Norwegian orthopyroxene eclogites: petrogenesis and implications for metasomatism and crust-mantle interactions during subduction of continental crust

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#### Abstract

## The University of Manchester Alexandra Quas-Cohen Doctor of Philosophy

Norwegian orthopyroxene eclogites: petrogenesis and implications for metasomatism and crust-mantle interactions during subduction of continental crust

#### November 2014

This study investigates the ultrahigh pressure (UHP) metamorphic and metasomatic processes involved in the transient subduction-exhumation of continental crust to over 100km depths during a collisional orogeny and the implications for the evolution of the continental crust and crust-mantle interaction at depth. The study focuses on garnet websterites (orthopyroxene eclogites) and zoned, clinopyroxenite-garnetite veining features present in a range of eclogite-facies, crustal metamafic-ultramafic bodies hosted within the Western Gneiss Region (WGR), western Norway. The structural occurrences and textures of some of these crustal garnet websterites are seemingly unique to the WGR but little research has focused on their origin or from a metasomatic perspective. Based on field and petrographic observations, a metasomatic origin is attributed to vein-associated garnet websterites at Kolmannskog, Myrbærneset and Svartberget. A metamorphic origin is attributed to body domains at Nybø, Kolmannskog, Båtneset and Myrbærneset and a combined metamorphic-metasomatic origin is attributed to garnet websterite body domains at Årsheimneset and Remøysunde and inferred at Grytting and Eiksunddal. UHP P-T conditions are obtained from garnet websterites of ~3.7GPa, 740°C at Nybø, ~3.1GPa, 670°C at Grytting, ~3.5GPa, 700°C at Årsheimneset, ~3.6GPa, 815°C at Remøysunde, ~3.0GPa, 750°C at Kolmannskog and ~3.85GPa, 790°C at Svartberget. On this basis, it is proposed the Nordøyane UHP domain be extended eastwards to incorporate the Kolmannskog locality which lies outside its currently defined boundary. Constructed P-T paths suggest the northern Nordøyane UHP domain experienced ~100°C higher temperatures than the southern Nordfjord-Stadlandet UHP domain but experienced similar pressures implying a lower regional P-T gradient than previously established. P-T paths also suggest UHP, vein-forming metasomatism occurred prior to peak temperatures. U-Pb isotopic dating of zircon and monazites in garnetite vein cores dates UHP metasomatism at 414±5.6Ma at Årsheimneset and 410±2.6Ma at Svartberget. The fluid responsible for UHP metasomatism is considered to be a Si-Al-K-H<sub>2</sub>O-rich supercritical liquid produced in the surrounding country rock associated with the breakdown of phengite with a Na-LILE-LREE-HFSE-P enrichment signature. The major element composition of the fluid added to the Svartberget body is calculated to be 48-60% SiO<sub>2</sub>, 17-27% Al<sub>2</sub>O<sub>3</sub>, 3-11% K<sub>2</sub>O, <10% MgO, CaO and FeO, 3-6% Na<sub>2</sub>O, <4% P<sub>2</sub>O<sub>5</sub>, <1% TiO<sub>2</sub> and MnO with an overall, undersaturatedsaturated sialic, syenitic character hybridised through interaction with the garnet peridotite body margins. The continental fluid-mafic-ultramafic rock systems studied imply a zoned metasomatic unit forms at the interface between subducted continental crust and above mantle wedge at depths of ≥120-130km and along any fluid pathways penetrating into the mantle transferring abundant alkalis, water and trace elements into the mantle. Fluidmantle interaction is proposed to form abundant biotite and amphibole and zones of garnet websterite, biotite websterite and biotite clinopyroxenite with lenses of eclogite and/or accessory phase (rutile, zircon, monazite, apatite, xenotime)-rich garnetite ±glimmerite selvages where residual fluids accumulate. Subcontinental mantle metasomatism may be associated with UHP, supercritical liquids derived from subducted, eclogite-facies, continental crust rather than oceanic crust as the continental crust is a greater source of the Si, alkalis, trace elements and water which characterise mantle metasomatism.

# **Declaration**

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## **Dedication**

To those I lost whilst studying

Chris Agnew Sniffer Nanny Apple Jo Byers-Brown Mr Flint

"Geologists have a saying - rocks remember"

Neil Armstrong

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## The Author

Alexandra obtained a 1<sup>st</sup> Class MESci Hons Geology degree from Cardiff University in 2008. Previous research experience includes her Masters research dissertation which focused on the petrography and whole rock chemistry of samples of a gabbroic body exposed at 15°45'N along the Mid Atlantic Ridge within the outside corner of a ridge-transform intersection of the Fifteen-Twenty Fracture Zone to gain insight into the formation and geochemical stratigraphy of the lower oceanic crust.

# **Acronyms and Abbreviations**

10010 0.110.7	
BGS	British Geological Society
BSE	back-scattered electron
CaEs	Ca-Eskola
CCSD	Chinese Continental Scientific Drilling
CIPW	Cross, Iddings, Pirsson, Washington
CL	cathodoluminescence
DHMS(s)	dense hydrous magnesium silicate(s)
DMM	Dora Maira Massif
EHT	extra-high tension
EMP	electron microprobe
EMPA	electron microprobe analysis
FAP	feldspathoid-alkali feldspar-plagioclase feldspar
FOV	field of view (image width)
HFSE(s)	high field strength element(s)
HP(s)	high pressure(s)
HREE(s)	heavy rare earth element(s)
G.T.R.D.	Dr Giles Droop (main supervisor)
K	distribution coefficient
I A-MC-ICP	laser ablation multiple collector-inductively coupled plasma
$I \parallel F(s)$	large ion lithonhile element(s)
	loss on ignition
LBFF(s)	light rare earth element(s)
ENEL(3)	magnification
Mate	Mg_Tshermaks
MDEE	middle rare earth elements
	multiphase solid inclusions
	maniphase solid inclusions
	mass spectrometry
	nominally annudrous minoral(c)
	Notural Environment Desearch Council
NERC	
NIGL	NERC Isolope Geosciences Laboratory
NGU	Norges Geologiske Undersøkelse (Norwegian Geological Society)
oct	octanedral
ptu	per formula unit
PPL	plane polarised light
P <sub>max</sub>	maximum (peak) pressure
ppm	parts per million
P-T	pressure-temperature
P-T-t	pressure-temperature-time
P-T-X	pressure-temperature-composition
QAP	quartz-alkali feldspar-plagioclase feldspar
REE(s)	rare earth element(s)
SCEP	second critical end point
SCLM	subcontinental lithospheric mantle
SEAES	School of Earth, Atmospheric and Environmental Sciences
SEM	scanning electron microscope
tet	tetrahedral
T <sub>max</sub>	maximum (peak) temperature
TTG	tonalite-trondhjemite-granodiorite
UHP(s)	ultrahigh pressure(s)
WGC	Western Gneiss Complex
WGR	Western Gneiss Region
XPL	cross polarised light
XRF	X-ray fluorescence

Table 0.11a: Alphabetical list of abbreviations and acronyms used in the text

and schmid (2007) as recommended by the British Geological society (BGS)					
Abbreviation	Mineral	Abbreviation	Mineral		
ab	albite	kely	kelyphite		
act	actinolite	kfs	k-feldspar		
aeg	aegirine	ky	kyanite		
afs	alkali feldspar	mag	magnetite		
alm	almandine	mc	microcline		
aln	allanite	mca	mica		
als	aluminosilicate	mgs	magnesite		
am	amphibole (calcic)	mnz	monazite		
an	anorthite	ms	muscovite		
and	andalusite	oam	orthoamphibole		
ank	ankerite	ol	olivine		
ann	annite	omp	omphacite		
ар	apatite	ор	opaque		
aug	augite	орх	orthopyroxene		
brt	barite	or	orthoclase		
bt	biotite	pgt	pigeonite		
cal	calcite	phg	phengite		
cb	carbonate	phl	phlogopite		
cel	celadonite	pl	plagioclase		
chl	chlorite	prg	pargasite		
coe	coesite	prp	pyrope		
срх	clinopyroxene	рх	pyroxene		
crn	corundum	qtz	quartz		
cum	cummingtonite	rt	rutile		
CZO	clinozoisite	sd	siderite		
di	diopside	sdg	sadanagaite		
dia	diamond	ser	sericite		
dol	dolomite	sid	siderophyllite		
ed	edenite	sil	sillimanite		
en	enstatite	spl	spinel		
ер	epidote	sps	spessartine		
etn	'eastonite'*	srp	serpentine		
fa	fayalite	symp	symplectite		
fo	forsterite	te	tephroite		
fs	ferrosilite	tlc	talc		
fsp	feldspar	tr	tremolite		
gr	graphite	ttn	titanite		
grs	grossular	ts	tschermakite		
grt	garnet	tur	tourmaline		
hbl	hornblende	usp	ulvöspinel		
hd	hedenbergite	wmca	white mica		
idg	iddingsite	wo	wollastonite		
ilm	ilmenite	xtm	xenotime		
jd	jadeite	ZO	zoisite		
ka	kanoite	zrn	zircon		

Table 0.11b: Alphabetical list of mineral abbreviations used in the text after Siivola and Schmid (2007) as recommended by the British Geological Society (BGS)

\* eastonite is no longer used as an end-member for biotite but an alternative name is not specified

### Map/Diagram/Traverse Key

Table 0.12: Key to field maps, diagrams and traverses ---- exposure (exp) boundary \_\_\_\_\_ contact \_\_\_ inferred contact ..... within unit change exposure surface shape  $\mathcal{N}$ water vegetation veg bimineralic eclogite garnet websterite garnet peridotite/olivine garnet websterite gneiss/schist interbanded schist-eclogite retrogressed garnet websterite amphibolitised eclogite amphibolite clinopyroxenite garnetite garnetite-clinopyroxenite-websterite peg pegmatitic vein V symplectitic (s) b biotite-rich carbonate-rich С garnet-rich g k kyanite-rich ks K-feldspar-rich orthopyroxene-rich 0 phengite-rich р quartz-rich q rutile-rich r sillimanite-rich s 'spinifex'-textured garnet Х s sulphide-bearing t talc (A)sample location

# <u>Graph Key</u>

Locality	Marker Shape	Example	
Nybø	ex	x	
Grytting	diamond	$\diamond$	
Årsheimneset	circle	Ο	
Remøysunde	square		
Eiksunddal	long dash	-	
Kolmannskog	star	*	
Båtneset	short dash	-	
Myrbærneset	cross	+	
Svartberget	triangle	Δ	
Rock Type	Marker Line/Line Colour	Example	
garnet peridotite/olivine garnet websterite	dark green	Δ	
eclogite	bright green	Δ	
garnet websterite domain	brown	Δ	
websterite vein selvage	orange	Δ	
clinopyroxenite vein selvage	light green	Δ	
eclogite vein selvage/core	red with bright green fill		
glimmerite (biotite) selvage	black	٨	
garnetite vein core	red	^	
quartz-white mica vein core	hlue	^	
gneiss	dark grev	Δ	
	light grey		
ignoous reference rock	ing it grey		
		•	
Mineral Textural Location	Marker Fill	Example	
inclusion	emply lighter shade of line		
poikiloblastic	same as line		
late, retrogressive	grev		
biotite-rich	black^		
apatite-rich	yellow		
alkali feldspar-rich	pink		
Element Crystal Profiles (Chpt 6)	Line Colour	Example	
Si	dark blue		
Ті	dark red		
Al	green		
Cr	purple		
Fe <sup>3+</sup>	bright blue		
Fe <sup>∠+</sup>	orange		
Mn	light blue		
Mg	pink		
Са	light green		
Na	light purple		
K	baby blue		

Table 0.13: Key describing marker and line formats used in graphs

^ - if symbol is a cross, star or ex the line and fill are exchanged to aid visibility as shown

#### Chapter 1 – Introduction

#### 1.1. Research Statement

The research carried out in this study investigates the ultrahigh pressure (UHP) metamorphic and metasomatic processes involved in the transient subduction-exhumation of continental crust to over 100km depths during a collisional orogeny and the implications for the evolution of the continental crust and crust-mantle interaction at depth. The study focuses on the features of a range of eclogite-facies, garnet websterite-bearing, mafic-ultramafic bodies hosted within the UHP terrane of the continental Western Gneiss Complex (WGC) in the Western Gneiss Region (WGR), western Norway.

The characteristics and settings of some of the garnet websterites and veining features studied currently appear to be unique to the WGC having not been observed in other UHP terranes worldwide. Whilst some petrographic and thermobarometric research has been carried out on these rocks, relatively little focus has been given to their origin: it is not yet established how they formed. The structure of garnet websterites at one recently discovered garnet peridotite body Svartberget (Vrijmoed *et al.*, 2005, 2006), where they occur in association with diamond-bearing veins, clearly indicates they are metasomatic in origin and formed at UHP conditions. Garnet websterites at other known and previously studied localities resemble those at Svartberget but are not obviously metasomatic. Studying the petrographic and geochemical characteristics of a range of different garnet websterites will help identify processes involved in their formation; potentially characterising previously unknown or unconsidered deep fluid processes which may have significant implications for other Earth system processes and our understanding of them.

#### 1.2. Ultrahigh Pressure Metamorphism

UHP metamorphism is an extension of the high pressure (HP), low temperature metamorphic facies series of Miyashiro (1961) with a lower limit defined in pressure-temperature (P-T) space by the quartz-coesite transformation boundary: coesite is a high pressure polymorph of silica (*Fig. 1.1*) (e.g. Krabbendam and Wain, 1997; Poli and Fumagalli, 2003; Cuthbert *et al.*, 2000). Its lower P-T conditions are mostly restricted to the eclogite-facies field limited by the lowest possible geotherm assumed to be 5°C/km



Figure 1.1: P-T diagram showing metamorphic facies and phase stability fields. Facies fields after Yardley (1989) as recommended by the British Geological Society (BGS) (Robertson, 1999). A = amphibolite facies; BS = blueschist facies; E – eclogite facies; G – granulite facies; GS – greenschist facies; Z – zeolite facies. Phase abbreviations as listed in Table 0.11b. Dashed line =  $5^{\circ}$ C/km geotherm (from Zheng, 2012). Solid lines = polymorph boundaries. Al<sub>2</sub>SiO<sub>5</sub> polymorph boundaries from Holdaway (1971). Quartz-coesite boundary from Bose and Ganguly (1995). Graphite-diamond boundary from Day (2012). Albite = jadeite + quartz equilibria from Holland (1980). Shaded areas represent metamorphic facies series after Maruyama et al. (2010) and references cited therein: red = high P-T series (high grade); blue = intermediate high P-T series; green = medium P-T series (medium grade); purple = intermediate low P-T series; orange = low P-T series (low grade).

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(Schreyer, 1988). The upper limit is not defined and transitions into mantle P-T conditions. The UHP field was first defined in the mid-1980s following the discovery of coesite in exhumed, subducted crustal rocks from the Dora Maira Nappe, western Alps by Chopin (1984). Coesite was first reported in the WGR shortly after by Smith (1984). Prior to this, such high pressures were not thought to be experienced by continental rocks during subduction: coesite was only known from impact craters and subcontinental lithospheric mantle xenoliths (e.g. Chao et al., 1960; Smyth, 1977). Since the mid 1980's discoveries of coesite (or convincing evidence of prior coesite) have been made on every continent in orogenic belts which represent past convergent margins (Fig. 1.2) and major UHP terranes lie within Phanerozoic orogens such as the Caledonian-Appalachian and the Himalayan-Alpine belts (e.g. Hacker, 2007). UHP conditions are confirmed primarily from the presence of coesite, convincing evidence of prior coesite such as palisade-textured and polycrystalline guartz (e.g. Smyth, 1977), the presence of diamond or inferred from reliable thermobarometric estimations. Other mineralogical indicators of high pressures include the exsolution of pyroxene from garnet and vice versa, orientated needles of silica from supersilicic clinopyroxene and K-rich silicates such as K-wadeite [K<sub>2</sub>ZrSi<sub>3</sub>O<sub>9</sub>] (e.g. Smith, 1984; van Roermund and Drury, 1998; Ye et al., 2000a; Ogasawara et al., 2002; Berger et al., 2010). Geochemical indicators of high pressures include grossular  $[Ca^{viii}_{3}AI^{vi}_{2}Si^{iv}_{3}O_{12}]$  and majorite  $[Mg^{vii}][Mg,Si]^{vi}Si^{iv}_{3}O_{12}]$  components in garnet (Drury *et al.*, 2001; Wain *et al.*, 2001; Kiseeva et al., 2012), jadeite [NaAlSi<sub>2</sub>O<sub>6</sub>] and K-rich diopside components in clinopyroxene (e.g. Harlow, 1997; Ogasawara et al., 2002) and low-Al enstatite orthopyroxene (van Roermund, 2009). Structural UHP indicators include TiO<sub>2</sub> with  $\alpha$ -PbO<sub>2</sub> structure and high pressure C2/C clinoenstatite (Ye et al., 2000a). In the absence of evidence for UHP conditions rocks are assigned to the HP, eclogite-facies field (Wain et al., 2000). Their absence however does not preclude the attainment of UHP conditions as the appropriate phases may not occur; for example, diamond can be abundant in metapelitic gneisses and dolomitic marbles but absent in adjacent eclogites due to unfavourable chemical conditions (Katayama et al., 2006 and references cited therein).

The eclogite-facies is split into three main sub-facies of quartz, coesite and diamond with the two latter forming the UHP field. Eclogitisation involves the transformation of framework and layer silicates to chain and orthosilicates and other crystal structures based on closest-packing arrangements which are more stable at higher pressures (Ernst *et al.*, 1997; Chopin and Ferraris, 2003). Polymorphic transformations occur as compounds try to overcome instabilities and attain a structure that allows further compression (Chopin and Ferraris, 2003). Both transformations increase density by reducing interatomic distances, increasing coordination numbers (Chopin and Ferraris, 2003; Schertl and O'Brien, 2013) and reducing volume; for example, the quartz to coesite transformation has a ~10% volume decrease (e.g. Wain *et al.*, 2001).

#### 1.2.1. Metamorphic Reactions at UHP Conditions

The relatively low temperatures and great pressures experienced by rocks in this geological setting are not favourable conditions for metamorphic reactions. Diffusion and prograde reactions become sluggish and conditions promote metastability as assemblages do not react if conditions are not kinetically favourable: competent, impermeable, anhydrous rocks can retain low pressure parageneses (Ernst et al., 1997). Therefore, peak conditions experienced may not be recorded. Rates of reactions depend on kinetics, chemical potential gradients, element mobility and transport, grain size, mineral nucleation and growth, the degree of reaction overstepping, the ease of mineral reaction, deformation and fluid availability and are positively dependent on temperature (Carmichael, 1969; Mearns, 1986; Mørk, 1986; Austrheim et al., 1997; Engvik et al., 2000; Krabbendam et al., 2000; Wain et al., 2001; Rubbo and Bruno, 2003; Wang and Tang, 2013). The absence of deformation and/or fluids is seen to prevent re-equilibration (Ernst et al., 1997; Molina et al., 2002; Carswell and Cuthbert, 2003a; Bingen et al., 2004; Massonne, 2009); for example, gabbros are known to transform to eclogite only if infiltrated by fluids (Kessel et al., 2005), partial transformations are evident along fractures in dry lithologies with very abrupt reaction fronts (e.g. Austrheim, 1990; Bingen et al., 2001) and igneous protoliths are able to persist even on a geological timescale (Austrheim et al., 1997; Carswell and Compagnoni, 2003).

Variations in the degree of eclogitisation over cm to dm-sc distances demonstrate there are local controls over the transformation (Mørk, 1986). Incomplete eclogitisation is thought to be due to rate-limiting, low diffusive mobility of Na, Al and Si to reaction sites (Mørk, 1986). The growth of coronas of garnet, for example, in the early stages of eclogitisation, acts to isolate reactant minerals from each other and reduce compositional gradients (Mørk, 1986). The diffusion of species through

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crystals is orders of magnitude slower than transportation via a connected intergranular fluid (Mørk, 1986; Schwarz *et al.*, 2011 and references cited therein); if such a fluid is absent, reaction rates and crystal growth can be incredibly slow (e.g. Glodny *et al.*, 2008).

Hydrous fluids are extremely effective catalysts: they aid species transport and exchange, quicken transport rates, promote deformation, change the reaction pathway or mechanism, lower the activation energy, enhance chemical gradients or aid nucleation and growth (Rubie, 1986 and references cited therein; Thompson, 1992; Ernst et al., 1997; Hermann and Green, 2001; Jahn et al., 2003). They also act as reactants in metamorphic reactions and can affect the stability of the assemblage e.g. by lowering the melting temperatures of rocks (Rubie, 1986; Krabbendam et al., 2000). They can also raise diffusion rates during recrystallisation allowing eclogites to achieve a much coarser grain-size than would be possible in the anhydrous state (Brueckner, 1977). The duration of fluid-present conditions is the main factor controlling or limiting the extent of reactions and reaction rates (Rubie, 1986; Carswell and Compagnoni, 2003; Ernst, 2005). The presence of a free fluid may occur for limited periods of time or not at all if water-conserving reactions occur forming hydrous phases stable at UHP conditions such as phengite and zoisite (Carswell and Cuthbert, 2003a) or if dissolved into grain boundaries (Rubie, 1986). Pervasive eclogite-facies metamorphism is unlikely to occur and textural and chemical equilibrium unlikely to be attained unless eclogitisation occurs under fluid-present conditions (e.g. Hermann, 2002; Bingen et al., 2004). Furthermore, an intergranular fluid is thought to be a requirement for the transition to go to completion at low temperature conditions (Mørk, 1985).

Slow reaction kinetics and the absence of fluid and deformation, whilst promoting chemical and textural disequilibrium, enable the metastable preservation of UHP conditions at the Earth's surface and interpretation of deep geological processes and evolution (e.g. Jahn *et al.*, 2003). HP-UHP parageneses are best preserved in metamafic-ultramafic rocks (Carswell and Compagnoni, 2003; Chopin and Ferraris, 2003) due to their greater resistance to deformation and more anhydrous nature (e.g. Cuthbert *et al.*, 2000 and references cited therein). Shear stresses on rheologically-contrasted materials result in stress partitioning concentrating into particular horizons and heterogeneous strain distribution (e.g. Raimbourg *et al.*,

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2007). Felsic rocks are therefore more likely to transform to an UHP paragenesis but they also re-transform more rapidly during retrogression (Griffin, 1987; Schmädicke, 1991; Carswell and Compagnoni, 2003). The preservation of felsic, eclogite-facies assemblages occurs only in low-strain zones such as pressure shadows of maficultramafic pods (Krabbendam *et al.*, 2000) demonstrating a reduced level of deformation is required for their preservation. The preservation of UHP minerals typically only occurs within a mechanically strong mineral such as garnet or zircon, acting as a pressure vessel preventing volume increase or back transformation and protecting them from metamorphic fluids and deformation (e.g. Chopin and Ferraris, 2003; Smit *et al.*, 2008). Preservation is also aided by volume increase upon transformation which acts to reduce porosity and permeability (Rubie, 1986).

#### 1.2.2. Eclogite-Facies Assemblages

All eclogite-facies rocks are eclogitic but the name 'eclogite' refers in particular to those with a mafic bulk composition. The main mineralogical distinction of the eclogite-facies is the absence of stable plagioclase feldspar (e.g. Hermann, 2002). Different bulk compositions can contain remarkably similar assemblages but with differing mineral abundance and highly variable compositions (e.g. Schmidt et al., 2004; Tropper and Manning, 2008 and references cited therein). Na and Fe do not form additional phases but substitute into Ca-Mg phases (Hermann, 2002). Eclogitefacies garnet and clinopyroxene have no available structural sites for large alkalis such as K leading to either expulsion from the rock during transformative dehydration (Jamtveit, 1987a) or the formation of K-bearing phases stable at HP-UHP conditions: this is often phengite or biotite or, in granitoids compositions, K-feldspar (e.g. Droop et al., 1990). Assemblages in mafic eclogites are dominated by garnet and omphacite (intermediate jadeite-diopside clinopyroxene), which are essential phases, and contain varying amounts of silica (quartz/coesite), orthopyroxene, kyanite, rutile (titanite), carbonate, apatite and hydrous phases including white mica (phengite/paragonite), biotite, Na-rich amphibole (barroisite/glaucophane/nyböite), clinozoisite/zoisite, epidote, ellenbergite, lawsonite and talc (e.g. Krogh, 1977; Smith, 1988; Ernst et al., 1997; Wain et al., 2001; Krogh Ravna and Paquin, 2003; Engvik et al., 2007). Mafic eclogites have been divided into two lineages - Mg-rich
orthopyroxene-bearing and Al-rich kyanite-bearing (Griffin and Brueckner, 1980; Gebauer *et al.*, 1985) dependent on protolith bulk compositions.

It was generally assumed that hornblende and phlogopite were not part of the eclogite-facies assemblage (Forbes, 1965); however, if Ca and Na are abundant, calcic amphibole can occur (Poli and Schmidt, 2002) and biotite is known in (and confined to) mafic-ultramafic eclogite-facies rocks (Lappin and Smith, 1978). It is difficult to determine the conditions of formation of amphiboles or the source of the fluid due to complex distributions, textures and compositional relationships (e.g. Lappin and Smith, 1978). Volatile-bearing phases commonly form post-peak pressure, 'early' phases with stable, mutual contacts or replacement relationships (Lappin and Smith, 1978).

Ultramafic eclogitic rocks have assemblages dominated by garnet, diopside, enstatite, forsterite and phlogopite (e.g. Krogh Ravna and Paquin, 2003). Felsic eclogitic rocks dominantly contain assemblages of garnet, quartz/coesite, phengite, biotite, jadeite/omphacite, kyanite, K-feldspar and zoisite (e.g. Wain *et al.*, 2001; Compagnoni and Rolfo, 2003). Calcareous eclogitic rocks (marbles) dominantly contain dolomite or aragonite and magnesite (which replace dolomite above 5GPa) with minor coesite, chlorite, clinopyroxene, phengite, garnet, zoisite and rutile depending upon the composition of any impurities (Compagnoni and Rolfo, 2003; Hacker, 2007).

Retrogression of eclogitic rocks typically involves the hydration of anhydrous phases, breakdown of unstable hydrous phases and transformation of coesite to quartz. Hydration promotes preservation but dehydration produces rate-enhancing water promoting further reactions and transformations (e.g. Carswell and Compagnoni, 2003; Ernst, 2005; Engvik *et al.*, 2007). Mafic-ultramafic bodies are typically dominantly anhydrous and so are less prone to retrogression than felsic lithologies which often contain abundant hydrous phases such as micas and zoisite (Heinrich, 1982; Cuthbert *et al.*, 2000 and references cited therein). Common retrograde transformations include the creation of delicate diopside and albite symplectites after omphacite as the jadeite component breaks down to albite in the presence of quartz (e.g. Mysen and Griffin, 1973; Koenermann, 1993; Carswell and Cuthbert, 2003a,b; Hacker, 2007). Further retrogression involves alteration of clinopyroxene to

amphibole, production of plagioclase from the grossular component of garnet, alteration of rutile to titanite or ilmenite, breakdown of phengite to muscovite and the celadonite component to biotite and plagioclase, transformation of kyanite to sillimanite and alteration of zoisite to epidote (e.g. Cuthbert *et al.*, 1983, 2000; Carswell and Cuthbert, 2003b; Compagnoni and Rolfo, 2003; Engvik *et al.*, 2007).

