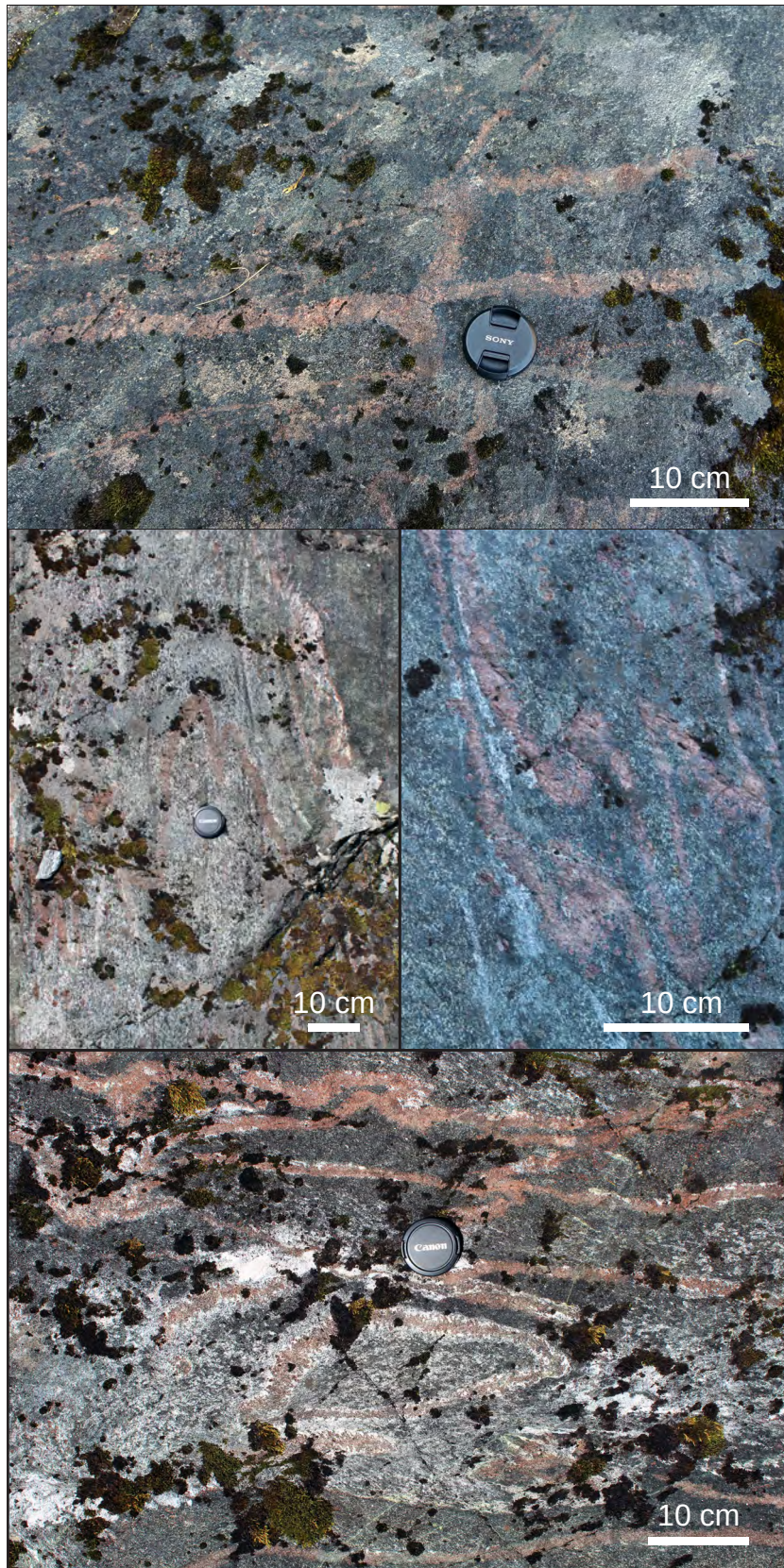
(a) Randomly orientated tabular prismatic clinozoisite (pale; 25-35 vol.%) and hornblende (dark; 55-65 vol.%) in the ultrabasic granofels. Note garnetite bands (~10 vol.%) in upper part of photograph.

(b) Contact (dashed line) between hornblende-clinozoisite-rock (upper) and hornblende (lower) with garnetite bands within each.

(c) Hornblende with folded garnetite bands.



Figure 2.7



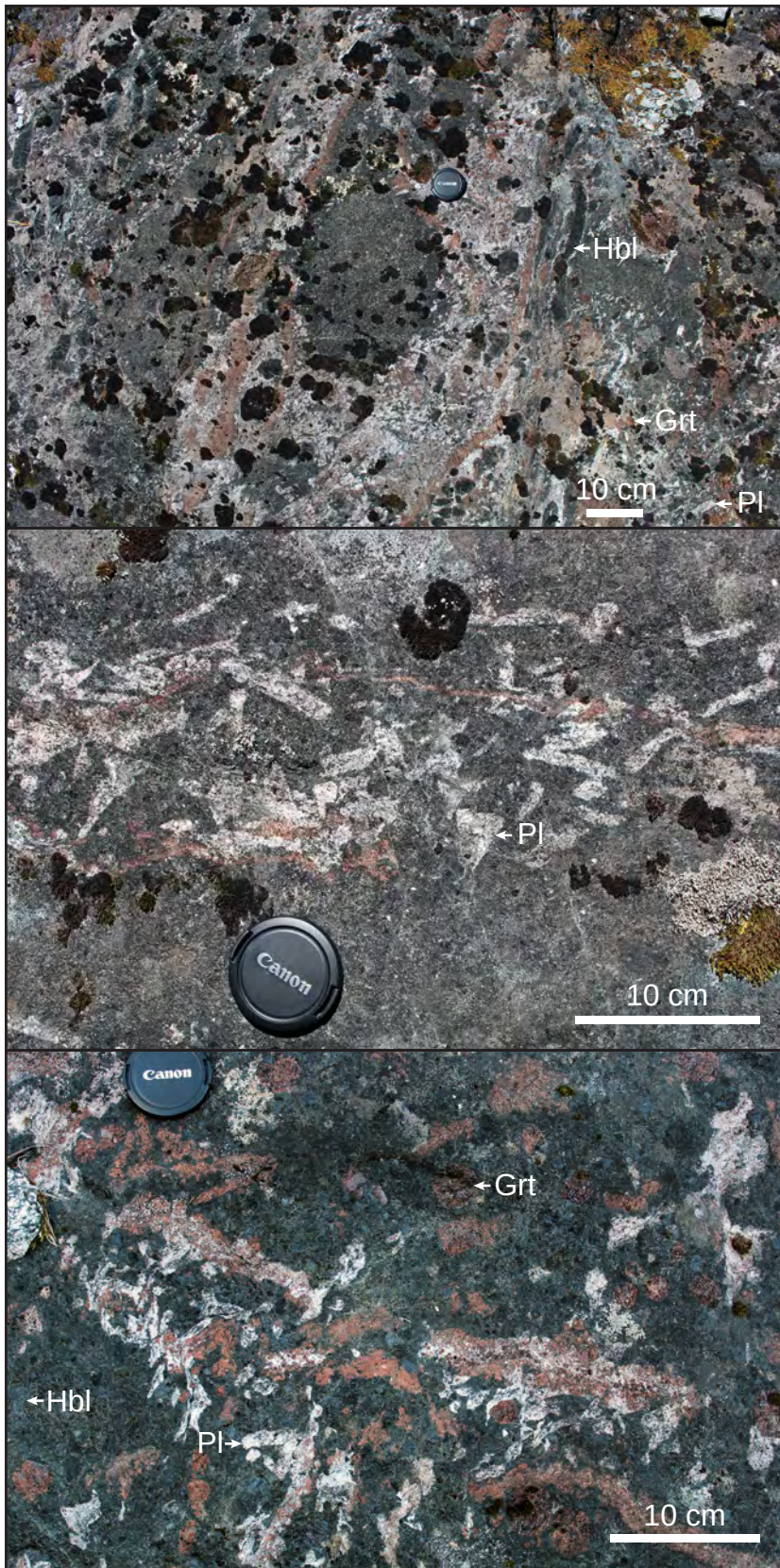
(a) Low strain parts of the ultrabasic granofels are characterised by garnetite bands that preserve the rectilinear pattern of garnet reaction zones (see Fig. 2.2c).

(b) Intermediate to high strain parts of the transition zone (left) and ultrabasic granofels (right) are characterised by open to tight folds of the garnetite bands.

(c) High strain parts of the ultrabasic granofels are characterised by sheath folds.



Figure 2.8



(a) Pegmatite with coarse hornblende (Hbl), garnet (Grt) and plagioclase (Pl). The ultra-basic granofels is dominated by hornblende-clinozoisite rock.

(b) Euhedral tabular plagioclase (Pl) grains within the hornblende-clinozoisite rock.

(c) Dismembered garnetite bands in hornblende (Hbl), plagioclase (Pl), garnet (Grt) pegmatite.



Areas of high strain are concentrated along the contact between the transition zone and the ultrabasic granulites (Figure 2.2a,b,c). High strain textures indicate local sinistral shearing. Towards the centre of the ultrabasic granulites, low strain textures are common.

### **3. Petrography**

Petrographical analysis was completed on polished thin sections. Mineral identification and microstructures of the samples were determined using optical microscopy using plane and polarized light (Macquarie University, Sydney, Australia). Back-scatter electron (BSE) imaging was performed on a Hitachi Desktop Scanning Electron microscope (SEM) (OptoFab node of the Australian National Fabrication Facility, Macquarie University). The operating conditions of the SEM were low vacuum and 15kv accelerating voltage. Modal mineral abundance was determined using image-processing software ImageJ (Rasband 1997-2015) applied to BSE images. The modal area % is inferred to equate to vol. % with an error of 3 area % (Rasband 1997-2015). Mineral abbreviations follow that of Whitney and Evans (2010).

#### **Gabbroic Gneiss**

Coarse-grained (1-2 mm) clinopyroxene (~15 vol. %) is more abundant than orthopyroxene (~10 vol. %, 0.2-1mm). It is common for orthopyroxene to be located within clinopyroxene grains (Figure 3.1a) as well as individual grains. Clinopyroxene and orthopyroxene are commonly surrounded by fine-grained inner coronae of pargasite-quartz symplectite clusters (~15%) (Figure 3.1a). This is sometimes surrounded by fine-grained outer corona of anhedral pargasite (~ 5%) (Figures 3.1a, b). Clinopyroxene commonly exhibit exsolution bands of orthopyroxene and oxides (Figure 3.3a, b). These mafic clusters (two-pyroxene  $\pm$  pargasite + quartz) are elongate to define the  $S_1$  gneissic foliation (Figures 3.3a, b; 2.2b).

Plagioclase typically forms ~45 vol. % and is partially recrystallised. Grain size ranges typically between 0.2-1 mm, where smaller grains have been more extensively recrystallised. Grains exhibit various deformation features, such as tapered twinning, undulose extinction, kink bands and sutured grain boundaries.

Abundant randomly orientated, fine-grained clinozoisite and kyanite grains are present within the plagioclase-rich areas. The abundance of clinozoisite and kyanite is 5-7

**Table 1. Mineral modal proportions for samples of this study, determined using ImageJ**

| Rock Type/Sample No.                              | Modal percentage mineralogy (vol. %)                                                                                      |
|---------------------------------------------------|---------------------------------------------------------------------------------------------------------------------------|
| Host Gabbroic Gneiss<br>PV1506 & PV1507           | Cpx - 15 %<br>Opx - 10 %<br>(Par + qz) sym - 15 %<br>Par - 5 %<br>Pl - 45%<br>Cz + ky 5-7 %<br>Minors (ap, rt, ilm) 3-5 % |
| Transition Zone<br>PV1508                         | (Par + qz) sym - 30 %<br>Par - 15 %<br>Pl - 45%<br>Cz + ky 5-7 %<br>Minors (ap, rt, ilm) 3-5 %                            |
| Hornblendite<br>PV1514 & PV1515                   | Par - 90 %<br>Pl - 7 %<br>Minors (bt, rt, ilm) - 3%                                                                       |
| Clinozoisite-rich Hornblendite<br>PV1509 & PV1518 | Par - 60-70 %<br>Cz - 25-35 %<br>Pl - 5<br>Minors (bt, rt, ilm) - <3 %                                                    |
| Pegmatoidal felsic domains                        | Pl - 40 %<br>Sc - 40%<br>Par - 20%<br>Garnet - 0-20 %                                                                     |

vol. % (Figure 3.1c). Commonly, clinozoisite forms colourless, elongate crystals. They are either euhedral, singular grains or exhibit a 'splayed' texture when clustered (Figure 3.1c). Kyanite is tabular and elongate, commonly seen as single euhedral grains. In some cases, kyanite can be associated closely with clinozoisite (Figure 3.1c).

Accessory minerals within the host include ilmenite, rutile and apatite and account for 3-5 vol. %. Rutile and ilmenite are present as individual grains. Ilmenite commonly exhibits exsolution textures with rutile (Figure 3.1c).

The garnet-rich areas, termed garnet reaction zone (GRZ) exhibits garnet in close association with clinopyroxene and quartz, with or without biotite (Figure 3.2a). Garnet is intergrown with areas of garnet + clinopyroxene + quartz  $\pm$  biotite as well as garnet + quartz and garnet + clinopyroxene areas. Biotite, where present, has poorly formed crystal faces (Figure 3.2a).

### Transition zone

Transition zone samples are similar in microstructure to the gabbroic gneiss. The most significant difference is a shift in the modal mineral composition of the rock, where clinopyroxene and orthopyroxene are not present (Table 1; Figure 3.2a, b). The modal



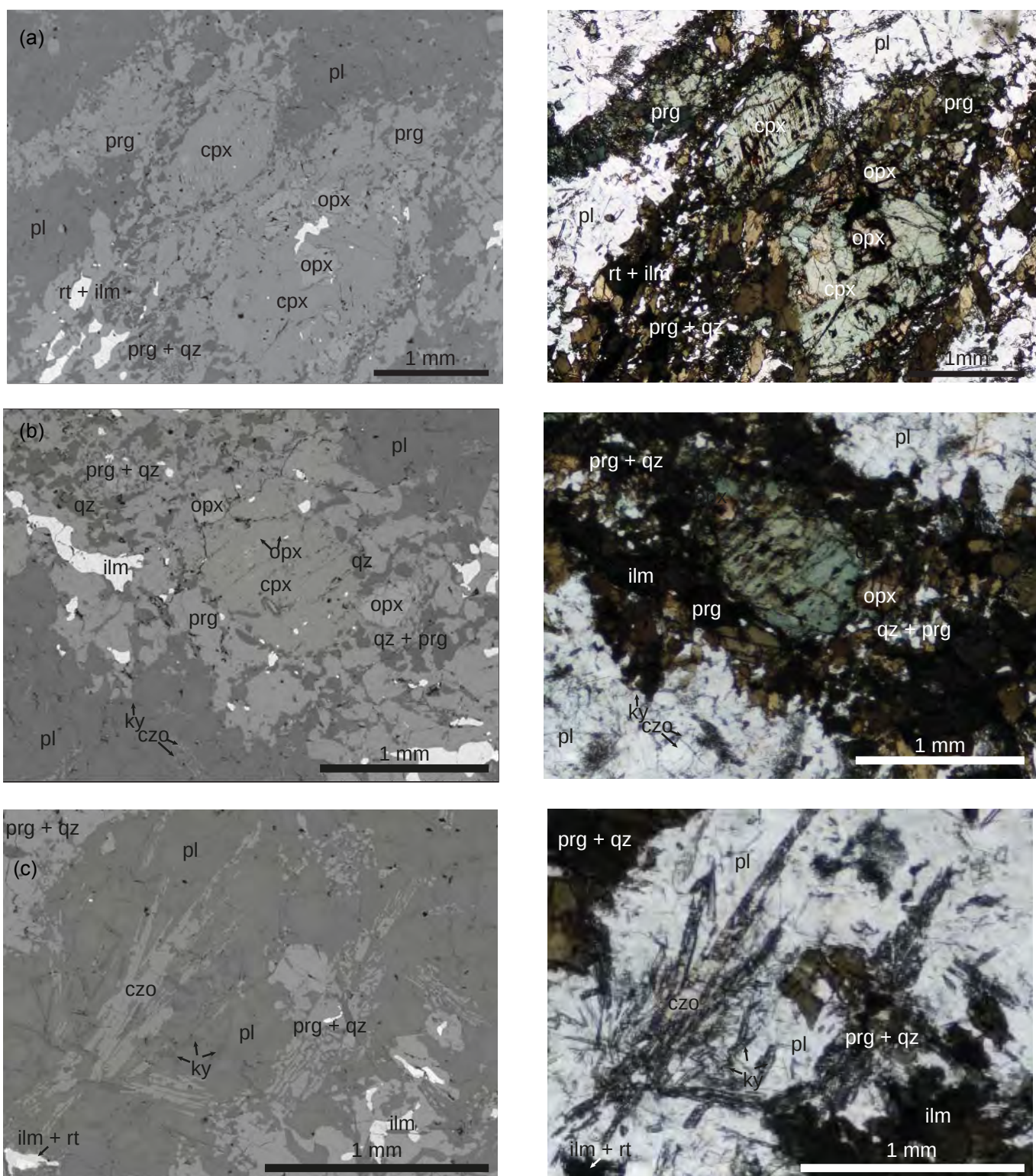


Figure 3.1

(a) Gabbroic gneiss (GG) BSE image (left) and light microscopy (right) showing 2-pyroxene (clinopyroxene = cpx and orthopyroxene = opx) + pargasite (prg) + quartz (qz) defining a strong foliation. The matrix is predominantly plagioclase (pl).

(b) Gabbroic gneiss (GG) BSE image (left) and light microscopy (right) showing 2-pyroxene (clinopyroxene = cpx and orthopyroxene = opx) + pargasite (prg) + quartz (qz) defining a strong foliation. The matrix is plagioclase (pl) + minor clinozoisite (czo), rutile (rt), ilmenite (ilm) and apatite (ap).

(c) Gabbroic gneiss (GG) BSE image (left) and light microscopy (right) showing a close-up examination of the plagioclase-rich (pl) matrix. Fibrous clinozoisite (czo) needles as either single grains or splayed clusters are common. Tabular kyanite (ky) crystals are intermittent within the matrix.



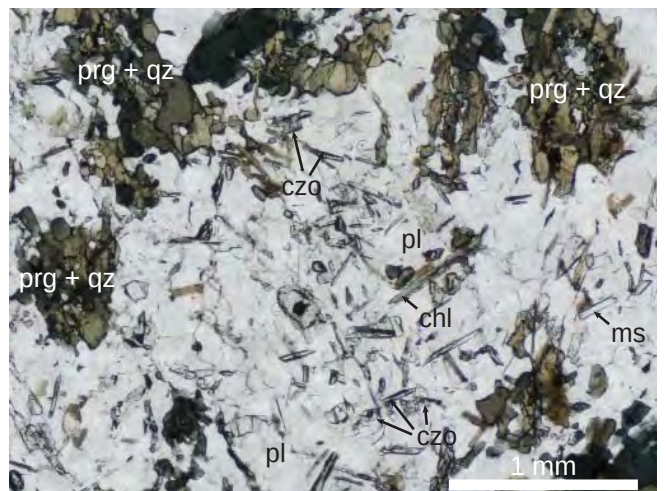
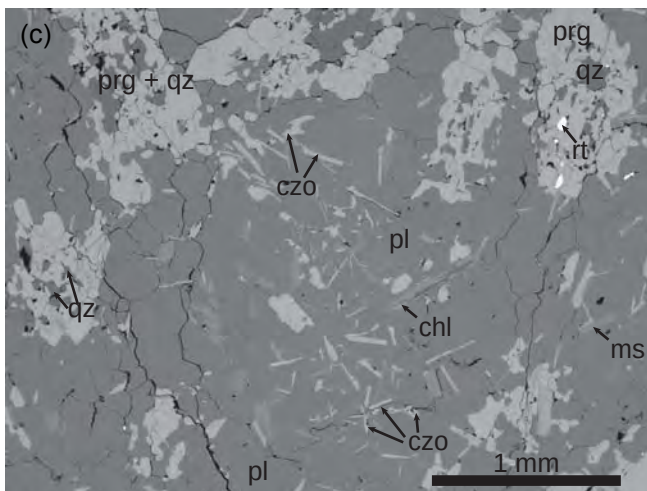
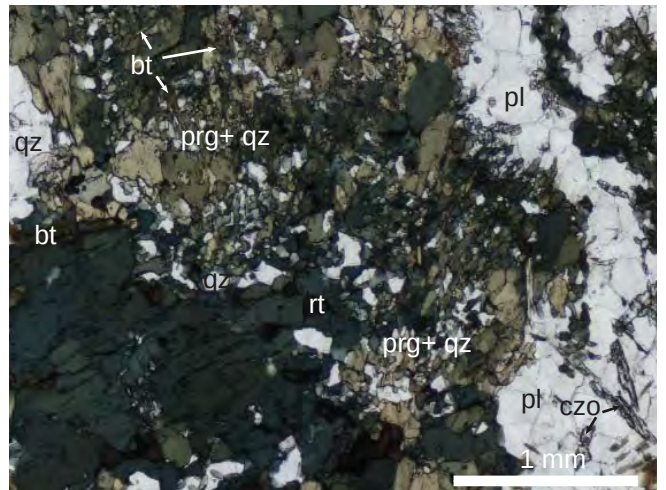
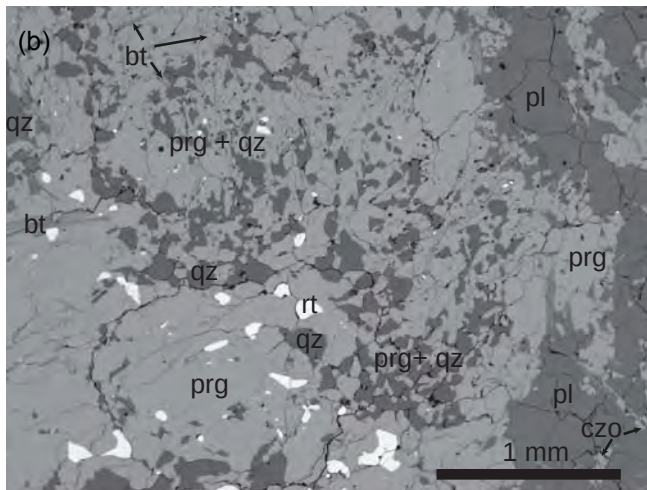
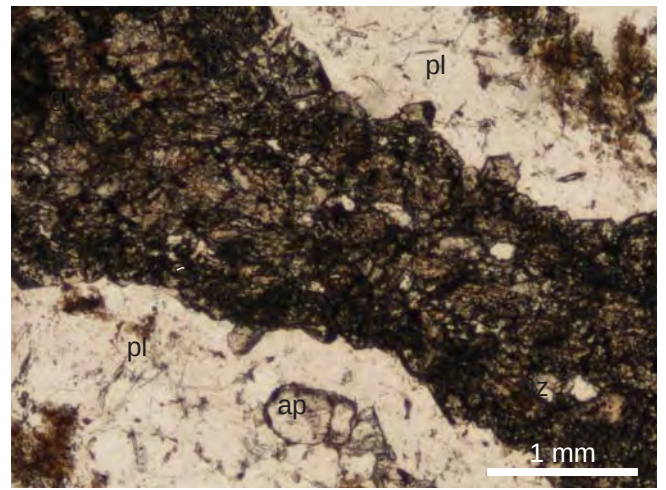
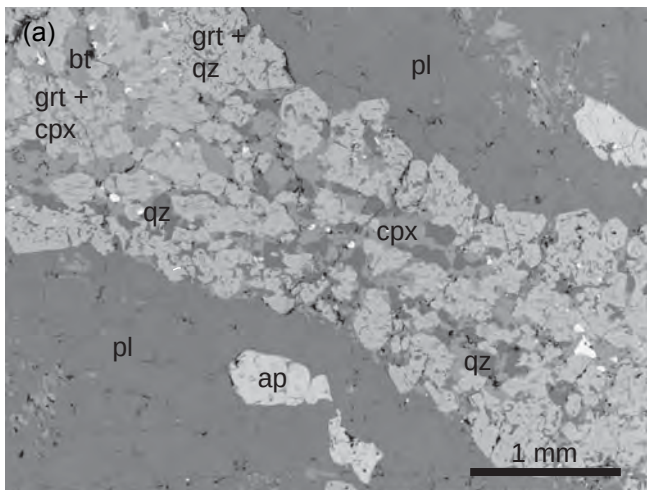


Figure 3.2

(a) GRZ within the host GG BSE image (right) and light microscopy (left). GRZ is comprised of garnet (grt), quartz (qz) and clinopyroxene (cpx) symplectites and define the S1 foliation. Less common clinozoisite (czo) and kyanite (ky) present within the plagioclase-dominant (pl) matrix. Minor apatite (ap) also found within the host gneiss.

(b) Transition zone mafic clusters of inner corona pargasite (prg) + quartz (qz) symplectites and outer corona pargasite. Inclusions of biotite (bt), rutile (rt) common.

(c) Transition zone matrix is plagioclase-dominant with randomly oriented, elongate clinozoisite (czo), muscovite (ms) with evidence of chloritisation (chl).

proportions range ~15 vol. % pargasite, ~ 30 vol. % pargasite-quartz symplectite defining the foliation and ~45 vol. % plagioclase, 5-7 vol. % clinozoisite and muscovite within the matrix. Accessory minerals, rutile, ilmenite and apatite are still present, accounting for 3-5 vol. % (Table 1). The clusters of pargasite  $\pm$  quartz with minor rutile, ilmenite, biotite and apatite retain the S<sub>1</sub> gneissic foliation (Figure 3.2a,b).

Plagioclase exhibits recrystallisation textures where the grain size has become homogenised (~0.5 mm). Grains commonly exhibit sutured or bulging grain boundaries. In rare cases, deformation twins and kink bands are also present. Plagioclase domains contain clinozoisite, as in the gabbroic gneiss. However, kyanite is no longer present. In small abundances, muscovite is present (Figure 3.2c).

## **Ultrabasic granofels**

### ***Hornblendite***

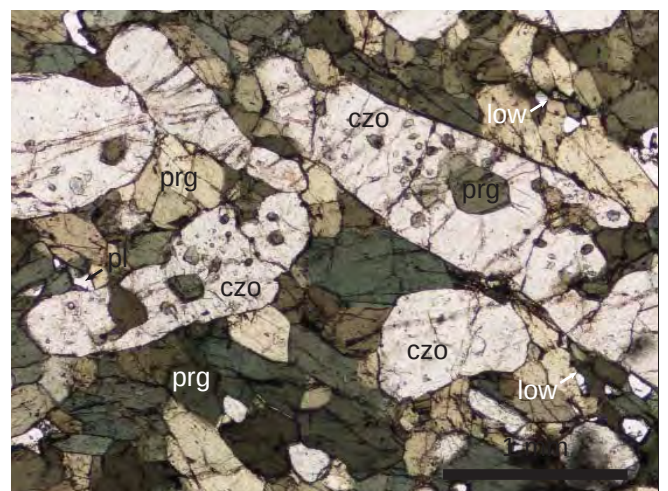
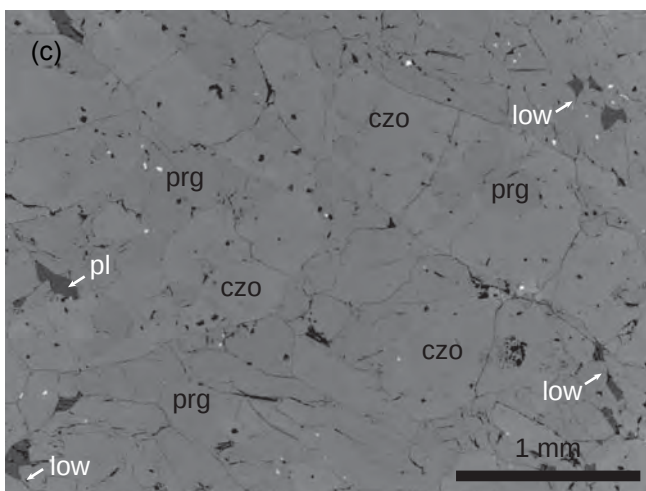
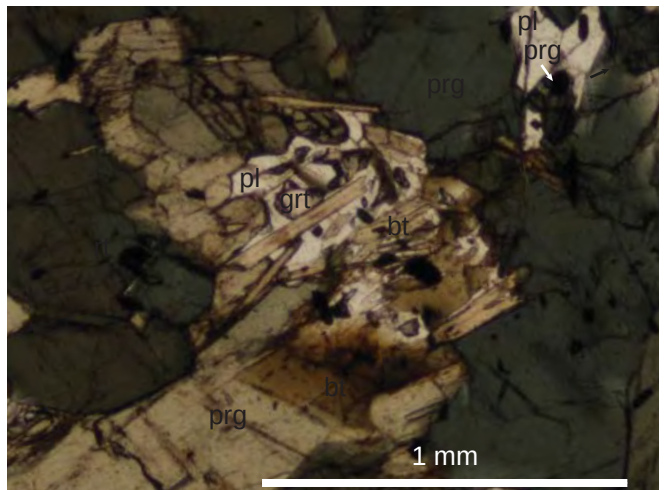
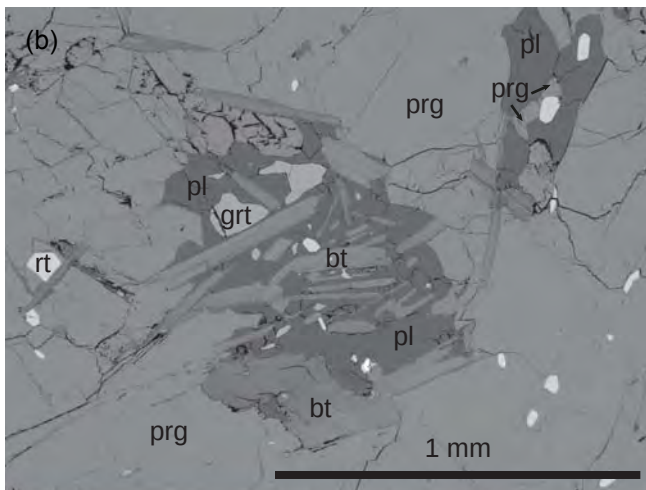
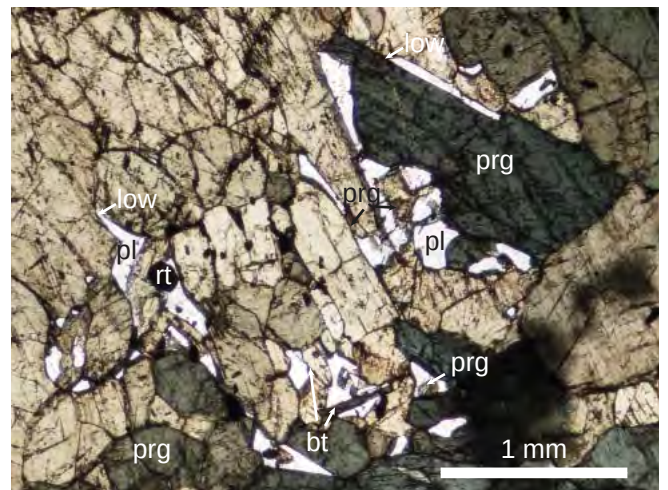
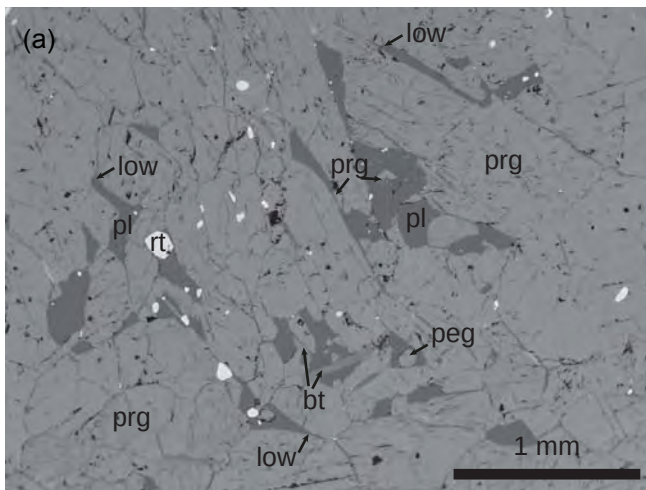
Hornblendite is predominantly comprised of pargasite (90-95 vol. %) with minor interstitial plagioclase (~ 5 vol. %). Pargasite is coarse-grained (2-10 cm) (Figure 3.3a). Interstitial plagioclase (<0.5 mm) is commonly seen between coarse-grained pargasite (Figure 3.3a) with low dihedral angles below 10° (Figures 3.3a, b, c).