# 1.3. Deep Subduction of Continental Crust

Due to the relatively rare nature of identified, well preserved, exposed and accessible continental UHP terranes worldwide and the complex tectonometamorphic evolution of orogenies, the mechanisms of their subduction and exhumation are still not well understood or constrained (e.g. Kylander-Clark et al., 2007; Zeng et al., 2009; Zhang et al., 2010; Chen et al., 2013). Subduction is evoked as the 100km-plus depths recorded cannot be achieved by continental underthrusting or crustal stacking and no other tectonic setting can produce the high pressures and relatively moderate temperatures (e.g. van Roermund et al., 2002 and references cited therein). In addition, subduction of continental crust to depths greater than that of the geophysical Moho is shown to occur under modern collision zones (Andersen et al., 1991a; Brueckner and van Roermund, 2004). The most well known and studied UHP terranes from which tectonic models are developed include those of the WGR, the Dora-Maira Massif (DMM), western Alps, the Kokchetav Massif, Kazakhstan, the Dabie-Sulu and North Qaidam terranes in China and the Bohemian Massif in central Europe. Other occurrences of UHP rocks are shown in Figure 1.2. A present day analogue of continental crust subduction is the eastern end of the Indonesian island arc where the Australian Plate is subducting beneath the southeast corner of the Eurasian Plate (Ernst, 2005).

HP-UHP terranes are typically found juxtaposed against low pressure, lithotectonic, allochthonous units (Ernst *et al.*, 1997; Compagnoni and Rolfo, 2003; Brueckner and van Roermund, 2004; Ernst, 2005) forming a metamorphic core complex to orogens and exposed by erosion and/or extensional tectonic regimes: juxtaposition occurs through considerable crustal thinning across such crustal-scale shear zones (Hartz *et al.*, 1994). Crustal UHP terranes are mostly known from Alpine-type continental collisional margins where continental lithosphere in a passive margin follows oceanic lithosphere down the subduction channel (Ernst, 2005) (*Fig. 1.3*). In comparison, two of the recognised UHP



*Figure 1.2: Map of worldwide coesite- and diamond-bearing UHP terranes. Image taken from Zheng (2012).* 



*Figure 1.3: Diagram illustrating deep subduction of continental crust during continental collision. SCLM = subcontinental lithospheric mantle.* 

terranes formed in Pacific-type collisional zones involving only the subduction of oceanic lithosphere (Carswell and Compagnoni, 2003): this is rare and requires an atypical tectonic regime (Carswell and Compagnoni, 2003).

UHP terranes consist of isolated mafic-ultramafic bodies found to record UHP conditions over an area of less than 1000km<sup>2</sup>, are relatively tabular or sheet-like and are mostly found in sialic, cratonic crust dominated by quartzofeldspathic and metapelitic lithologies characterised by a pervasive, exhumation-related, amphibolite-facies paragenesis dominated by exhumation-related structures (e.g. Ernst, 2005). The mafic-ultramafic bodies are thought to originate either as larger bodies formed by the fractional crystallisation of basaltic magma in crustal magma chambers (e.g. Griffin and Brueckner, 1980; Cuthbert *et al.*, 1983; Zhang *et al.*, 2010), as dolerite dykes/sheets, as bands of supracrustal volcaniclastic sediments or as part of the mantle wedge entrained during subduction (e.g. Cuthbert *et al.*, 1983; Jahn *et al.*, 2003).

Following the discovery of coesite in crustal rocks, collisional models had to be amended to incorporate the subduction and exhumation of basement rocks to UHP depths. Density is considered to be a controlling factor in all aspects of subduction-exhumation tectonics as changes in petrophysical properties exert control over orogenic tectonics (e.g. Dewey et al., 1993; Austrheim et al., 1997; Wain et al., 2001; Carswell and Cuthbert, 2003a). The amount of residual buoyancy in subducted terranes determines their ability to be exhumed and depends upon their lithological composition and the degree of metamorphic transformation. Eclogitisation of mafic lithologies results in a density increase from  $\sim$ 3.0g.cm<sup>-3</sup> to  $\sim$ 3.74g.cm<sup>-3</sup> at depths of  $\sim$ 100km (Ernst *et al.*, 1997) which is higher than the average upper mantle which has a density of  $\sim$ 3.2 g.cm<sup>-3</sup> at depths of  $\sim$ 100km (Ernst *et al.*, 1997; Mibe et al., 2003). Eclogitised oceanic crust is therefore negatively buoyant preventing its exhumation and promoting its detachment and sinking into the deeper mantle and is responsible for the general absence of exhumed, eclogite-facies oceanic terranes at the Earth's surface (e.g. Ernst et al., 1997; Ernst, 2005). Felsic lithologies remain positively buoyant upon eclogitisation having a density of ~3.03g.cm<sup>-3</sup> at depths of ~100km (e.g. Ernst et al., 1997; Ernst, 2005) as long as the abundance of garnet remains low (as its formation causes the greatest change in density) (Zheng et al., 2009; Peterman et al., 2009) and the proportion of hosted, eclogitised mafic-ultramafic lithologies is small (e.g. England and Holland, 1979). The density of the continental crust is thought to only approach that of

the mantle at depths of ~200km (Ye *et al.*, 2000a) or ~250km (Wu *et al.*, 2009). Felsic lithologies are more susceptible to retrogression and therefore continental terranes become less dense more readily during exhumation promoting further exhumation. The low abundances (5-15%) of eclogite-facies, mafic-ultramafic rocks which characterise UHP terranes worldwide may be due to selective exhumation with terranes with higher proportions of eclogitised mafic-ultramafic lithologies not buoyant enough to return to the Earth's surface (Zheng *et al.*, 2009). Metamorphism and lithology are therefore important controls on the evolution and characteristics of orogenic belts (e.g. Engvik *et al.*, 2000).

The presence of UHP terranes within worldwide orogenic belts suggests that deep continental subduction is a fundamental process in continental collision (Hacker, 2007). The maximum depth of subduction or ability to exhume may have changed over time in response to planetary cooling, continent formation and lithospheric evolution (e.g. Ernst et al., 1997; Carswell and Compagnoni, 2003; Spengler et al., 2009). It is thought high geothermal gradients in the Precambrian hampered plate subduction to blueschist-facies and eclogite-facies P-T conditions precluding either the development or preservation of UHP rocks or continental subduction all together (Ota et al., 2004; Zheng et al., 2009). The absence of UHP terranes older than ~650Ma may reflect this or may simply be due to subsequent reworking, burial, erosion or re-subduction (Carswell and Compagnoni, 2003). P-T paths in subduction zones depend on a number of factors including the structure and age of the subduction zone, convergence rate, basal heat flux, thermal conductivity, convection in the mantle wedge, the initial thermal gradient, the depth reached within the subduction channel and the mechanisms of exhumation (e.g. Peacock, 1990, 1992). Deep subduction-related P-T paths typically form clockwise loops (Peacock, 1992) with maximum temperatures (T<sub>max</sub>) after maximum pressure (P<sub>max</sub>) followed by near-isothermal decompression during exhumation and rapid cooling (e.g. Krogh Ravna and Paquin, 2003).

## 1.3.1. <u>Tectonic Model of the Subduction-Exhumation of Continental Crust</u>

Subduction of continental crust is driven by the push of converging continents, the drag of underlying mantle and the pull by previously subducted, attached, sinking, heavy oceanic lithosphere (Chemenda *et al.*, 1995, 1996; Ernst *et al.*, 1997; Wain *et al.*, 2001; Koons *et al.*, 2003; McClelland *et al.*, 2006 and references cited therein; Robinson *et al.*, 2003; Brueckner and van Roermund, 2004; Raimbourg *et al.*, 2007). It is resisted by frictional forces along the upper surface of the subduction channel and

by the positive buoyancy of the shallow crust relative to the mantle (Chemenda *et al.*, 1995; Ernst *et al.*, 1997; Wain *et al.*, 2001). The extent of continental subduction is dependent upon a range of factors such as the velocity of convergence, evolution of the collision, age of oceanic and continental lithosphere, age and thermal structure of the subduction zone and the strength of coupling to the oceanic crust (Peacock, 1990; Koons *et al.*, 2003). The timing of continental subduction relative to peak convergence and crustal thickening is poorly constrained but is thought that subduction to and exhumation from UHP depths occurs relatively early in continental collision (e.g. Zheng, 2012) and the rocks that reach UHP conditions and are exhumed are those near the transition from oceanic to continental lithosphere (Zheng *et al.*, 2009).

Detachment of the continental crust from the oceanic crust, 'slab break-off', and from the subcontinental lithospheric mantle, 'delamination', removes the pull force of subduction, induces isostatic rebound of the crust due to its positive buoyancy and enables the continental material to be exhumed to crustal levels back along the subduction channel (Cuthbert et al., 1983 and references cited therein; Andersen et al., 1991a; McClelland et al., 2006 and references cited therein; Zheng, 2012). Detachment may also induce asthenospheric upwelling aiding exhumation and causing the post-peak pressure thermal maximum experienced by many UHP terranes such as the WGR and the Bohemian Massif (Schmädicke et al., 2010) and the widespread magmatic activity in the Dabie-Sulu Orogen and European Variscides (Massonne and O'Brien, 2003; Ernst, 2005; Webb et al., 2006). Detachment probably occurs due to differences in physical properties and composition (Zheng *et al.*, 2009). It may originate in eclogite-facies shear zones due to changes in volume during phase transition and consequent increased porosities and infiltration of fluid, due to transition to different mechanisms of deformation or be facilitated by partial melting (Austrheim 1987, 1990 and references cited therein; Austrheim et al., 1997 and references cited therein; Hermann and Green, 2001). There is little evidence with which to constrain the length of residence of the continental crust at UHP conditions (Spengler et al., 2009). Long residence times (>15Myr) are indicated by geochronological studies but are not proposed or supported by tectonic models (Hacker, 2007; Spengler et al., 2009). Subduction and residence for a relatively brief period of time and subsequent exhumation has been proposed and termed 'dunk' tectonics by Brueckner and van Roermund (2004).

The subducted crust undergoes exhumation in a bid to attain a normal crustal thickness and an equilibrium with the mantle in terms of density (e.g. Massonne and O'Brien, 2003). Rapid exhumation and cooling is evident in all terranes indicated by the short time interval between UHP and low pressure conditions (suggesting this may be a requirement for their exhumation) (Massonne and O'Brien, 2003; Ernst, 2005). They are typically characterised by isothermal exhumation P-T paths, such as recorded in the Bohemian Massif and the Dabie-Sulu terranes (Massonne and O'Brien, 2003; Ernst, 2005; Leech et al., 2006). These do not lend themselves to accurate thermochronometry: only overall rates can be distinguished but it is widely thought that exhumation rates vary in relation to the changing mechanisms of exhumation (e.g. Rubatto et al., 2003; Ernst, 2005; Kylander-Clark et al., 2008). Magmatism and migmatism, which are expected during exhumation due to the degree of decompression, may influence or further facilitate exhumation through reducing the lithological strength of the terrane and lubricating contacts (e.g. Hermann and Green, 2001; Labrousse et al., 2002; Chen et al., 2013 and references cited therein). Deformation during exhumation is strongly partitioned into the weaker sialic, ductile, felsic gneisses and schists (e.g. Austrheim et al., 1997; Carswell and Compagnoni, 2003) and localised within shear zones (e.g. Gerya and Stöckhert, 2002; Spandler and Hermann, 2006). The higher survival rates of UHP mineralogies are due to the less penetrative deformation suffered by the more structurallycompetent mafic-ultramafic bodies (e.g. Carswell and Cuthbert, 2003a). Break up of large mafic-ultramafic bodies is thought to occur during exhumation as extensional deformational features are concordant with exhumation-related structures in the country rocks (e.g. Hartz et al., 1994).

Buoyancy-driven exhumation, termed 'eduction' by Andersen *et al.* (1991a), is the most comprehensive model and widely accepted mechanism for driving exhumation (e.g. Andersen *et al.*, 1991a; Brueckner and van Roermund, 2004) although it is not accepted by all to explain the complete exhumation of terranes (Hacker, 2007). An alternative model is the 'expulsion-extrusion' model where subducted material is 'squeezed' or forced back up to the surface in the subduction channel due to continued convergence at depth and extension in the upper regions of the orogen (Platt, 1993 and references therein; Gerya and Stöckhert, 2002). Eclogite-facies continental crust is not buoyant with respect to continental crust at the Moho which has a density of 2.7g.cm<sup>-3</sup> (Hacker, 2007). Rapid exhumation can therefore only occur

to the base of the Moho where large positive buoyancy forces no longer apply (e.g. Brueckner and van Roermund, 2004; Hacker, 2007). Exhumation above the Moho through the crust must be driven by a different physical property or driving mechanism: extensional deformation is probably responsible (e.g. Platt, 1986, 1993). Tectonic denudation on extensional faults and large detachments in the mid-upper crust is a well-documented process for syn-post-collisional exhumation of highpressure rocks in orogenic belts (Andersen *et al.*, 1994). The relative removal of overlying units has been proposed to occur by tectonic imbrication, uplift and erosion, strike-slip faulting ('escape' tectonics) and gravitational collapse of the orogen aided by isostatic rebound and asthenospheric upwelling (e.g. Platt, 1993 and references therein; Ernst *et al.*, 1997; Koons *et al.*, 2003; Brueckner and van Roermund, 2004; Ernst, 2005; Kirkland *et al.*, 2006). Erosion is considered responsible for the final ≤20km exhumation of the terranes to the Earth's surface (e.g. Platt, 1993; Ernst, 2005).

The scale and coherency of HP-UHP crustal terranes are also poorly constrained but a variety of evidence suggests large terranes can be subducted and exhumed coherently (e.g. Cuthbert *et al.*, 2000; Ye *et al.*, 2000b; Hirajima and Nakamura, 2003; Jahn *et al.*, 2003; Shatsky and Sobolev, 2003; Kikuchi and Ogasawara, 2006; Xu *et al.*, 2006; Hacker, 2007; Hacker *et al.*, 2010; Zheng, 2012). However, not all subducted continental crust may be exhumed (Brueckner and van Roermund, 2004); some may remain in the upper mantle having been exchanged into the mantle wedge or torn from the edges of the terrane due to frictional forces or some may sink deeper into the mantle attached to the oceanic crust or subcontinental lithospheric mantle towards the "depth of no return" ca. 250km (Wu *et al.*, 2009).

## 1.4. UHP Subduction Zone Fluids

The large-scale process of volatile and element recycling at subduction zones occurs through the release of fluids from the subducting slab towards the slab-mantle interface, their infiltration of the overlying mantle wedge which they metasomatise, hydrate and of which they induce partial melting causing extensive magmatic activity resulting in arc development and volcanic outgassing (e.g. Peacock, 1990; Thompson, 1992; Philippot, 1993; Poli and Schmidt, 1995; Manning, 2004; Maruyama *et al.*, 2009). Other fluid-related structures such as mud volcanoes and serpentine diapirs also form in the forearc due to fluid flowing towards the surface along the slab-mantle interface (Philippot, 1993). The efficiency of the recycling of volatile phases at subduction zones is poorly constrained due to highly variable regimes worldwide and difficulties in quantifying inputs and outputs: estimations vary widely. However, only a small volume of fluid actually released from the oceanic slab is thought to reach the surface in arc magmas (Becker *et al.*, 2000; Prouteau *et al.*, 2001).

Fluid release from a subducting slab is thought to be a continuous process which occurs over a large depth interval from shallow burial up to 300km through compaction, metamorphic devolatisation and partial melting (e.g. Poli and Schmidt, 1995; Becker et al., 2000; John et al., 2008). Subduction zone fluids over ~25km depths are thought to be dominantly hydrous with low salinities and CO<sub>2</sub> contents (e.g. Manning, 2004; Bebout, 2007):  $CO_2$  tends to disappear from the free fluid phase when carbonates become stable and CO<sub>2</sub> reacts with silicate minerals (Touret, 2009; Frost, 2012). Dehydration reactions are strongly controlled by temperature and form steep, univariant lines on P-T diagrams over a restricted temperature window and therefore can occur rapidly under steep thermal gradients (Poli and Schmidt, 2002; Ernst, 2005). Some dehydration reactions are discontinuous implying a focused release of fluid at particular horizons but natural, complex systems are dominated by solid solutions with extremely large compositional fields and are therefore more continuous causing fluid release over a range of P-T conditions and depth (Poli and Schmidt, 2002). The capacity of a subducted slab to produce fluid is thought to diminish with depth (Manning, 2004) and is reflected in cross-arc variations (Bebout, 2007 and references cited therein).

Metamorphism is thought to occur episodically in subduction zones in distinct stages in relation to fluid production and/or deformation events (e.g. Austrheim *et al.*, 1997 and references cited therein; Carswell and Cuthbert, 2003a). The amphibolite-facies and blueschist-facies to eclogite-facies transition at relatively shallow levels of ~50-80km represents a major dehydration horizon (e.g. Katayama *et al.*, 2006; Spandler and Hermann, 2006 and references cited therein; John *et al.*, 2008; Zhang *et al.* 2008 and references cited therein; Sheng *et al.*, 2013): it is thought that most (90%) fluids released from the oceanic crust are done so in association with eclogitisation (Philippot, 1993). The dominant release of fluids at shallower depths than the sub-arc source regions of arc magmas (below the forearc) suggests complex mechanisms are required to transport components laterally (e.g. Philippot, 1993; Poli and Schmidt, 1995; Sorensen *et al.*, 1997; Zack *et al.*, 2001).

Original volatile contents may be redistributed locally by contemporaneous dehydration and hydration reactions and transformation into volatile-bearing minerals stable at higher P-T conditions carrying water deeper into the mantle past the source region of arc magmas resulting in no considerable net expulsion (Peacock, 1990; Poli and Schmidt, 1995; Manning, 2004); these include lawsonite, high pressure amphiboles, chlorite, chloritoid, staurolite, talc, phengite and zoisite-clinozoisite (Pawley and Holloway, 1993; Philippot, 1993; Pawley, 1994; Poli and Schmidt, 1995). In mature, cold subduction zones hydrous phases can remain stable to considerable depths; for example, dense, hydrous magnesium silicates (DHMSs) can transport water deep into the mantle to the 660km discontinuity (Thompson, 1992; Philippot, 1993; Xiao *et al.*, 2002 and references cited therein; Song *et al.*, 2009 and references cited therein).

Continental crust subducted to eclogite-facies conditions is able to retain a greater amount of  $H_2O$  at HP-UHP conditions due to the stability of major hydrous phases such as phengite and does not release large volumes of fluid (Hermann, 2002; Zheng, 2009). The absence of magmatic arcs above modern day continental subduction zones also suggests not enough fluid is produced to induce partial melting in the mantle wedge to cause arc magmatism (Zheng, 2004, 2012; Zhao *et al.*, 2007). An example of this is the Indonesian Arc which is characterised by a suprasubduction zone magmatic arc to the west where oceanic crust is subducted but not to the east where continental crust is subducted (Ernst, 2005); although, the absence of arc magmatism may be due to other reasons such as the angle and rate of subduction (Zheng *et al.*, 2009).

Compositions of the fluids derived from the slab are likely to change with depth (e.g. Becker *et al.*, 2000 and references cited therein; Manning, 2004) as different metamorphic reactions form highly variable saline fluids (Liebscher, 2010 and references cited therein). Dehydration can remove water-soluble and mobile elements from the subducted slab leaving insoluble, less mobile elements and anhydrous assemblages (Jahn *et al.*, 2003; Zheng *et al.*, 2009); unless elements are retained by newly formed phases. Arc basalts are typically and distinctively enriched in large ion lithophile elements (LILE) (K, Ba, Rb, Cs) and water-soluble elements (B, Be, Li, Pb, Th, U), which are considered to be a component from the slab, and depleted in P, high field strength elements (HFSE) and heavy rare earth elements (HREE) (Schmidt, 1996; Sorensen *et al.*, 1997; Zack *et al.*, 2001; Thrane, 2002; Kessel *et al.*, 2005; Scambelluri *et al.*, 2006; Zhao *et al.*, 2007; Malaspina *et al.*, 2009; Zheng *et al.*, 2013). Carbonate phases are stable in a wide range of pressure-

temperature-composition (P-T-X) conditions so CO<sub>2</sub> is not greatly transferred from slab to overlying mantle as a volatile species (Poli and Schmidt, 2002). Changing compositions related to progressive element loss and partial melting are reflected in cross-arc variations (Bebout, 2007 and references cited therein); for example, K and light rare earth elements (LREE) increase in arc magmas with increasing depth of the subduction zone below the arc (Schmidt, 1996 and references cited therein).

#### 1.4.1. Fluid Phases

'Fluid' is a word with varying scientific and non-scientific meaning. In this thesis the term 'fluid' is used to refer to a phase with fluid movement i.e. a liquid or vapour/gas with a non-specified composition or viscosity. 'Aqueous fluids' are low viscosity liquids with compositions dominated by H<sub>2</sub>O and solute contents of only a few wt% (Kessel *et al.*, 2005). 'Silicate melts' are more viscous liquids with compositions dominated by H<sub>2</sub>O and solute solute contents of only a few wt% (Messel *et al.*, 2005). 'Silicate melts' are more viscous liquids with compositions dominated by a silicate liquid with water contents of less than 15wt% (e.g. Ferrando *et al.*, 2005; Kessel *et al.*, 2005).

The solubility of silicate ions in aqueous fluids increases significantly with increasing pressures through silica polymerisation (e.g. Molina et al., 2002; Manning, 2004; Frezzotti et al., 2007 and references cited therein). At high pressures (>1GPa) the mutual solubility between aqueous fluids and silicate melts continues to increase until there is no physical difference between them (Liebscher, 2010) and partition coefficients converge to unity (Bureau and Keppler, 1999) (see Fig. 1.4). Complete miscibility occurs above a critical curve (e.g. Hermann et al., 2006; Liebscher, 2010) and the intersection of the water-saturated solidus and liquidus phase boundaries with the critical curve defines a critical point in the system where all water is dissolved in the melt (Bureau and Keppler, 1999; Ferrando et al., 2005 and references cited therein; Frezzotti et al., 2007; Hermann et al., 2013): this is referred to as the second critical end-point (SCEP) (e.g. Schmidt et al., 2004) (Fig. 1.4). Above the critical curve, there is no isothermal phase boundary and rocks coexist with the fluidmelt whose composition can vary in a continuous way (Ferrando et al., 2005). The location of the SCEP is greatly dependent on the composition of the system (Fig. 1.5) (e.g. Poli and Schmidt, 2002; Schmidt et al., 2004; Ferrando et al., 2005; Kessel et al., 2005); although the addition of carbonates is expected to have only a small effect



Figure 1.4: P-T-X phase diagrams for a simple binary system, rock- $H_2O$  showing the location of the second critical end-point ( $CP^2$ ) and supercritical liquids (grey shaded area). Image from Hermann et al. (2006).



Figure 1.5: P-T phase diagrams for experiments of simple systems showing the location of the wet solidus and second critical end-point. Image taken Hermann et al. (2006). a) granite- $H_2O$  system; b) different compositional systems.

(Kessel et al., 2005).

The aqueous fluid-silicate melt phase has been termed a 'supercritical fluid' (e.g. Schmidt *et al.*, 2004); not to be confused with the 'supercritical fluid' of the physics world which refers to a fluid beyond the critical point between liquid and gaseous phases and describes a change of structure but not chemical composition. To avoid confusion, in this thesis they are referred to as 'supercritical liquids' (e.g. Spandler and Picard, 2013) to reaffirm their position within the liquid phase field: it is a compositional miscibility not a phase miscibility *s.s.*. Supercritical liquids are characterised by a lower viscosity than silicate melts and higher solute content than aqueous fluids which make them effective metasomatising agents (Liebscher, 2010).

It is thought supercritical liquids exist at upper mantle P-T conditions for a wide range of compositions (Bureau and Keppler, 1999) although unmixing should occur in the shallowest parts of the mantle (Bureau and Keppler, 1999). Phase relationships between aqueous fluids and silicate melts are of crucial importance for understanding material transport and magma generation at convergent plate margins (Bureau and Keppler, 1999). Experiments to determine the stability fields and SCEPs in P-T space of different chemical systems have only been carried out on relatively simple systems (e.g. H<sub>2</sub>O-silica, H<sub>2</sub>O-NaCl, H<sub>2</sub>O-albite) (e.g. Liebscher, 2010). The location of the SCEP has been found to range from 1.6GPa in feldspathic crustal compositions to over 10GPa in ultramafic mantle compositions (Kessel *et al.*, 2005). The presence of supercritical liquids in UHP rocks is often interpreted by the presence of multiphase solid inclusions (Ferrando *et al.*, 2005; Malaspina *et al.*, 2006; Lang and Gilotti, 2007; Zhang *et al.* 2008).

# 1.4.2. Fluid Flow at UHP Conditions

In the crust, fluids are thought to be transported via two end-member mechanisms: rapid, channelised flow along pathways and pervasive flow along grain boundaries. Common pathways include fractures, faults, zones of rheological contrast and lithological contacts (e.g. Zheng *et al.*, 2009; Spandler *et al.*, 2011; Sheng *et al.*, 2013). Fractures provide pathways for focussed flow and can operate on a range of scales. Transport veins are characterised by sharp contacts with the host rock as rapid fluid flow usually prevents substantial fluid-rock interaction leading to sharp interfaces (John *et al.*, 2008 and references cited therein). This enables the fluid to retain more of its initial source composition and geochemical signature (Manning, 2004; John *et al.*, 2011). Pervasive grain boundary flow occurs when fluid films on grain boundaries are relatively thick and connected (Rubie, 1986). It is dependent on the porosity and permeability of the rock and therefore varies with depth and confining pressure as well as deformation and reaction history. It can be a considerable contributor to fluid flow in foliated rocks even at depth and is favoured by uniform rheology, high permeability and high fluid flux (Oliver, 1996). It results in a greater degree of chemical interaction with the host rock and fluid compositions reflect their integrated flow history (John *et al.*, 2011). Pervasive microfractures along grain boundaries represent the combination of these two mechanisms: grain boundaries are common sites for microcracking (Oliver, 1996).

A third mechanism, intracrystalline diffusion is also thought to operate but compared to the rates of the other mechanisms, it is a negligible or absent component of mass transport, particularly at low temperatures (<600-700°C) (e.g. Molina *et al.*, 2002). Diffusive element migration and exchange arises through chemical potential gradients due to concentration differences of elements between rock domains or wall-rock and fluid and aims to reduce chemical gradients and attain a chemical equilibrium (e.g. Carmichael, 1969; Rubbo and Bruno, 2003; Wang and Tang, 2013 and references cited therein). Grain boundary diffusion occurs when there is no free fluid phase lining grain boundaries (Rubie, 1986) but it too is not considered to be a dominant mechanism of mass transport during metamorphism (Glodny *et al.*, 2008).

Fluids travel in the direction of the principal pressure and thermal gradients: on a crustal scale, this is upwards in the metamorphic pile (Zheng, 2004). Thermally-driven fluid flow is largely related to thermal expansion (decreasing density) and relative buoyancy (e.g. Burg and Philippot, 1991; Oliver, 1996). In subduction zones this is upwards towards the slab-mantle wedge interface due to buoyancy effects (Oliver, 1996 and references cited therein). Inverted isotherms above the interface, which act against the buoyant rise of fluids, and the large directional stresses experienced may promote pooling and lateral migration (e.g. Philippot, 1993).