In some areas, clusters of finer-grained euhedral to anhedral minerals are observed within single plagioclase grains (Figure 3.3b). The clusters consist of biotite, amphibole, garnet and oxides (rutile/ilmenite). Biotite has well formed crystal faces and is commonly elongate; it may be either completely enclosed within plagioclase, or in most cases, extends out into the hornblendite (Figure 3.3b). Amphibole, garnet and oxide inclusions are commonly enclosed within the plagioclase. In rare cases, amphibole grains may be present towards the edge of the plagioclase in contact with other grains, but garnet grains are invariably included within plagioclase (Figure 3.3b).

### ***Clinozoisite-rich Hornblendite***

This rock type mineralogy is similar to the hornblendite, however, clinozoisite is present in significant amounts (25-35 vol. %), with the remainder comprised of pargasite (60-70 vol. %) and interstitial plagioclase (~5 vol. %) (Figure 3.3c; Table 1). Pargasite and plagioclase exhibit textures common to the hornblendite, including low dihedral angles (< 10°) of plagioclase (Figures 3.3c and 3.4a).





**Figure 3.3**

(a) Hornblende BSE image (left) and light microscopy (right) showing pargasite (prg) plagioclase (pl) and minor oxide mineralogy. Small inclusions within pl of biotite (bt) and rutile (rt).

(b) Plagioclase with abundant inclusions of biotite (bt), garnet (grt), pargasite (prg) and rutile (rt). Grt is completely surrounded by pl. Bt commonly extends into the pargasite matrix as elongate, tabular crystals. Prg in contact with rt or edges of pl grain.

(c) Clinozoisite-rich hornblende with prismatic clinozoisite (czo) grains with rounded ends. Several czo grains have prg inclusions. Pl exhibits low dihedral angles (arrow).

Clinzoisite grains are typically between ~0.5-2 cm length and ~0.2-1.5 mm in width, and are colourless (Figure 3.3c). Their grain shape and habit is different to clinzoisite grains present within the gabbroic gneiss and transition zone gabbro. Clinzoisite grains common to the gneissic rock are elongate. However, clinzoisite within this rock type are prismatic with rounded ends (Figures 3.1c and 3.3c). They are commonly randomly orientated, except in samples taken from high strain areas (Figure 3.3c). Grains frequently have abundant inclusions of pargasite (Figure 3.3c).

### ***Garnet trains***

Garnet trains account for approximately ~10 vol. % of the hornblende and hornblende-clinzoisite rocks and occur in both rock types. In garnet trains (Figure 2.7a,b,c), garnet comprises >95 vol. % (Figures 3.2a and 3.4a). The remaining 5 vol. % predominantly consists of pargasite inclusions with rare biotite, rutile, ilmenite and plagioclase inclusions (Figure 3.4a). The modal proportion of garnet has significantly increased from GRZ in the host and transition, once entering the ultrabasic granofels unit (Table 1).

### ***Felsic domains***

Pegmatoidal felsic domains found within the ultrabasic granofels unit show high percentages of coarse-grained, felsic material, with large porphyroblasts of garnet, amphibole and plagioclase (Figure 2.8a). Thin sections in this study are fine-grained interstitial materials to coarse-grained pegmatoidal domains (Figure 2.2a, b,c).

The felsic material within these domains is comprised of 30 vol. % plagioclase and 30 vol. % scapolite with the remaining mineralogy comprised of variably sized garnet, pargasite and rare oxides (Table 1; Figure 3.4b, c). Two thin sections of pegmatoidal domains were examined with different minerals present within the pegmatoidal areas.

Sample PV1511B has abundant garnet (20 vol. %; Figure 3.4b). Garnet are euhedral, 0.5-1 mm diameter, and are rarely in contact with each other (Figure 3.4b). The remaining 20 vol. % is pargasite that is commonly intergrown with scapolite (Figure 3.4b). Sample PV1513 is predominantly plagioclase and scapolite with the remaining 50 vol. % comprised of pargasite only (Figure 3.4c). Pargasite exhibits poor crystal faces and commonly is intergrown with scapolite.



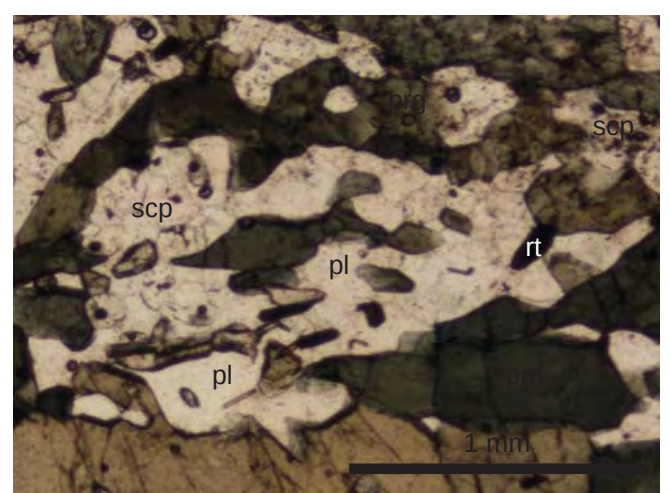
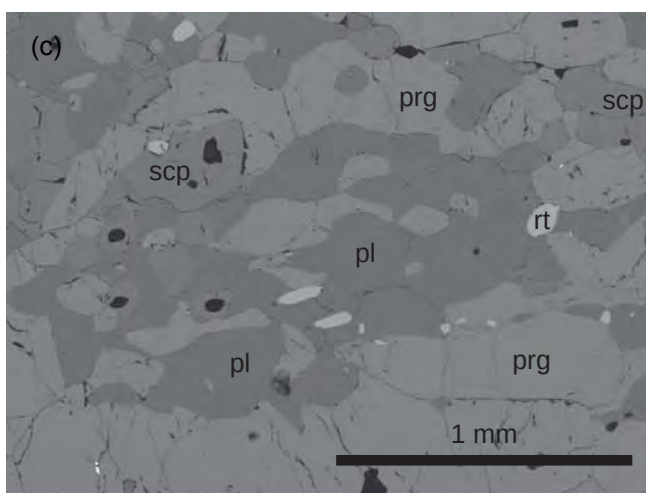
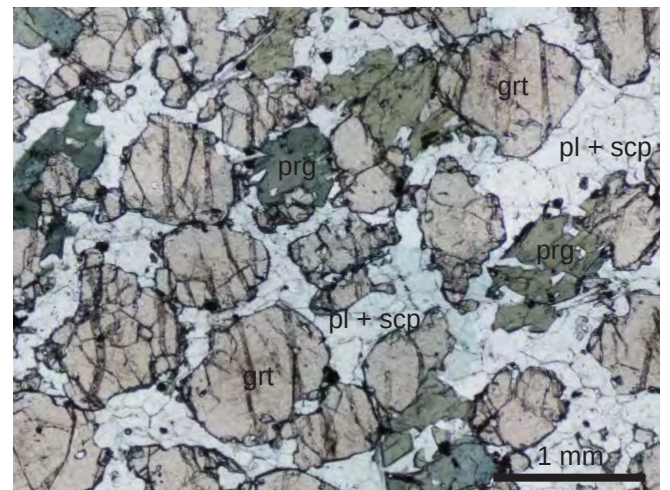
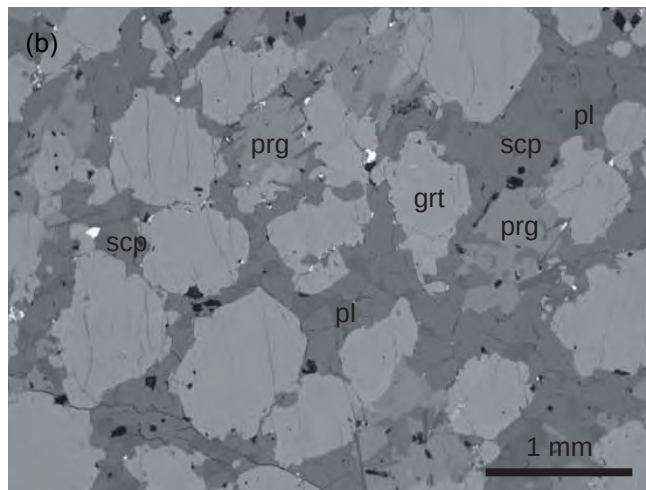
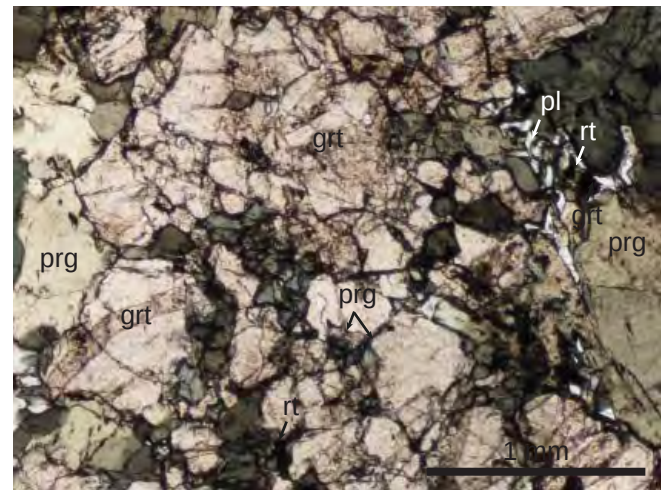
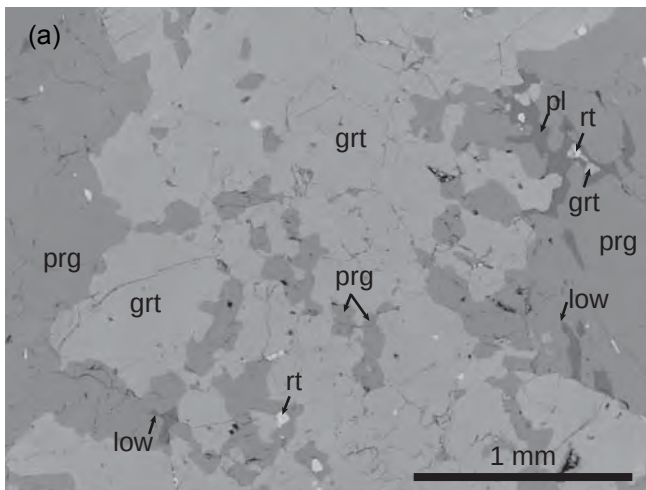


Figure 3.4

(a) Garnet train within ultramafic granofels exhibit the same characteristics in both hornblende and clinozoisite-rich hornblende domains. Garnet (grt) has abundant inclusions of pargasite (prg). Plagioclase with inclusions of rutile (rt) and grt with low dihedral angles (arrow).

(b) Felsic-rich domains comprised of plagioclase (pl) and scapolite (scp), which is interstitial to the pegmatite. Euhedral garnet (grt), rarely in contact with each other. Pargasite (prg), sometimes intergrown with scp.

(c) Felsic-rich domain, as in (b) lacking garnet. Predominantly scp, pl and prg, with minor rutile (rt).

## **4. Whole rock geochemistry**

### **Methods**

Whole rock geochemical analyses of representative samples of gabbroic gneiss (1 sample), transition zone (1 sample), hornblendite (3 samples) and clinozoisite-rich hornblendite (3 samples) are presented in Table 2.

Fresh, unaltered samples were initially crushed using a hydraulic press with tungsten carbide plates until rock fragments were  $< 1\text{cm}^3$ . Samples fragments were reduced to a fine powder using a tungsten carbide barrel and ring mill.

### **X-ray fluorescence (XRF)**

Whole rock major and trace element concentrations were determined at the Mark Wainwright Analytical Centre at the University of New South Wales, Kensington, Australia, using a PW2400 WDXRF Spectrometer. For major element analysis, whole rock powders were fused into glass discs using a lithium borate flux at  $1050^\circ\text{C}$ . Trace elements Pb, Sr, Rb, Y, Zr, Cu, Zn, Ni, Ga, Co, U, Sn, As, Sb, Mo, Cd, Ba, Cr, V, Nb, Th and Ce were determined on a 10 % w/w wax pressed powdered disc with boric acid backing. Loss on ignition (LOI) was established by heating the samples to  $1050^\circ\text{C}$  for 1 hour.

### **Inductively coupled plasma-mass spectrometry (ICP-MS)**

Samples were analysed for bulk trace and rare earth elements by inductively coupled plasma-mass spectrometry (ICP-MS) at the Geochemical Analysis Unit, Macquarie University, Sydney, Australia. Powders were digested with concentrated  $\text{HNO}_3$  and HF and measured using 1/1000 and 1/5000 dilutions in 2%  $\text{HNO}_3$  + 0.25% HF.

Diluted rock solutions were analysed by ICP-MS using an Agilent 7500cs ICP-MS Octopole reaction system with a self-aspiring micro-flow nebulizer. He was used as the collision gas. External calibration standards (BCR-2a) and US Geological Survey reference samples (BIR-1 and BHVO-2) were run at the start and end of both runs. Precision is typically better than 5 % for each element.

### **Isocon methods**

To evaluate the mass transfer of elements between the three rock types, an isocon analysis was applied to the bulk chemistry data, following the method by Grant (1986). This was based on composition-volume equations proposed by Gresens (1967). For this method, it is necessary to fix one or a group of elements as immobile, often aluminum



Table 2. Whole rock major (XRF) and trace (ICP-MS) geochemical analyses (Hbl = hornblende, Czo-rich Hbl= clinozoisite-rich hornblende. Zr results from XRF).