Fluid mobility and fluxes are limited and inefficient during both deep subduction and exhumation due to disconnected pore spaces except for during widespread devolatisation (Zanetti et al., 1999; Frezzotti et al., 2007). Deformation plays a crucial role in triggering, promoting and controlling the flow of fluids at depth dominating over thermally driven flow and can produce complicated flow patterns (Oliver, 1996). It connects isolated fluids (e.g. Mørk, 1985; Mibe et al., 2003 and references cited therein), increases the surface area for grain boundary diffusion and flow through dynamic recrystallisation, (Mørk, 1985), causes dissolution and migration of species (Philippot, 1993) and increases connectivity by enhancing crystallographic anisotropy, creating foliated fabrics (Rampone and Morten, 2001) and increasing the degree of channelisation and fluid influx (e.g. Oliver, 1996; Zack et al., 2001). Change in fluid flow behaviour in subduction zones may relate to changes in the mechanisms of deformation (Philippot, 1993). Dilatancy is a requirement for fluid movement on a range of scales (Oliver, 1996). Negative solid volume change during metamorphic reactions can also facilitate fluid flow and reduce pore-fluid pressures (Philippot, 1993). Fluids also actively affect deformation (e.g. Spandler and Hermann, 2006): for example, hydraulic fracturing or 'hydrofracturing' occurs in rocks of low permeability and high competency and is driven by high gradients in fluid pressure or greater fluid pressures than the confining, lithostatic pressures (Kostenko et al., 2002 and references cited therein; Kessel et al., 2005 and references cited therein). It is thought mafic-ultramafic lithologies can remain brittle to depths of 100km (Philippot, 1993) and hydrofracturing and dehydration embrittlement may be responsible for intermediate-deep earthquakes in the Benioff Zone (e.g. Kessel et al., 2005). Furthermore, wetted grain boundaries can reduce the strength of the crust by enabling effective grain-boundary sliding (Engvik et al., 2000; Butler et al., 2013 and references cited therein).

Flow along grain boundaries depends upon the wetting and packing characteristics of the constituent minerals (Oliver, 1996). Wetting characteristics depend on the dihedral angle ( $\theta$ ) which determines the distribution and connectivity of the fluid phase (Mibe *et al.*, 2003 and references cited therein). It is the angle of a pore between two intersecting crystal walls (Mibe *et al.*, 2003 and references cited therein). Below 60° hydrous fluid can fill the channel between the crystals and efficient percolation is possible: above 60° and pores become disconnected pockets at triple junctions and complete percolation is not possible (Mibe *et al.*, 2003). The

size of the angle depends on the crystal phases, pressure, temperature and the fluid composition (Mibe *et al.*, 2003). A change to the flow regime may also occur at depth due to changes in the dihedral angle (Mibe *et al.*, 2003) as rocks at high lithostatic pressures have angles above 60° so any free fluid remaining from eclogitisation is effectively trapped (e.g. Philippot, 1993; Kostenko *et al.*, 2002 and references cited therein; Mibe *et al.*, 2003; Spandler and Hermann, 2006 and references cited therein). Trapped fluids occur as tubes along crystal boundaries or at crystal triple junctions (e.g. Liu *et al.*, 2009; Zheng, 2012) and also as fluid inclusions, within crystal fractures and within imperfections in crystal lattices (Gong *et al.*, 2007). Undeformed eclogites could transport trapped fluids to considerable depths (e.g. Mibe *et al.*, 2003 and references cited therein). Fluid flow over cm-dm-scales at UHP conditions therefore requires permeability-creating or enhancing mechanisms and channelisation probably by some deformation-related dilation process such as hydrofracturing (Kostenko *et al.*, 2002; Kessel *et al.*, 2005): far-travelled fluids are thought to be insignificant (Oliver, 1996 and references cited therein).

Flow also requires fluid pressures to keep the gaps in the rock open and connected. Pore fluid pressures increase through fluid-producing dehydration reactions or porosity-reducing crystal growth or deformation (e.g. Burg and Philippot, 1991 and references cited therein; Philippot, 1993; Oliver, 1996). However, upon rock failure, expansion of the fluid into a larger volume reduces pore pressures (Burg and Philippot, 1991). Following the release of pressure, fluids will move freely until such a time that confining and lithostatic pressures begin to build up and close the gaps; at which point, the cycle begins again (e.g. Burg and Philippot, 1991; John *et al.*, 2008 and references cited therein). This cyclic nature of pressure build up and release results in the rapid, transient, episodic intrusion of fluids separated by periods of restricted interconnectivity (e.g. Philippot, 1993; Oliver, 1996).

## 1.4.3. Potential Fluid Sources at UHP Conditions

Different fluid sources have different influences on the timing of fluid availability and the metamorphic evolution of the slab (Engvik *et al.*, 2000). Local fluid sources are evoked for HP-UHP conditions for reasons described above. A considerable abundance of free water at HP-UHP is not envisaged at upper mantle depths because it would cause extensive melting of all lithologies (as the wet solidii for mafic and felsic rocks are remarkably close to each other at HP-UHP conditions) and this is not observed (Zheng, 2009 and references cited therein). A trapped free fluid component is not considered to be a significant source of fluids at UHP depths in continental crust due to low initial free fluid contents (e.g. Rubie, 1986; Zhao *et al.*, 2007; Zheng, 2009).

The main source of fluids at HP-UHP conditions is thought to be the breakdown of hydrous phases such as lawsonite, paragonite, phengite, serpentine, epidote-zoisite, talc, Ti-clinohumite, topaz-OH and phlogopite (e.g. Hermann and Green, 2001; Engvik *et al.*, 2007; Zhang *et al.* 2008; Zheng, 2009 and references cited therein; Zheng, 2012). This can occur on the prograde and retrograde paths (Ferrando *et al.*, 2005; Massonne, 2009) and depends on phase stability which is greatly dependent on temperature and the bulk composition (e.g. Jamtveit, 1987a; Hermann and Green, 2001). Dehydration melting of fertile components in the continental crust at HP-UHPs can occur under fluid-absent and fluid-present conditions (Patiño Douce, 2005; Auzanneau *et al.*, 2006; Zhao *et al.*, 2007; Massonne, 2009; Zeng *et al.*, 2009 and references cited therein); although fluid-absent melting occurs at higher temperatures (Tumiati *et al.*, 2007) and phases may remain metastable until the fluid-absent solidus is significantly overstepped (Patiño Douce, 2005).

There are increasing lines of evidence that host gneisses partially melt during continental crust subduction-exhumation although a range of timings has been proposed: during prograde subduction (e.g. Scambelluri *et al.*, 2006 and references cited therein), after peak pressures but still at UHP conditions (e.g. Lang and Gilotti, 2007), during the retrograde transition from UHP to HP conditions if temperatures are over 750°C (Zheng, 2009 and references cited therein) or during slab exhumation (e.g. Zhao *et al.*, 2007 and references cited therein). Phengite  $[K_2(Al,Mg,Fe)_4AlSi_7O_{20}(OH)_4]$ , a Si-rich white mica, is thought to be a key hydrous phase in the generation of fluids in continental crust at UHP conditions (e.g. Hermann and Green, 2001; Wunder and Melzer, 2003; Schmidt *et al.*, 2004; Patiño Douce, 2005; Auzanneau *et al.*, 2006; Scambelluri *et al.*, 2006 and references cited therein); although it can remain stable to over 8.5GPa (Schmidt *et al.*, 2004). The breakdown of phengite with increasing temperatures produces a K-rich fluid (Schmidt, 1996;

Auzanneau *et al.*, 2006) and various mineral products depending upon the chemical system and reactants.

Another significant source of fluid are nominally anhydrous minerals (NAMs). Hydroxyl groups (OH<sup>-</sup>) and molecular water originally present in a free fluid in the protolith are progressively dissolved into anhydrous minerals such as omphacite, garnet and rutile during prograde metamorphism as the solubility increases systematically with pressure leaving even less free fluid active at UHP conditions (e.g. Becker et al., 2000 and references cited therein; Katayama et al., 2006; Zheng, 2009; Zheng, 2012). Water is preferentially dissolved into omphacite over garnet (e.g. Thompson, 1992 and references cited therein; Katayama et al., 2006) and it is thought host gneisses are capable of storing more water than mafic eclogites at UHP conditions (Zheng et al., 2009). Estimates of minimum contents in omphacite vary up to 1300ppm structural OH<sup>-</sup> and up to 3020ppm dissolved H<sub>2</sub>O (Zheng, 2004 and references cited therein; Zheng, 2012 and references cited therein). Minimum estimates for garnet and rutile include 2470ppm and 3280ppm, respectively (Zheng, 2012 and references cited therein). In the early stages of exhumation water is thought to be released from NAMs rapidly due to an abrupt decrease in water solubility with decompression (e.g. Zheng, 2004; Zeng et al., 2009 and references cited therein): this may result in highly localised retrograde reactions or locally induce partial melting.

# 1.4.4. Fluid Composition at UHP Conditions

Compositions of internally-derived fluids are strongly controlled by dehydration and hydration reactions occurring in the immediate rock (Oliver, 1996; Svensen *et al.*, 2001) and can undergo distinct changes during progressive metamorphism (Franz *et al.*, 2001). Fluids produced through dehydration during prograde metamorphism are internally buffered with compositions in equilibrium with the host rocks (Touret and Frezzotti, 2003). The rarity of carbonates at UHP conditions indicates a subordinate role for  $CO_2$  (Frezzotti *et al.*, 2007) and carbonic fluids cannot be responsible for the metasomatic enrichment of incompatible elements in the mantle (Matsumoto *et al.*, 2000 and references cited therein); although, both H<sub>2</sub>O- and CO<sub>2</sub>-dominated inclusions have been found in eclogite-facies rocks in continental terranes (Andersen *et al.*, 1990 and references cited therein; Smit *et al.*, 2008). Crystallisation of hydrous phases increases the  $CO_2/H_2O$  in the residual fluid and may lead to the late-stage formation of a  $CO_2$ -rich fluid and formation of carbonate phases (Zanetti *et al.*, 1999; Frezzotti *et al.*, 2007). Fluids inclusions of pure N<sub>2</sub>, CH<sub>4</sub>, saline brines and mixtures have also been found in eclogites in continental terranes (Andersen *et al.*, 1990 and references cited therein; Andersen *et al.*, 1991b and references cited therein; Becker *et al.*, 1999a and references cited therein; Xiao *et al.*, 2002; Manning, 2004 and references cited therein).

Element mobility in subduction zones depends on the nature and chemistry of the fluid phase, fluid-rock ratios, fluid flow, partition coefficients and the P-T conditions of the slab (e.g. Spandler *et al.*, 2004; John *et al.*, 2008; Skora and Blundy, 2012 and references cited therein). Water-soluble and mobile elements (e.g. K, LILE, LREE, Be, B) are strongly partitioned into the fluid phase and removed efficiently from the crust during dehydration reactions and particularly by supercritical liquids (e.g. Schmidt, 1996; Becker *et al.*, 2000; Kessel *et al.*, 2005; Zhao *et al.*, 2007 and references cited therein; Skora and Blundy, 2012). Al is thought to be virtually insoluble in shallow fluids (Manning, 2004) and HFSE and HREE, particularly Nb and Zr, are considered to be immobile during diagenetic, hydrothermal alteration processes and lower P-T metamorphic dehydration (e.g. Becker *et al.*, 2000; Sheng *et al.*, 2012) due to a low solubility of middle REE (MREE), HREE and HFSE in aqueous fluids (Zheng, 2009; Sheng *et al.*, 2013). However, the assumption of immobility does not apply as these elements have clearly been mobilised under eclogite-facies conditions (e.g. Spandler and Hermann, 2006 and references cited therein; Spandler *et al.*, 2011).

It is thought the breakdown of hydrous phases particularly control the LILE trace element budget of fluids transferred into the overlying mantle and arc magma source region (e.g. John *et al.*, 2008; Zhang *et al.* 2008). They may be retained in newly formed minerals during metamorphic reactions (Becker *et al.*, 2000; Spandler *et al.*, 2004 and references cited therein; Zhao *et al.*, 2007 and references cited therein) but released if the new phases formed do not contain the crystallographic sites in which the elements can be admitted (Gresens, 1967). If LILE-hosting phases such as phengite are abundant in the continental crust, LILE may not be efficiently removed during dehydration and can be transported to UHP depths where they could characterise UHP fluids (Becker *et al.*, 2000; Hermann and Green, 2001 and references cited therein; Bebout, 2007; Zhang *et al.* 2008).

Aqueous fluids are typically enriched in water-soluble elements and contain components like salts, non-polar gases and rock components (Liebscher, 2010): they are rarely pure at elevated P-T conditions (Liebscher, 2010). Water's polar character, high dielectric constant and greater tendency to dissociate make it a substantially more powerful solvent than other volatiles (Manning, 2004; Liebscher, 2010). Dissociation, which is favoured at high pressures (Skora and Blundy, 2012), increases the abundance of chloride (Cl<sup>-</sup>) ions which promote metal-chloride complexing and transport (e.g. Svensen *et al.*, 1999; Manning, 2004). However, ligands such as Cl<sup>-</sup> may be irrelevant at UHP conditions (Frezzotti *et al.*, 2007) if their concentration is low.

Silicate melts are typically enriched in incompatible elements and water-insoluble, HFSE which are more strongly partitioned into silicate melts than aqueous fluids (Rampone and Morten, 2001; Zhao *et al.*, 2007). Polymerised solute may play an important role in controlling fluid composition at depth (Manning, 2004): the polymerisation of silica in melts enhances transport of Ca, Na, Al and nominally insoluble elements that generally have low solubility in aqueous fluids (Manning, 2004; Frezzotti *et al.*, 2007 and references cited therein).

As the compositions of supercritical liquids span those of aqueous fluids and silicate melts (Schmidt *et al.*, 2004) they have high capacities to dissolve and transport both water-soluble and water-insoluble elements and have silicate melt-like and aqueous fluid-like trace element contents (Kessel *et al.*, 2005; Zhao *et al.*, 2007; Zeng *et al.*, 2009; Zheng, 2012 and references cited therein). Their compositions are strongly dependent on phases in their source region which can retain particular elements; for example, garnet and rutile in eclogitic residues retain HFSEs and HREEs (e.g. Okamoto *et al.*, 2006; John *et al.*, 2008). Internal UHP fluids produced in continental crust are thought to be intermediate alumina-silicate solutions with low water activities (Frezzotti *et al.*, 2007). Fluids derived from phengite breakdown will be rich in alkalis, Al, Si, LILE and LREE and Mg-poor (e.g. Wunder and Melzer, 2003; Schmidt, 1996; Liu *et al.*, 2013 and references cited therein) and be high silica rhyolites (Schmidt *et al.*, 2004). Similarly, granitic melts formed by partial melting of felsic gneiss are characterised by low Fe, Mg, Ca, Cr and Ni and high Si, Na, K, LILE and LREE (Zheng *et al.*, 2009 and references cited therein).

#### 1.5. Metasomatism

Metasomatism is the chemical alteration of a rock typically by a fluid although diffusive exchange in a dry system can also occur. Most metamorphic reactions are not isochemical as mineral reactions typically involve at least volatile losses or gains (Bailey, 1982 and references cited therein) and should be considered metasomatic. Following infiltration of a reactive metasomatic agent, exchange of material occurs by intergranular and intragranular diffusion driven by chemical gradients established across the fluid-rock interface due to different element concentrations (e.g. Carmichael, 1969; McInnes *et al.*, 2001; Marocchi *et al.*, 2009, 2010; White and Powell, 2010; Wang and Tang, 2013 and references cited therein). Without chemical exchange or infiltration of the fluid far into the wall-rock, for example if wall-rocks are impermeable, there may be no significant record of fluid infiltration (Oliver, 1996). Furthermore, fluids buffered by the rock composition may be in chemical and stable isotopic equilibrium with their immediate surroundings and so leave little trace of their presence (e.g. Oliver, 1996; Zanetti *et al.*, 1999). Metasomatism by internal fluids often occurs at lithological contacts where a change in bulk composition occurs and fluid flow is channelised (e.g. Zhao *et al.*, 2007; Marocchi *et al.*, 2010).

The degree of metasomatism is affected by the fluid-rock ratio, the degree of chemical disequilibrium and element concentration profiles, fluid mobility, the duration of fluid-rock interaction and the rate of diffusion (e.g. Carmichael, 1969; Touret, 2009; Kiseeva *et al.*, 2012) and will be greatest where chemical gradients are steep and prolonged (e.g. Tumiati *et al.*, 2007). The fluid may be rapidly consumed forming volatile-bearing phases (Philippot, 1993) preventing extended periods of fluid-rock interaction. Furthermore, the crystallisation of metasomatic phases at fluid-rock interfaces reduces the chemical difference across the contact and reduces porosity and permeability (Bodinier *et al.* 2008) which both work to reduce further metasomatic interaction.

Two types of metasomatism are defined: modal metasomatism involving the crystallisation of new minerals; and cryptic metasomatism involving chemical change of present minerals and is not readily visible in the field unless changes in colour are produced. Complex zoning and chemical heterogeneity are common in metasomatic minerals indicating a spectrum of replacement conditions occur (Bailey, 1982).

Modal metasomatic features include planar veins, veinlets and irregular segregations. They are usually associated with structures which facilitate fluid infiltration. Alteration haloes along fractures and rinds on blocks demonstrate the outward migration of infiltrating fluids into the wall-rock through diffusion (e.g. Oliver, 1996; Bebout, 2007 and references cited therein; Milke et al., 2011 and references cited therein). Altered rocks can be characterised by incomplete reaction textures, replacement and inclusion of original crystals in new phases, interstitial volatile-bearing phases, reaction coronas and chemical zoning (Oliver, 1996). Partially altered lithologies provide the clearest fingerprint of the geochemical alteration attributable to subduction zone metamorphism (Bebout, 2007). Metasomatism changes both the composition of the rock and the fluid as chemical exchange progresses and an equilibrium is approached. Hot, metasomatic fluids can also thermally erode or resorb solid minerals locally forming a mixed fluid and diluting the original component (e.g. Simon and Chopin, 2001; Manning, 2004; Wang and Tang, 2013). The distribution of highly incompatible elements is particularly affected by fluid-rock interactions (e.g. Takahashi, 1992). Compositional evolution of the fluid-rock system can produce lateral zones of different assemblages as even though chemical gradients may be smooth, mineralogical changes are abrupt (Rivalenti et al., 1997). Alternatively, a zoned appearance may result from fractures or other planes of weakness being repeatedly used by fluids with different compositions with a progressive metasomatic sequence being established (e.g. Burg and Philippot, 1991; John et al., 2008 and references cited therein). Furthermore, zonation may result from the different diffusion rates of elements with slowdiffusing elements such as Si, Al and Ti forming phases closer to the fracture (Mørk, 1986 and references cited therein; Proyer et al., 2009). Zoned veins have a symmetrical pattern if the wall-rock is the same either side of the fluid pathway.

Phlogopite and amphibole are minerals considered diagnostic of mantle metasomatism (Prouteau *et al.*, 2001) and an enrichment of mobile, incompatible elements such as LILE is often considered to be a geochemical signature of metasomatism by fluids (Malaspina *et al.*, 2006; Song *et al.*, 2009). However, if metasomatic phases formed do not host such trace elements, such as orthopyroxene, they remain in the fluid (e.g. Malaspina *et al.*, 2006). Al is considered as the least mobile of the major components of metasomatic systems (Carmichael, 1969) although the limit of its migration may be larger at higher P-T conditions (Carmichael, 1969).

### 1.5.1. Fluid-Rock Interaction at UHP Conditions

It is thought that fluids expelled from subducted sediments and the oceanic crust during the early stages of subduction metasomatise the overlying lithospheric mantle wedge forming hydrous phases such as phlogopite and amphibole and pyroxene-rich lithologies in lateral bands along the slab-mantle wedge interface or forming a sequence of cumulate layers along the walls of fluid channels (e.g. Sekine and Wyllie, 1983; Peacock, 1990; Becker *et al.*, 1999b; Prouteau *et al.*, 2001; Malaspina *et al.*, 2006; Zheng, 2012). The first products of metasomatism, close to the slab-mantle wedge interface, should be rich in trace elements compatible in the crystallising metasomatic phases (Manning, 2004). Remaining fluid, with a partial compositional signature of the slab, is not expected to migrate far into the mantle wedge due to water-consuming metasomatic reactions (Manning, 2004; Kiseeva *et al.*, 2012 and references cited therein).

A similar process of fluid-mantle interaction is evoked for fluids expelled at UHP conditions providing pathways exist along which the fluids can flow and fluid is abundant enough to reach the slab-mantle interface. The absence of arc magmatism above continental subduction zones does not necessarily mean that the mantle wedge is not metasomatised by fluids but that partial melting is not induced or magma does not reach the surface.

Considerable chemical exchange and extensive reaction zones are likely to occur at the interface between felsic crustal rocks and fluids and mafic-ultramafic rocks in the crust or mantle wedge due to their drastically different bulk compositions inducing change of their petrophysical properties (e.g. Jamtveit, 1987a; Massonne, 2009; Marocchi *et al.*, 2009, 2010). For example, orogenic, mantle wedge garnet peridotites in the Ulten Zone, eastern Alps, interacted with a hydrous partial melt generated in the hosting migmatised continental rocks forming a compatible element-rich, orthopyroxene-rich layer and residual aqueous fluid (Scambelluri *et al.*, 2006) with contacts marked by mica-rich aggregates and amphibole-phlogopite rocks known as "blackwalls" and an abundance of accessory phases (Marocchi *et al.*, 2009, 2010). Another example is metasomatism of the Zhimafang mantle peridotite (Jiangzhuang, China) in the Sulu UHP terrane by a silica-rich melt during exhumation causing distinct enrichment of LILE, LREE, Zr and K and growth of zircon and volatile-bearing

minerals (Zhang *et al.*, 2011). Hydrous, silica-rich fluids interacting with and refertilising the mantle wedge might produce, depending on the P-T conditions, orthopyroxene, phlogopite, clinopyroxene, pyrope-rich garnet, zircon, Ti-clinohumite and/or apatite and consume olivine (Sekine and Wyllie, 1983; Rapp *et al.*, 1999; Kiseeva *et al.*, 2012; Zheng, 2012). Successive layers might include monomineralic phlogopite-rock (glimmerite) and combinations of phlogopite-pyroxenes and quartz (Sekine and Wyllie, 1983). Hybridised, residual liquids are thought to be high Mg# adakitic liquids (Sekine and Wyllie, 1983; Rapp *et al.*, 1999).

The relatively geologically-rapid rates of subduction and exhumation of HP-UHP continental rocks suggests interaction of exhumed terranes with mantle fluids may be limited (Shui *et al.*, 2006). A mantle-derived component in veins in HP-UHP rocks from Donghai and the Chinese Continental Scientific Drilling (CCSD) Project is nearly negligible (Shui *et al.*, 2006). However, continental crust which is not exhumed is envisaged to interact considerably with the surrounding mantle.

Mantle xenoliths indicate there are major metasomatising agents in the bulk of the mantle away from subduction zones (Thompson, 1992). Dramatic modal metasomatism producing high modal orthopyroxene is often observed in xenoliths from kimberlites in African and Siberian cratons (Zanetti *et al.*, 1999 and references cited therein; Kiseeva *et al.*, 2012 and references cited therein). Mica-amphibole-rutile-ilmenite-diopside (MARID)-type metasomatism is thought to be related to kimberlitic fluids (Lorand and Gregoire, 2010 and references cited therein). Fe-Ti oxides are not normal components of the shallow upper mantle source of basalts (Lorand and Gregoire, 2010) and must result from metasomatism of Fe-Ti-rich, K-rich, H<sub>2</sub>O-rich liquids (Delaney *et al.*, 1980). Xenolith suites suggest the subcontinental lithospheric mantle is metasomatised by enriched, K-rich, silica-rich fluids (e.g. Kiseeva *et al.*, 2012).

## 1.5.2. Vein Formation

Veins represent products of fluid-rock interaction (Spandler and Hermann, 2006) precipitated from saturated solutions (Zhang *et al.* 2008 and references cited therein) and are usually inferred to represent the pathways of compositionally distinct fluids with diffusional exchange into the wall-rock (Oliver, 1996). However, metamorphic

segregation via diffusional creep and short-range mass transport of internally-derived fluids can also form veins (Philippot, 1993) and pseudotachylyte eclogite-facies veins are thought to form by partial melt-inducing frictional heating during faulting (Hwang *et al.*, 2001). Metamorphic segregation veins containing material derived from their adjacent wall rock are abundant in all types of eclogite from the European Alps (Philippot, 1993).

The assemblage of a vein does not directly reflect the composition of the fluid but the components left behind after reaction with the wall-rock (Hermann *et al.*, 2013). Large, euhedral crystals containing abundant primary fluid inclusions and only cocrystallising phases are thought to have crystallised freely from the fluid (Zhang *et al.* 2008) and phases concentrated along the edge of veins crystallised early/first (Lorand and Gregoire, 2010).

Determining the absolute timing of metasomatism and vein formation is difficult as the timing of fluid infiltration may not correspond to the final equilibration of new minerals to their ambient P-T conditions. Many veins in HP-UHP terranes are thought to form during eclogitisation (Becker *et al.*, 1999a; John *et al.*, 2008). These are surrounded by dehydration haloes and leach zones (John *et al.*, 2008 and references cited therein). Partial melting of eclogitic mafic-ultramafic bodies is not thought to be a mechanism of vein formation as equilibration conditions are below the watersaturated solidus of eclogite and quartz and hydrous phases, present in many eclogites, are unlikely to remain in the residue during melting (Becker *et al.*, 1999a; Spandler *et al.*, 2011).

Quartz veins are a common feature in metamorphic rocks (Franz *et al.*, 2001). They are interpreted to be products of mineral precipitation from a significant flux of aqueous fluids (Sheng *et al.*, 2012) and to form syn-metamorphically under high fluid pressures caused by prograde dehydration, bulk volume reduction during eclogitisation (Franz *et al.*, 2001) or during decompression during exhumation (e.g. Zheng, 2004; Shui *et al.*, 2006). Veins produced during exhumation may be associated with the breakdown of hydrous phases or exsolution of structural hydroxyl and molecular water from NAMs (Zhao *et al.*, 2007 and references cited therein) and the decripitation of fluid inclusions (Zheng, 2004) and may contribute to quartz veining (Zhao *et al.*, 2007 and references cited therein).

#### 1.6. Western Gneiss Region

The WGR of western Norway delineates the exposure of the rocks of the WGC (*Fig. 1.6*) previously known as the Basal Gneiss Region and Basal Gneiss Complex (Carswell, 1973 and references cited therein). The WGC is a composite tectonometamorphic terrane on the outermost, western part of the Fennoscandian Shield, the northwest segment of the Baltic Shield and is the lowest structural level (root zone) of the Scandinavian Caledonides (*Fig. 1.7*) (e.g. Cuthbert *et al.*, 1983; Andersen and Jamtveit, 1990; Cuthbert *et al.*, 2000; Andersen and Austrheim, 2008). It is exposed in a tectonic window through the fartravelled nappes of the Scandinavian Caledonides (Caledonian Allochthon) (*Fig. 1.6*) which were emplaced onto Baltica during the Caledonian Orogeny following closure of the Iapetus Ocean between Baltica and Laurentia in the early Palaeozoic.

#### 1.6.1. <u>Geographical Characteristics</u>

The Baltic Shield forms Norway, Sweden, Finland, northwest Russia, northern Poland and underlies the Baltic Sea (Bingen *et al.*, 2008). The Scandinavian Caledonides are a ~2000km-long belt stretching from Finnmark, north of Arctic Circle, to Stavanger in southern Norway and have a maximum width of 400km from hinterland to foreland (Wilks and Cuthbert, 1994 and references cited therein; Fossen, 2000; Nordgulen *et al*, 2008). The WGC has an estimated geographical extent of at least 50,000km<sup>2</sup> (Kylander-Clark *et al.*, 2008). Its eastern edge runs along the Jostedal Mountains in central Norway (e.g. Roberts and Gee, 1985; Nordgulen *et al.*, 2008) and plunges below the Trondheim Nappe Complex to the north and Jotun Nappe Complex to the south (Koenermann, 1993) (see *Fig. 1.8*). It is one of the best preserved, best exposed, largest and accessible deep crustal provinces in the world: its orogenic setting has been well established and it is well-known for exposures of stunning eclogites and garnetiferous peridotites (Cuthbert *et al.*, 2000; Walsh *et al.*, 2007; Andersen and Austrheim, 2008). It has a glacial landscape and the terrain is characterised by dense forests, rocky shores and steep-sided fjords.

## 1.6.2. <u>Geological History</u>

The Baltic Shield (or the East European Craton), formed and evolved over three billion years from the mid Archaean to the Palaeozoic particularly during orogenies



*Figure 1.6: Map of Norway showing main geological units. Amended from NGU (2014).* 



Figure 1.7: Diagrammatic reconstruction of the three-armed system of the Caledonian Orogeny (Woudloper, 2008) formed through E-W collision of Laurentia and Baltica and N-S convergence with Avalonia forming Laurussia (Euramerica). Laurussia collided with Gondwana during the Hercynian-Variscan Orogeny in the final formation of the Mesozoic supercontinent Pangaea (e.g. Piper, 1985a).

such as the Svecofennian (1.9-1.75Ga), Gothian (1.75-1.5Ga) and Sveconorwegian (Grenvillian) (1.3-0.9Ga) and associated magmatic and plutonic events involved in the formation and break-up of supercontinent Rodinia and opening of the lapetus Ocean ca. 700Ma (e.g. Mearns, 1986 and references cited therein; Roberts, 2003; Bingen *et al.*, 2008; Lapen *et al.*, 2009).

The Caledonian Orogeny (*Fig. 1.7*) resulted from oblique convergence between Baltica and Laurentia (e.g. Ziegler, 1985; Torsvik *et al.*, 1996; Roberts, 2003; Bottrill *et al.*, 2014) following the anticlockwise rotation of Baltica away from Siberia to face Laurentia in the Lower Palaeozoic (Roberts, 2003; Brueckner and van Roermund, 2004). Ocean closure occurred mostly via the subduction of the Baltic Plate underneath the Laurentian Plate producing short-lived volcanic arcs, I-type plutonic suites and basins off the Laurentian continental margin which were later incorporated into the units of the Caledonidian Allochthon (Higgins *et al.*, 2004). Five main stages of the Caledonian Orogeny have been defined with an age range of 500Ma to ~390Ma. It was a Himalayan-scale orogen and formed a major suture in the Pangaean landscape (Piper, 1985b; Andersen *et al.*, 1991a; Higgins *et al.*, 2004). Its structure is similar to that of the Alps; dominated by thrust-nappe tectonics (Griffin and Brueckner, 1980; Roberts and Gee, 1985) and extends to western and central Europe (Ziegler, 1985).

The main collisional and last stage, the Scandian Orogeny, began in the late Silurian, ca. 438-435Ma and continued until the mid Devonian (ca. 390Ma) and consisted of various contractional and extensional phases (e.g. Nordgulen *et al.*, 2008). Scandian structures and metamorphism dominate the Scandinavian Caledonides having overprinted most previously formed structures and parageneses (Terry and Robinson, 2003).

The WGC experienced short-lived, northwestwards, deep subduction to UHP conditions underneath the Laurentian Plate during the Scandian and was exhumed during the late Devonian in the footwall of a system of low-angle, normal detachment faults such as the Nordfjord-Sogn Detachment (*Fig. 1.8*) during late-post orogenic extension (e.g. Cuthbert, 1991; Koenermann, 1993; Cuthbert *et al.*, 2000; Carswell and Cuthbert, 2003b). Its basal contact is not exposed and therefore its

structural context and thickness is not known. There is a high variation in the extent of structural and mineralogical reworking of pre-exhumation (e.g. eclogite-facies) structures due to heterogeneous deformation and the partitioning of later strain which resulted in high- and low-strain domains (Koons *et al.*, 2003; Engvik *et al.*, 2007). Late folding formed large (20-30km-diameter), imposing domes folding the entire crust including the Scandian basal thrusts and detachments, metamorphic isograds and earlier structures (Koenermann, 1993). The folds expose UHP rocks in their antiformal culminations and host greenschist-facies molasse basins, such as the Hornelen Basin (*Fig. 1.8*), in their synforms (e.g. Krabbendam and Wain, 1997; Hacker, 2007). The latest structures in the WGR record deformation associated with post-Caledonian normal, extensional faults preserved in Permian graben-rift structures and which offset Devonian unconformities (Bockelie and Nystuen, 1985; Hartz *et al.*, 1994).

## 1.6.3. Lithologies and Metamorphism

The WGC is a predominantly amphibolite-facies terrane (equilibrated to 650-800°C, 0.5-1.5GPa (Peterman et al., 2009)) with metastable, igneous and HP-UHP metamorphic bodies preserved in low-strain zones (e.g. Krabbendam and Wain, 1997; Austrheim et al., 2003; Robinson et al., 2003). It is estimated 95% of the exposed crust transformed to an amphibolite-facies paragenesis during Scandian exhumation (e.g. Cuthbert et al., 2000; Engvik et al., 2007). Only weak, local greenschist-facies overprints are evident to the East (Brueckner, 1977; Terry and Robinson, 2003). It consists of reworked, Proterozoic-Phanerozoic, autochthonous/para-autochthonous and locally imbricated, basement orthogneisses and interfolded, supracrustal, late Proterozoic, metasedimentary cover (Cuthbert and Carswell, 1990; Carswell and Cuthbert, 2003b; Vrijmoed et al., 2006; Walsh et al., 2007; Hacker et al., 2010). It is estimated 90% of the WGC is composed of migmatitic, banded gneisses with a granodioritic composition (e.g. Krabbendam and Wain, 1997). The upper continental crust is thought to be composed of granitic-granodioritic magmatic rocks (Zheng, 2012 and references cited therein) suggesting the WGR forms part of upper crust of Baltic Shield. The remaining 10% contains a wide variety of minor lithologies including metamorphosed anorthosites, large syeniticmonzonitic (mangerites) and other plutonic, acidic intrusions; bands and lenses of calc-silicate and pelitic paraschists and paragneisses, marbles and quartzites; and



Figure 1.8: Simplified tectonic map of the Western Gneiss Region and surrounding area. Image modified from Brueckner and Cuthbert (2013). T = Tafjord enclave. Thin dotted lines represent exhumation-related shear sense trajectories. Thick solid lines are approximate positions of extensional faults. HSZ = Hardangerfjord Shear Zone; KF = Koppera Fault; LGFZ = Lærdal-Gjende Fault Zone; OF = Olestøl Fault; RF = Røranger Fault. Thick dashed lines are approximate locations of exhumation-related detachments. AD = Åre Detachment; BASZ = Bergen Arcs Shear Zone; JDZ = Jotunheimen Detachement Zone; MTFZ = Møre Trondelag Fracture Zone; NSD = Nordfjord-Sogn Detachment Zone; RD = Rørange Detachment.

mafic-ultramafic sheets and bodies (e.g. Koenermann, 1993; Cuthbert *et al.*, 2000 and references cited therein). The WGC is characterised by a regional gradational change in both its structural affinity to Caledonian-related deformation (e.g. Hacker *et al.*, 2010) and subduction-related metamorphic P-T conditions increasing from the southeast to the northwest (e.g. Cuthbert *et al.*, 2000) (*Fig. 1.9*).

Eclogite-facies conditions have a regional distribution occurring in mafic-ultramafic and felsic lithologies and have Caledonian, mostly Scandian ages (Krabbendam and Wain, 1997; Brueckner and van Roermund, 2004; Hacker, 2007). It is proposed that a large proportion of the country rocks did not transform to an eclogite-facies assemblage as there is a widespread lack of evidence for densification (Peterman *et al.*, 2009 and references cited therein); although this maybe be a result of strong overprinting (Peterman *et al.*, 2009). The rare eclogite-facies, felsic rocks are invariably adjacent to or near mafic-ultramafic bodies or interlayered with eclogite (Peterman *et al.*, 2009 and references cited therein). Partial melting appears to have accompanied decompression but some melting at UHP conditions cannot be ruled out (Butler *et al.*, 2013 and references cited therein). It appears to be most intense in the northern WGC (Auzanneau *et al.*, 2006 and references cited therein) but may not be related to Scandian exhumation and decompression (e.g. Gordon *et al.*, 2011); although Scandian ages are recorded in zircons in migmatitic leucosomes near UHP eclogites (Gordon *et al.*, 2011).

Eclogitic mafic bodies hosted directly in the country rocks ('external eclogites') occur as variably-sized (centimetre- to kilometre-scale) lenticular pods, tabular sheets and isolated boudins and are distributed throughout the WGC often in linear belts along particular horizons and folded within the major structures (Lappin and Smith, 1978; Griffin and Brueckner, 1980; Krabbendam and Wain, 1997; Cuthbert *et al.*, 2000 and references cited therein; Austrheim *et al.*, 2003; Carswell and Cuthbert, 2003a). They appear to be associated with anorthosite horizons and supracrustal units in which they are more numerous (Cuthbert *et al.*, 1983; Griffin, 1987). It is thought the bodies originally formed large, low pressure, layered mafic-ultramafic intrusions and mafic (doleritic) dyke swarms and sills associated with prior episodes of crustal formation and continental break-up (e.g. Griffin and Brueckner, 1985; Mearns, 1986 and references cited therein; Griffin, 1987; Jamtveit, 1987b and references cited therein; Koenermann, 1993; Andersen *et al.*, 1994; Austrheim *et al.*, 2003; Engvik *et* 



Figure 1.9: Map of the WGC showing UHP domain, the regional temperature and pressure gradients and age data. Image modified from Kylander-Clark et al. (2008) with data from Hacker et al. (2010). Blue dashed lines are approximate, "poorly constrained" locations of isobars based on eclogite pressures from Hacker et al. (2010), Figure 1. Circles are approximate locations of mineralogical indicators of UHP metamorphism from Vrijmoed et al. (2008), Figure 1. Light green circles = polycrystalline quartz; bright green circles = coesite; red circles = microdiamond; purple circle = majorite garnet.

*al.*, 2007) and as volcaniclastic sediments and layers in supracrustal sequences (Griffin, 1987; Jamtveit, 1987a,b and references cited therein). Bodies are generally little affected by strain showing only weak lineations and foliated margins (Andersen *et al.*, 1994). Fabrics in the surrounding gneisses generally truncate eclogite-facies structures although some are concordant (Krabbendam and Wain, 1997). The mineralogy of eclogitic bodies can vary considerably within a single body and modal banding is thought to reflect original magmatic variation (Andersen *et al.*, 1994; Cuthbert *et al.*, 2000). <sup>87</sup>Sr/<sup>86</sup>Sr isotope ratios of bodies suggest they have been pervasively contaminated by crustal fluids (Brueckner, 1977; Griffin and Brueckner, 1985). Bodies are seen to retrogress completely or partially (depending upon their size and deformation) to symplectitic eclogite and eventually to garnet-free amphibolite (e.g. Krabbendam and Wain, 1997).

Garnet peridotite bodies and their retrogressive equivalents are rare but do occur throughout the WGC and typically in linear belts or concentrations such as the Nordfjord area where they decorate basement-cover contacts (e.g. Cuthbert *et al.*, 2000 and references cited therein; Reverdatto *et al.*, 2008). They occur in the same tectonostratigraphic units as the eclogitic bodies and are seemingly spatially associated with them (van Roermund and Drury, 1998; Lapen *et al.*, 2009). Two types were distinguished by Carswell *et al.* (1983): Fe-Ti-rich crustal peridotite and Mg-Cr mantle-derived peridotite. Garnet websterite lithologies are associated with both eclogite and garnet peridotite bodies.

The UHP terrane consists of three discrete UHP domains named Nordfjord-Stadlandet, Sørøyane and Nordøyane within the most coastal portions of the WGR exposed in antiformal culminations (*Fig. 1.9*) (e.g. Cuthbert *et al.*, 2000). They are thought to be linked beneath the intervening HP domain (e.g. Root *et al.*, 2005). Its size is estimated at ~5000km<sup>2</sup> with a width of <40km (Cuthbert *et al.*, 2000; Carswell and Cuthbert, 2003b). The southern boundary of the UHP terrane is well defined along the north shore of Nordfjord (e.g. Krabbendam and Wain, 1997; Vrijmoed *et al.*, 2006; Hacker, 2007; Andersen and Austrheim, 2008; Kylander-Clark *et al.*, 2008) however its northern and eastern limits are not: their boundaries are subject to change with new discovery of UHP metamorphism. It lies structurally below the eclogite-facies HP terrane, represents the deepest structural levels exposed in Norway and of the Scandinavian Caledonides (Hacker, 2007; Andersen and Austrheim, 2008) and is at least 20km-thick; its lower contact is not exposed so the true thickness or character of the contact is unknown (e.g. Root *et al.*, 2005). All three domains have similar P-T conditions but do show an increase northwards: the northern Nordøyane domain is the only domain where diamond has been observed (*Fig. 1.9*) suggesting it was the most deeply subducted (Carswell and Cuthbert, 2003b; Hacker, 2007).

Cores and rims of zircons from eclogites across the study area yield Precambrian and Caledonian (440-400Ma) ages, respectively (e.g. Walsh and Hacker, 2004; Walsh *et al.*, 2007). It is not clear whether the spread of Caledonian ages represent discrete or simultaneous events as the errors are too large to resolve any detail (Walsh *et al.*, 2007). Peak, Scandian UHP conditions are thought to have occurred ca. 410-400Ma and is estimated to have reached 3.6GPa, 750-800°C (Carswell *et al.*, 2003a; Hacker, 2007; Walsh *et al.*, 2007; Hacker *et al.*, 2010). Cooling ages are younger in the west than in the east (Kylander-Clark *et al.*, 2008) representing later exhumation of deeper rocks and are not simultaneous along the orogenic belt due to the oblique nature of collision.

Subduction-related fluids are thought to have had two dominant compositions: N<sub>2</sub>-CO<sub>2</sub>-H<sub>2</sub>O fluids (which developed into residual N<sub>2</sub>-CO<sub>2</sub> fluids) and saline brines (Svensen *et al.*, 2001 and references cited therein). There appears to be no systematic regional pattern in their distribution (Svensen *et al.*, 2001 and references cited therein). Peak conditions are associated with N<sub>2</sub>-bearing aqueous fluid and CO<sub>2</sub>rich fluid is associated with later retrogression (Andersen *et al.*, 1990 and references cited therein).

## 1.6.4. Overview of Historical Research

Eclogites in the WGR were first mentioned by Irgens and Hiortdahl (1864) and Reusch (1877) (Krabbendam *et al.*, 2003 and references cited therein). Pioneering work was carried out in the late 19<sup>th</sup> and early 20<sup>th</sup> centuries by C.F. Kolderup on the tectonostratigraphy of the Scandinavian Caledonides (Andersen and Austrheim, 2008 and references cited therein) and by P. Eskola who published a now landmark paper "On the occurrence of eclogites" in 1921 (Eskola, 1921) which included detailed descriptions of eclogites in the WGR (Krabbendam *et al.*, 2003). In the 1960s research

was focused on the Nordfjord area with detailed geological mapping and descriptions of lithology and structure such as presented by Bryhni (1966) and Lappin (1966), respectively (Krabbendam et al., 2003). In the 1980s, research was very much focused on the mechanisms of formation of the Devonian molasse basins, the structure of the WGR, the emplacement of the nappes, and the general tectonic history of the Scandinavian Caledonides (e.g. Cuthbert et al., 1983; Roberts and Gee, 1985; Seranne and Seguret, 1987). As mentioned, the first discovery and confirmation (by Raman spectroscopy) of coesite in the WGR was made by Smith (1984). Since, much like other terranes, research has centred on defining the extent of the UHP terrane which has been regularly expanded, the origin of the maficultramafic bodies, the coherency and tectonic structure of the HP-UHP terrane and defining the regional trend in P-T conditions. During the 1980s-1990s there was intense debate over the origin of the HP-UHP garnet peridotite and eclogite bodies between an *in situ* origin or tectonic emplacement into a lower pressure terrane (e.g. Smith 1980, 1984; Cordellier et al., 1981; Gebauer et al., 1985). The Geological Survey of Norway (NGU) carried out comprehensive mapping of the Nordfjord region in the 1990s (Krabbendam et al., 2003) and the first discovery and confirmed identification of microdiamond in the WGR was by Dobrzhinetskaya et al. (1995). In the 2000s, focus changed to characterising HP-UHP mineralogy and metamorphism, better constraining metamorphic isograds, UHP domains and metamorphic transitions. The first geothermobarometric estimations of metamorphism in the WGR were widely divergent due to complex petrography, unsuitable calibrations and erroneous extrapolations (e.g. Cuthbert et al., 2000).

Previous research into fluid-rock interaction is well established in the WGR although not especially at UHP localities. Vein systems, shear zones and other fluid-rich environments have been studied in detail with respect to eclogite-facies metamorphism, deformation and reaction kinetics (e.g. Austrheim, 1987; Andersen *et al.*, 1991b; Wain *et al.*, 2001). Vein systems such as those present in the WGR and studied in this thesis are rare and there is no report of similar features in other UHP terranes.

## 1.7. Previous Research of Continental UHP Terranes

Over the last three decades the number of crustal UHP rocks identified worldwide has increased and the awareness of their importance and significance in understanding crustal, orogenic and mantle processes has grown (e.g. Kylander-Clark *et al.*, 2009). Rocks from UHP terranes all over the world are being studied to expand our knowledge of these terranes and address the lack of data with which to answer questions. The compilation and comparison of data has aided the characterisation of UHP terranes. Considerable research is currently being undertaken in China on rocks of the Dabie-Sulu orogenic belt particularly due to the CCSD Project which provides a core through the terrane.

Previous and recent study of has focused on a number of research fields and topics including: finding mineralogical evidence of UHP metamorphism to define the areal extent of UHP domains and determine the scale and coherency of subduction-exhumation; structural studies to constrain the tectonic model of continental terrane subductionexhumation; petrological studies to constrain the metamorphic evolution, preservation and retrogression of the rocks; geochemical studies to determine the characteristics and evolution of typical and chemically unusual rocks; petrological and thermobarometric studies to estimate the peak P-T conditions experienced; geochronological studies using a range of isotopic systems (e.g. U-Pb, Lu-Hf, Sm-Nd, Rb-Sr, Re-Os, Ar-Ar) to determine the timing and duration of UHP metamorphism, constrain the pressure-temperature-time (P-Tt) paths followed and determine exhumation rates; fluid inclusion studies to determine the provenance of fluids and parameters of fluid-rock interaction; solid phase inclusion studies in minerals such as zircon, coesite and diamond to understand the chemical environment and assemblages during peak conditions; stable and radiometric isotopic studies (e.g. O and Sr isotopes) to determine fluid provenance, the extent of fluid-rock interaction and the areal extent of UHP domains; and chemical, structural and experimental studies to define new UHP phases and structures.

The research field also caused the expansion of experimental research and modelling of crustal rocks to higher pressure conditions such as models of subduction, the physics and structures of crustal minerals, experimental phase relations, deformation mechanisms, development of new thermobarometers and thermobarometric calibration equations, thermodynamic datasets and the kinetics of metamorphic reactions and fluid flow at such
conditions. This type of research supports the study of the rocks and enables different analytical tools to be used.

#### 1.8. Reasons for Researching Continental UHP Terranes

Subducted-exhumed, continental UHP terranes provide us with rocks which preserve the highest pressure conditions that crustal rocks are known to experience and are therefore a very valuable resource for geological research. Their study can provide us with insight into Earth systems and processes at upper mantle depths, enable us to constrain and inspire tectonodynamic models of subduction-exhumation and continental collision and expand our knowledge and understanding of the deep subduction of continental crust, the evolution of subduction zones and continents and the generation and mechanisms of deep earthquakes (e.g. Griffin, 1987; Mibe *et al.*, 2003; Korsakov *et al.*, 2004; Manning, 2004; Vrijmoed *et al.*, 2008; Zhang *et al.* 2008; Zhang *et al.*, 2010; Zheng, 2012). They are, however, not perfect windows as UHP assemblages, textures and structures in the majority of the terranes are pervasively overprinted during exhumation: most insight into deep level processes is obtained through the study of the mafic-ultramafic bodies hosted within the retrogressed felsic rocks, which better preserve the HP-UHP conditions experienced (e.g. Wain *et al.*, 2000).

Such bodies provide us with a greater structural and lithological context than mantle xenoliths, our other geological insight into mantle conditions and processes, as their larger-scale characteristics and relationships are preserved in the field at outcrop scale. Furthermore, xenoliths are rarely sourced from subduction zones or source zones of arc magmas (Sorensen *et al.*, 1997) and so they provide us with geological evidence of a different part of the mantle of which there is little geological evidence at the Earth's surface (e.g. Trønnes, 2010). The exhumed terranes also provide us with our best insight into the fate of subducted oceanic and continental crust that remains in the mantle (e.g. Hermann, 2002; Patiño Douce, 2005; Bebout, 2007). The amount of un-exhumed continental crust is not known so the importance and extent of its influence is not constrained (e.g. Liou and Tsujimori, 2013). The contacts between mafic-ultramafic bodies and host felsic lithologies are considered to be a lithologically close analogue of the crust-mantle wedge interface during deep subduction of the continental crust (e.g. Marocchi *et al.*, 2009). The study of such rocks therefore yields vital information regarding elusive metasomatic interaction between the lithologies at depth (e.g. Spandler and Hermann, 2006).

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There is also little known of the transport, behaviour, availability physiochemical parameters and compositional evolution of fluids during deep subduction and in mantle systems and little understanding of the recycling, distribution, sources, transfer, fluxes, compounds and controls of elements or compounds of socioeconomic and geological interest (e.g. Thompson, 1992; Becker *et al.*, 1999a, 2000; Zack *et al.*, 2001; Molina *et al.*, 2002; Spandler *et al.*, 2004, 2011; Patiño Douce, 2005; Zhang *et al.* 2008; Sumino *et al.*, 2010; Zheng, 2012; Hermann *et al.*, 2013). Knowledge of such processes are crucial to understanding the Earth's evolution and development of accurate geodynamic models as they have implications for arc magmatism, the global recycling and behaviour of volatile and trace elements, the evolution of the mantle and mantle wedge and the tectonics of collisional margins which they actively influence (e.g. Becker *et al.*, 2003; Korsakov *et al.*, 2004; Manning, 2004; Katayama *et al.*, 2006; Spandler and Hermann, 2006; Zhao *et al.*, 2007; Zhang *et al.* 2008; Marocchi *et al.*, 2009; Zheng, 2012; Chen *et al.*, 2013; Hermann *et al.*, 2013).

# 1.9. Aims and Objectives

A primary aim of this study is to investigate and compare the garnet websterites at Svartberget, where they are related to veining and clearly metasomatic in origin, to the garnet websterites present at other localities to determine whether they have a similar origin and evolution. Another main aim is to characterise different aspects of the metasomatism recorded in the HP-UHP bodies studied to gain insight into fluid activity, behaviour and parameters at mantle depths and identify the processes occurring within the mafic-ultramafic bodies and/or the surrounding country rocks related fluid production. Research is undertaken to determine: the products of and chemical change caused by fluidrock interaction; the source, composition and phase of the metasomatising agent; the P-T conditions of metasomatism; and the timing of metasomatism. The evolution of the maficultramafic bodies and country rocks is also investigated to support the interpretation of the origins of the garnet websterites and parameters of fluid-rock interaction.

With respect to the 'bigger picture', the aims of this study are: to contribute to the understanding of the Earth's systems and processes during deep subduction-exhumation of the continental crust, particularly with respect to fluid-rock interaction at depth; to gain insight into the fate of continental crust retained in the mantle; to contribute new, reliable

data to the datasets of the WGR to constrain and refine geodynamic models of its evolution; and to characterise deep, crustal fluid systems. Furthermore, studying the orthopyroxene-rich products of fluid-rock interaction at mantle depths in a structural setting in the field can contribute to understanding the genesis of orthopyroxene-rich mantle xenoliths which are thought to be of metasomatic origin. Studying these crustal bodies will also aid the better characterisation of the lithological structure of the mantle and mantle wedge with respect to metasomatic processes.

A range of analysis techniques were employed to fulfil the aims. Field, petrographic, bulk rock and mineral chemistry studies were used to determine the origin and evolution of the mafic-ultramafic bodies, garnet websterites and HP-UHP fluid-related features and to determine the chemical change incurred due to fluid-rock interaction and fluid parameters. To determine the P-T conditions of metasomatism, a thermobarometric study was carried out using quantitative chemical data of mineral phases from a range of orthopyroxenebearing lithologies. To determine the timing of metasomatism, evolution of the fluid-rock system and constrain P-T-t paths, a geochronological study was carried out using radiometric U-Pb data from a range of minerals from a range of lithologies.

### 1.10. Thesis Structure

Following this introductory chapter, research carried out is presented by analysis technique. Field and petrographic observations are presented in Chapters 2 and 3, respectively. Whole rock data and associated mass balance calculations are presented in Chapters 4 and 5, respectively. Mineral chemistry data and associated thermobarometric estimations are presented in Chapters 6 and 7, respectively and radiometric U-Pb data are presented in Chapter 8. The various research themes are then discussed and concluded in Chapters 9 and 10 and Chapter 11 details suggested further work to progress the research carried out in this thesis.

Additional information and data are presented on the Appendix CD, the contents of which are detailed on pages 19-20 and include overviews of the key localities studied with plots of locality-specific geochemical data, full methodologies of data acquisition techniques, hand sample assemblage tables and descriptions, geochemical data and additional graphs. References made to Appendix sections and figures are prefixed with their appropriate letter (A-M excluding I).

Abbreviations and acronyms used are listed in *Tables 0.11a,b* in the preliminary pages (pages 27-28) for reference. A key for locality maps, diagrams and traverses is presented in *Table 0.12* (page 29) and marker and line formats of geochemical graphs are described in *Table 0.13* in the preliminary pages (page 30). Throughout the following chapters references to localities and hand samples are written in [] brackets.

# Chapter 2 – Field Observations

# 2.1. Introduction

This chapter details the fieldwork carried out for the research project and reports observations made in the field from the UHP localities selected for study. The observations focus on the general characteristics, differences and similarities of the various mafic-ultramafic bodies, fluid-related features and the structural relationships of orthopyroxene-bearing lithologies. Whether the orthopyroxene-bearing rocks are garnet websterites or 'orthopyroxene eclogites' *s.s.* requires geochemical distinction and is investigated in Chapter 6: until then they are referred to as garnet websterites.

## 2.2. Aims and Objectives

The aims and objectives of carrying out fieldwork were: to obtain a general overview, familiarity and understanding of the lithologies, localities and structure of the WGC; to observe and describe in detail the occurrences of garnet websterites at selected HP-UHP localities particularly their structural relationships; to observe and describe in detail selected localities from the perspective of fluid-rock interaction particularly looking for features indicative of fluid flow and metasomatism; and to collect rock samples for further petrographic and geochemical study.