|                                    | Gabbroic<br>gneiss | Transition<br>Zone | Hbl          | Hbl          | Hbl          | Czo-rich<br>hbl | Czo-rich<br>hbl | Czo-rich<br>hbl |
|------------------------------------|--------------------|--------------------|--------------|--------------|--------------|-----------------|-----------------|-----------------|
| <b>SiO<sub>2</sub></b>             | 52.1               | 52.6               | 41.6         | 41.2         | 42.0         | 40.4            | 40.3            | 42.2            |
| <b>TiO<sub>2</sub></b>             | 0.9                | 0.7                | 1.4          | 1.9          | 1.8          | 1.6             | 1.3             | 1.4             |
| <b>Al<sub>2</sub>O<sub>3</sub></b> | 19.5               | 19.4               | 15.6         | 14.5         | 12.9         | 17.8            | 17.9            | 15.3            |
| <b>Fe<sub>2</sub>O<sub>3</sub></b> | 8.5                | 8.2                | 15.3         | 15.3         | 16.2         | 13.4            | 14.7            | 16.5            |
| <b>Mn<sub>3</sub>O<sub>4</sub></b> | 0.2                | 0.1                | 0.1          | 0.2          | 0.2          | 0.1             | 0.2             | 0.3             |
| <b>MgO</b>                         | 4.3                | 4.4                | 10.5         | 11.0         | 11.4         | 8.4             | 8.3             | 9.7             |
| <b>CaO</b>                         | 8.3                | 8.2                | 10.1         | 10.9         | 10.4         | 13.1            | 12.0            | 10.1            |
| <b>Na<sub>2</sub>O</b>             | 5.3                | 5.1                | 3.5          | 3.1          | 2.3          | 2.7             | 2.9             | 3.2             |
| <b>K<sub>2</sub>O</b>              | 0.5                | 0.6                | 0.7          | 0.6          | 0.7          | 0.5             | 0.4             | 0.5             |
| <b>P<sub>2</sub>O<sub>5</sub></b>  | 0.1                | 0.2                | 0.0          | 0.0          | 0.0          | 0.2             | 0.4             | 0.2             |
| <b>L.O.I.</b>                      | 0.4                | 0.9                | 1.4          | 1.4          | 2.0          | 1.4             | 1.5             | 1.4             |
| <b>Total</b>                       | <b>100.1</b>       | <b>100.4</b>       | <b>100.2</b> | <b>100.1</b> | <b>100.0</b> | <b>99.6</b>     | <b>100.0</b>    | <b>100.7</b>    |
| <i>Li</i>                          | 4.1                | 11.5               | 47.7         | 6.4          | 9.4          | 6.8             | N/A             | 8.4             |
| <i>Be</i>                          | 0.8                | 0.8                | 1.1          | 1.0          | 0.9          | 1.1             | N/A             | 0.9             |
| <i>Sc</i>                          | 21.7               | 21.0               | 36.3         | 60.0         | 62.7         | 42.0            | N/A             | 42.9            |
| <i>V</i>                           | 207.9              | 196.4              | 419.7        | 441.8        | 536.2        | 461.4           | N/A             | 353.7           |
| <i>Cr</i>                          | 20.7               | 22.3               | 44.2         | 197.7        | 133.8        | 46.8            | N/A             | 104.9           |
| <i>Co</i>                          | 56.9               | 54.5               | 61.5         | 71.4         | 70.0         | 89.0            | N/A             | 73.4            |
| <i>Ni</i>                          | 16.8               | 16.7               | 39.6         | 78.6         | 69.9         | 38.9            | N/A             | 76.5            |
| <i>Cu</i>                          | 31.8               | 71.1               | 16.1         | 73.9         | 54.4         | 11.2            | N/A             | 392.6           |
| <i>Zn</i>                          | 72.5               | 76.6               | 132.7        | 120.1        | 127.9        | 126.8           | N/A             | 122.5           |
| <i>Ga</i>                          | 19.2               | 19.2               | 19.7         | 20.4         | 21.5         | 28.6            | N/A             | 18.5            |
| <i>Rb</i>                          | 2.2                | 4.3                | 2.2          | 3.2          | 5.7          | 1.5             | N/A             | 2.3             |
| <i>Sr</i>                          | 827.4              | 876.9              | 111.2        | 176.5        | 90.3         | 556.3           | N/A             | 432.4           |
| <i>Y</i>                           | 9.2                | 9.3                | 4.2          | 19.2         | 20.5         | 21.8            | N/A             | 16.4            |
| <i>Zr (XRF)</i>                    | 42.0               | 44.3               | 25.1         | 31.8         | 41.2         | 37.9            | 38.2            | 35.6            |
| <i>Nb</i>                          | 1.2                | 1.0                | 0.7          | 1.9          | 2.4          | 2.1             | N/A             | 0.9             |
| <i>Mo</i>                          | 0.2                | 0.2                | 0.1          | 0.2          | 0.2          | 2.0             | N/A             | 0.3             |
| <i>Cd</i>                          | 0.0                | 0.0                | 0.0          | 0.0          | 0.0          | 0.0             | N/A             | 0.0             |
| <i>Cs</i>                          | 0.1                | 0.2                | 0.1          | 0.1          | 0.3          | 0.0             | N/A             | 0.1             |
| <i>Ba</i>                          | 421.5              | 262.8              | 76.4         | 83.8         | 111.7        | 86.0            | N/A             | 83.3            |
| <i>La</i>                          | 5.5                | 6.1                | 0.1          | 1.3          | 3.6          | 15.2            | N/A             | 20.3            |
| <i>Ce</i>                          | 11.8               | 13.3               | 0.5          | 6.0          | 14.4         | 33.0            | N/A             | 46.6            |
| <i>Pr</i>                          | 1.5                | 1.7                | 0.1          | 1.2          | 2.6          | 4.6             | N/A             | 6.4             |
| <i>Nd</i>                          | 6.8                | 7.6                | 0.6          | 7.7          | 14.2         | 20.2            | N/A             | 28.5            |
| <i>Sm</i>                          | 1.6                | 1.8                | 0.3          | 2.9          | 4.3          | 5.4             | N/A             | 5.7             |
| <i>Eu</i>                          | 0.9                | 0.9                | 0.2          | 0.9          | 1.2          | 2.5             | N/A             | 1.6             |
| <i>Tb</i>                          | 0.2                | 0.2                | 0.1          | 0.5          | 0.6          | 0.8             | N/A             | 0.6             |
| <i>Gd</i>                          | 1.6                | 1.8                | 0.5          | 3.4          | 4.2          | 5.5             | N/A             | 4.5             |
| <i>Dy</i>                          | 1.4                | 1.5                | 0.7          | 3.3          | 3.6          | 4.4             | N/A             | 3.1             |
| <i>Ho</i>                          | 0.3                | 0.3                | 0.1          | 0.7          | 0.7          | 0.7             | N/A             | 0.6             |
| <i>Er</i>                          | 0.8                | 0.9                | 0.4          | 1.8          | 1.8          | 1.6             | N/A             | 1.5             |
| <i>Yb</i>                          | 0.8                | 0.8                | 0.3          | 1.5          | 1.4          | 0.8             | N/A             | 1.3             |
| <i>Lu</i>                          | 0.1                | 0.1                | 0.0          | 0.2          | 0.2          | 0.1             | N/A             | 0.2             |
| <i>Hf</i>                          | 0.4                | 0.3                | 0.6          | 0.9          | 1.1          | 0.7             | N/A             | 0.7             |
| <i>Ta</i>                          | 0.1                | 0.1                | 0.0          | 0.1          | 0.1          | 0.2             | N/A             | 0.1             |
| <i>Th</i>                          | 0.1                | 0.2                | 0.0          | 0.0          | 0.0          | 0.5             | N/A             | 1.2             |
| <i>U</i>                           | 0.0                | 0.1                | 0.0          | 0.0          | 0.0          | 0.2             | N/A             | 0.2             |

(Al<sub>2</sub>O<sub>3</sub>), zirconium (Zr) or sometimes titanium (TiO<sub>2</sub>) (Grant 2005). Commonly, aluminum is fixed as immobile. However, Al<sub>2</sub>O<sub>3</sub> is known to be mobile in partial melting and the separation of leucosome (Grant, 2005). Therefore, for this study, zirconium has been chosen as a more appropriate 'immobile' element within the ultrabasic granofels system, to be fixed for the isocon analysis.

## **Results**

### **Major elements**

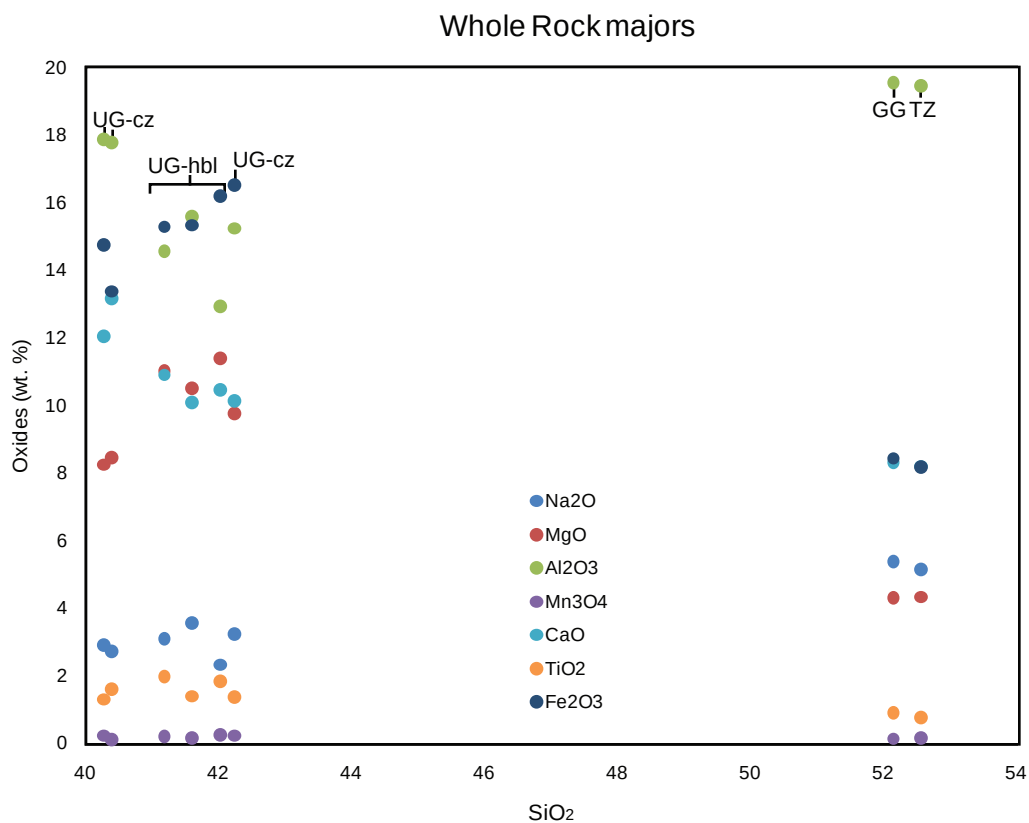
All major element whole-rock chemistry data is presented in Table 2. The major oxides of the host gabbroic gneiss, the transition zone and ultrabasic granofels (separated into hornblende and clinozoisite-rich hornblende) are plotted against SiO<sub>2</sub> in Figure 4a. The host and transition zone bulk compositions are similar, and variations are within error. SiO<sub>2</sub> content varies between 52.13-52.58 wt.%. SiO<sub>2</sub> content is less (41.17-42.03 wt.%) for hornblende and clinozoisite-rich hornblende samples (40.41-42.25 wt.%). All major elements with the exception of Mn show significant differences between the host/transition samples and the ultrabasic granofels samples (Figure 4a). Al<sub>2</sub>O<sub>3</sub> content is lowered from 19.53 wt.% (gabbroic gneiss) and 19.43 wt.% (transition zone) to 12.90-15.57 wt.% for hornblende and 15.26-17.85 wt.% for clinozoisite-rich hornblende samples. Na<sub>2</sub>O decreases from 5.11-5.35 (gabbroic gneiss and transition zone) to 2.34-3.53 wt.% (hornblende) and 2.69-3.22 wt.% (clinozoisite-rich hornblende). MgO increases from 4.29-4.36 wt.% to 10.48-11.37 (hornblende) and 8.26-9.72 wt.% for clinozoisite-rich hornblende. CaO increases from 8.22-8.29 wt.% in gabbro and transition zone to 10.06-10.92 wt.% hornblende 10.12-13.13 wt.% in clinozoisite-rich hornblende. Fe<sub>2</sub>O<sub>3</sub> increases from 8.16-8.46 wt.% in gabbro and transition zone to 15.31-16.17 wt.% hornblende 13.40-16.51 wt.% in clinozoisite-rich hornblende. Among the ultrabasic granofels samples, Al<sub>2</sub>O<sub>3</sub> and CaO weakly correlate negatively with increases in SiO<sub>2</sub> content, while MgO and Fe<sub>2</sub>O<sub>3</sub> define a weak negative correlation.

### **Trace and rare earth elements**

Trace and rare earth element concentrations are presented in Table 2. Trace element concentrations normalised to primitive mantle values from McDonough and Sun (1995) are graphed in Figure 4b. Ni contents are lowest in gabbroic gneiss and transition zone (16.7-16.8 ppm), and higher in hornblende (39.6-78.6 ppm) and clinozoisite-rich hornblende (38.9-76.5 ppm). Sr is high in gabbroic gneiss and transition zone (827.4-876.9 ppm) and lower in hornblende (90.3-176.5 ppm). Clinozoisite-rich samples have Sr



4a



4b

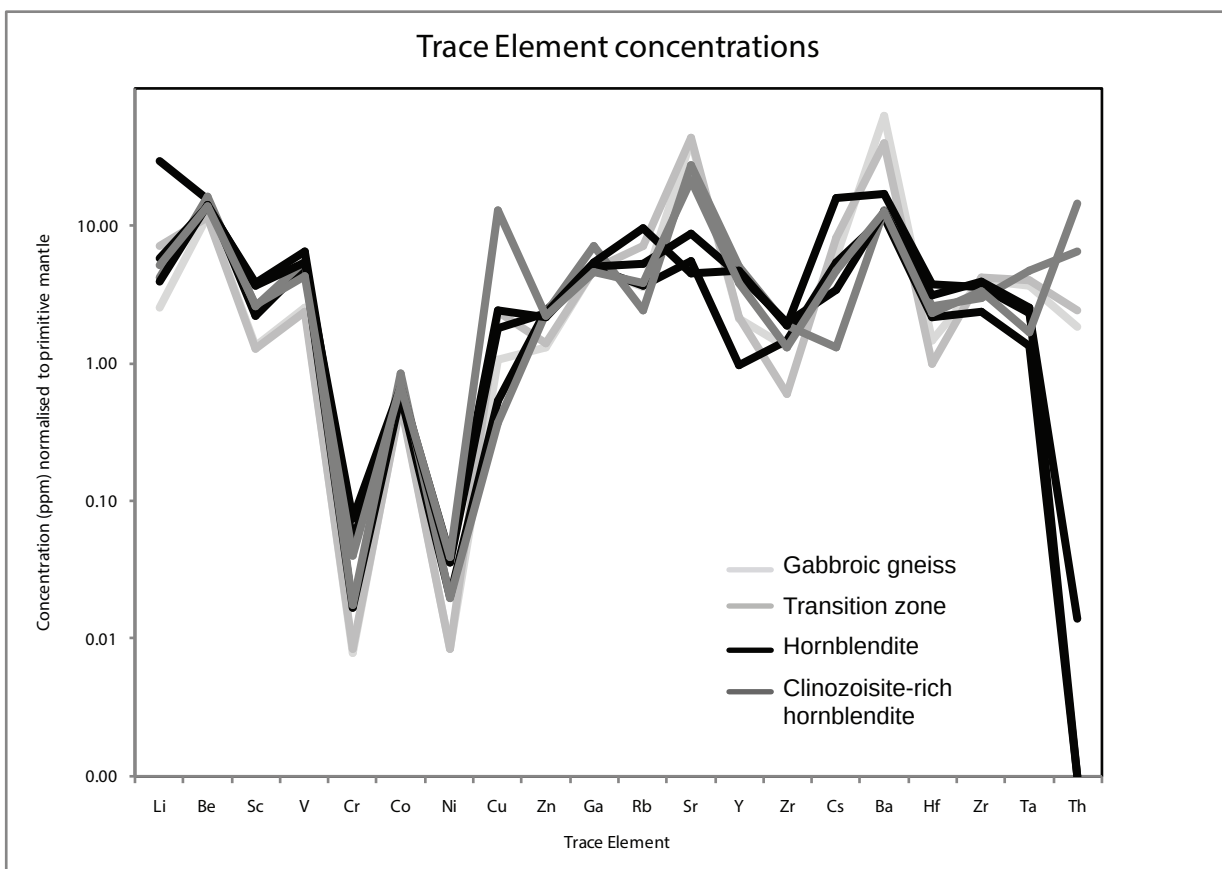


Figure 4a. Whole rock major oxides (wt. %) versus SiO<sub>2</sub> (wt. %) for gabbroic gneiss (GG), transition zone (TZ) and ultrabasic granofels (UG) separated into clinozoisite-rich (UG-cz) and hornblende (UG-hbl). 4b. Trace element concentrations normalised to primitive mantle values for selected trace elements.

content between 432.4-556.3 ppm. Ba content decreases from 421.5 ppm in gabbroic gneiss, to 262.8 ppm in transition zone, to 76.4-111.7 ppm in hornblendite and 86.0-83.3 in clinozoisite-rich hornblendite. Rb < 10 ppm for all samples. Zr does not dissolve well within solution for ICP-MS, so data is taken from XRF trace elements. Zr contents ranges 25.1-44.3 ppm for all samples.

Rare earth element concentrations normalised to C1 chondrite values from McDonough and Sun (1995) are shown in Figure 4c. Gabbroic gneiss and transition zone samples have similar bulk REE compositions, where the transition zone sample is slightly enriched in concentrations. They both show light-REE enrichment and decreasing towards the heavy-REE, with a pronounced deflection in Sm and a sharp Eu positive anomaly. Clinozoisite-rich hornblendite samples show similar REE normalized patterns to the gabbroic gneiss and transition zone, with higher concentrations. Hornblendite samples show chondrite-normalised patterns with increasing light REE values and decreasing REE values, with an inflection point at Eu.

### **Isocon analysis**

For isocon construction, zirconium (Zr) is fixed as it is considered an immobile element (Grant 2005). Enriched elements lie above the solid line and depleted elements below. There is no significant mass change in elements between the gabbroic gneiss and transition zone (Figure 4d), except for a marked decrease in Ba content. Between the transition zone and hornblendite, major oxides  $\text{Al}_2\text{O}_3$ , FeO,  $\text{SiO}_2$ , MgO, MnO,  $\text{K}_2\text{O}$  and  $\text{TiO}_2$  are enriched. Major oxides  $\text{P}_2\text{O}_5$  and trace elements Sr and Ba are depleted. Between the transition zone and clinozoisite-rich hornblendite, major oxides  $\text{Al}_2\text{O}_3$ , FeO,  $\text{SiO}_2$ , CaO, MgO,  $\text{P}_2\text{O}_5$  and  $\text{TiO}_2$  and trace element Zn are enriched. Major oxides  $\text{SiO}_2$ ,  $\text{Na}_2\text{O}$ , MnO,  $\text{K}_2\text{O}$  and trace elements Sr and Ba are depleted. Those lying close to the isocon do not represent significant mass change. The slope of the line is changed in the hornblendite and clinozoisite-hornblendite samples, and they are no longer fixed at a 1:1 ratio.

## **5. Mineral chemistry**

Mineral chemistry is characterised for the 1) host gabbroic gneiss, and 2) transition zone units, and for four domains within the 3) ultrabasic granofels unit, 3a) hornblendite, 3b) clinozoisite-rich hornblendite, 3c) pegmatoidal felsic domains and 3d) crystallised melt





pocket. A focus of this study is to characterise the major element chemistry of the minerals within 8 representative samples. Mineral geochemical analyses are presented in Table 3.