# 2.3. Data Acquisition

*Figure 2.1* shows the location of all sites visited on a two-week excursion to the WGR in June-July 2010. The fieldtrip was attended by Dr Giles Droop, Dr Simon Cuthbert and myself. Locality information and reference codes (referred to in [] brackets in the text) are listed in *Table 2.1*. The localities visited were selected by my supervisors from those identified and studied: no previously unknown localities were visited; and those in the UHP domains were chosen for their occurrences of garnet websterites, established vein systems or other important features. The order in which they were visited during the fieldtrip was designed to reflect the regional pattern of increasing P-T conditions as established from previous studies.



Figure 2.1: Simplified geological map showing fieldwork localities. Image modified from Walsh et al. (2013). Blue markers – field localities. Numbers – locality 'QC' numbers. Red markers – accommodation.

At each locality, lithological and structural observations were noted and features photographed. Hand samples were obtained at each locality where permitted to do so to represent each lithology present and to capture veining features in order to determine the evolution of the rocks, determine their chemical compositions and to compare the compositions of the various vein zones.

Of the HP-UHP localities visited, seven were chosen for detailed study in the field due to their occurrences of garnet websterite, exposure, freshness and metasomatic features. Diagrams and maps were made at these seven localities to provide a visual and concise record of their characteristics and relevant features. Compass-and-pace maps of the entire bodies were made at four localities – Nybø [QC15], Grytting [QC25], Remøysunde [QC29] and Myrbærneset [QC34] due to their horizontal exposure on shoreline platforms and size. Due to a more vertical nature of exposure a diagrammatic traverse was made at Kolmannskog [QC32] and at Årsheimneset [QC26] a previously taken photograph of the body was annotated. A gridded map of a portion of the body at Svartberget [QC36] was made considering its larger size, the time available, the level of detail required and the detailed map already made by Vrijmoed (2009) of the whole body (see *Fig. A9.2* in the Appendix).

# 2.4. Data Presentation

The characteristics, main rock types and major features of the localities visited are detailed in *Table 2.1*. Rocks are classified according to Robertson (1999). 'Ultramafic' is used to describe olivine-bearing lithologies (peridotites, olivine websterites) and 'mafic' is used to describe pyroxene-dominant lithologies without olivine (eclogites, garnet websterites). The Mg-Cr – Fe-Ti classification of meta-ultramafic rocks is after Carswell *et al.* (1983) and refers to mantle-derived and intrusive crustal protoliths, respectively. In addition to the seven localities selected for detailed study in the field, two further garnet websterite-bearing localities, Eiksunddal [QC30] and Båtneset [QC33] were selected for petrographic study. These nine localities are the focus of petrographic study and geochemical analysis and are referred to as the 'key' localities. Brief introductions to the key localities are presented below with digitised versions of field maps, traverses and diagrams (*Figs. 2.3-2.9*). More detailed overviews of the key localities are presented in the Appendix [A1-A9] including descriptions of their structural relationships and lithologies.

Tuble 2.	ia. Betans of localities	co visited on j	relatinp						
No.	Name	Munici- pality	Map^	Long/Lat	UHP Domain	Exposure Type	Classification	No.of Samples	Details
QC1	Brensdal	Hyllestad	а	[03]0500, [67]7802	-	roadside cliff face (20m)	country rocks - orthogneiss	-	folded, amphibolite-facies, stromatic, foliated, migmatitic gneiss
QC2	Kolgrov/Hyllestad	Hyllestad	а	[03]0122, [68]8696	-	roadside cliff face	country rocks - metapelite	3	garnet-bearing schists
QC3	Hovlandsvatnet	Fjaler	b	[03]0608, [67]9816	-	roadside cliff face (3m)	country rocks - orthogneiss	-	amphibolite-facies, stromatic, migmatitic gneiss
QC4	Kvia, Breidvatnet	Fjaler	b	[03]0546, [68]0142	-	roadside cliff face (3-4m)	country rocks - mafic gneiss	4	retrograde amphibolite gneiss with white mica segregations and quartz- carbonate veins
QC5	Gjølanger	Fjaler	b	[02]9863, [68]0311	-	roadside cliff face (3m)	meta-ultramafic sheet	4	olivine-spinel websterite sheet with ~2m-long pyroxenite-garnetite veins; hosted in tonalite-diorite gneiss with retrograde contacts
QC6	Gjølanger	Fjaler	b	[02]9823, [68]0274	-	roadside	metamafic pods	4	veined eclogite pods in pyroxenite; hosted in tonalitic gneiss
QC7	Sørdal Mine, Kyia	Fjaler	b	[03]9713, [68]0174	-	cliff face, fallen boulders	'Fe-Ti' metamafic body	4	partially retrogressed, banded, foliated white mica-zoisite-kyanite (whiteschist) eclogite
QC8	Kvineset	Førde	с	[03]2908, [68]2259	-	roadside	metamafic pods	6	~13m-wide phengite-rich eclogite body with amphibolitised, mica-rich margins hosted in banded, granitic gneiss
QC9	Naustdal Old Bridge	Naustdal	d	[03]2556, [68]2496	-	roadside cliff face (20m)	metamafic pods	4	partially retrogressed, metadoleritic eclogite pods hosted in gneiss with quartz- carbonate veins
QC10	Naustdal river	Naustdal	d	[03]2562 <i>,</i> [68]2496	-	riverbank	metamafic pod	1	metadoleritic eclogite with omphacite veins
QC11	Engjabøneset	Naustdal	с	[03]1011, [68]2280	-	roadside	metamafic pod	4	eclogite pods and bands with retrogressed margins hosted in granitic gneiss
QC12	Verpeneset‡	Vågsøy	е	[03]0067, [68]6940	Nordfjord- Stadlandet	roadside cliff face (10m)	metamafic pods	1	metagabbroic pods of medium grained, lineated quartz eclogite and coarse grained, lineated-foliated, banded, white mica-zoisite-quartz-kyanite (whiteschist) eclogite; hosted in migmatitic gneiss
QC13	Almenningen	Vågsøy	e	[03]0327, [68]7007	Nordfjord- Stadlandet	roadside cliff face (10m)	metamafic pods	3	series of partially retrogressed eclogite pods hosted in banded, migmatitic gneiss; ~20m-long coarse grained, phengite-bearing eclogite body with amphibolitised margins
QC14	Kroken	Vågsøy	е	[03]0806 <i>,</i> [68]7049	Nordfjord- Stadlandet	roadside cliff face	metamafic pods and bands	8	carbonate-rich eclogite pods and metamafic bands in schists and migmatitic gneisses with gradational contacts and quartz-rich veins
QC15	Nybø	Vågsøy	e	[03]0167, [68]7279	Nordfjord- Stadlandet	shoreline	metamafic body	14	~28m-wide partial exposure of medium-grained, granoblastic eclogite with orthopyroxene-rich and lineated domains; hosted in gneiss with retrogressed contacts

#### Table 2.1a: Details of localities visited on fieldtrip

^ Local Maps and Codes: a - Risnesøya (M711-1117-2); b - Dale (M711-1117-1); c - Bygstad (M711-1217-4); d - Naustdal (M711-1218-3); e - Måløy (M711-1118-1) ‡ protected site; samples taken from fallen fragments

No.	Name	Munici- pality	Map^	Long/Lat	UHP Domain	Exposure Type	Classification	No.of Samples	Details
QC16	Vetrhuset	Selje	e	[03]0527, [68]7711	Nordfjord- Stadlandet	roadside	metamafic pods	1	eclogite pods hosted in metapelitic schists between folded, migmatitic orthogneiss
QC17	Nordpollen	Selje	е	[03]0512, [68]7723	Nordfjord- Stadlandet	shoreline	metamafic pods	4	eclogite pods in metapelitic schist within migmatitic gneiss
QC18	Flatraket	Selje	e	[03]0326, [68]7774	Nordfjord- Stadlandet	roadside	country rocks	-	mangerite with megacrystic K-feldspar and heterogeneous shear zones forming augen gneiss and mylonite
QC19	Gangeskardneset	Selje	е	[03]0150, [68]7760	Nordfjord- Stadlandet	shoreline	metamafic pods in country rocks	4	eclogite pods with various veining features and amphibolitised margins; hosted in stromatic, migmatitic gneiss
QC20	Almklovdalen	Vanylvan	f	[03]2142, [68]8153	Nordfjord- Stadlandet	olivine mine	'Mg-Cr' meta- ultramafic body	-	partially retrogressed garnet/chlorite/serpentine peridotite (Iherzolite-harzburgite-dunite) with garnet websterite and internal eclogite domains and sheets
QC21	Lien	Vanylvan	f	[03]2267, [68]7939	Nordfjord- Stadlandet	hilltop crags	'Mg-Cr' meta- ultramafic body	-	folded, layered garnet peridotite (wehrlite, dunite, pyroxenite), garnet websterite and internal eclogite sheets within chlorite peridotite (Almklovdalen body)
QC22	Raudkleivane	Vanylvan	f	[03]2090, [68]7939	Nordfjord- Stadlandet	roadside	metamafic sheet	-	veined/banded internal eclogite sheet in chlorite dunite (Almklovdalen body) with black-wall reaction zones at contact
QC23	Nybø beach	Vågsøy	е	[03]0184, [68]7272	Nordfjord- Stadlandet	bedrock	metamafic pod	2	~2m-long phengite eclogite pod
QC24	Grytting shore	Selje	f	[03]0902, [68]8554	Nordfjord- Stadlandet	shoreline	metamafic pods	2	amphibolite and retrograde eclogite pods and bands with quartz veins in schist in migmatitic gneiss
QC25	Grytting‡	Selje	f	[03]0900, [68]8568	Nordfjord- Stadlandet	shoreline	metamafic body	7	~30m-wide, medium-coarse grained eclogite body with a coarse-ultracoarse garnet websterite domain; hosted in garnet-rich, schistose layer in migmatitic gneiss
QC26	Årsheimneset	Selje	f	[03]1396, [68]8743	Nordfjord- Stadlandet	roadside cliff face (6m)	metamafic body	10	~6m-high coarse grained quartz eclogite body with garnet websterite domains and laterally-zoned veins; hosted in gneiss with retrogressed contacts
QC27	Årsheimneset shore	Selje	f	[03]1393, [68]8750	Nordfjord- Stadlandet	shoreline	metamafic pods	6	banded, retrograde eclogite and amphibolite pods hosted in schist within migmatitic gneiss

Table 2.1b: Details of localities visited on fieldtrip

^ Local Maps and Codes: e - Måløy (M711-1118-1); f - Vanylvan (M711-1119-3)
‡ protected site; samples taken from surrounding rocks

Table 2	.1c: Details of loca	lities visited on	fieldtrip						
No.	Name	Munici- pality	Map^	Long/Lat	UHP Domain	Exposure Type	Classification	No.of Samples	Details
QC28	Eggesbø	Bergsøya, Herøy	g	*[03]2641, [69]1473	Sørøyane	roadside cliff face (6m)	country rocks	5	retrograde eclogitic metamafic bands with undeformed leucosomes in migmatitic gneiss
QC29	Remøysunde	Remøya, Herøy	g	*[03]2582, [69]2104	Sørøyane	shoreline	metamafic body	15	15m-wide, coarse grained eclogite body with domains of coarse-ultracoarse garnet websterite with patches rich in biotite and carbonate; short, zoned veins are present in the garnet websterite; hosted in migmatitic gneisses
QC30	Eiksunddal~	Hareidland -et, Ulstein	h	*[03]4092, [69]0830	Sørøyane	fallen boulders	'Fe-Ti' metamafic body	5†	boulders of migmatitic gneiss with garnet websterite and eclogite and orthopyroxene-biotite-rich veins
QC31	Remøysunde road	Remøya, Herøy	g	*[03]2586, [69]2088	Sørøyane	roadside	metamafic pods	1	retrograde eclogite and amphibolite pods with leucosomes in migmatitic gneiss
QC32	Kolmannskog	Fræna	i	*[04]1093, [69]7724	-	roadside	'Fe-Ti' meta- ultramafic body with metamafic pods	7	40m-long, medium-coarse grained garnet peridotite-garnet websterite body and eclogite pods in gneiss with rare laterally-zoned veins and retrogressive features
QC33	Båtneset	Otrøy, Midsund	j	*[03]8378, [69]5883	Nordøyane	shoreline	metamafic pod	2	3m-long, lineated-foliated, finely banded and biotite-rich, medium-coarse grained garnet websterite pod hosted in garnet-bearing gneiss
QC34	Myrbærneset	Otrøy, Midsund	k	*[03]8341, [69]5895	Nordøyane	shoreline	'Fe-Ti' metamafic- ultramafic body	7	15m-long elongate body with domains of medium grained, granoblastic garnet peridotite, garnet websterite and lineated eclogite and a few laterally- zoned veins; pod is split in two (at the surface) by a ~60cm-wide pegmatitic sheet; hosted in gneiss
QC35	Ugelvik‡	Otrøy, Midsund	k	*[03]8240, [69]5454	Nordøyane	quarry	'Mg-Cr' meta- ultramafic body	-	seams and layers of garnet pyroxenite in garnet/spinel/serpentine peridotite (dunite, harzburgite) with orthopyroxene megacrysts with exsolved garnet; hosted in gneiss
QC36	Svartberget	Fræna	i	*[03]9858, [69]7058	Nordøyane	shoreline	'Fe-Ti' meta- ultramafic body	18	~50m-wide garnet peridotite body with network of laterally-zoned veins and vein-related garnet websterite domains; hosted in migmatitic gneiss; portion of HP-UHP contact evident

~ locality not visited by A. Quas-Cohen

<sup>^</sup> Local Maps and Codes: g - Fosnavåg (M711-1119-4); h - Ålesund (M711-1119-1); i - Molde (2698); j - Vestnes (M711-1220-2); k - Brattvåg (M711-1220-3)
<sup>\*</sup> UTM requires correction of 81m westward and 207m southward

‡ protected site

+ fallen boulders

Following the locality introductions are observations made focusing on features which relate to the research themes of this thesis. These are presented by lithology to aid comparison between the different localities. Photographs of features of the bodies and country rocks are presented in *Figures 2.10-2.17*.

# 2.4.1. Key Localities

Details of the key localities are presented in *Table 2.1* and their locations are shown in *Figure 2.2*. Topographic and detailed geological maps showing their locations are presented in the locality overviews in the Appendix [A1-A9]. Digitised field maps, traverses and images are presented in *Figures 2.3-2.9*: a full diagram key is presented in *Table 0.12*.



Figure 2.2: Map showing key HP-UHP localities selected for study. Image modified from Walsh et al. (2013). Blue markers – seven localities selected for detailed study in the field, petrographic and geochemical study. Orange markers – localities selected for petrographic and geochemical study. Bracketed numbers – locality 'QC' reference codes.

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The bodies at all these key localities are considered to be crust-derived lithologies (part of the continental crust before subduction) and are classified as 'external' eclogites (after e.g. Griffin and Brueckner, 1980) and Fe-Ti garnet peridotites (after Carswell *et al.* (1983)). The Nybø (*Fig. 2.3*), Årsheimneset (*Fig. 2.4*), Remøysunde (*Fig. 2.5*), Grytting (*Fig. 2.6*), and Eiksunddal localities are eclogite bodies with garnet websterite domains. Garnet peridotite bodies with garnet websterite domains are found at Kolmannskog (*Fig. 2.7*) and Svartberget (*Fig. 2.8*) and one body with eclogite, garnet peridotite and garnet websterite domains occurs at Myrbærneset (*Fig. 2.9*). The Båtneset locality is characterised by a garnet websterite body without eclogite or garnet peridotite.

# 2.4.2. <u>Country Rocks</u>

The majority of the country rocks have a pervasive, amphibolite-facies paragenesis dominated by quartz, plagioclase feldspar and biotite: their modal distribution defines various scales of banding (*Figs. 2.10a,b*). Bulk compositional variation is dominantly represented through alkali feldspar-rich and sillimanite-rich units. Gneisses and schists are locally garnetiferous, contain abundant white mica and sporadic kyanite porphyroblasts (*Fig. 2.10c*) or form kyanite schists at localities such as Remøysunde and Grytting. Partially retrogressed, eclogite-facies felsic rocks with relic clinopyroxene (omphacite) form lineated bands/sheets within the country rocks, such as observed at the Årsheimneset shore [QC27] (*Fig. 2.10d*). Fresh and partially retrogressed eclogitic mafic bands also occur such as at Kroken [QC14] and Eggesbø [QC28]. The mafic bands and small pods contain garnet-quartz-rich, clinopyroxene-rich, garnet-rich and quartz-rich bands.

Generally the fabric and banding of the host gneisses gently wrap around boudinaged, mafic-ultramafic pods (<1m-scale) and bodies (>1m-scale) (*Fig. 2.10e*): this is seen to occur at all scales. However, deformation is highly heterogeneous with strain often severely concentrated into shear zones where even the more competent pods are drawn out (*Fig. 2.10f*) and mylonites are formed. Tight ptygmatic and isoclinal folding is also observed. Within the pressure shadows of the larger bodies the country rocks exhibit lower degrees of strain with less well-defined fabrics. Within some of these zones of larger bodies, for example at Remøysunde, higher pressure assemblages are retained.



Figure 2.3: Compass-and-pace map of the Nybø locality.



*Figure 2.4: Diagrammatic photograph trace of the Årsheimneset locality. Photograph taken by Tony Carswell (from Carswell and Cuthbert, 2003b).* 



Figure 2.5: Compass-and-pace map of the Remøysunde locality [QC29].



Figure 2.6: Compass-and-pace map of the Grytting locality [QC25].









Figure 2.7: Traverse and field sketch of the Kolmannskog locality [QC32].





Figure 2.8: Compass-and-pace map of a 5x6m area of the Svartberget garnet peridotite body [QC36].



Figure 2.9: Compass-and-pace map of the Myrbærneset locality [QC34].



Figure 2.10: Photographs showing features of the country rocks observed in the field. a) alkali feldsparrich, banded gneiss [Årsheimneset, QC27D]; b) host gneiss [Årsheimneset, QC26A]; c) garnet-kyanitebearing gneiss [Vetrhuset, QC17], photo courtesy of G.T.R.D.; d) retrograde felsic eclogite/garnetbearing gneiss [Årsheimneset, QC27B]; e) banded gneisses wrapping metamafic pods [Båtneset, QC33]; f) drawn out mafic boudins in strongly sheared gneiss [Båtneset, QC33]; g) leucosome with mafic selvages [Brensdal, QC1]; h) progressive break-up of boudins by intrusion and squeezing of gneiss [Båtneset, QC33].

The gneisses at both the HP and UHP localities are typically stromatic migmatites: thin granitic leucosomes with sharp hornblende-biotite selvages (melanosomes) (*Fig. 2.10g*), granitic, alkali feldspar-rich bands or lenses and folded streaks of quartz and anorthosite are observed. Unlike the gneisses, no features indicative of migmatisation are observed in the schistose metapelite horizons which typically host the mafic pods at the HP localities or at the southern UHP localities visited. However, migmatitic features are observed at the northern UHP localities visited such as in sillimanite-rich schistose rocks at Remøysunde. Furthermore, undeformed, leucocratic bands and sweats are observed in retrogressed mafic lithologies within the gneisses at Eggesbø and Remøysunde and late, felsic pegmatites cutting into meta-ultramafic bodies are observed at Myrbærneset and Svartberget: these are described in more detail below.

The large, crustal, mafic-ultramafic bodies were broken up under tensional forces associated with exhumation-relation deformation. Gneiss is seen being squeezed along fractures and between the more competent boudins (*Fig. 2.10h*) demonstrating a fluid, ductile nature and an active involvement in this process. The dismemberment of bodies is seen at all stages in the field from initial breakup of larger angular bodies to the complete isolation of smoothly shaped boudins.

# 2.4.3. Mafic-Ultramafic Bodies

No relict igneous or prograde, plagioclase-bearing assemblages or textures are observed demonstrating they are pervasively equilibrated to eclogite-facies conditions. The larger bodies show minimal amphibolite-facies retrograde metamorphism: alteration is typically only pervasive and complete in their outermost edges. Amphibolite-facies retrogression within the more interior portions of the bodies is typically associated with fine fractures with thin, grey, amphibolitised margins or amphibole-rich, ultracoarse veins. Some patches of eclogite, such as at Remøysunde, show alteration of clinopyroxene to a paler green phase and garnets have black coronas (*Fig. 2.11e*). Of the key localities studied, the garnet peridotite body at Kolmannskog shows the greatest degree of retrogression. Most small pods (<1m-sc) are completely retrogressed (or at least appear to be) and metre-scale pods often have partially retrogressed or fresh eclogite-facies cores.



Figure 2.11: Photographs showing features of host bodies observed in the field. a) pastel-coloured eclogite with clinopyroxenite domain [Nybø, QC15A]; b) deeply coloured bimineralic eclogite with beaded string garnets [Remøysunde]; c) garnet-rich band (segregation) in bimineralic eclogite [Grytting]; d) modal banding in an eclogite pod [Verpeneset, QC12]; e) porphyroblastic garnets [Remøysunde]; f) porphyroblastic, mauve garnets in clinopyroxenite band [Nybø, QC15F]; g) garnet peridotite [Svartberget, QC36A]; h) garnet peridotite [Myrbærneset, QC34B].

Three major mineralogies of eclogite s.s. were observed at the key localities: bimineralic eclogite at Nybø, Grytting, Eiksunddal, Remøysunde and Myrbærneset, quartz eclogite at Arsheimneset and biotite eclogite at Nybø. Quartz-kyanite-zoisitebearing, phengite-bearing, epidote-bearing and carbonate-bearing eclogites were also observed at other UHP localities. Their appearances vary from medium grained, pale-bright, pastel coloured with granoblastic-interstitial textures (e.g. Nybø; Fig. 2.11a) to coarse grained, deeply coloured with decussate textures (e.g. Remøysunde, Eiksunddal, Årsheimneset; Fig. 2.11b). Lineation fabrics are observed in some eclogites such as at Nybø and Myrbærneset: these are both medium-grained, more pastel-coloured eclogites. The lineation at Nybø is patchy and varies in its strength. Modal variation of major and minor phases within eclogite bodies is observed with garnet-rich and clinopyroxene-rich domains (Fig. 2.11a) and segregations (Fig. 2.11c) and well defined modal banding is also observed in other major phases (Fig. 2.11d). For example, at Nybø, garnet and clinopyroxene occur in fine monocrystalline bands which become a dominant feature in some areas. Textural variation within bodies also occurs through changes in grain size and the form of garnet: a range is observed from large, individual porphyroblasts (Fig. 2.11e) to fine, sub-grain aggregates interstitial with clinopyroxene (Fig. 2.11a). Two types of garnet are found in close proximity to each other (cm-scale) at Nybø: mauve garnet porphyroblasts in clinopyroxenite bands occur within domains of orange-pink garnet interstitial with clinopyroxene (Fig. 2.11f).

The peridotite bodies and domains are medium grained, deeply coloured, with a dominantly granoblastic texture (*Figs. 2.11g,h*) and a spaced foliation defined by biotite when present. The abundance of olivine and clinopyroxene vary producing garnet-bearing olivine websterites at Svartberget and Myrbærneset and garnet-bearing harzburgites *s.s.* at Kolmannskog, respectively: the scale and extent of this heterogeneity is not determined so the ultramafic bodies are still referred to as garnet peridotites.

# 2.4.4. Garnet Websterites

Garnet websterites (garnet-clinopyroxene-orthopyroxene assemblages) are observed in three main structural relationships and with various textures and mineralogical associations. They occur as small-to-large domains within mafic-ultramafic bodies, as

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isolated bodies and in association with laterally zoned veins. The observed occurrences at the key localities, their assemblage and characteristics are listed in *Table 2.2*.

There are four types of body domain garnet websterite. Firstly, large, irregular, deeply coloured, decussate, coarse-ultracoarse grained domains at some edges of eclogite bodies such as Grytting, Årsheimneset and Remøysunde (*Figs. 2.4-2.6, 2.12a*). Secondly, medium-ultracoarse grained, deeply coloured decussate-textured blocks within the garnet peridotite body at Svartberget (*Figs. 2.8, 2.12b*); although these blocks may just be apparent rather than a pervasive domain and actually represent a surface coating or selvage. Thirdly, deeply-coloured, fine-medium grained, granoblastic-textured domains within ultramafic bodies at Myrbærneset and Kolmannskog (*Figs. 2.9, 2.7*). Like the host peridotite, some garnet websterite body domains at Myrbærneset are also biotite-rich. Lastly, paler, pastel coloured, medium grained patches with the eclogite body at Nybø (*Fig. 2.12c*); here the orthopyroxene-rich domains have granoblastic textures and lineated fabrics similar to the eclogite and occur in no well defined domains; although exposure at Nybø is limited so delineating any such domain is difficult.

The coarse-ultracoarse garnet websterite domains in eclogite bodies have sharp but irregular contacts with their host. Elongate, decussate pyroxene crystals locally reach over 10cm long. The domains, and particularly their boundary zones, are characterised by patches rich in volatile-bearing phases such as biotite, carbonate and amphibole: this was first observed by Eskola in 1921. The eclogite side of the contacts at Arsheimneset and Remøysunde are also often characterised by patches rich in carbonate, biotite and amphibole. Biotite-rich patches are also seen in the garnet websterites at Nybø. Portions of the garnet websterite at Remøysunde seemingly contain two sub-domains: garnet, clinopyroxene and rutile in one with textures and grain sizes similar to the host eclogite and orthopyroxene, biotite, amphibole and carbonate in the second (see Fig. 2.12a). In comparison, the orthopyroxene in the garnet websterite at Grytting is well integrated with garnet and clinopyroxene throughout the domain (see Fig. 2.12d) (although its concentration and size are variable) and the websterite is considerably coarser grained than in the eclogite; the distributions of carbonate, amphibole and biotite are still patchy however. At Arsheimneset the domains contain short and long laterally-zoned veins,



Figure 2.12: Photographs showing features of garnet websterite domains and pods observed in the field. a) deeply coloured garnet websterite with two mineralogical sub-domains [Remøysunde]; b) garnet websterite with ultracoarse pyroxene and finer garnet crystals [Svartberget], photo courtesy of G.T.R.D; c) pastel coloured, medium grained garnet websterite [Nybø, QC15M]; d) coarse garnet websterite with orthopyroxene well distributed amongst garnet and clinopyroxene [Grytting]; e) main pod [Båtneset]; f) lineated fabric and fine banding of garnet websterite pod [Båtneset]; g) fine modal banding of garnet websterite [Nybø, QC15N];

at Remøysunde they contain short laterally-zoned veins and segregations and at Grytting they contain no veining features (although a considerable portion of the domain has been eroded away).

The second type of structural garnet websterite is seen at the Båtneset locality where garnet websterite forms the entire body (of what can be observed). It is a medium grained, deeply coloured, lineated, finely modally banded, biotite-bearing garnet websterite with a granoblastic-interstitial texture and forms a 2m-wide, 3m-long ovoid isolated body (*Figs. 2.12e-g*). Similar lineated fabrics are found in the garnet websterite at Nybø (*Fig. 2.12h*) where similar fabrics are also shared by the associated eclogite. Lineations are defined by elongate pyroxenes and garnet which develop into fine modal banding. Foliations are defined by biotite where present as a major phase. The fabrics are truncated against the gneissose fabrics at the body-country rock contact.

The third structural relationship, in association with veining features, occurs in three forms. Firstly as coarse-ultracoarse, deeply coloured, decussate-textured, irregular-restricted domains particularly at vein junctions (*Fig. 2.14a*) or where there is a high density of irregular veins or well-established, thick, complexly-zoned veins with dm-scale widths at Svartberget (*Fig. 2.8, 2.14b*). Secondly as marginal vein selvages to laterally-zoned, garnetite-clinopyroxenite veins at Kolmannskog, Myrbærneset and particularly at Svartberget where there is a network of such veins. The width of the garnet websterite selvages vary and those immediately adjacent to the clinopyroxenite selvage at Svartberget are garnet-poor, biotite-rich websterites. A 1m-thick zone of biotite websterite is also present at the southeastern contact of the Svartberget body: this is discussed in more detail below. Lastly, garnet websterites also occur as thin, cm-scale-thick veins within the host body without other vein zones although in close proximity to zoned veins such as observed at Kolmannskog (see *Fig. 2.7*).

There is some overlap of the different structural types: for example, at Årsheimneset, the garnet websterite domains contain continuous, laterally zoned veins and are therefore also vein-marginal; and the blocks at vein junctions at Svartberget may be vein selvage websterites.

Locality	Number	Main Body Type	Sample	Websterite Structure	Websterite Characteristics	Websterite Assemblage <sup>^</sup>	Websterite Type
Nybø	QC15	bimineralic eclogite	J,L,M,N Δ1308	body domains(?) unsure (poor exposure)	similar texture to eclogite, lineated, pastel coloured	grt-opx-cpx, am-bt-mgs (rt)	metamorphic
Grytting	QC25	bimineralic eclogite	Ø16, A650	body domain	coarse-pegmatitic, brightly coloured, decussate	grt-cpx-opx-bt-cb-am (op, rt-zrn- ap-hc)	metamorphic- metasomatic
Årsheimneset	QC26	quartz eclogite	G,P,A347,A 634,Δ1311, Δ1312,Z4A	body domains	deeply coloured, decussate, medium-coarse	grt-cpx-opx ±bt ±cb ±am, qtz (rt- ap-zrn-op-spl-mnz) <i>am-symp-idg-</i> <i>cal</i>	metamorphic- metasomatic
Remøysunde	QC29	bimineralic eclogite	A,N,P	body domain	decussate, deeply coloured, pegmatitic, two sub-zones	cpx-grt-opx-bt-cb-am	metamorphic- metasomatic
Eiksunddal	QC30	eclogite	A-E	body domain(?)	decussate, deeply coloured, coarse	grt-cpx-opx ±bt, ap-am ±chu (rt- op-zrn ±spl), <i>am-pl-symp-op</i>	metamorphic- metasomatic
		garnet	F	vein selvage	coarse, dark	opx-cpx-grt-bt-am	metasomatic
Kolmannskog	QC32	harzburgite/		veins	coarse	opx-cpx (?)	metasomatic
		peridotite	А	body domain	fine-medium, cpx-rich, edge of body	cpx-grt-opx, bt (rt-op)	metamorphic
Båtneset	QC33	garnet websterite	A,B	body domain	lineated, banded	grt-cpx-opx-bt (rt-op)	metamorphic
		eclogite/oliv-	В	body domain	fine-medium, deeply coloured, granoblastic, (schistose)	grt-opx-cpx-ol-bt-op, am-cb-spl (zrn-aln)	metamorphic
Myrbærneset	QC34	ine garnet websterite/gar	C,D,E	body domain	fine-medium, deeply coloured, granoblastic	grt-opx-cpx ±bt, opq-am-cb (rt- zrn-spl)	metamorphic
		net websterite	F,G	vein selvage	fine-medium, deeply coloured, granoblastic	grt-opx-cpx, op-bt-amp-cb (rt- zrn-spl)	metasomatic
			A-C,T	main body	fine, granoblastic, dark	opx-cpx-grt-ol, cb-am-bt-op ±chu (zrn-ap-spl) <i>idg-am</i>	metamorphic
Svartberget	QC36	garnet peridotite/oli-	D	body domain	pegmatitic, decussate, deeply coloured	cpx-opx-grt-bt-am (rt-zrn-op) <i>am-</i> <i>idg</i>	metasomatic
C C		vine garnet websterite	E,G	vein selvages	coarse, decussate, deeply coloured	cpx-opx-bt, am ±grt (rt-zrn-mnz- op) <i>am-op-idg</i>	metasomatic
			S	contact	foliated, coarse	bt-opx-cpx, am (op-zrn) <i>cb-idg</i>	metasomatic

Table 2.2: Structural, textural and mineralogical characteristics	of websterites at the nine key localities	. Phase abbreviations as listed in Table 0.11b.
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^ phases: major, minor (accessory) retrograde

Locality	Number	Host Rock	Sample	Type^	Zones*	Abundance	Major Assemblage
Krakan	0014	eclogite pod	-	v	-	rare	cpx-qtz-wmca
KTOKETI	QC14	metamafic pod (contact)	-	S		common	qtz-wmca
Nuba	0015	garnet websterite	L,M	S	-	rare	opx-am-cb-bt
Νύρα	QCIS	bimineralic eclogite	В,К	s,vl	g/c	common	grt/cpx-bt
Cangaskardnasat	0.010	eclogite	A,D	vl	-	common	grt
Gangeskaruneset	QCI9	eclogite	С	S	-	rare	grt-qtz
Crutting	0025	bimineralic eclogite	-	s/vl	-	rare	grt
Grytting	QC25	metamafic pod	-	S	-	rare	grt-qtz
		garnet websterite	H,J	v,s	g/c	common	grt-qtz-bt-wmca/cpx-bt
Årshoimnosot	0026	quartz eclogite	F	v	-	rare	qtz-phg (pl-bt)
Arsheimneset	QC20	quartz eclogite	-	S	-	rare	qtz ±grt ±bt
		quartz eclogite	-	S	g/c	common	grt/cpx ±qtz
		eclogite	-	v	-		срх
Årsheimneset shore	QC27	eclogite	-	s/vl/v	-		qtz ±grt ±rt
		eclogite	-	vl	-	rare	rt
		garnet websterite	М	s/v	g/c/gc	rare	grt-qtz-bt-rt/cpx-bt/grt-cpx-rt
Pomovsundo	0020	metamafic pods	-	vl	-	common	grt ±qtz
Remøysunde	QC29	eclogite	L	S	-	common	grt (beaded strings, tubes)
		eclogite	-	v	-	common	cpx ±qtz
Fiksunddal	0C20	eclogite	-	v	-	rare(?)	opx-bt
LIKSUHUUdi	QC30	eclogite	-	v	-	rare	срх
Kolmannskog	0022	garnet peridotite	F	v	gw/c/g	rare	cpx-opx-bt-am/cpx-bt-am/grt ±bt ±cpx
Kulthallinskug	QC32	garnet peridotite	-	v	-	rare	opx-cpx ±rt
		garnet peridotite/olivine	F	v	awlela	rare	grt-cpx-opx-bt-op/cpx-op/grt-op ±ap
Myrbærneset	QC34	websterite	I	v	gw/c/g	Tale	±cpx
		eclogite	А	vl	-	common	rt
		garnet peridotite/olivine	FGU	v	gw/bw/bc/	common,	cpx-opx-bt-am/cpx-bt-am/bt/grt ±ap
Svartberget	QC36	websterite	L,0,0	v	±gl/g	interconnecting	±bt ±cpx
		biotite clinopyroxenite	Q	р	-	rare	qtz-afs-mc-bt-wmca ( <i>after qtz-phg</i> )

Table 2.3: Locations and characteristics of eclogite-facies fluid-related features in the mafic-ultramafic bodies at UHP and key localities. Mineral abbreviations as listed in Table 0.11b.

^ v = vein, vl = veinlet, s = segregation, p = pegmatite

\* g = garnetite, gl = glimmerite, c = clinopyroxenite; bc = biotite clinopyroxenite, gc = garnet clinopyroxenite, gw = garnet websterite, bw = biotite websterite

Locality	Number	Sample	Host Rock	Type^	Abundance	Major Assemblage
Verpeneset	QC12		eclogite	v	rare	qtz (bt-cb)
Almenningen	QC13		eclogite	v	common	qtz
Nordpollen	QC17		metamafic pods	v	common	qtz
Crytting shore	0024		retrograde mafic pods	V	common	chl
Grytting shore	QC24		retrograde mafic pods	V	common	qtz
			garnet websterite	v	rare	cb-chl-qtz-am
Grytting	QC25		garnet websterite	v	common	am-tlc
			eclogite	s/v	common	qtz
Pomovcundo	0020		eclogite	s/v	common	qtz-cb
Kemøysunde	QCZ9	L	eclogite	vl	common	am
Remøysunde road	QC31		retrograde metamafic pods		common	qtz-pl, bt-am
			garnet peridotite	р		qtz-fsp-bt
Kalmannskag	0022		garnet websterite	v	common	tlc-tr-bt
KUIIIaiiiisKug	QC52	E	garnet websterite	v	common	act-oam-chl
			peridotite-gneiss contact	S	rare	hbl-bt ('blackwall')
Svartberget	QC36		garnet peridotite	v		am

Table 2.4: Locations of retrograde fluid-related features (veins, veinlets and segregations) in the UHP and key locality bodies and their structural, textural and major mineralogical characteristics. Mineral abbreviations as listed in Table 0.11b.

^ v = vein, vl = veinlet, s = segregation, p = pegmatite

Locality	Number	Host Rock	Sample	Type^	Zones	Abundance	Major Assemblage
Verpeneset	QC12	gneiss		I	-		qtz-kfs-wmca
Almenningen	QC13	gneiss (pod contact)		S	-	common	grt-qtz-wmca
Nybø	QC15	gneiss		I	-		qtz-kfs-wmca
Vetrhuset	QC16	gneiss		I	-		qtz-kfs-wmca
Nordpollen	QC17	schist		v	-	rare	qtz-tur
Gangeskardneset	QC19	gneiss		I	-		qtz-fsp, bt
Grutting shore	0024	gneiss		I	-	common	qtz-kfs-pl, bt-am
dividing shore	QC24	gneiss		v	-	common	qtz
Crutting	0025	gneiss		I	-		qtz-kfs-wmca
Grytung	QCZJ	gneiss		v	-	common	qtz
Årsheimneset	QC26	gneiss	D	S	-	rare	tur-qtz-pl
Årsheimneset shore	QC27	gneiss		I	-		qtz-kfs-pl ±wmca
Eggesbø	QC28	mafic gneiss		I	-	common	qtz-pl ±kfs ±wmca, bt-am
		gneiss	С	S	-	one	ky-bt-pl (after ky-phg)
Remaveunde	0020	garnet-bearing gneiss		s/vl	-	common	qtz ±grt ±phg
Nemøysunde	QCZS	gneiss		v	-	rare	grt-qtz-ky
		gneiss		I		common	qtz-pl-kfs-wmca
Eiksunddal	QC30	gneiss		I	-		qtz-kfs-wmca
Remøysunde road	QC31	gneiss		I	-	common	qtz-pl-kfs-bt
Kolmannskog	0022	gneiss		I	-	common	qtz-pl-kfs, bt
KUIIIaiiiiskug	QCSZ	gneiss		v	-	common	qtz
Båtneset	QC33	gneiss		I	-	common	qtz-pl-kfs-wmca, bt-am
Myrbærneset	QC34	gneiss		р	-	rare	qtz-pl-bt
Suarthorgot	0026	gneiss		Ι	-	common	qtz-pl-kfs-bt ±wmca
Svartberget	QC30	gneiss		р	-	common	qtz-pl-kfs-bt

Table 2.5: Locations of fluid-related features in the country rocks at UHP and key localities and their structural, textural and major mineralogical characteristics. Mineral abbreviations as listed in Table 0.11b.

^ v = vein, vl = veinlet, s = segregation, l = leucosome/migmatitic, p = pegmatite

The textures of garnet websterites relative to their host body varies between localities. The garnet websterites at Svartberget are coarser grained with a more decussate texture with pyroxenes forming large, elongate laths as opposed to the finer, more equant blasts seen in the medium grained, granoblastic matrix of the host body. In contrast, the garnet websterite domains at Årsheimneset have similarly interstitial-decussate textures with porphyroblastic garnets as the main quartz eclogite body. Finally, the garnet websterite body domains and vein selvages in the Myrbærneset body also have similar textures to the host peridotite domains but they are medium grained with a granoblastic texture.

# 2.4.5. Fluid-Related Features

*Tables 2.3 and 2.4* list eclogite-facies and retrograde fluid-related features observed in the mafic-ultramafic bodies at the UHP localities visited, respectively, and those observed in the country rocks are listed in *Table 2.5*. Fluid-related features are classified into three types based on their size and shape: segregations, veinlets and veins with the two former being more common than fully developed veins. All three are seen in close proximity to each other in eclogite bodies (see *Fig. 2.13a*) but veinlets and segregations are not observed in garnet peridotite bodies or domains.

# 2.4.5.1. Segregations

Segregations are well-defined but irregularly-shaped concentrations with coarser grain sizes than the matrix and without any internal structure or well-defined zoning and they range in size from cm-wide to m-wide domains (*Figs. 2.13a-d*). Mineralogies observed in segregations in mafic-ultramafic bodies include: garnet ±quartz, clinopyroxene ±biotite ±quartz, quartz ±white mica ±rutile and orthopyroxene-amphibole-biotite ±carbonate. Observed quartz and pegmatitic carbonate-quartz segregations are classified as lower P-T features.

Coarse segregations are seen in the host gneisses close to the bodies such as a cluster of kyanite porphyroblasts with large, interstitial intergrowths of plagioclase-biotite (after phengite) at Remøysunde (*Fig. 2.13d*) ~20cm from the websteritic body domain. Other segregations in the country rocks have quartz ±garnet ±white mica mineralogies and white mica- or biotite-rich segregations

±quartz often mark mafic pod-country rock contacts such as seen at Gjølanger [QC6], Kvineset [QC8] and Kroken and contain coarser mica than the matrix.

# 2.4.5.2. Veinlets

Veinlets are discontinuous (cm-dm-scale in length), typically thin (<0.5cm-wide) concentrations of phases found in the host rock. Veinlets may have sparser, more disseminated textures and coarser grain sizes than the matrix (*Figs. 2.13e,f*). As mentioned, such smaller-scale features are not observed in the ultramafic domains or bodies.

Observed veinlet assemblages in mafic bodies include: quartz ±garnet ±rutile, garnet ±quartz, zoned clinopyroxene-garnet and disseminated, rutile-rich trails often extensions of quartz-rutile veinlets. The dominant fluid-related feature seen in eclogites is continuous and discrete, mm-cm-thick garnet-rich veinlets ±quartz with adjacent garnet-poor bands such as observed at Gangeskardneset [QC19] (*Fig. 2.13e*) and Årsheimneset. Garnet-rich veinlets range from disseminated trails to beaded strings to pure garnetite. Disseminated rutile veinlets or trails are also present along the margins of some garnet-rich vein cores. Veinlets in the host gneisses typically have quartz ±garnet mineralogies (*Fig. 2.13f*).

### 2.4.5.3. Veins

Veins are more continuous than veinlets (>10cm in length), thicker (>1cm-wide), may connect to lithological boundaries, may contain phases not represented in the host rock and polyphase veins may be laterally zoned (e.g. *Figs. 2.14e,f*). They are typically coarse grained with decussate textures. Structural characteristics range from wavy to straight and parallel to interconnecting: veins at Svartberget form a stockwork-like network dividing the body into straightsided, dm-m-scale blocks (*Figs. 2.8, 2.14a, A9.1c,d, A9.2*) whilst some of those at Årsheimneset are wavy and follow similarly orientated paths through the body (*Fig. 2.4*).



Figure 2.13: Photographs showing fluid-related features observed in the field. a) image showing types of fluid-related features. s – segregation, vl – veinlet, v – vein. Photo courtesy of G.T.R.D. [Gangeskardneset]; b) large garnet-rich domain (segregation) although it may be an oblique cut through a zoned vein [Remøysunde]; c) coarse, orthopyroxene-rich segregation in garnet websterite [Nybø, QC15L]; d) kyanite-rich segregation near the main body [Remøysunde]; e) garnet veinlet [Gangeskardneset, QC19]; f) garnet-quartz veinlets in the host gneiss [Remøysunde].



Figure 2.14: Photographs showing features of laterally zoned veins features observed in the field. gw = garnet websterite. a) network of zoned veins around garnet websterite blocks [Svartberget], photo courtesy of G.T.R.D.; b) wide, complex vein with pegmatitic garnet websterite domain to either side [Svartberget], photo courtesy of G.T.R.D.; c) zoned vein [Kolmannskog, QC32F]; d) garnetite-cored zoned vein in garnet peridotite with websteritic and clinopyroxenite selvages [Myrbærneset]; e) zoned vein with clinopyroxene-rich garnetite [Svartberget, QC36E]; f) biotite-cored, quartz-bearing, zoned vein [Årsheimneset, QC26H]; g) wide, complex, laterally zoned vein with repetition of zones [Svartberget]; h) zoned vein junction with irregular mass of zones, websterite blocks and weathered down websterite-peridotite selvages [Svartberget].

Mineralogies of veins observed in UHP mafic-ultramafic bodies include: garnet ±quartz, clinopyroxene ±quartz ±phengite ±rutile, quartz-phengite, orthopyroxene-biotite, orthopyroxene-clinopyroxene ±rutile and laterally zoned garnetite-clinopyroxenite-websterite veins. The two latter are the only vein types found in garnet peridotite bodies. Assemblages vary both between veins in the same body, along the lengths of veins and between host body types: a particular observation is the distinction of the presence of quartz in veins within eclogite bodies and their absence from those in garnet peridotite bodies.

The laterally-zoned veins typically consist of a garnet- and/or biotite-rich websteritic selvage, a clinopyroxenite selvage ±biotite and a garnet-rich core ± clinopyroxene or ±biotite ±quartz ±white mica ±apatite (Figs. 2.14a-e, A9.4a-e). In eclogite bodies, these veins mostly occur within garnet websterite domains and so they do not strictly have a separate garnet websterite selvage. The zoning structure can vary along the length of vein from well-defined to indistinct as is observed at Årsheimneset. One vein at Årsheimneset has an established quartzwhite mica vein core: it contains coarse bands of granoblastic quartz with seams of foliated, schistose white mica and occasional, tiny rose pink garnets (Fig. 2.15d). At Svartberget, zones are reasonably well structured although they grade into irregularly zoned garnetite-clinopyroxenite-websterite masses at vein intersections and they can become more diffuse when curved or expanded, particularly the garnetite core. The zoning pattern across the vein is typically well defined, simple and symmetrical although the widths of selvages either side can differ and complex patterns with the repetition or forking of core zones with intervening selvages is observed in thick (>1m-wide) veins at Svartberget (Fig. 2.14b, 2.14g). No cross-cutting relationships are observed at vein junctions (Fig. 2.14a) or in multiply zoned, wide veins (Fig. 2.14b). Vrijmoed et al. (2006, 2013) described more mineralogically distinct vein zones in the veins at Svartberget but did not describe a clinopyroxenite ±biotite zone between the websterite and garnetite zones despite being present in all the veins observed (Fig. A9.2). The various zoning patterns are illustrated in Figure 2.15.

Contacts between hosts and vein selvages are typically sharp but irregular: this is reflected by their preferential separation whilst sampling. At Svartberget the veins are often selectively weathered down (*Fig. 2.14h*). The transitions of

clinopyroxenite selvages to garnetite cores vary both along and between veins: observed transitions include: <1cm-wide, pure biotite 'glimmerite' selvages; garnet clinopyroxenite selvages; or no intervening zone with a sharp transition to garnetite. Garnetite vein cores show variable textures and assemblages ranging from pure garnet ±apatite to clinopyroxene-rich or quartz-biotite-rich garnetite with decussate-interstitial textures. Again, this variation is seen along the length of veins and in different veins. Biotite typically accompanies garnet in both garnet peridotite and eclogite bodies and can form continuous, central segregations or an interstitial matrix around garnet. As mentioned, quartz is common in garnetite cores in eclogite bodies but is absent in veins observed in garnet peridotite bodies. Conversely, clinopyroxene is mostly absent from garnetite cores in eclogite bodies but can form a major component in those in garnet peridotite bodies.

### Vein zoning patterns at Årsheimneset

grt +bt +amp	ht	grt	qtz	qtz-bt	grt
±crb ±qtz	Di	8''			qtz

#### Vein zoning patterns at Svartberget

art	h.t.			bt
git tht	ы. 	DL	grt	срх
100	±grt	±amp	±bt	±ap

#### Vein zoning patterns at Myrbærneset



Figure 2.15: Diagrammatic representation of zoning patterns of well established veins found at key localities. Zone widths vary between veins so widths are representative. Additional minerals are major phases and are as listed in Table 0.11b.
The zoned veins and garnetite segregations found at Myrbærneset (*Fig. 2.14d*) differ from those found elsewhere: they are medium grained with granoblastic textures like their host bodies and contain no additional major phases not found in the host body. Furthermore, they are characterised by a reduction in the amount of biotite rather than an increase and the clinopyroxenite selvages are very thin unlike at other localities. They are however similar in their lateral structure and zonation pattern.

More irregularly zoned clinopyroxenite-garnetite veins not accompanied by websterite selvages or other major phases are observed in mafic pods at Årsheimneset shore (*Fig. 2.16a*). Clinopyroxenite veins without garnet, seen at localities such as Remøysunde have no lateral zoning and are typically pure clinopyroxene (*Fig. 2.16b*). Those observed at Remøysunde are over 10cm-wide, have sporadic accompanying phases such as quartz or rutile and have sharp contacts with the host rock which have been preferentially weathered out (*Fig. 2.16b*). They occur in fractures which reach the body contact and cut into body at a range of angles (see *Fig. 2.5*); they do not continue into the host gneisses. Orthopyroxene-rich veins include websterite veins without clinopyroxenite or garnetite cores observed at Kolmannskog which occur in clusters following fine, parallel fractures in the peridotite and as sheens on exposed surfaces; and ~5cm-wide orthopyroxene-biotite veins observed in the garnet websterite-eclogite boulders at the Eiksunddal locality (*Fig. 2.16c*).

Vein types observed in the host gneisses include quartz ±garnet ±kyanite or ±tourmaline and a retrogressed clinopyroxene-phengite-quartz-rutile vein. Tourmaline occurs as numerous, aligned, elongate crystals in quartz veins (e.g. Nordpollen [QC17]) or single pegmatitic crystals in quartz lenses (e.g. Årsheimneset).

#### 2.4.5.4. Retrograde Features

Retrograde veining features are detailed in *Table 2.4*. Quartz veins and veinlets are the most common; they cut into small boudins, are concentrated into boudin necks and form streaks aligned parallel to the boudin margins.



Figure 2.16: Photographs showing other fluid-related veining features observed in the field. a) irregular clinopyroxenite vein in garnetite [Årsheimneset shore, QC27]; b) clinopyroxenite vein [Remøysunde]; c) biotite-orthopyroxene-rich vein within garnet websterite [Eiksunddal boulder]; d) white mica-quartz vein core [Årsheimneset, QC26F]; e) retrogressive vein [Kolmannskog, QC32E]; f) late carbonate-chlorite vein with amphibolitised margins [Grytting]; g) late, fine fractures with amphibolitised walls [Nybø, QC23B]; h) late granitic, pegmatitic veins breaking apart a body [Svartberget]; i) leucocratic 'sweats' in retrograde metamafic pods [Eggesbø, QC28].

The processes involved in retrogressive fluid-rock interaction may be analogous to those which occur at higher P-T conditions so their features are worth noting. Very sharp transitions between fresh and pervasively retrogressed bands are observed at Kolmannskog with rafts of relict material (*Fig. 2.16e*). Retrograde veins at Grytting have diffuse alteration haloes either side of vein-filled fractures (*Fig. 2.16f*). Clustered and isolated, fine fractures cut through rocks such as at Nybø (*Fig. 2.16g*). Late, undeformed pegmatites and granitic veins which have intruded into the body at Svartberget form a network of veins breaking up the body (*Fig. 2.16h*); the pegmatites are usually swept away into the gneissose foliation once over the pod margin demonstrating they formed during exhumation-related deformation. Regularly spaced chlorite-filled fractures are seen cutting through mafic boudins and tapering quickly into the host gneisses.

Undeformed, branching, leucocratic 'sweats' and veins with diffuse edges are observed within mafic bands such as at Eggesbø (*Figs. 2.16i*), in boudin necks often crossing into the host rocks and towards the centres of retrograde mafic pods such as at Remøysunde road [QC31]: such pods also show more gradational contacts with the host rocks. The 'sweats' have plagiogranitictrondhjemtic compositions consisting of plagioclase-quartz and veins also contain biotite.

#### 2.4.5.5. Svartberget Contact

The southeast contact of the Svartberget body is characterised by a ~3m-wide, zoned unit between the typical host gneiss and the garnet peridotite *s.s.* (*Fig. 2.17a*) and appears to be a metasomatic feature with zones similar to the veins in the interior of the body: the unit is not present at other visible contacts. Adjacent to the typical peridotite is a ~1m-wide, concentric zone of foliated biotite websterite (*Fig. 2.17b*) followed by a similarly wide zone of biotite clinopyroxenite which becomes more biotite-rich further towards the contact (*Fig. 2.17c*). Within the clinopyroxenite are quartz-biotite-feldspar-rich, clinopyroxene-absent bands (*Fig. 2.17d*) and occasional garnet-biotite segregations (*Fig. 2.17e*). Within the biotite websterite is a 30cm-wide lens which consists of coarse, pegmatitic quartz and a fine grained mica-feldspar



Figure 2.17: Images of the southeastern contact at Svartberget and associated rock types. a) zoned, southeastern contact; b) biotite websterite [QC36S]; c) biotite clinopyroxenite [QC36P]; d) felsic material in biotite clinopyroxenite zone [QC36N]; e) zoned garnet glimmerite-clinopyroxenite vein in biotite clinopyroxenite zone [QC36R]; f) felsic lens in clinopyroxenite zone [QC36Q]; g) body-country-rock contact s.s. to right and pegmatitic lens to left; h) leucocratic gneiss [QC36L].

intergrowth (after phengite) (*Fig. 2.17f*) and lies concordant to the gneiss-body contact with thin, irregular offshoot veinlets which extend into the biotite websterite. Quartz in the vein appears blue when the intergrowth is just below it. The edges of this lens and bands within it are finer grained and foliated. The outer biotite clinopyroxenite is immediately adjacent to a band of leucocratic gneiss (*Figs. 2.17g,h*) which hosts rafts of clinopyroxenite and which becomes more foliated and biotite-rich away from the contact towards the typical host gneisses. Feldspar-biotite offshoots from this band extend into the clinopyroxenite (across the contact) thinning into biotite-rich streaks.

#### 2.5. Data Analysis

## 2.5.1. Country Rocks

The country rocks are the expected source of metasomatic fluids at UHP conditions. It is therefore important to determine the peak paragenesis present in the gneisses and schists and identify any fluid-producing processes which may have occurred. The pervasive amphibolite-facies assemblage largely prevents this. However, the fresh and retrogressed eclogite-facies felsic and mafic bands in the gneisses indicate high pressure conditions were experienced by the country rocks and their assemblages transformed, at least locally, to an eclogite-facies paragenesis. A UHP nature cannot be determined in the field.