## Methods

Backscatter electron (BSE) imaging was performed on all samples using a Hitachi Benchtop scanning electron microscope (SEM) with energy dispersive spectrometer (EDS). Operating conditions were low vacuum and 15 kv accelerating voltage. BSE images were used to target analysis of major and minor element chemistry that was determined using a Cameca SX100 electron microprobe (EMP) at the Geochemical Analysis Unit (GAU) in the Australian Research Council National Key Centre for Geochemical Evolution and Metallogeny of Continents (GEMOC), at Macquarie University, North Ryde, Australia. Operating conditions were 15 kV accelerating voltage and 20 nA beam current using a 1-2  $\mu\text{m}$  focussed beam.

## Results

Plagioclase analyses are recalculated for 8 oxygens (Deer et al., 1992). The anorthite ( $\text{CaAl}_2\text{Si}_2\text{O}_8$ ) end-member content ( $X_{\text{An}}=\text{Ca}/(\text{Ca}+\text{Na}+\text{K})$ ) is calculated. Albite ( $\text{NaAlSi}_3\text{O}_8$ ;  $X_{\text{Ab}}=\text{Na}/(\text{Ca}+\text{Na}+\text{K})$ ) and orthoclase ( $\text{KAlSi}_3\text{O}_8$ ;  $X_{\text{Or}}=\text{K}/(\text{Ca}+\text{Na}+\text{K})$ ) end-member calculations were made to plot on a plagioclase classification ternary diagram. Garnet analyses are recalculated for 12 oxygens (Deer et al., 1992). The end-member proportions of pyrope ( $X_{\text{Py}}=100*\text{Mg}/(\text{Fe}+\text{Mg}+\text{Ca}+\text{Mn})$ ), almandine ( $X_{\text{Alm}}=100*\text{Fe}/(\text{Fe}+\text{Mg}+\text{Ca}+\text{Mn})$ ), grossular ( $X_{\text{Grs}}=100*\text{Ca}/(\text{Fe}+\text{Mg}+\text{Ca}+\text{Mn})$ ), and spessartine ( $X_{\text{Sp}}=100*\text{Mn}/(\text{Fe}+\text{Mg}+\text{Ca}+\text{Mn})$ ) are calculated. Clinopyroxene and orthopyroxene are recalculated for 6 oxygens to determine their end member chemistry, and plotted using the spreadsheet PX-NOM (Sturm, 2002); the classification diagram of Ca-Mg-Fe pyroxenes by Morimoto et al. (1988) is presented. Amphibole analyses are recalculated for 23 oxygens following the classification of Leake et al. (1997). Biotite and phengite analyses are recalculated for 22 oxygens, clinozoisite for 25 oxygens, chlorite for 28 oxygens, kyanite for 5 oxygens and ilmenite for 3 oxygens (Deer et al., 1992). Scapolite analyses are recalculated for 25.1-25.2 oxygens. The meionite (Me) end-member content [ $\text{Me}=100(\text{Ca}+\text{Mg}+\text{Fe}+\text{Mn}+\text{Ti})/(\text{Na}+\text{K}+\text{Ca}+\text{Mg}+\text{Fe}+\text{Mn}+\text{Ti})$ ] is calculated (Deer et al., 1992). Mg number, ( $\text{Mg\#}=(\text{Mg}/(\text{Mg}+\text{Fe}))$ ) and  $X_{\text{Fe}}=\text{Fe}/(\text{Fe}+\text{Mg})$  where reported.

Table 3. Representative mineral analyses from EMP data (GG=gabbroic gneiss, GRZ=garnet reaction zone, TZ=transition zone, UG=ultrabasic granofels, cpx = clinopyroxene, opx= orthopyroxene, prg=pargasite, pl=plagioclase, cz=clinozoisite, grt=garnet, bt=biotite, ms=muscovite).

|                                | GG            |               |              |               |              | GRZ           |               |               |              |              | TZ            |              |               |              |              | UG            |              |              |               |               |
|--------------------------------|---------------|---------------|--------------|---------------|--------------|---------------|---------------|---------------|--------------|--------------|---------------|--------------|---------------|--------------|--------------|---------------|--------------|--------------|---------------|---------------|
|                                | Cpx           | Opx           | Prg          | Pl            | Cz           | Ky            | Grt           | Cpx           | Pl           | Bt           | Ky            | Prg          | Pl            | Cz           | Ms           | Grt           | Prg          | Cz           | Pl            | Grt           |
| SiO <sub>2</sub>               | 52.35         | 50.08         | 40.81        | 60.40         | 38.42        | 37.92         | 52.96         | 52.85         | 60.84        | 37.65        | 37.44         | 41.85        | 61.44         | 37.90        | 46.58        | 37.66         | 40.71        | 37.94        | 61.62         | 38.48         |
| TiO <sub>2</sub>               | 0.22          | 0.04          | 1.19         |               | 0.08         |               | 0.16          | 0.38          |              | 4.33         |               | 0.56         |               | 0.06         | 0.72         | 0.08          | 0.86         | 0.23         |               | 0.08          |
| Al <sub>2</sub> O <sub>3</sub> | 3.52          | 2.59          | 14.07        | 25.68         | 28.20        | 62.37         | 3.02          | 7.16          | 24.27        | 14.99        | 62.42         | 15.92        | 24.47         | 27.02        | 33.86        | 20.86         | 15.85        | 27.88        | 24.43         | 21.70         |
| Fe <sub>2</sub> O <sub>3</sub> | 2.85          |               |              | 0.37          |              | 1.14          | 3.72          | 1.58          | 0.00         |              | 0.71          |              |               | 8.85         |              | 2.42          |              | 7.59         | 0.00          | 2.70          |
| FeO                            | 6.64          | 28.13         | 17.83        |               |              |               | 3.47          | 3.25          | 0.00         | 12.10        |               | 15.33        |               | 0.00         | 2.89         | 25.42         | 14.34        | 0.00         | 0.20          | 21.02         |
| MnO                            | 0.25          | 0.75          | 0.20         |               | 0.13         |               |               | 0.02          |              |              |               | 0.21         |               | 0.14         |              | 2.92          | 0.14         | 0.09         |               | 1.20          |
| MgO                            | 11.98         | 18.21         | 8.90         |               | 0.06         |               | 14.11         | 12.07         |              | 15.98        |               | 9.82         |               | 0.08         | 1.29         | 4.39          | 10.68        | 0.26         |               | 6.21          |
| CaO                            | 20.83         | 0.35          | 10.58        | 7.20          | 23.87        |               | 21.90         | 19.70         | 6.07         | 0.10         |               | 10.92        | 5.80          | 23.70        | 0.03         | 7.12          | 10.86        | 23.81        | 4.98          | 10.05         |
| Na <sub>2</sub> O              | 1.80          |               | 2.22         | 7.92          |              |               | 1.57          | 3.02          | 8.05         | 0.24         |               | 1.75         | 8.97          |              | 1.33         |               | 2.97         |              | 9.31          | 0.03          |
| K <sub>2</sub> O               |               |               | 1.06         | 0.17          |              |               |               |               | 0.28         | 9.33         |               | 0.56         | 0.07          |              | 9.34         |               | 0.45         |              | 0.06          |               |
| Cl                             |               |               | 0.02         |               |              |               |               |               |              | 0.51         |               | 0.03         |               | 0.11         |              |               | 0.06         | 0.24         |               |               |
| F                              |               |               |              |               |              |               | 0.04          |               |              | 0.09         |               |              |               |              |              |               |              |              |               |               |
| SO <sub>3</sub>                |               |               |              |               |              |               |               |               |              |              |               |              |               |              |              |               |              |              |               |               |
| <b>Total</b>                   | <b>100.15</b> | <b>100.15</b> | <b>96.88</b> | <b>101.37</b> | <b>97.79</b> | <b>101.32</b> | <b>100.58</b> | <b>100.03</b> | <b>99.51</b> | <b>95.32</b> | <b>100.50</b> | <b>96.95</b> | <b>100.75</b> | <b>96.97</b> | <b>96.04</b> | <b>100.63</b> | <b>96.92</b> | <b>97.28</b> | <b>100.60</b> | <b>101.20</b> |
| <i>XAn</i>                     |               |               |              | 0.33          |              |               |               |               | 0.29         |              |               |              | 0.26          |              |              |               |              |              | 0.23          |               |
| <i>XAb</i>                     |               |               |              | 0.66          |              |               |               |               | 0.69         |              |               |              | 0.73          |              |              |               |              |              | 0.77          |               |
| <i>XOr</i>                     |               |               |              | 0.01          |              |               |               |               | 0.02         |              |               |              | 0.00          |              |              |               |              |              | 0.00          |               |
| <i>XPyr</i>                    |               |               |              |               |              |               | 0.31          |               |              |              |               |              |               |              |              | 0.17          |              |              |               | 0.46          |
| <i>XAlm</i>                    |               |               |              |               |              |               | 0.51          |               |              |              |               |              |               |              |              | 0.56          |              |              |               | 0.24          |
| <i>XGrs+</i>                   |               |               |              |               |              |               | 0.18          |               |              |              |               |              |               |              |              | 0.27          |              |              |               | 0.31          |
| <i>sps</i>                     |               |               |              |               |              |               |               |               |              |              |               |              |               |              |              |               |              |              |               |               |
| <i>Mg#</i>                     |               |               | 0.47         |               |              |               |               |               |              |              |               | 0.53         |               |              |              |               | 0.57         |              |               |               |
| <i>XFe</i>                     |               |               |              |               |              |               |               |               |              |              |               |              |               |              |              |               |              |              |               |               |

| Melt          |               |              |              |              | Peg           |              |
|---------------|---------------|--------------|--------------|--------------|---------------|--------------|
| Pl            | Grt           | Prg          | Bt           | Chl          | Pl            | Scp          |
| 63.06         | 38.49         | 41.74        | 35.65        | 26.69        | 63.09         | 48.66        |
|               | 0.09          | 0.75         | 1.82         | 0.10         |               |              |
| 23.80         | 21.73         | 15.75        | 17.87        | 22.63        | 24.03         | 23.55        |
|               | 2.47          |              | 0.00         |              |               | 0.00         |
| 0.17          | 20.70         | 13.95        | 14.59        | 17.15        |               | 0.12         |
|               | 0.90          | 0.14         | 0.08         | 0.13         |               |              |
|               | 6.18          | 10.91        | 14.88        | 21.31        |               |              |
| 4.68          | 10.52         | 10.03        | 0.15         |              | 4.86          | 14.66        |
| 9.51          | 0.03          | 2.96         | 0.22         |              | 9.38          | 5.58         |
|               |               | 0.65         | 8.36         |              | 0.05          |              |
|               |               | 0.06         | 0.03         |              |               | 0.15         |
|               |               |              |              |              |               | 3.61         |
| <b>101.22</b> | <b>100.86</b> | <b>96.94</b> | <b>93.65</b> | <b>88.01</b> | <b>101.41</b> | <b>96.33</b> |
| 0.21          |               |              |              |              |               | 0.22         |
| 0.79          |               |              |              |              |               | 0.78         |
| 0.00          |               |              |              |              |               | 0.00         |
|               | 0.45          |              |              |              |               |              |
|               | 0.24          |              |              |              |               |              |
|               | 0.31          |              |              |              |               |              |
|               |               | 0.58         |              |              |               |              |

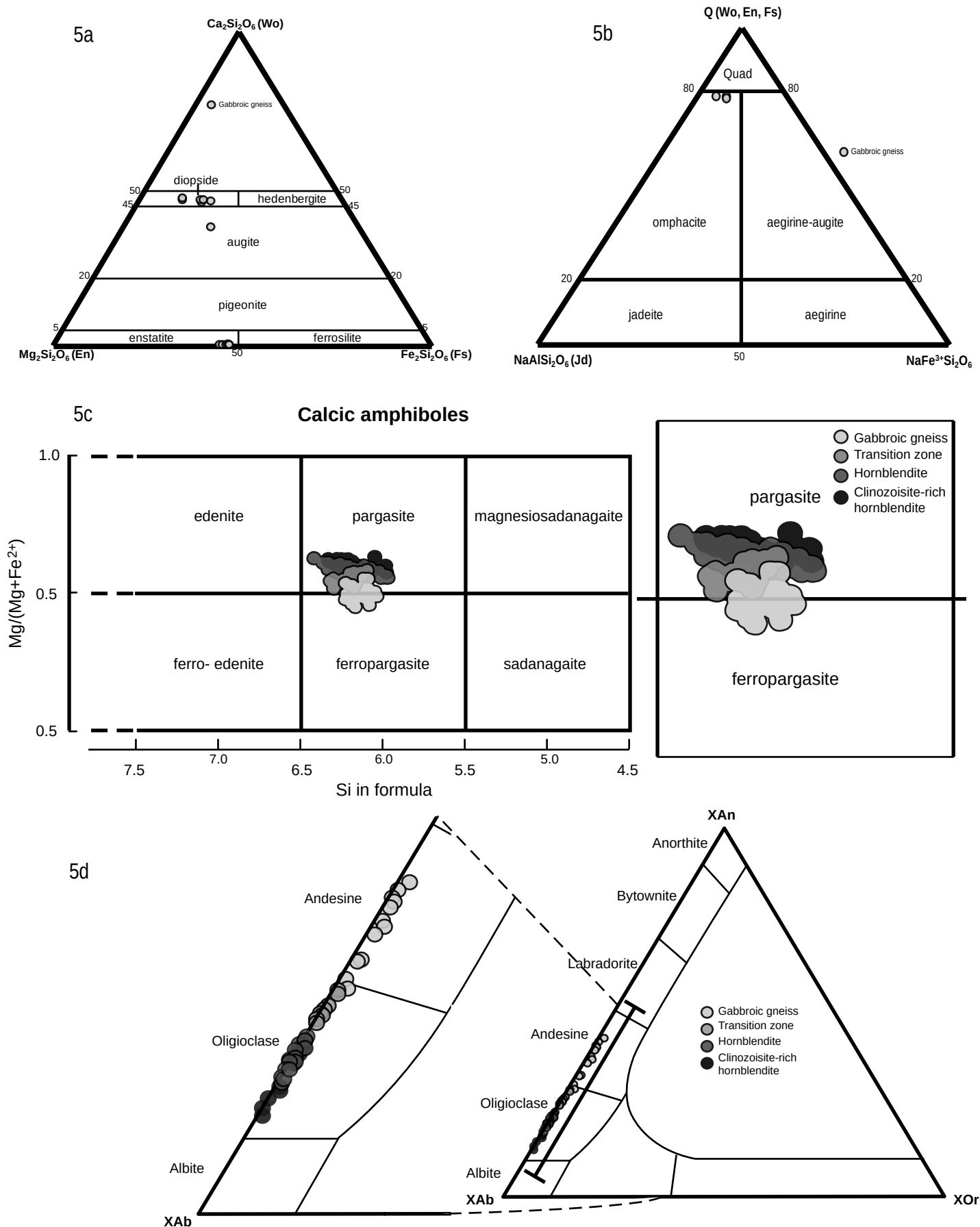


Pyroxene mineral chemistry is presented for samples of the host gabbroic gneiss (Figure 5a,b). Clinopyroxene grains are diopside and classify as aluminan sodian diopside and less commonly aluminian ferrian sodian or aluminian ferrian diopside. One grain classifies as aluminan sodian augite. Three grains classify as omphacite.  $X_{Fe}$  ranges 0.18-0.37. Orthopyroxene grains are aluminan enstatite, and has  $X_{Fe} = 0.44-0.47$ .

Amphibole is common to all units of this study, and all are calcic amphiboles (Leake et al., 1997) and is presented in Figure 5c. Within the host gabbroic gneiss, amphibole is predominantly associated with inner coronae amphibole + quartz symplectite clusters surrounding the two-pyroxene assemblages as well as outer coronae amphibole. Using their Mg# (0.45-0.54) versus Si (cations), samples range from ferropargasite to pargasite (Figure 5c). Within the transition zone, the two-pyroxene mineralogy is absent, and the mode of pargasite has increased (Table 3). The grains in the transition zone are also pargasite with slightly higher Mg# = 0.51-0.57 (Table 3, Figure 5c). The amphibole within the ultrabasic granofels is also pargasite in both the hornblende and clinozoisite-rich hornblende units (Mg#=0.55-0.62 and 0.56-0.62, respectively; Figure 5c, Table 3). Pargasite within hornblende-clinozoisite unit has a slightly increased median (Mg# 0.60) than hornblende samples (Mg# 0.58), accounting for the overall trend between the two rock types (Figure 5c).

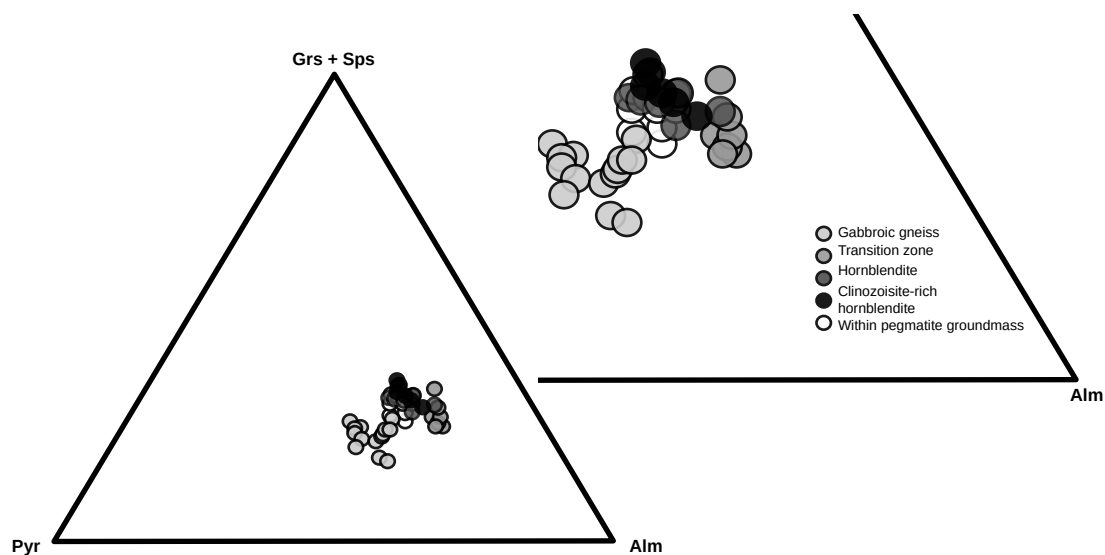
Plagioclase mineral chemistry of the host gabbroic gneiss straddles the andesine-oligoclase boundary ( $An_{29}-An_{43}$ ; Figure 5d), while plagioclase in the transition zone is oligoclase ( $An_{25}-An_{28}$ ). Rare plagioclase in the ultrabasic granofels is oligoclase ( $An_{17-23}$  for hornblende and  $An_{13-21}$  clinozoisite-rich hornblende). Plagioclase in all samples is not zoned. Plagioclase grains within the pegmatoidal felsic domains have a similar composition to plagioclase in the ultrabasic granofels ( $An_{16}-An_{24}$ ; Table 3). Intergrown with the plagioclase within these domains is scapolite (Figure 3.4c). The scapolite is enriched in  $SO_2$ , ranging between 3.3-4.3 wt. % (Table 3). Scapolite analyses Me content between  $Me_{58}-Me_{64}$ , classifying as intermediate seapolite (mizzonite) (Table 3).

Garnet from the host gneiss is within a GRZ, and is almandine-pyrope with the composition  $Alm_{40-51}Pyr_{26-36}Grs_{17-26}Sps_{0-1}$  (Figure 5e). Garnet is typically intergrown with clinopyroxene or quartz. Garnet located within transition zone is neoblastic and is unzoned with the composition  $Alm_{52-58}Pyr_{15-19}Grs_{17-27}Sps_{6-9}$  (Figure 5e). Garnet data from ultrabasic granofels units are taken from garnet trains existing within the units and range between

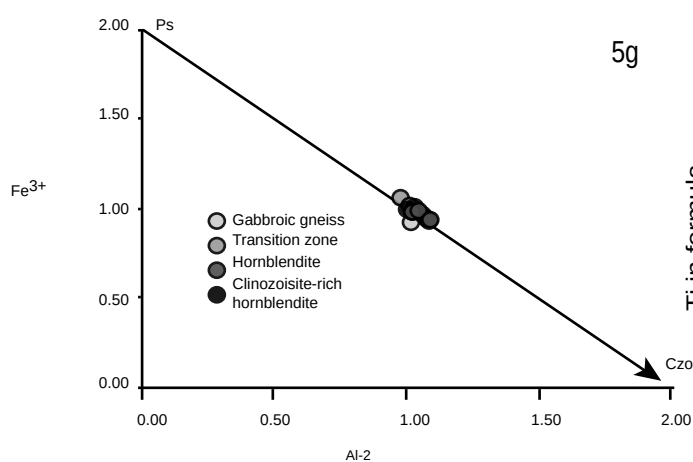


Figures 5a.,b Clinopyroxene and orthopyroxene mineral data from the gabbroic gneiss plot using PX-NOM. 5c. Calcic amphiboles for all samples from Leake et al. 1997. 5d Classification plot for plagioclase (An = anorthite, Ab = albite, Or = orthoclase) ternary for all samples.

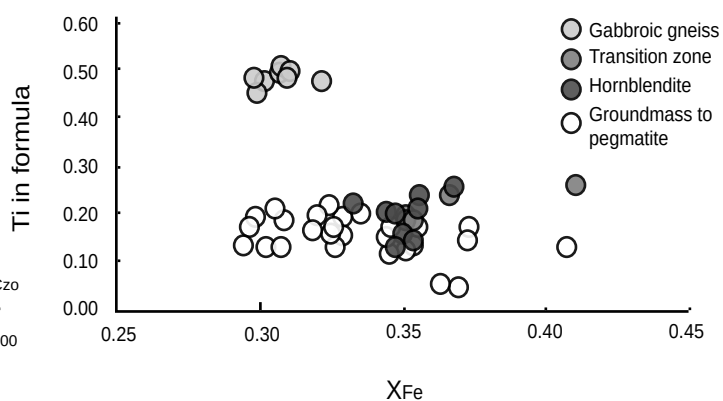
5e



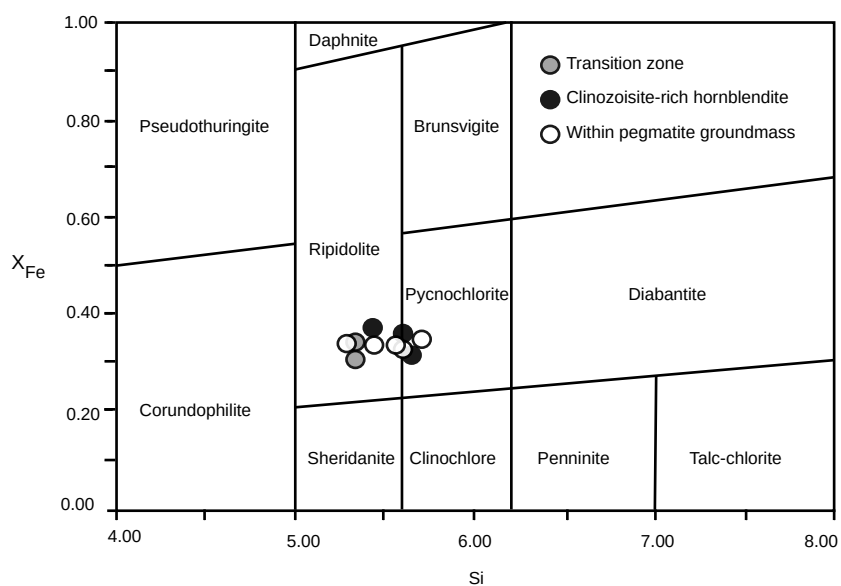
5f



5g



5h



Figures 5e. Garnet ternary classification (pyr= pyrope, alm = almandine, grs = grossular and sps = spessartine) for all samples. 5f. Clinozoisite classification (ps = pistacite, czo = clinozoisite). 5g. Biotite XFe versus Ti in formula for all samples. 5h. Chlorite classification diagram for samples containing chlorite.



Alm<sub>45-55</sub>Pyr<sub>17-24</sub>Grs<sub>19-29</sub>Sps<sub>2-5</sub> for hornblendites and Alm<sub>44-51</sub>Pyr<sub>20-23</sub>Grs<sub>27-32</sub>Sps<sub>2-3</sub> for clinozoisite-rich hornblendites (Figure 5e). Large garnet grains within interstitial petgamtoidal felsic domains have a similar composition to the garnet trains of the ultrabasic granofels, Alm<sub>44-50</sub>Pyr<sub>20-26</sub>Grs<sub>23-31</sub>Sps<sub>2-3</sub> (Table 2). Overall, there are 3 domains of garnet compositions, GRZ that are more pyrope (Mg) enriched, transition zone neoblasts that are more almandine (Fe) enriched and ultrabasic granofels variations of garnet where compositions are indistinguishable (Figure 5e). No symmetric or asymmetric chemical zoning was observed in any samples.

Clinozoisite shows moderate substitution of Al by Fe<sup>3+</sup> with Fe<sup>3+</sup> cations ranging 0.84-1.12 for all samples (for 25 oxygens). Host gabbro samples range between Fe<sup>3+</sup> = 0.86-0.91, transition zone Fe<sup>3+</sup> = 0.86-1.12 and clinozoisite-rich hornblendites Fe<sup>3+</sup> = 0.89-1.09 (Figure 5f). As a result, there is little to no variation within or between samples in terms of clinozoisite grain chemical composition.

Biotite (Figure 5g) occurs in a range of textural settings within the three rock types. Within the host gabbroic gneiss, biotite is located within GRZ, surrounded by garnet, with irregular crystal faces. Analyses have X<sub>Fe</sub> = 0.30-0.32 and high Ti contents, 0.46-0.50 cations per formula unit (22 oxygens). Cl and F of this type of biotite range Cl = 0.50-0.56 and F <0.09 wt. % (Table 3). Biotite grains within the transition zone are found as euhedral, tabular grains within pargasite-quartz symplectites. They have no apparent orientation or mineral preference. X<sub>Fe</sub> = 0.35-0.41 is higher and Ti content (Ti = 0.17-0.26) is lower than biotite within the GRZ. Cl content is significantly lower (Cl = 0.03-0.05 wt. %) and F is below detection limit (Table XX). Biotite within the ultrabasic granofels are found within singular plagioclase grains, or in rarer cases, as interstitial grains within pargasite or pargasite-clinozoisite. X<sub>Fe</sub> = 0.33-0.37, Ti = 0.15-0.25 cations and Cl = 0.03-0.08 wt. % for biotite associated within plagioclase grains, where F is also below detection limit (Table 3). Muscovite is present within the transition zone (Table 3).

Kyanite is present within the host gabbroic gneiss plagioclase-dominant matrix with clinozoisite, and has 0.7-1.3 wt. % Fe<sub>2</sub>O<sub>3</sub> (Table 3). Kyanite is not present within the transition zone. There is also chlorite within the plagioclase matrix, which classifies as ripidolite (Figure 2). Significant proportions of chlorite were present in close association to interstitial pegmatoidal felsic domains (Figure 3.5a, b). The chemical composition of the chlorite found in these pockets ranges between ripidolite to pynochlorite (Figure 5h).

## 6. Thermobarometry

P-T estimates were obtained by applying a combination of thermobarometric techniques using the mineral assemblages of the samples. The results are summarised in Table 3. Calculations for garnet-clinopyroxene temperature estimates follow Ellis and Green (1979), Powell (1985) and Krogh (1988). Adjacent clinopyroxene-garnet pairs from the gabbroic gneiss garnet reaction zone (GRZ) indicate temperatures of ~650-700 °C, assuming  $P=14$  kbar, following estimates from previous GRZ estimates from Clarke et al. (2000). For  $T=600$  °C, compositions of adjacent grains of garnet, clinopyroxene and plagioclase give  $P=16-18$ .

P-T estimates using paired amphibole-plagioclase grains within the ultrabasic granofels units are presented in Table 3. Temperature estimates by Holland and Blundy (1994) from their HB2 calibration reaction edenite + albite = richterite + anorthite. Pressures determined using the method of Anderson and Smith (1995). Hornblendite estimates range  $P=9.4-9.7$  kbar for  $T=710-720$  °C and clinozoisite-rich hornblendite samples range  $P=8.9-9.6$  kbar and  $T=697-714$  °C. Temperature estimates based on Ti in amphibole following the method of Otten (1984) give estimates of 645-652°C (hornblendite) and 666-668°C (clinozoisite-rich hornblendite) (Table 3).

## 7. Discussion

The studied rocks in Fiordland, New Zealand provide an opportunity to examine processes occurring within a deeply exposed Phanerozoic arc root. Most rock units (excluding the ultrabasic granofels) within the Pembroke Granulite have been extensively studied, providing insight into the deformational and metamorphic processes occurring within the lower crust (Clarke et al., 2000, Daczko et al., 2001a, Daczko et al., 2001b, Daczko et al., 2002, Schröter et al., 2004, Clarke et al., 2005, Stevenson et al., 2005, Daczko and Halpin, 2009, Allibone et al., 2009a, Smith et al., 2015). However, the field relationships and characteristics of the ultrabasic granofels has not been reported and discussed.

**Table 3. Thermobarometry calculations using methods 1. Ellis and Green (1979), 2. Powell (1985), 3. Krogh (1988), 4. Eckert et al. (1991) 5. Newton and Perkins (1982), 6. Holland and Blundy (1994), 7. Anderson and Smith (1995) (grt = garnet, pl = plagioclase, cpx=clinopyroxene, prg = pargasite, qz = quartz, czo=clinozoisite, ky = kyanite, bt = biotite).**

| Rock                           | Assemblage           | Assumed P (kbar) | Calculated T (°C)  | Calculated P (Kbar) | Method  |
|--------------------------------|----------------------|------------------|--------------------|---------------------|---------|
| GRZ in gabbroic gneiss         | grt-cpx-qz -bt (GRZ) | 14               | 694, 674, 651      |                     | 1, 2, 3 |
|                                | pl-czo-ky (matrix)   | 14               | 598, 573, 521      |                     | 1, 2, 3 |
| Hornblendite                   | prg-pl               |                  | 710, 719, 720      | 18.04, 16.1         | 4, 5    |
|                                |                      |                  |                    |                     | 6       |
| Clinozoisite-rich hornblendite | prg-czo-pl           |                  | 697, 701, 712, 714 | 9, 10, 9            | 7       |
|                                |                      |                  |                    |                     | 6       |
|                                |                      |                  |                    | 9, 9, 10, 9         | 7       |

### **The origin of the ultrabasic granofels: Igneous cumulate or reaction product from metasomatism**

In principle, ultrabasic rocks are commonly interpreted as igneous cumulates. However, field relationships reported here are not consistent with this interpretation. In the following, the possibility of an igneous versus reaction product origin of the ultrabasic rock is discussed.

Igneous cumulate rocks formed by fractionating parental magmas are commonly ultramafic to mafic, or ultrabasic to basic in composition. Amphibole grains in cumulate rocks are thought to have been derived from one of two mechanisms, (i) primary crystallisation and accumulation from a magma; and (ii) production by reaction between early-forming phases (e.g. olivine) that are consumed in the presence of hydrous fluids or melts (Larocque and Canil, 2010, Daczko et al., 2012).

The ultrabasic granofels units' mineralogy, predominantly comprised of pargasite amphibole  $\pm$  clinozoisite with rare plagioclase is consistent with an igneous cumulate origin. Furthermore, whole-rock trace element data for ultrabasic granofels samples show a hump-shaped middle rare earth element (REE) pattern (Figure 4c), with an increase from La to Nd-Sm, characteristic of amphibole REE partitioning signatures (Bottazzi et al., 1999, Davidson et al., 2007). This is coupled with relatively high Cr content and low Rb content, also consistent with a cumulate origin (Daczko et al., 2012). Clinozoisite, a Ca-bearing rock-forming mineral, preferentially incorporates light rare earth elements (Bottazzi



et al., 1999), and clinozoisite-rich hornblendite samples reflect this with their enriched light REE patterns, again consistent with an igneous origin. Furthermore, heavy-REE depletion patterns common to all samples is consistent with minerals crystallising from a heavy-REE depleted magma. This supports the interpretation that the rocks formed by igneous/cumulate processes from a parental magma similar to the Western Fiordland Orthogneiss (Allibone et al., 2009a, Daczko et al., 2012).

Petrological observations show that pargasite and clinozoisite crystallised in equilibrium, as they exhibit no overprinting or recrystallisation textures (Figure 3.3c). Grains in contact, show no evidence of reaction, and pargasite inclusions within clinozoisite indicate they crystallised together. The chemistry of the inclusions is the same composition as seen throughout the ultrabasic granulites, indicating chemical equilibrium. This further supports an igneous origin where phases are in equilibrium.