Features indicative of partial melting in the country rocks during UHP Scandian conditions would provide evidence of fluid production for metasomatism. The lack of migmatitic features in the metapelitic units at the HP and southern UHP localities suggests the timing of migmatisation of the adjacent gneisses was pre-Scandian and prior to the incorporation of supracrustal units into the basement orthogneisses as the more hydrous, volatile-rich schistose metasediments are expected to melt more easily. In contrast, the migmatitic features in the metapelitic rocks at Remøysunde, the undeformed, *in situ* nature of features indicative of partial melting in the gneisses and late pegmatites cutting ultramafic bodies in the northern part of the WGR suggest partial melting occurred in association with the last deformational and thermal event i.e. the Scandian. Previous migmatitic features of the gneisses in this

area analogous to those in the HP terrane would be overprinted by the later event. The relationship between partial melting in the country rocks and Scandian peak pressure conditions is not considered to be well established (e.g. Gordon *et al.*, 2013). However, migmatitic features are seen to increase in abundance towards to the northwest following the regional gradient in P-T conditions (Hacker *et al.*, 2010; Gordon *et al.*, 2013) and studies provide evidence to suggest widespread partial melting occurred post-eclogitisation during Scandian decompression and exhumation and also at high pressure conditions (e.g. Labrousse *et al.*, 2002; Hacker *et al.*, 2010; Labrousse *et al.*, 2011; Gordon *et al.*, 2013). The widespread association of deformational structures in the gneisses and schists with the dismembered, retrogressed mafic-ultramafic pods and bodies demonstrates their structures and fabrics are related to exhumation and decompression under amphibolite-facies conditions. None of the migmatitic and fluid-related features observed in the gneisses and schists can therefore be reliably associated with UHP conditions.

#### 2.5.2. <u>Mafic-Ultramafic Bodies</u>

The presence of both eclogite and peridotite within the same pod at Myrbærneset suggests all the mafic-ultramafic bodies may have had an associated origin and their protoliths are different products in the evolution of larger magmatic bodies and their associated intrusions which, for the most part, have become subsequently separated during dismemberment. This suggests a suite of bulk compositions may be represented by the pods and bodies: this will be investigated in Chapter 4. Alternatively, the eclogite domains at Myrbærneset may represent a later intrusion of mafic magma into the ultramafic body which remained together during dismemberment; however the irregular nature of the domains (see *Fig. 2.9*) suggests this is not the case.

The mineralogical variation of the eclogites probably reflects small-scale compositional variation of their protoliths. Modal variation may too reflect primary compositional heterogeneity such as original layering; although fine scale banding may be due to metamorphic segregation. Their textural variation, including the different forms of garnet, may reflect textural variation of their protoliths and the local chemical environment during metamorphism which controlled the rate of growth. Colour variation may be a product of the varying grain sizes or may reflect compositional variation: chemical distinction between the two types is investigated in Chapter 4.

The granoblastic textures of the Svartberget and Myrbærneset garnet peridotites and Nybø eclogite demonstrate they are well-equilibrated rocks (e.g. Hermann, 2002). The textures may have formed due to high nucleation rates or formed through dynamic recrystallisation of a coarser grained rock. The lineated fabrics of the eclogites at Nybø and Myrbærneset indicate they experienced plastic deformation during metamorphic equilibration. The lack of fabric in the coarser grained eclogites at Grytting, Årsheimneset and Remøysunde and garnet peridotites at Kolmannskog and Svartberget, however, indicates this deformation did not affect all maficultramafic bodies during their equilibration to HP-UHP conditions. The development of deformational fabrics in the bodies also indicates strain was not wholly partitioned into the more ductile and deformable country rocks at this time. A lower competency contrast may have existed between the lithologies perhaps in response to the higher pressure conditions and development of an eclogite-facies paragenesis in the country rocks.

The retrogression seen at the margins of the bodies demonstrates the bodies were as they are now prior to late amphibolitisation in order to acquire the retrogression pattern they possess. This suggests dismemberment took place at an early stage of exhumation. Amphibolitisation of the dominantly anhydrous, mafic-ultramafic lithologies requires H<sub>2</sub>O: the complete retrogression of only small pods indicates retrograde fluid availability was minimal. Alternatively it may reflect a short duration of fluid availability and the slower reaction rates of mafic rock hydration relative to the dehydration of felsic lithologies (Heinrich, 1982).

## 2.5.3. Garnet Websterites

The range in mineralogical, textural and structural characteristics of the garnet websterites suggests a range of processes and origins may be involved in their formation. The similarities shared by some however suggest the repetition of processes and common origins. The main distinction to be made is whether they are metamorphic or metasomatic in origin: the range in their characteristics indicates both types are present. The structural relationships of the vein-associated garnet websterite selvages suggest they are associated with fluid-rock interaction and are therefore considered to be metasomatic in origin. The origins of the other structural types of garnet websterites, however, are less certain. In this section, features of the garnet websterites are analysed to determine their origin: the interpreted origin of each websterite is presented in the final column of *Table 2.2*.

Firstly, the garnet-bearing, plagioclase feldspar-free nature of websterites and their later overprinting by amphibolite-facies parageneses demonstrate all the garnet websterite domains formed at and/or equilibrated to eclogite-facies conditions (>500°C). Furthermore, all occurrences of garnet websterite are found within or near UHP domains: Kolmannskog is located just outside the current boundary of the northernmost Nordøyane domain but such boundaries are subject to change should UHP natures be established. This suggests their formation may be related in some way to the higher P-T conditions experienced by the rocks in these areas, particularly those of metasomatic origin. The various colour and textural types of the garnet websterites, as with the eclogites, may be related to bulk chemical composition, their grain sizes and/or variation of their protoliths. A chemical distinction is investigated in Chapter 4.

The extraordinary sizes and random orientations of pyroxene crystals in some body domains and vein-associated garnet websterites suggest they did not equilibrate whilst undergoing plastic deformation and have not suffered dynamic recrystallisation. Coarse grain sizes in metamorphic rocks grow under slow heating rates for prolonged periods of time so as to not greatly overstep prograde reaction equilibria and cause high nucleation rates (producing numerous, small crystals) (Barker, 1998). The overall medium-coarse coarse grain sizes of the garnet websterites and eclogites suggest ambient temperatures did not change rapidly. The localised nature of ultracoarse crystal sizes within the garnet websterite domains suggests development of very large crystals was due to a local process in addition to the ambient conditions; for example the prolonged presence of catalytic fluids.

In comparison, the fine-medium grained, granoblastic textures of the lineated garnet websterite domains, at Nybø and particularly at Myrbærneset suggest they underwent a different evolution and one shared by the whole body considering the similar textures of the host body eclogite and peridotite domains, respectively. They

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may have formed under rapid temperature increase during eclogitisation and high nucleation rates or through dynamic recrystallisation of a coarser grained body under stress. The latter process is proposed to produce similar textures in similar rocks from the island of Fjørtoft in the northern UHP domain and to have occurred during retrogression (Carswell *et al.*, 2006). Considering the association of deformation, fluid-activity and metamorphic reaction kinetics (see Chapter 1), recrystallisation probably occurred still under HP-UHP conditions or their eclogite-facies assemblages would not have survived. Furthermore, the truncation of the garnet websterite lineated fabrics against the foliation in the country rocks, such as at Båtneset, demonstrate they are eclogite-facies fabrics and are not equivalent to those in the amphibolite-facies gneisses; consequently nothing can be inferred from their orientation.

The common presence of patches rich in volatile-bearing minerals in the garnet websterite domains of eclogite bodies is suggestive of local fluid activity or a patchy distribution of phases in the stable prograde-peak metamorphic paragenesis due to the P-T-X conditions of equilibrium and/or low H<sub>2</sub>O and CO<sub>2</sub> activities: the dominantly anhydrous nature of the adjacent eclogites suggests the bodies were reasonably efficiently dehydrated during eclogitisation supporting the former. The apparent presence of sub-assemblages in the garnet websterite from Remøysunde suggests the formation of the garnet websterite was associated with a volatile-bearing fluid which metasomatised the host eclogite. The presence of laterally-zoned veins in garnet websterite domains also suggests they have a fluid-related evolution. However, the heterogeneously distributed, rare or absent nature of the veins within the garnet websterite domains, such as at Remøysunde and Grytting, and the patchy occurrence of the volatile-bearing phases suggest the dominant process involved in the formation of anhydrous garnet websterite at these localities may not be related to metasomatism or vein formation but are metamorphic lithologies.

Strictly, weight should not be placed on the position of garnet websterite domains within the eclogite bodies as they occur today considering their exhumation-related modification and dismemberment. However, as shown, the destructive gneisses made use of any plane of weakness within the body during dismemberment, whether an existing fracture or lithological boundary. Such planes of weakness are likely to have been exploited in a similar way by metasomatising fluids under HP-UHP

conditions. Therefore, the occurrence of garnet websterites towards the edges of bodies as they exist today, such as at Årsheimneset and Grytting, may indicate that they formed along a plane of weakness, fracture or a lithological boundary which was subsequently exploited upon pod breakup.

The garnet websterite body at Båtneset is thought to represent portions of larger metamafic bodies dismembered along lithological boundaries (or plane of weakness within an even larger garnet websterite body). Assuming dismemberment occurred during exhumation both pods of metamorphic and metasomatic garnet websterite may occur. However, the medium grain size, fine modal banding and absence of fluid-related features in the body at Båtneset suggest it has a metamorphic, not UHP metasomatic origin. The abundance of biotite however, suggests either a K-rich protolith or a pre-/syn-deformation, pervasive influx of K-rich hydrous fluids. Its lineated fabric suggests it underwent deformation during metamorphic equilibration to HP-UHP conditions like the bodies at Nybø and Myrbærneset. The proximity of Båtneset and Myrbærneset (*Fig. 2.2*) and K-rich nature of both garnet websterites suggest they may have developed in the same crustal body.

In comparison, the vein-bounded websterite blocks at Svartberget are considered to have a metasomatic origin being metasomatised peridotite blocks as an extension of the vein selvage websterite. As shown in *Figure 2.14a*, some small blocks are completely websteritic indicating the metasomatic reaction front on the various sides have moved inwards and joined up in the centre totally transforming the peridotite. On a larger scale and with subsequent dismemberment, metre-scale blocks may form isolated bodies where the garnet websterite has a metasomatic origin.

## 2.5.4. Fluid-Related Features

The rocks studied are considered an analogue for crust-mantle interaction involving the transfer of material from one chemical system to another. It is therefore important to distinguish features related to the activity of internal fluids and those associated with the infiltration of external fluids. This section considers how the fluidrelated features described above may have formed.

The range in scales of the fluid-related features in eclogites and garnet websterites suggests they may develop over time and the close proximity of different sized and shaped features demonstrates a spatial association. The lack of segregations and veinlets in garnet peridotite domains however suggests processes forming these smaller features did not operate and a gradual evolution of the veins present in this lithology did not occur. This indicates different vein-forming mechanisms must operate and that the physiochemical properties of the host body have some control over these.

The textural and structural characteristics of the segregations may be due to a range of origins. The garnet-rich bands which occur at Grytting (*Fig. 2.11c*) are within a band of coarser eclogite near to the ultracoarse garnet websterite domain and are not observed away from this domain. They may have a fluid-related origin but may also relate to compositional or textural variation of the protolith or form through metamorphic segregation. In comparison, the irregular, coarser grain sizes, decussate textures and concentration of phases of the orthopyroxene-amphibole-biotite segregations in medium grained, anhydrous garnet websterites at Nybø (*Fig. 2.13c*) are clearly related to fluid activity. The mineralogies of segregations in eclogites reflect those of the host eclogite although minerals such as quartz may be more concentrated. This suggests external fluids are not a requirement for their formation but they form through redistribution of internal fluids and metamorphic segregation. The common association of garnet and quartz in segregations and veins may reflect the increased mobility of Al in Si-rich fluids at high pressures via aluminosilicate polymers (Manning, 2004).

Like segregations, veinlets in eclogite bodies contain no phases not present in their host rock. This, together with their isolated and small nature suggests they formed via internal processes without the addition of external components. The structural nature of veinlets is indicative of forming along microfractures which may form through hydrofracturing should pore fluid pressures exceed the constrictional stress. Internal sources of fluids include trapped pore water produced through dehydration during eclogitisation, water exsolved from NAMs upon decompression or the breakdown of unstable hydrous phases. The pervasive equilibration of the maficultramafic bodies to eclogite-facies conditions suggests fluids were abundant during prograde metamorphism otherwise they would not have transformed: igneous

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mafic-ultramafic lithologies still persist in the WGR such as the Hustad Igneous Complex and transitional gabbros on Flemsøy through the absence of catalytic water (e.g. Mørk, 1985, 1986; Rubie, 1986; Austrheim *et al.*, 2003). Therefore, a considerable volume of trapped fluid might persist in the bodies. Alternatively, increased porosity and permeability may develop in quartz eclogites during the transformation of matrix quartz to coesite on the prograde path which involves a ~10% decrease in volume (e.g. Carswell and Compagnoni, 2003). Conversely, the retrograde transformation of coesite to quartz might induce fracturing during volume increase. The lack of veinlets observed in garnet peridotite bodies suggests fractureand space-inducing processes and/or fluid-producing processes did not occur, perhaps associated with the absence of free silica or a lower trapped water content.

Disseminated, interstitial rutile trails in eclogites may reflect the path of late, Ti-rich fluids or the residue left after activity of Ti-insoluble fluids. As eclogites are rutile-rich the trails probably represent the internal redistribution of Ti rather than interaction with an external fluid. The formation of garnetite veinlets, clinopyroxenite veins and irregular garnetite-clinopyroxenite segregations may also form through variable solubility in internally-derived fluid and are a type of metamorphic segregation. Features of clinopyroxenite veins like those at Remøysunde, which have metre-scale lengths, sharp contacts with their host and lack of alteration zones, suggest they are transport veins for chemically equilibrated fluids.

The continuous nature of the laterally-zoned veins suggests their development is also associated with channelised fluid flow along transport pathways. However, the presence of phases absent or not found in any great abundance in the host bodies suggests the veins formed from externally-derived fluids and the addition of components, such as K to form the abundant biotite. Furthermore, no 'leach' zones in the adjacent host bodies are evident indicating zone-forming components are not sourced from the adjacent wall rock. At Svartberget, the loss of olivine from the host body to the garnet websterite selvage suggests silica is also added and reacts with olivine to form orthopyroxene  $[Mg_2SiO_4 + SiO_2 = Mg_2Si_2O_6]$ . The quantitative loss and addition of components across vein zones is investigated in Chapters 4 and 5.

The straight, interconnected, common, cm-m-wide nature of fractures at Svartberget suggests these are related to rapid, brittle macro-scale hydrofracturing in a large

stress field. In comparison, the wavy, thinner, sporadic nature of the veins at Årsheimneset suggest these veins developed through gradual propagation of smaller fractures navigating around heterogeneities exploiting planes of weakness and connecting established cracks and microfractures. Both the lengths and widths of these vein types vary suggesting they develop laterally and lengthways over time. The proportional variation of zone widths suggests all zones develop at the same time and widen together with their final widths dependent on the amount and duration of fluid availability. Variation may have arisen through the duration of time that fractures were open either through the timing of their formation or closure or the fractures had different widths and therefore fluid fluxes. The lack of cross-cutting relationships at vein junctions at Svartberget indicates the veins all formed contemporaneously and fluid injection was not episodic on a long-term scale. The occurrence of multiple garnetite zones in the wider veins suggests these were characterised by fracture zones.

The transitional and localised nature of different vein assemblages within the same body indicates the composition of the chemical system varied locally and the similarity of the laterally-zoned veins at different localities suggests their formation is genetically related. The loss of zone structure may reflect a change in the shape of the space in which the vein has to form: for example, at Svartberget, vein cores become more diffuse or expand in thickness when they curve and veins become irregular masses at vein intersections where the distance between fracture walls is likely to be greater.

The presence and absence of quartz in the laterally zoned veins in eclogite and garnet peridotite bodies, respectively, suggests that the lithology has some control over their core zone mineralogy. Quartz in a 'leucosome' zone in the centre of some veins at Svartberget is however noted by Vrijmoed *et al.* (2013): this suggests it is related to the evolution of the local fluid-rock system. The presence of quartz in veins at Årsheimneset and Remøysunde therefore may reflect the relatively more felsic nature of mafic eclogite bodies compared to ultramafic, crucially olivine-bearing peridotite bodies and the lesser geochemical gradients between them. Alternatively, the metasomatising fluids may have had different compositions at different localities. White mica also has a distinct occurrence being absent from all veins observed within garnet peridotite bodies but sporadically present within veins in garnet websterites

and eclogites. Like quartz and the Si content, this may reflect a lower Mg content of the eclogite bodies causing phengite to be the stable mica. The quartz-white mica vein core at Årsheimneset suggests the infiltrating fluids are Si-K-Al-H<sub>2</sub>O-rich. The absence of  $CO_2$ -bearing phases from all laterally-zoned veins observed suggests the  $CO_2$  activities of the fluids remain low.

The coarse grain sizes of the laterally zoned veins at Svartberget, Årsheimneset and Remøysunde, like their garnet websterites, suggests the crystals grew during a time of gradual temperature change. The typical decussate textures of the vein zones suggest textural equilibrium was maintained and kept pace with the changing conditions. The textures also suggest the bodies did not experience internal deformation syn- or post-formation demonstrating the efficiency of strain partitioning into the surrounding country rocks during exhumation. Biotite-rich zones are often foliated however. In comparison, the laterally-zoned veins at Myrbærneset have a medium grained, granoblastic texture similar to that of the host body suggesting they have undergone a different evolution to the veins at the other localities. The relative decrease in the abundance of biotite and low abundance of clinopyroxene and the higher abundance of orthopyroxene suggests the fluid-rock chemical system also differed. This will be investigated further in Chapters 3, 4 and 6.

The orthopyroxene-biotite-rich veins observed at Eiksunddal also demonstrate that Si-K-Al-rich fluids were mobile. The mica-rich seams or selvages that line the contacts of some mafic pods in the gneisses suggest that flow of Si-K-Al-H<sub>2</sub>O-rich fluids were concentrated at the lithological contacts during their dismemberment but they may not be HP-UHP features. The presence of the kyanite-(ex)phengite segregation in the host gneisses near the garnet websterite domain in the body at Remøysunde demonstrates the presence of Al-K-Si-rich fluids in the gneisses. Unless the segregation lay within a pressure shadow of the body in which the gneisses were not ripped away, there is no evidence to suggest the segregation is actually associated with the UHP metasomatism of the body. Tourmaline is known to occur in UHP rocks in other UHP terranes such as the Kokchetav Massif and Erzgebirge (e.g. Marschall *et al.*, 2009; Shimizu and Ogasawara, 2013) but its occurrence in veins observed in the amphibolite-facies gneisses does not suggest an UHP origin.

The undeformed, leucocratic 'sweats' with diffuse edges observed within retrograde mafic bands and pods at some localities demonstrate temperatures were high enough to induce partial melting of mafic lithologies during the Scandian event and static, low-strain conditions were present in the cores of even small pods during this time. The presence of veins cutting into the gneisses indicate it occurred after ductile deformation or in areas external to the pods shadowed from strain by the more competent mafic material. These features demonstrate partial melting occurred in the host gneisses during the Scandian. In addition, the undeformed, granitic pegmatites and veins in bodies which are swept away into the gneissose foliation and the concentration of quartz veinlets into boudin necks demonstrate granitic and felsic fluids were present and mobile during exhumation-related deformation and the intrusion of such fluids helped to break up the pods. Whilst not directly relevant to HP-UHP fluid processes, they are analogous. The sharp nature of the hydrous, retrograde bands in the garnet websterite domain in Grytting demonstrates that pervasive, fluid-related alteration can occur behind a sharp reaction front. The more diffuse alteration haloes expanding from late fractures demonstrate diffusive transport of fluid occurred perpendicular to the fracture wall. The fine fractures with diffuse margins demonstrate the scale of alteration relative to fracture width and finally, the brittle fractures in mafic pods within plastically-deformed gneisses demonstrates the contrasting response of the lithologies to stress.

# 2.5.5. <u>Svartberget Contact</u>

The parallel bands in the zoned unit at the southeast contact have a similar zoning pattern to the veins in the interior of the body. The absence of such bands at other visible contacts suggests they are not related to retrograde fluid-rock interaction during exhumation otherwise all contacts would be characterised in this way. Its characteristics therefore suggest the unit is related to fluid-rock interaction at HP-UHP conditions. Compared to the vein zones inside the body, there is no garnet websterite zone adjacent to the garnet peridotite or well-established garnetite zone. Instead, only a biotite websterite zone and small lenses of biotite garnetite within the biotite clinopyroxenite domain occur, respectively. These differences indicate the chemical composition of the fluid-rock system was different to that in the interior of the body inferring the composition of the fluid which infiltrated the body differed from that which metasomatised the contact. However, the abundance of biotite within the contact zones and relative lack of it in the host peridotite and removal of olivine indicates that the fluid was still Si-K-H<sub>2</sub>O-rich like that of the infiltrating fluid. Again, the lack of white mica in the metasomatic zones suggests biotite was the stable mica and related to the more Mg-rich environment of the peridotite relative to the adjacent gneisses.

The~2m thickness of the metasomatic unit suggests metasomatism occurred for an extended period of time and/or with a greater fluid abundance: this would require the accumulation or a continuous supply of fluids at the contact. Accumulation may have occurred at the contact against the more competent garnet peridotite body and within UHP pressure shadows of the body as these would be areas of relatively lower stress towards which fluids may have flowed along pressure gradients. As mentioned, micaceous seams are often observed along the contacts of small mafic pods hosted in the gneisses reflecting the channelling of hydrous fluids along the lithological boundaries during dismemberment: an analogous process may have occurred at UHP conditions. Continued arrival of fluids at the contact would increase the fluid pressure: this may have induced the hydrofracturing of the body. The contact unit may have both formed and been preserved in a pressure shadow. The leucocratic gneisses at the contact are foliated and have an amphibolite-facies paragenesis and so they may be associated with exhumation-related partial melting. However the concentration of leucocratic material against the contact is analogous to the inferred UHP process.

The pegmatitic quartz-(ex-)phengite lens within the biotite websterite may represent a crystallised portion of the fluid which metasomatised the body contact. Its finer grained outer margins suggest they may have cooled more quickly analogous to chilled margins of dyke intrusions. The undeformed nature of the offshoot veinlets from the lens and other felsic segregations in the metasomatic unit and their extension at an angle to the gneissose foliation and plane of the contact demonstrate the unit is undeformed. The large sizes (cm-scale) of the pseudomorphic feldsparmica intergrowths indicate the original crystals were ultracoarse indicating the lens formed in the presence of a fluid and under slow rates of temperature change.

#### 2.6. Data Interpretation

#### 2.6.1. Origins of Garnet Websterites

Considering the range of structural locations, textures and mineralogies of garnet websterites at the localities visited is it conceivable that each had a unique origin and evolution. The vein selvages and vein-associated, small, irregular domains at Svartberget are most clearly metasomatic in origin. The small blocks between veins and larger domains are also considered to be metasomatic in origin resulting from encasing vein selvages and long-term metasomatism, respectively. The contact zone at Svartberget is considered to be a UHP metasomatic feature but formed in a slightly different chemical system to the interior veins.

The garnet websterite veins and selvages at Kolmannskog and vein selvages at Myrbærneset are also considered to be metasomatic in origin. Those at Myrbærneset are thought to have developed with a K-poor fluid, formed at a different time and undergone subsequent dynamic recrystallisation. The websterite veins at Kolmannskog are thought to represent metasomatism associated with shortlived, minimal fluid flow without developing further lateral zones.

The texture, fabric and lack of volatile-bearing phases in the garnet websterites at Nybø and their indistinct boundaries or structure with the adjacent eclogite and absence of established fluid-related features suggest they are metamorphic in origin and result from portions of the mafic protolith body with a more ultramafic character. A metamorphic origin is also tentatively assigned to the body domains at Kolmannskog, Myrbærneset and the Båtneset body. The greater abundance of biotite particularly at Båtneset and Myrbærneset, however, suggests a K-H<sub>2</sub>O-rich component has been involved in their evolution and they may be deformed metasomatic domains.

The final four localities Årsheimneset, Remøysunde, Grytting and Eiksunddal all contain coarse-ultracoarse, decussate-textured, volatile-bearing phase-bearing garnet websterites in body domains of eclogite bodies. The origins of these body domains are less clear, particularly at Eiksunddal where the large-scale structural relationships of garnet websterites and veining features are not established. The presence of abundant volatile-bearing phases, ultracoarse patches and veining features within the garnet websterites suggest a fluid-related development but their patchy, heterogeneous distribution suggest fluid-activity is localised. Furthermore, HP-UHP veining features are not observed at all within or near the websterite domain at Grytting. Considering the four localities together, the features are indicative of a metamorphic garnet websterite heterogeneously metasomatised during its prograde evolution by external fluids. The protolith of the garnet websterite is interpreted to be a more ultramafic portion of the mafic body: the metasomatising fluids are shown to be Si-rich so metasomatism of the eclogite domains would not result in a more ultramafic lithology.

#### 2.6.2. Fluid-Rock Interaction

As the objective of this study is to investigate metasomatic fluid-rock interaction, the origin and fluid parameters of veins and domains which require externally-derived fluid to form are the focus of this section and of future study. These include the laterally-zoned veins at various localities and the southeast contact at Svartberget. Veinlets and most segregations are interpreted to have evolved through redistribution internal of components and locally sourced fluids and are therefore not strictly metasomatic in nature.

Most models of vein formation during HP-UHP metamorphic conditions in other HP-UHP terranes are associated with fluid escape from the rocks during prograde devolatisation and the transition to eclogite-facies conditions (e.g. Becker *et al.*, 1999a; Spandler and Hermann, 2006; John *et al.*, 2008 and references cited therein; Spandler *et al.*, 2011 and references cited therein). However, the veins observed occur within HP-UHP assemblages and so are either metamorphosed lower pressure veins or formed following the transformation of the host body lithologies to eclogitefacies conditions (e.g. Simon and Chopin, 2001).

The mineralogies of all the vein types in garnet peridotite, garnet websterite and eclogite domains considered to have developed through infiltration of external fluids indicate it was a Si-K-Al-rich fluid; except for the texturally and mineralogically distinct zoned veins at Myrbærneset. These are thought to have formed from a fluid

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of different composition possibly at a different stage along the P-T path via a different method of generation or source.

The variation in assemblage, zoning pattern and structure between and along the laterally-zoned veins is interpreted to reflect the local chemistry of the system controlled by the local physical and chemical characteristics of the host rock, the composition and evolution of the fluid and the geometry of the fracture system (lengths, widths, intersections etc). For example, the general absence of quartz in veins in garnet peridotite is considered to reflect silica-undersaturation following reaction of the host rock. The leucosome zones seen by Vrijmoed *et al.* (2013) are noted to be only present in wide veins at Svartberget. They are interpreted to represent the pooling and crystallisation of trapped, remaining fluid following the smoothing of chemical gradients and inability of components to diffuse towards the reaction front.

The similarity of laterally-zoned veins at Kolmannskog to those at Svartberget with selvage garnet websterite zones suggests they may have similar models of formation. As diamond has been found at Svartberget (Vrijmoed *et al.*, 2006, 2008), a nearby locality in the northernmost UHP domain, demonstrating a UHP nature, and considering the similar lithologies to Svartberget and the eastern extension and shape of the southernmost UHP domain (e.g. Hacker *et al.*, 2010) it is proposed that the Kolmannskog locality too experienced UHP conditions: this is investigated in Chapter 7. The occurrence of metasomatic garnet websterite within UHP domains suggests their formation may be related to the increased P-T conditions experienced by the terrane: the inclusion of metasomatic garnet websterites at Kolmannskog into the UHP domain would further support this association.