However, field relationships indicate the ultrabasic granulites unit formed in the late stages of the metamorphic history of the Pembroke Granulite, as its boundaries cross-cut previously mapped regional D<sub>4</sub> shear zones (Figure 2.1c) (Daczko et al., 2001b). As such, it is difficult to explain the presence of the ultrabasic granulites as a cumulate. Hence, the possibility of producing the ultrabasic granulites by metasomatism and fluid- or melt-rock interactions is explored in the following.

Sharp field contacts (Figure 2.3a,b) of the order of one grain diameter, show a marked decrease in plagioclase (45 vol.% to <5% vol.%) and quartz (5 vol.% to <0.01 vol.%) from the transition zone into the ultrabasic granulites. This is coincident with a bulk rock chemical change of between 9.9-12.3 wt.% SiO<sub>2</sub>. This could be consistent with a cumulate origin forming ultrabasic cumulate layers or 'pods' of accumulation. However, the structural nature of this contact follows an irregular pattern, which protrude along pre-existing garnet granulite reaction zones (GRZ) (Figure 2.3 a,b), which is interpreted to be formed as a reaction between a metasomatising agent and the host gabbroic gneiss. GRZ have been mapped from the gabbroic gneiss, into the ultrabasic granulites and are significantly modified from their original texture (garnet + two-pyroxene + pargasite + quartz), reported by Blattner (1976), Daczko et al. (2001a) and Smith et al. (2015), to garnet ± pargasite in the ultrabasic granulites (Figure 3.3a,b,c). Due to their fine grain size and structural weakness, it is inferred they acted as faster fluid conduits to the metasomatising agent.

The transition zone spans a width of 2-7 m, and surrounds the ultrabasic granofels (Figure 2.2a). It comprises pargasite + quartz mafic clusters, where the original two-pyroxene assemblage is no longer present and is interpreted to be due to 'hydration' from a metasomatising agent. Large, porphyroblastic garnets surrounded by leucosome are common to the transition zone, and are interpreted to be peritectic grains that formed during fluid- or melt-fluxing (Figure 2.4a). GRZ are modified to garnet surrounded by leucosome, pargasite, or both, often with small (cm-sized) nucleations of new garnet, a reaction product of the fluxing fluid or melt.

Petrological observations within the ultrabasic granofels are consistent with fluid or melt-rock interactions due to fluxing. The low dihedral angles ( $<10^\circ$ ) of plagioclase against pargasite and clinozoisite grains (Figure 3.3a,b,c) are textures common to melt-filled pores that are later pseudomorphed by minerals (Holness and Sawyer, 2008). Plagioclase-dominated pockets with biotite, garnet and amphibole are interpreted to have nucleated from the fluid or melt agent, and are inferred to represent trapped and crystallised pockets of the metasomatising agent (Holness and Sawyer, 2008).

Similarly, coarse-grained pegmatoidal domains with coarse grains of garnet, pargasite and plagioclase (Figures 2.8a,b,c), are inferred to represent crystallised parts of the final melt in the system. The presence of high-sulphur scapolite with plagioclase in domains of finer-grained pegmatoidal areas (groundmass to the coarse-grained felsic domains) strongly indicates a metasomatic environment (Stolz, 1987). This end-member of scapolite is thought to be formed by a reaction with a relatively  $\text{CO}_2$ -rich fluid phase, calcic plagioclase and accessible sulphides (Stolz, 1987), which points to a metasomatic origin for the ultrabasic granofels.

As previously discussed, whole-rock REE patterns display signatures expected from amphibole crystallised within igneous cumulates, with light-REE enrichment for clinozoisite-rich samples due to preferential uptake of those elements in clinozoisite. However, the level of enrichment within clinozoisite-rich hornblendite samples is significant, suggesting a metasomatic source of light-REE (Downes, 2001). This is supported by the isocon analysis using whole-rock chemistry for major and trace elements. The slope of the "constant mass" line shifts for ultrabasic granofels samples and this

indicates a mass increase (Grant, 1986) (Figure 4d). As a result, a volume increase is inferred to be likely from a fluid or melt flux through the system.

Whole rock and mineral chemistry, as well as the ultrabasic mineral assemblage of the rock points to an igneous origin. In addition, on the basis of field relationships, and mineral textural relationships, the ultrabasic granofels unit situated within the Pembroke Granulite is interpreted to be emplaced via fluid or melt-rock interactions with the garnet reaction zone bearing host gabbroic gneiss.

### **P-T conditions of GRZ formation and ultrabasic rock formation**

Pressure and temperature estimates for the Pembroke Granulite have been previously reported to be  $T = 662\text{--}699$  at  $P = 14.7\text{--}16.6$  (Clarke et al., 2000). The garnet reaction zone (GRZ) mineral assemblage (garnet + cpx + quartz  $\pm$  biotite with closely associated plagioclase and kyanite) are used to estimate P-T conditions of GRZ within the weakly modified gabbroic gneiss of this study. Garnet-clinopyroxene pairs gave estimates of  $T = 598\text{--}694$  °C from the methods of Ellis and Green (1979) and assuming 14 kbar pressure (using estimates previously made for the Pembroke Granulite by Clarke et al. 2000). The textures within the GRZ from this study are similar to those reported as "domain 3" by Smith et al. (2015), including the addition of biotite. These conditions are lower temperature compared to previously published data on GRZ formation of  $P = 12\text{--}16$  kbar at  $T = 720\text{--}890$  °C in the Western Fiordland Orthogneiss (Daczko and Halpin, 2009).

Within the ultrabasic granofels, local pockets of amphibole-plagioclase are used to apply paired thermobarometry from Holland and Blundy (1994) (temperature) and Anderson and Smith (1995) (pressure). The results show equilibration within the ultrabasic granofels occurred at  $T=696\text{--}720$  and  $P=9\text{--}10$  kbar (Table 3). These conditions are interpreted to represent the  $P$ - $T$  conditions at which the ultrabasic unit formed.

### **How to make an ultrabasic granofels by melt-rock interaction**

Chemically and petrographically, the ultrabasic granofels unit can be argued as a typical, igneous fractionation and accumulation origin, as well as a reaction product from fluid or melt-rock interactions. The second hypothesis, one that is being argued for here, is based on a combination of field relationships, atypical petrography and some geochemical

signatures. This section explores how melt-rock interaction may produce a rock that has textural and chemical characteristics much like an igneous cumulate within the lower crust. At the same time, if melt-rock interaction can produce a “phantom” cumulative, igneous rock, this study suggests that other ultrabasic rocks may have a similar origin.

A study by (Kelemen et al., 1992) investigated the conversion of lherzolite into harzburgite through melt-rock reactions within the upper mantle. They proposed this as an alternative mechanism for creating harzburgite samples that were too enriched in SiO<sub>2</sub> (orthopyroxene) and had too high light to heavy REE ratios to be residues of partial melts derived from primitive mantle. They report similarly sharp outcrop boundaries to those described here, where abrupt mineralogical changes from dunite (clinopyroxene-rich and no orthopyroxene) to harzburgite (>15 vol. % orthopyroxene and <2 vol.% clinopyroxene). This is analogous to the modification reported here from gabbroic gneiss (45 vol.% plagioclase, 5 vol.% quartz to and 15 vol.% pargasite) to ultrabasic granofels (<5 vol. % plagioclase, no quartz and 60-90 vol. % pargasite). The abrupt contact boundaries are inconsistent with mechanisms involving advective fluxing of migrating melts or advective heating of host rock by melt. The expected outcome from these mechanisms would produce a gradational change in the proportions of the depleted mineral as a function of distance. Instead, they infer this is consistent with phase saturation in a cooling, migrating magma.

Kelemen et al. (1992) inferred a reaction where clinopyroxene was dissolved and a larger mass of orthopyroxene + olivine was produced, in fixed proportions from an orthopyroxene-saturated liquid. The saturation process begins with a liquid saturated only with olivine interacting with peridotite to form dunite, producing increasingly orthopyroxene-rich liquid compositions. Between certain temperature and pressure conditions, the liquid will produce harzburgite by reaction with lherzolite indefinitely, and upon cooling, the liquid will evolve to clinopyroxene-saturation (Kelemen et al., 1992). Using this as a template for understanding the ultrabasic granofels system, the liquid (melt) is pargasite + clinozoisite-saturated which migrates through the host gabbroic gneiss. The melt-rock reaction dissolves plagioclase + quartz and precipitates pargasite in fixed proportions relative to clinozoisite, with a total mass greater than the mass of the solid reactants.

Naney (1983) performed experiments on stability relationships of common rock-forming ferromagnesian silicate phases within model granite and granodioritic systems.



They were investigated as a function of temperature with variable water content at 2 and 8 kbar pressure. Hornblende (pargasite in this study) was only observed in granodioritic systems, and epidote (clinozoisite in this study) was only observed in experiments of 8 kbar. The mineralogy, pargasite + clinozoisite + biotite + plagioclase + liquid was shown to be stable between  $T \sim 675-720$  °C and 8-11 wt. %  $H_2O$  at 8 kbar. These results are used to estimate the stability range of the ultrabasic granofels system, which has formed from a precursor igneous gabbroic-gabbronotite composition. Figure 7.1, modified after Naney (1983), highlights the stability range of co-existing pargasite + clinozoisite. This is coupled with pargasite-plagioclase paired thermobarometry calculations from the ultrabasic granofels which has  $T=696-720$  °C and 9-10 kbar pressure (Figure 7.1). This suggests that fluxing a hydrous, siliceous melt through a rock system within these P-T constraints, will allow crystallisation of ferromagnesian phases such as pargasite and clinozoisite while simultaneously dissolving the other phases in the host gabbroic gneiss.

Zen and Hammarstrom (1984) investigated co-existing magmatic epidote and hornblende following the experiments of Naney (1983), using natural epidote-bearing plutonic samples. Using microprobe data, they constructed the reaction where hornblende + plagioclase + epidote + biotite would be solid components and alkali feldspar, magnetite and quartz would be wholly or partially in the melt phase. Their results are coincident with Naney (1983). In comparison with the ultrabasic granofels system analysed here, pargasite and clinozoisite may have crystallized from the melt, but plagioclase remained in the melt, potentially being fluxed out of the system. This would suggest that the fluxing melt producing the pargasite-clinozoisite would not have to exhibit extreme chemical composition but instead could be a chemically “normal” granodiorite.

A simplified model for the formation of the ultrabasic granofels is presented in Figure 7.2. The host rock, i.e. the Pembroke Granulite, is characterized by a distinct  $S_1$  foliation, presence of three distinct sets of GRZ and some thrust faulting (Stage 1). Following this, initial ductile localized shearing creates local permeability in the rock, allowing melt ingress into the gabbroic gneiss  $\pm$  GRZ and continuous melt fluxing (Stage 2). The localised strain is documented in the field by the presence of GRZ deflection on either edge of the high strain zone, with the GRZ in the host remaining in low strain conditions. As the pargasite + clinozoisite-saturated melt percolates into the system, the process of dissolving plagioclase and quartz and precipitating pargasite + clinozoisite begins. This results in the formation of a pargasite-clinozoisite rock while plagioclase and

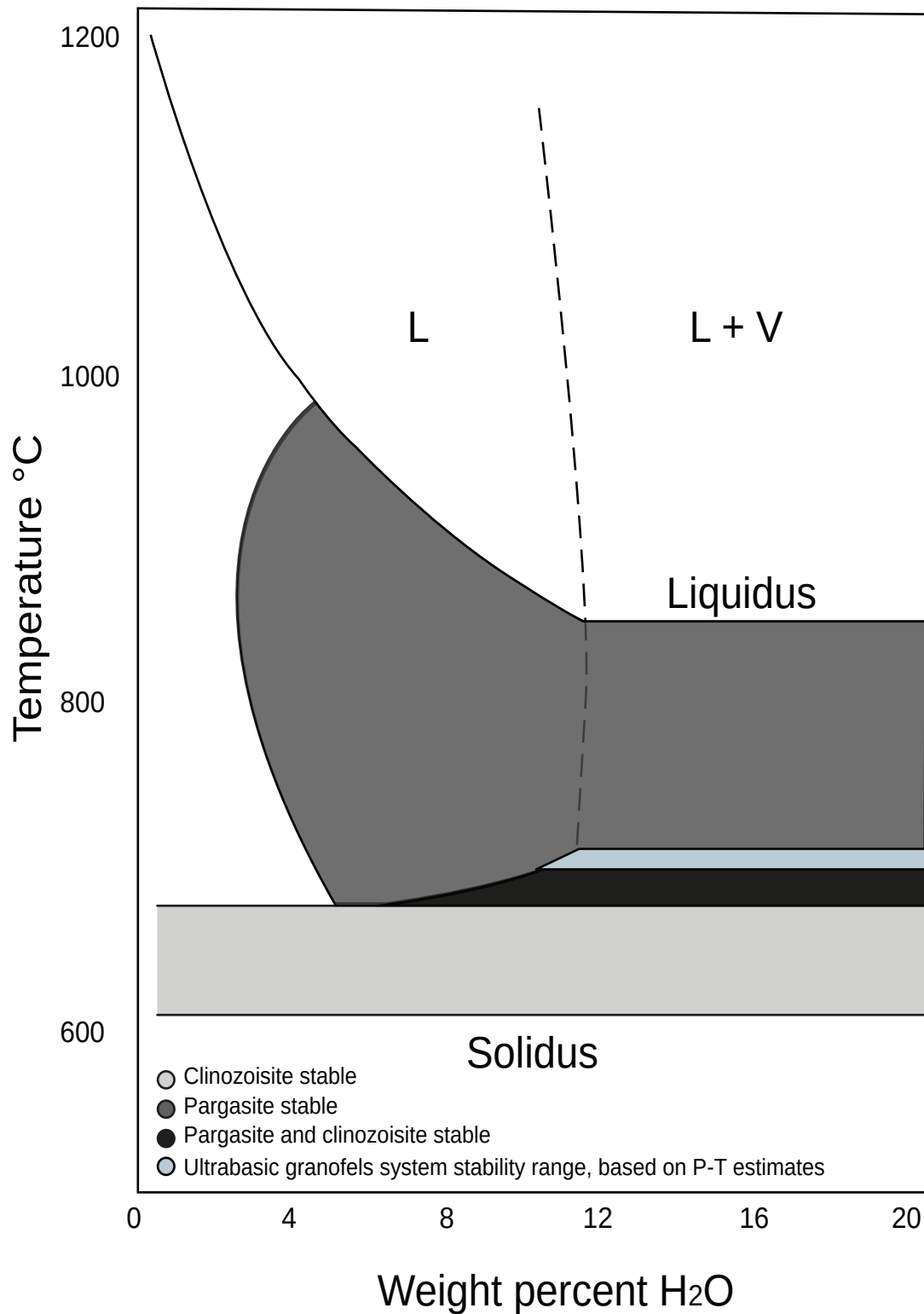


Figure 7.1. Using experimental data after Naney (1983), the stability ranges for pargasite, clinozoisite and pargasite + clinozoisite were determined for 8 kbar. From P-T data from this study, an estimated stability range for the ultrabasic granofels system is indicated.

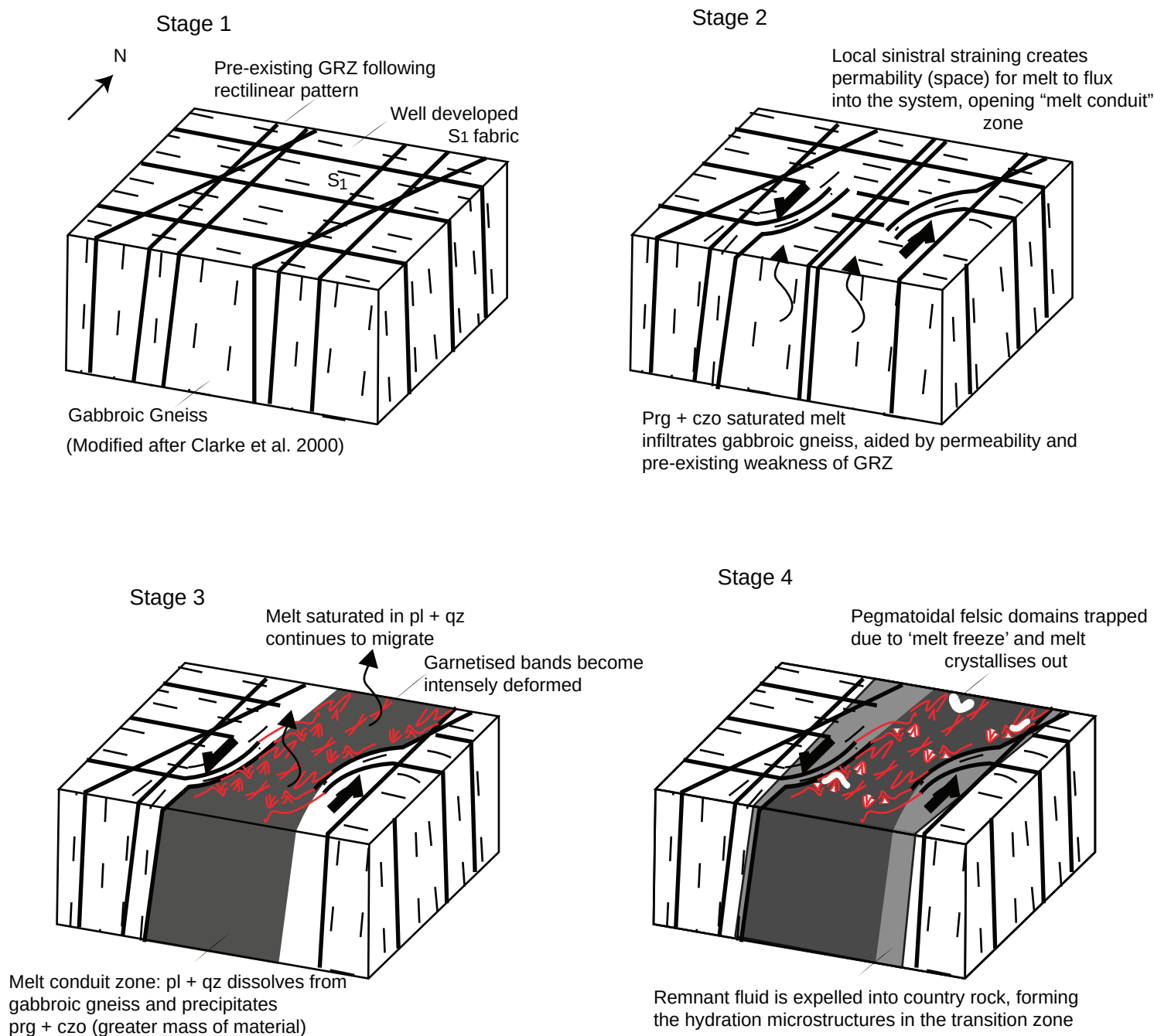


Figure 7.2. A simplified model of the ultrabasic granofels system.

quartz become increasingly rare. Continued shearing as well as transient melt transport (Bons et al., 2008), causes intense deformation of GRZ (dismembered garnetite bands, sheath folds (Figure 2.7a,b,c), collapse type structures (Figure 2.7c), as well as creating more permeability for melt to infiltrate (Stage 3). The reaction continues until the fluxing rate of melt is insufficient to maintain sufficiently high temperatures in the melt “conduit” zone. Hence, the system cools (Stage 4) and the melt “freezes” in remnant pockets. This results in the crystallisation of plagioclase-bearing melt pockets with low dihedral angles. As a result of further crystallization of the now trapped melt, fluid-rich pegmatoidal material is formed with large garnet, pargasite and plagioclase (Figure 2.2a). At final crystallization, remnant fluid is expelled into the country rock, producing the microstructures seen in the transition zone. These include, reaction induced modification of GRZs (Figure 2.4b), hydration of the rock resulting in very local partial melting recognized by the growth of large peritectic, porphyroblastic garnet surrounded locally by plagioclase-rich leucosome (Figure 2.4a).

### **Ultrabasic rocks as tracers of deep crustal melt fluxing**

The model presented here for the formation of the ultrabasic rocks outcropping in the deep crustal section of the Pembroke Valley, New Zealand, has several implications for the possible interpretation and use of “phantom” igneous rocks observed in such deep crustal sections. If the model is correct, then such ultrabasic rocks may be one of the very few tracers of melt transport in the lower crust. These results possibly go a long way to resolving the cryptic pathways of melt migration from the deep to shallow crust. The ultrabasic granofels unit characterised in this study and possibly other amphibole-rich units globally would represent melt “conduits” similar to “conduits” known from the mantle (Kelemen et al. 1992) but not reported in the deep crust. Further detailed studies on similar rocks in other deep crustal sections are needed to confirm if such a scenario is unusual or potentially more common than previously recognized.

## **8. Conclusions**

This study characterised an ultrabasic granofels unit in terms of structural field and petrographical relationships, whole-rock and mineral geochemistry and pressure and temperature conditions. From this information the following conclusions were made:



- 1) The ultrabasic granofels unit is a reaction product from melt-rock interaction, and not formed by igneous cumulate processes. The reaction consists of a pargasite + clinozoisite-saturated melt interacting with gabbroic gneiss ( $\pm$  GRZ). This creates a melt "conduit" zone which dissolves plagioclase and quartz and precipitates pargasite  $\pm$  clinozoisite.
- 2) Melt percolation was enhanced from localised ductile shearing and pre-existing GRZ acting as fluid conduits. The reaction continues until the melt fluxing rate is insufficient to maintain sufficiently temperatures in the melt "conduit" zone, causing the system to cool. This causes the melt to "freeze" and crystallise, trapping the melt in remnant pore space.
- 3) Final remnants of melt expelled into the system causes late-stage features such as pegmatoidal felsic domains and the formation of the transition zone.
- 4) Despite this model, some aspects of the rock appears to behave as a "phantom" igneous cumulate. Implications from this study of melt-rock interactions in the deep crust and "phantom" expressions of cumulate ultrabasic rocks require further detailed study of similar outcrops.

## References

- Allibone, A. H., Jongens, R., Turnbull, I. M., Milan, L. A., Daczko, N. R., DePaoli, M. C. and Tulloch, A. J. (2009a) 'Plutonic rocks of Western Fiordland, New Zealand: Field relations, geochemistry, correlation, and nomenclature', *New Zealand Journal of Geology and Geophysics*, 52(4), pp. 379-415.
- Allibone, A. H., Milan, J. A., Daczko, N. R. and Turnbull, I. M. (2009b) 'Granulite facies thermal aureoles and metastable amphibolite facies assemblages adjacent to the Western Fiordland Orthogneiss in southwest Fiordland, New Zealand', *Journal of Metamorphic Geology*, 27, pp. 349-369.
- Anderson, J. L. and Smith, D. R. (1995) 'The effect of temperature and oxygen fugacity on Al-in-hornblende barometry.', *American Mineralogist*, 80, pp. 549-559.
- Blattner, P. (1976) 'Replacement of hornblende by garnet in granulite facies assemblages near Milford Sound, New Zealand', *Contributions to Mineralogy and Petrology*, 55(2), pp. 181-190.
- Bons, P. D., Druguet, E., Castaño, L.-M. and Elburg, M. A. (2008) 'Finding what is now not there anymore: Recognizing missing fluid and magma volumes', *Geology*, 36(11), pp. 851-854.
- Bottazzi, P., Tiepolo, M., Vannucci, R., Zanetti, A., Brumm, R., Foley, S. F. and Oberti, R. (1999) 'Distinct site preferences for heavy and light REE in amphibole and the prediction of Amph/LDREE', *Contributions to Mineralogy and Petrology*, 137(1-2), pp. 36-45.
- Bradshaw, J. Y. (1989) 'Early Cretaceous Vein-Related Garnet Granulite in Fiordland, Southwest New Zealand: A Case for Infiltration of Mantle-Derived CO<sub>2</sub>-Rich Fluids', *The Journal of Geology*, 97(6), pp. 697-717.
- Brown, M. and Rushmer, T. (2006) *Evolution and differentiation of the continental crust*. Cambridge University Press.
- Clarke, G. L., Daczko, N. R., Klepeis, K. A. and Rushmer, T. (2005) 'Roles for fluid and/or melt advection in forming high- P mafic migmatites, Fiordland, New Zealand', *Journal of Metamorphic Geology*, 23(7), pp. 557-567.
- Clarke, G. L., Klepeis, K. A. and Daczko, N. R. (2000) 'Cretaceous high-P granulites at Milford Sound, New Zealand: metamorphic history and emplacement in a convergent margin setting', *Journal of metamorphic geology*, 18(4), pp. 359-374.
- Daczko, N. R., Clarke, G. L. and Klepeis, K. A. (2001a) 'Transformation of two-pyroxene hornblende granulite to garnet granulite involving simultaneous melting and fracturing of the lower crust, Fiordland, New Zealand', *Journal of Metamorphic Geology*, 19(5), pp. 549-562.
- Daczko, N. R., Clarke, G. L. and Klepeis, K. A. (2002) 'Kyanite-paragonite-bearing assemblages, northern Fiordland, New Zealand: rapid cooling of the lower crustal root to a Cretaceous magmatic arc', *Journal of Metamorphic Geology*, 20(9), pp. 887-902.
- Daczko, N. R., Emami, S., Allibone, A. H. and Turnbull, I. M. (2012) 'Petrogenesis and geochemical characterisation of ultramafic cumulate rocks from Hawes Head, Fiordland, New Zealand', *New Zealand Journal of Geology and Geophysics*, 55(4), pp. 361-374.
- Daczko, N. R. and Halpin, J. A. (2009) 'Evidence for melt migration enhancing recrystallization of metastable assemblages in mafic lower crust, Fiordland, New Zealand', *Journal of Metamorphic Geology*, 27(2), pp. 167-185.
- Daczko, N. R., Klepeis, K. A. and Clarke, G. L. (2001b) 'Evidence of Early Cretaceous collisional-style orogenesis in northern Fiordland, New Zealand and its effects on the evolution of the lower crust', *Journal of Structural Geology*, 23(4), pp. 693-713.

- Davidson, J., Turner, S., Handley, H., Macpherson, C. and Dosseto, A. (2007) 'Amphibole "sponge" in arc crust?', *Geology*, 35(9), pp. 787-790.
- Deer, W. A., Howie, R. A. and Zussman, J. (1992) *An Introduction to the Rock-Forming Minerals*. 2 edn. Essex, England: Longman Scientific & Technical.
- Downes, H. (2001) 'Formation and modification of the shallow sub-continental lithospheric mantle: a review of geochemical evidence from ultramafic xenolith suites and tectonically emplaced ultramafic massifs of western and central Europe.', *Journal of Petrology*, 42, pp. 233-250.
- Grant, J. A. (1986) 'The Isocon Diagram: A Simple Solution to Gresens' Equation for Metasomatic Alteration', *Economic Geology*, 81, pp. 1976-1982.
- Grant, J. A. (2005) 'Isocon analysis: A brief review of the method and applications', *Physics and Chemistry of the Earth*, 30, pp. 997-1004.
- Gresens, R. L. (1967) 'Compositional-volumes relationships of metasomatism', *Chemical Geology*, 2, pp. 47-55.
- Holland, T. and Blundy, J. (1994) 'Non-ideal interactions in calcic amphiboles and their bearing on amphibole-plagioclase thermometry.', *Contributions to Mineralogy & Petrology*, 116, pp. 433-447.
- Hollis, J. A., Clarke, G. L., Klepeis, K. A., Daczko, N. R. and Ireland, T. R. (2003) 'Geochronology and geochemistry of high-pressure granulites of the Arthur River Complex, Fiordland, New Zealand: Cretaceous magmatism and metamorphism on the palaeo-Pacific Margin', *Journal of Metamorphic Geology*, 21, pp. 299-313.
- Holness, M. B. and Sawyer, E. W. (2008) 'On the Pseudomorphing of Melt-filled Powers During the Crystallization of Migmatites', *Journal of Petrology*, 49(7), pp. 1343-1363.
- Kelemen, P. B., Dick, H. J. and Quick, J. E. (1992) 'Formation of harzburgite by pervasive melt/rock reaction in the upper mantle', *Nature*, 358(6388), pp. 635-641.
- Kelemen, P. B., Shimizu, N. and Salters, V. J. (1995) 'Extraction of mid-ocean-ridge basalt from the upwelling mantle by focused flow of melt in dunite channels', *Nature*, 375(6534), pp. 747-753.
- Larocque, J. and Canil, D. (2010) 'The role of amphibole in the evolution of arc magmas and crust: The case from the Jurassic Bonanza Arc Section, Vancouver Island, Canada.', *Contributions to Mineralogy & Petrology*, 159(475-792).
- Leake, B. E., Woolley, A. R., Arps, C. E. S., Birch, W. D., Gilbert, M. C., Grice, J. D., Hawthorne, F. C., Kato, A., Kisch, H. J., Krivovichev, V. G., Linthout, K., Laird, J., Mandarino, J., Maresch, W. V., Nickel, E. H., Rock, N. M. S., Schumacher, J. C., Smith, D. C., Stephenson, N. C. N., Ungaretti, L., Whittaker, E. J. W. and Youzhi, G. (1997) 'Nomenclature of amphiboles: Report of the subcommittee on Amphiboles of the International Mineralogical Association Commission on New Minerals and Mineral Names', *The Canadian Mineralogist*, 35, pp. 219-246.
- McDonough, W. F. and Sun, S. s. (1995) 'The composition of the Earth', *Chemical Geology*, 120(3-4), pp. 223-253.
- Morimoto, N., Fabries, J., Ferguson, A. K., Ginzburg, I. V., Ross, M., Seifert, F. A., Zussman, J., Aoki, K. and Gottardi, D. (1988) 'Nomenclature of pyroxenes', *American Mineralogist*, 62, pp. 53-62.
- Muir, R. J., Ireland, T. R., Weaver, S. D., Bradshaw, J. D., Evans, J. A., Eby, G. N. and Shelly, D. (1998) 'Geochronology and geochemistry of a Mesozoic magmatic arc system, Fiordland, New Zealand', *Journal of the Geological Society of London*, 155, pp. 1037-1053.
- Naney, M. T. (1983) 'Phase equilibria of rock-forming ferromagnesian silicates in granitic systems', *American Journal of Science*, 283, pp. 993-1033.
- Otten, M. T. (1984) 'The origin of brown hornblende in the Artfjället gabbro and dolerites.', *Contributions to Mineralogy & Petrology*, 86, pp. 189-199.

- Reverdatto, V. V., Selyatitskiy, A. Y. and Carswell, D. A. (2008) 'Geochemical distinctions between "crustal" and mantle-derived peridotites/pyroxenites in high/ultrahigh pressure metamorphic complexes', *Russian Geology and Geophysics*, 49(2), pp. 73-90.
- Schröter, F. C., Stevenson, J. A., Daczko, N. R., Clarke, G. L., Pearson, N. J. and Klepeis, K. A. (2004) 'Trace element partitioning during high-P partial melting and melt-rock interaction; an example from northern Fiordland, New Zealand', *Journal of Metamorphic Geology*, 22(5), pp. 443-457.
- Smith, J. R., Piaolo, S., Daczko, N. R. and Evans, L. (2015) 'The effect of pre-tectonic reaction and annealing extent on behaviour during subsequent deformation: insights from paired shear zones in the lower crust of Fiordland, New Zealand', *Journal of Metamorphic Geology*, 33, pp. 557-577.
- Stevenson, J. A., Daczko, N. R., Clarke, G. L., Pearson, N. J. and Klepeis, K. A. (2005) 'Direct observation of adakit melts generated in the lower continental crust, Fiordland, New Zealand', *Terra Nova*, 17(1), pp. 73-79.
- Stolz, A. J. (1987) 'Fluid activity in the lower crust and upper mantle: mineralogical evidence bearing on the origin of amphibole and scapolite in ultramafic and mafic granulite xenoliths', *Mineralogical Magazine*, 51, pp. 719-732.
- Stowell, H., Tulloch, A., Zuluaga, C. and Koenig, A. (2010) 'Timing and duration of garnet granulite metamorphism in magmatic arc crust, Fiordland, New Zealand', *Chemical geology*, 273(1-2), pp. 91-110.
- Sturm, R. (2002) 'PX-NOM - an interactive spreadsheet program for the computation of pyroxene analyses derived from the electron microprobe', *Computers and Geosciences*, 28, pp. 473-483.
- Taylor, S. R. and McLennan, S. M. (1995) 'The geochemical evolution of the continental crust', *Reviews of Geophysics*, 33(2), pp. 241-265.
- Wandres, A. M., Weaver, S. D., Shelley, D. and Bradshaw, J. D. (1998) 'Change from calc-alkaline to adakitic magmatism recorded in the Early Cretaceous Darran Complex, Fiordland, New Zealand', *New Zealand Journal of Geology and Geophysics*, 41, pp. 1-14.
- Whitney, D. L. and Evans, B. W. (2010) 'Abbreviations for names of rock-forming minerals', *American Mineralogist*, 95, pp. 185-187.
- Zen, E. and Hammarstrom, J. M. (1984) 'Magmatic epidote and its petrologic significance', *Geology*, 12, pp. 515-518.



## Supplementary material

There is no supplementary material for this thesis.