The zoned pattern of the veins may develop either sequentially as the fluid-rock system evolved with progressive interaction or contemporaneously with the growth of certain phases at certain points along the chemical gradient. The model of vein zone formation is investigated in Chapter 5. The general process of formation of garnetite (>50-60% garnet) is still debated: proposed mechanisms include UHP metamorphism, metamorphic differentiation, contact metamorphism, metasomatism and melting reactions (Rivalenti *et al.*, 1997; Vrána and Frýda, 2003; Mel'nik *et al.*, 2013). Garnetite is also the product of continued metamorphism of

eclogite at pressures of over 10GPa as pyroxene dissolves into garnet (Sekine *et al.*, 1986). Its structural relationships within vein cores at the localities studied provide a clear example of metasomatic garnetite. Its chemistry is investigated in Chapters 4-6 but initial field observations suggest it developed through the interaction of felsic and mafic-ultramafic components. A similar chemical system is indicated in the Ivrea-Verbano Zone, Alps where metres-thick layers of biotite-rich garnetite occur between granulite-facies metapelitic and mafic-ultramafic rocks (Rivalenti *et al.*, 1997) and between eclogite and migmatitic tonalite-trondhjemitic gneiss in the Belomorian Mobile Belt, Russia (Mel'nik *et al.*, 2013).

The veins are evidence of fluid infiltration into the mafic-ultramafic bodies. The southeastern contact at Svartberget is interpreted to be an original eclogite-facies body-country rock contact and to reflect the fluid-rock interaction at the lithological contacts of the parent bodies at UHP conditions. Consequently, this means the Svartberget body lay at the edge of its parent body as it was at UHP conditions. The difference between the zones produced at the contact compared to the veins in the body interior is considered to indicate fluids of differing composition. The quartz-(ex-)phengite lens is interpreted to be an intrusion of the UHP fluid responsible for the metasomatic alteration at the contact which cooled and crystallised before fully interacting with the ultramafic body. Its existence demonstrates an abundance of fluid infiltrated the body and more metasomatism may have occurred if conditions had permitted it.

## 2.7. Summary

Garnet websterites observed within metamafic-ultramafic bodies at nine key localities in the WGR, western Norway have various structural relationships and textures suggesting they have a range of origins and evolutions. They occur as small-to-large domains within eclogite and garnet peridotite bodies, as isolated bodies and in association with laterally zoned veins within garnet peridotite bodies. Their textures range from coarse-ultracoarse grained and decussate to medium grained, granoblastic and lineated. Garnet websterites which form vein selvages in the garnet peridotite domains in the bodies at Svartberget, Kolmannskog and Myrbærneset are confidently assigned metasomatic origins because of their structural association with fluid-related features; they are considered to have formed by alteration of the host ultramafic bodies. A metamorphic origin is inferred for the garnet websterite domains in the bodies at Nybø, Kolmannskog and Myrbærneset and the entire Båtneset body due to the absence of veining features or associated concentrations of volatile-bearing phases and textural similarity to other metamorphic domains of the bodies unlike at Svartberget. A combined origin is inferred for the garnet websterite domains of eclogite bodies at Grytting, Årsheimneset, Eiksunddal and Remøysunde due to the presence of veining features indicative of externally-derived fluids and/or concentrations of volatilebearing phases. These are considered to have formed by metasomatism of metamorphic garnet websterite domains.

A range of veining features related to both HP-UHP and retrograde conditions are classified into irregular segregations, small veinlets and well-established veins depending upon their sizes, shapes, geometries and mineralogies. The smaller-scale segregations and veinlets were found in the country rocks and within eclogite bodies and domains. Well-developed veins were found in both eclogite and garnet peridotite bodies and garnet websterite domains. Veinlets, most segregations and clinopyroxenite veins in the mafic-ultramafic bodies are thought to be associated with internally-derived fluids and component redistribution processes as no additional components were required for their formation. The dominant feature type considered to be associated with externally-derived fluids are laterally zoned, garnetite-clinopyroxenite-websterite veins which are seen in garnet peridotite and garnet websterite domains in the bodies at Remøysunde, Årsheimneset, Kolmannskog, Myrbærneset and Svartberget. Those at Årsheimneset and Svartberget are particularly well established and provide strong evidence of a metasomatic, HP-UHP origin of garnetite. Textural and mineralogical characteristics of those at Myrbærneset including their finer grained, granoblastic textures and absence of biotite and rutile suggest they have experienced a different evolution to those at the other localities.

The vein assemblages with increased garnet, pyroxene and addition of biotite suggest metasomatism was associated with Si-K-Al-H<sub>2</sub>O-rich fluids sourced from the adjacent country rocks. No features indicative of partial melting in the surrounding country rocks can be assigned confidently to HP-UHP conditions but relict phases of clinopyroxene and garnet suggest they transformed to eclogite-facies assemblages. Fluids are considered to have infiltrated the bodies along hydrofractures or propagated microfractures due to the vein geometries observed. The banded unit at the southeast contact at Svartberget is confidently interpreted to be an UHP feature formed through the metasomatism of the UHP body contact by Si-K-Al-H<sub>2</sub>O-rich fluids (of which a pegmatitic lens within it represents)

due to its concentric, banded appearance and the abundance of biotite relative to the garnet peridotite body. Differences in the assemblages (rarity of garnet, increased abundance of biotite) and zone widths of the metasomatic bands at the contact with those of the interior veins are inferred to indicate a difference in their fluid-rock chemical system.

## Chapter 3 – Petrography

## 3.1. Introduction

In this chapter I present petrographic descriptions of the rock types at the UHP localities visited and observations relevant to understanding the origin and evolution of the rocks and fluid-rock interaction. Observations were made of thin sections of hand samples collected in the field and of previously made thin sections using both optical and scanning electron microscopes.

## 3.2. Aims and Objectives

The aims of petrographic study were to: observe features not visible in the field in order to further determine the evolutionary pathway of the rocks; provide detailed context for geochemical data and thermobarometric and age estimates; determine the suitability of rock samples for geochemical analysis with respect to their degree of alteration and mineral assemblages; and to select particular minerals and crystals for geochemical analysis depending on their textural relationships, shapes and sizes.

To achieve the above aims, the objectives of petrographic study were to:

- identify the full mineral assemblage of each rock type and vein domain to produce comprehensive descriptions of the rocks
- identify inclusions and inclusion suites such as in garnet porphyroblasts to gain insight into the protolith and prograde evolution of the rock
- identify compositional change such as colour zoning and to determine evolutionary stages in the chemical system
- petrographically confirm UHP conditions were experienced by observing minerals indicative of UHP metamorphism such as coesite, diamond and phase exsolution
- identify rocks with peak parageneses showing textural and chemical equilibrium so to select crystals to yield reliable geochemical data for thermobarometric estimates
- determine structural and textural relationships and the relative timing of mineral growth to determine the deformation-metamorphic-metasomatic evolution of the rocks

- identify features indicative of fluid abundance such as fluid inclusions and hydrous multiphase inclusion assemblages to determine the relative timing of fluid activity, fluid parameters and fluid sources at UHP conditions
- identify retrogressive features such as mineral reactions, exsolved phases and alteration to determine the post-peak evolution of the rocks

Rock samples from all HP-UHP localities were studied with the additional aims of determining the common and particular experiences of the mafic-ultramafic bodies and country rocks within the UHP terrane and to further characterise differences between the textural types of eclogite and garnet websterite and the veins at the Myrbærneset locality.

## 3.3. Data Acquisition

30 micron-thick polished thin sections of all samples from UHP domains were prepared at various laboratories as detailed in the Appendix [B1]. Photographs of typical and atypical features in the thin sections were captured with a Lumenera Infinity 2 camera and Infinity Analyze software programme. A Jeol 6400 Scanning Electron Microscope (SEM) at the School of Earth, Atmospheric and Environmental Sciences (SEAES), University of Manchester was used to further image the thin sections and qualitatively identify phases. Sections were carbon coated as required for use in the SEM. It was operated in backscatter electron (BSE) mode with a 1.5nA beam current, a 20kV accelerating voltage and a 15mm working distance. Software programme SemAfore was used to capture the BSE images. Full methodologies are presented in the Appendix [B1-2]. A selection of samples and previously made thin sections provided by my supervisors G. Droop and S. Cuthbert from the localities visited was also incorporated into the study. These were particularly vital for the Grytting locality as collection of new samples is not permitted.

#### 3.4. Data Presentation

Assemblage tables of hand samples from all UHP localities are presented on the Appendix CD [C]. Petrographic descriptions of all hand samples collected on the field trip and thin sections analysed are also presented in the Appendix [D]. Petrographic descriptions of the lithologies observed at each key locality are presented in the locality overviews in the Appendix [A1-9]. Mineral assemblages written in Appendix A are in the format [major,

minor (accessory) *retrograde*]. Mineral abbreviations are as listed in *Table 0.11b*. Sample reference codes are written in [] brackets.

Below are petrographic descriptions and images of the features observed relevant to the aims and objectives of study. Additional images are presented in the Appendix [A1-9].

#### 3.4.1. <u>Country Rocks</u>

The dominant matrix of the gneisses and schists is a fine intergrowth of biotite and plagioclase (*Figs. 3.1a-b*) with or without quartz, alkali feldspar (microcline), garnet, epidote, kyanite or sillimanite, white mica, amphibole and accessory apatite, zircon, titanite, opaques, monazite, xenotime and allanite. Coarser crystals of biotite also form discrete seams often with sillimanite and white mica and when present define a spaced foliation. More schistose layers are richer in biotite, garnet, white mica, kyanite or sillimanite and contain rutile in addition to opaques. Not all phases are present in all gneisses adjacent to the main bodies at the localities visited. For example, those at Remøysunde are particularly kyanite/sillimanite-rich whilst others are not, those at Nybø are epidote-rich and those at Svartberget are alkali feldsparrich.

Biotite shows dark brown-green-straw pleochroism in the gneisses reflecting Fe-rich compositions (*Fig. 3.1a*) and orange-colourless pleochroism in the schists reflecting Mg-rich compositions (*Fig. A2.2e*). Plagioclase shows albite and pericline multiple twinning or no twinning and forms larger crystals as well as the fine intergrowth. Quartz typically forms seams between biotite-plagioclase-rich bands of large, granoblastic crystals, smaller, irregular crystals and sub-grain aggregates. Some larger crystals undergo undulose extinction and have irregular, lobate boundaries, multicrystalline textures and neoblasts at their edges. They contain numerous planar trails of fluid inclusions (see section 3.4.9.). Epidote forms small anhedral (irregular/atoll)-euhedral blasts in the matrix and is more common in biotite-rich bands. Microcline forms small, granoblastic crystals with crosshatched twinning. Sillimanite forms small crystals which often have a fine intergrowth of biotite and plagioclase along the rims and is typically present in pressure shadows of garnet blasts, included in garnet or within quartz-rich seams (*Figs. 3.1g,h*). Some white mica



Figure 3.1: Images of features of the country-rocks. a-b) photographs showing the fine intergrowth of biotite and plagioclase in the matrix of the host gneisses [Årsheimneset, QC26D-3A] x2.5mag, field of view (FOV) 4.2mm; c) photographs showing atoll-textured garnet in the host gneisses [Almenningen, QC13A-2A] x6.3 mag, FOV 1.64mm, PPL-XPL; e) photograph showing tiny inclusions in garnet [Årsheimneset, QC26C-5B] PPL, x6.3 mag, FOV 1.63mm; f) BSE SEM image showing tiny inclusions in garnet [Årsheimneset, QC26C-B2-A]; g-h) Photographs of white mica in a quartz seam showing partial alteration along its edges [Kroken, QC14C-3A] PPL-XPL, x2.5 mag, FOV 4.2mm.

crystals are surrounded by smaller flakes of biotite at their margins and along their cleavages (*Figs. 3.1g,h*).

Garnet in the gneisses typically form small (~<1mm-diameter), an-euhedral, isolated, inclusion-poor blasts in the matrix some of which have skeletal-atoll appearances with quartz ±mica cores (Fig. 3.1c,d). Those in the metapelitic schists are often larger inclusion-rich porphyroblasts with quartz and unfoliated white mica in their pressure shadows and foliated mica-rich bands wrapped around them. Crystals are most euhedral when in contact with quartz and feint concentric colour zoning is observed in some blasts. The garnets from Årsheimneset are Mn-rich. Inclusions observed include mono-, multi-crystalline quartz, biotite, white mica, biotite-plagioclase, amphibole, apatite, epidote and rutile or titanite. Some garnets also contain multiphase solid inclusions (MPSIs) containing feldspars, micas and titanite (see section 3.4.8.). No regular inclusion suites or patterns are observed although the larger, quartz-inclusion-rich garnet porphyroblasts in metapelitic schist [NFD 9334] have a ring of larger quartz inclusions in their middle zones. Some garnets in gneiss at Årsheimneset and Remøysunde [e.g. QC26C, QC29F] dominantly contain tiny, 10μmdiameter, euhedral, hexagonal-cubic, monocrystalline inclusions (Figs. 3.1e,f) with a variety of compositions including Ca (calcite?), Al (corundum?), Al-Si (kyanite?), Si-Al-Na (albite), Mn-Fe (Fe rhodochrosite?), Si (quartz), K, Ca-Al, Si-Al-K (white mica?) and Fe (magnetite?).

Apatite forms sub-euhedral crystals distributed within the matrix. Zircons are hosted in biotite and in the matrix and are typically <100µm in diameter. Zircon and monazite grains are described in more detail in Chapter 8. Xenotime forms tiny crystals typically attached to zircon crystals or euhedral ~50µm-diameter grains in the matrix. Opaque phases are Ti-Fe and Fe oxides (ilmenite, magnetite).

The partially retrogressed felsic, eclogite-facies rocks consist of quartz, garnet, a biotite-plagioclase intergrowth (after phengite) and an amphibole-plagioclase intergrowth after clinopyroxene of which skeletal relics persist in some sections (e.g. QC27B). Epidote and titanite are also present. Titanite forms irregular grains in the matrix and coronas around small rutile grains. Amphibole also forms an intergrowth with plagioclase ±biotite and an opaque phase after garnet.



Figure 3.2: Images of features in the garnet peridotite bodies. a-b) photographs showing the granoblastic matrix with inclusion-rich, anhedral garnets, opaque blebs and clinohumite [Svartberget, QC36C-10C] PPL-XPL, x2.5 mag, FOV 4.2mm; c-d) photographs showing the opaque blebs and blades in clinopyroxene in peridotite [Svartberget, QC36A-7B] PPL-XPL, x2.5 mag, FOV 4.2mm; e-f) photographs showing spongy, inclusion-rich garnets [Svartberget, QC36C-5B] PPL-XPL, x2.5 mag, FOV 4.2mm;g-h) photographs showing inclusion-poor, granoblastic garnets in biotite-rich olivine garnet websterite [Myrbærneset, QC34B-4B] PPL-XPL, x2.5 mag, FOV 4.2mm.

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# 3.4.2. Garnet Peridotites

Garnet peridotites are composed of clinopyroxene, orthopyroxene, garnet and olivine with or without dolomite-magnesite, minor biotite, amphibole, clinohumite and accessory opaques, spinel, apatite, allanite and zircon. The Kolmannskog peridotite has both olivine-orthopyroxene-rich, clinopyroxene-poor harzburgite domains and clinopyroxene-richer lherzolite domains. The thin sections from Svartberget and Myrbærneset show a lower abundance of olivine forming olivine garnet websterite assemblages. They have similar medium grained textures: orthopyroxene, clinopyroxene and olivine (and biotite and amphibole where present) form a polygonal, granoblastic matrix (Figs. 3.2a,b). Many pyroxene crystals contain tiny inclusions and opaque, ovoid-spherical 'blebs' and blades within their cores or throughout the crystals, particularly at Svartberget (*Figs. 3.2c-d*). They are dominantly composed of Fe with small amounts of Cr and Ti. Twinned clinopyroxene is common. Some orthopyroxene crystals contain fine exsolution lamellae visible in prismatic sections in extinction under cross-polarised light (XPL). Garnets in the olivine garnet websterites are irregularly shaped and contain many small, ovoid inclusions of clinopyroxene, orthopyroxene and olivine forming 'spongy' cores (Figs. 3.2a,b,e,f). There is no apparent optical continuity between inclusions. Garnets can also contain an-euhedral biotite, amphibole, carbonate, zircon, and opaques with spinel and have inclusion-poorer rims where present. Some form granular aggregates with inclusion-rich cores. Garnets in the Myrbærneset olivine garnet websterite also form granular aggregates of sub-euhedral grains which contain few, large inclusions of clinopyroxene (Figs. 3.2q,h). Those in the Kolmannskog peridotite are inclusionpoor and inclusion-rich and have anhedral forms but are considerably more retrogressed. Olivine forms small, an-euhedral, inclusion-poor grains with a heterogeneous distribution and also larger crystals in the Kolmannskog harzburgite. Most are characterised by diagnostic heavy fractures some of which contain an opaque phase. The largest crystals in the Svartberget olivine garnet websterite contain ovoid-euhedral inclusions of clinopyroxene, orthopyroxene, biotite and opaques.

Biotite, amphibole, carbonates and clinohumite have an irregular distribution. Biotite shows orange-pale green-colourless pleochroism (Mg-rich). It occurs as subhedral, granoblastic, similarly sized crystals in the matrix and as smaller, euhedral and

rounded crystals included in garnet. It is rare in the olivine garnet websterite from Svartberget but is a major phase in that from Myrbærneset. Amphibole is pale green coloured and weakly pleochroic (Mg-rich) and forms texturally-equilibrated, granoblastic crystals in the matrix associated with biotite. Contacts with garnet are less sharp and straight. A texturally late, more strongly coloured amphibole is also present (see section 3.4.6.). Carbonate phases occur as inclusions within garnet, clinopyroxene and orthopyroxene and within the matrix and typically consists of a core of magnesite and mantle of dolomite. Calcite often forms thin rims on matrix crystals. Clinohumite is only seen in one thin section of the olivine garnet websterite from Svartberget. It is strongly coloured with orange-beige pleochroism and forms anhedral-interstitial crystals with a heterogeneous distribution. It is partially replaced with fine, opaque, exsolved streaks and blebs of Ti-Fe (ilmenite) and Fe oxide or fully transformed to an olivine-ilmenite pseudomorphic intergrowth.

Zircons are very rare and tiny with 10-20µm diameters and are hosted in garnet. Apatite occurs sporadically in the matrix. Opaque phases occur throughout the matrix forming interstitial crystals with holly-star (concave) to convex shapes and consist of Fe-Ti, Ti-Fe, Fe, Fe-Cr and Fe-Cr-Ti oxides and Fe-Cu and Cu-Fe sulphides. Fe and Ti-Fe oxide compositions are often together in a single grain as domains and numerous Fe-Ti-Cr compositions occur together as exsolved streaks and lamellae producing tiger-stripe patterns at Svartberget. Spinel is dark green in plane-polarised light (PPL) (hercynite) and occurs only within opaque crystals always forming with an edge of the grain.

#### 3.4.3. Eclogites

Eclogites consist dominantly of garnet and clinopyroxene with or without quartz, white mica, zoisite, minor biotite, amphibole and carbonate and accessory rutile, apatite, zircon, monazite, allanite, epidote, opaques and baryte. No distinction of assemblage is made between textural types. Garnets in eclogite have a range of forms from euhedral porphyroblasts to anhedral, granular aggregates interstitial with clinopyroxene (*Fig. 3.3a,b*) to megacrystic aggregates (*Fig. A8.2a*). Variation is seen on a hand sample and thin section scale. Porphyroblasts (<4cm in diameter) are often full of large and small clinopyroxene and amphibole inclusions, especially in their cores (*Fig. 3.3a*), and also rutile, quartz, apatite, allanite, multiphase inclusions

and/or many tiny inclusions. Some quartz inclusions in samples from Arsheimneset have palisade and polycrystalline textures indicative of prior coesite (Fig. 3.3c). Many amphibole inclusions have a strong green colouration and are connected by fractures, some replace clinopyroxene inclusions but some are isolated and have a more turquoise colouration. Smaller garnet blasts generally just contain clinopyroxene, rutile and amphibole (Fig. 3.3a) and interstitial garnet is inclusionpoor. Inclusion distribution in the porphyroblasts is typically denser in cores and mid zones. Some from Nybø have concentric inclusion patterns (Fig. A1.3a). The cores and mid zones of the mauve porphyroblasts in clinopyroxenite bands at Nybø are full of amphibole and clinopyroxene inclusions (Fig. 3.3e). Some cores in garnets from the eclogite at Myrbærneset have 'spongy' appearances full of clinopyroxene and rutile inclusions (Fig. 3.3f,g). At this locality, the larger garnets contain few large inclusions but numerous tiny inclusions. Clinopyroxene textures range from decussate, granoblastic to interstitial depending upon garnet forms and grain size. They are typically inclusion-poor containing rutile and quartz. Quartz grains have a range of crystal sizes in the matrix and often show multicrystalline textures.

Amphibole is sporadically distributed throughout the matrix and is pale green coloured and weakly pleochroic (Mg-rich). Some grains have exsolution features including blebs or lamellae both inclined and concordant to cleavage. Coronas of more strongly coloured amphibole are also present (*Fig. 3.3a*) (see section 3.4.6.). Biotite shows orange-colourless pleochroism (Mg-rich) and has a heterogeneous distribution. Amphibole and biotite are typically found together.

Apatite is included in both garnet and clinopyroxene and found in the matrix. Some crystals contain short, aligned, black needles. Rutile is distributed throughout the thin sections of eclogites or along aligned, sparse, discontinuous trails with amphibole-plagioclase selvages (*Fig. 3.3f*) often along garnet boundaries (see *Fig. 3.3h*). Inclined opaque blades and fine lamellae of Ti-Fe oxide (ilmenite) are seen in some grains (*Fig. 3.3a*). Opaque phases are Fe and Ti-Fe oxides, Fe-Cu oxides and Fe, Fe-Cu and Cu-Fe sulphides. They occur in the matrix and along fine fractures. Zircon grains are typically hosted in garnets but are also found in clinopyroxene crystals. Allanite is hosted in garnets and in the matrix.



Figure 3.3: Images of features in eclogite bodies. a) photograph of a garnet porphyroblast in eclogite [Årsheimneset, A632-8A] PPL, x2.5 mag, FOV 4.2mm; b) photograph of interstitial garnet in eclogite [Årsheimneset, A631B-9B] PPL, x2.5 mag, FOV 4.2mm; c) polycrystalline quartz inclusion in garnet in eclogite [Årsheimneset, A605-3C] XPL, FOV 1.64mm; d) polycrystalline and palisade-textured quartz inclusion in zoisite in eclogite [Verpeneset, N70-Q10-1A] XPL, FOV 0.65mm; e) scan of thin section between cross polarised sheets showing inclusion-rich mauve porphyroblasts in clinopyroxenite band and inclusion-poorer interstitial garnet [Nybø, QC15F] FOV 2cm; f-g) photographs showing trails of rutile and inclusion-rich garnet cores [Myrbærneset, QC34A-2C] PPL-XPL, x2.5mag, FOV 4.2mm; h) photograph showing rutile grains along garnet-clinopyroxene contacts [Årsheimneset, A632-1A] PPL, x2.5 mag, FOV 4.2mm.

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# 3.4.4. Garnet Websterites

In general, garnet websterites are composed of clinopyroxene, orthopyroxene and garnet with or without quartz, biotite, carbonate, amphibole, clinohumite and accessory rutile, apatite, opaques, zircon, spinel, monazite and allanite. They have granoblastic, decussate and/or interstitial textures and varying grain sizes and modal abundances of phases. No distinction of assemblage is made between textural types. Garnets have a range of forms including small, inclusion-poor grains (e.g. at Svartberget; Figs. 3.4a,b), inclusion-rich porphyroblasts with or without inclusionpoor rims (e.g. at Årsheimneset; Figs. 3.4c,d), interstitial aggregates with or without large inclusions (e.g. at Arsheimneset; Fig. 3.4e), granoblastic grains and sub-grain aggregates (e.g. at Myrbærneset; Figs. 3.4f,g) and inclusion-poor, elongate, aligned crystals (e.g. at Nybø; Figs. 3.4h,i). Like eclogite garnets, a range of forms can be present at one locality and within one thin section. For example, the lineated websterite body at Båtneset has a bimodal size population of garnet having finer, streaked-out granular aggregates and occasional <1cm-diameter porphyroblasts. Porphyroblastic garnets contain combinations of clinopyroxene, orthopyroxene, amphibole, biotite, quartz, white mica, carbonate, apatite, rutile, zircon, opaques and multiphase inclusions (see section 3.4.8.). Rims typically only include rutile, biotite and zircon. At Årsheimneset, biotite is only included in the rims with the cores containing only clinopyroxene, amphibole, quartz and rutile. Pyroxenes in the coarse websterite domains form decussate, elongate laths which can be intergrown. Clinopyroxene also forms irregularly shaped crystals interstitial with garnet (Fig. 3.4e) or equant, granoblastic crystals such as at Myrbærneset (Figs. 3.4f,q). Large laths include garnet, orthopyroxene, biotite, carbonate, rutile, quartz and numerous planar trails of fluid inclusions (see section 3.4.9.). Orthopyroxenes can include clinopyroxene, biotite, quartz and fluid-inclusion trails and typically show fine exsolution lamellae in extinction under XPL. Some have kink bands and patchy extinction. In the finer grained websterite body domains (Myrbærneset, Båtneset, Nybø), pyroxenes form granoblastic, more equant crystals and contain fewer inclusions (Figs. 3.4f,g). Occasional coarser, elongate crystals of orthopyroxene and clinopyroxene with irregular boundaries occur at Myrbærneset and Nybø and one thin section from Myrbærneset has a coarser, more decussate texture. At Båtneset, orthopyroxene is more common in the garnet-rich bands.



Figure 3.4: Images of garnets in garnet websterites. a-b) small, inclusion-poor, anhedral grains [Svartberget, QC36D] PPL-XPL, x2.5mag, FOV 4.2mm; c-d) scans of thin section showing inclusion-rich garnet porphyroblast [Årsheimneset, QC26HC], PPL-XPL, FOV 2cm; e) photograph showing interstitial garnet [Årsheimneset, A634-4B], PPL, x2.5 mag, FOV 4.2mm; f-g) photographs showing granoblastic garnets [Myrbærneset, QC34E-5A], PPL-XPL, x2.5 mag, FOV 4.2mm; h-i) photographs showing the lineated fabric of garnet websterite [Δ1308] PPL-XPL, x2.5 mag, FOV 4.2mm-wide.

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Irregular biotite crystals within the matrix of garnet websterites are typically associated with orthopyroxene, garnet and poikiloblastic and an-euhedral, prismatic amphiboles. Biotite can be quite coarse and can define a sparse foliation where more abundant such as at Båtneset and Myrbærneset. It shows orange-colourless pleochroism (Mg-rich) and hosts numerous tiny zircons with radiation halos. It is included in all major phases including garnet (Figs. 3.5a,b). Amphibole forms pale green and weakly pleochroic crystals (Mg-rich) with sharp contacts often associated and intergrown with biotite or large poikiloblasts hosting orthopyroxene, clinopyroxene, garnet and biotite (e.g. *Figs. A2.3g,h*): contacts with garnet and biotite are often lined by a fine rim of plagioclase. Many crystals contain oblique, lower Ca exsolution lamellae (Fig. 3.5c). Mono- and multi-crystalline quartz is present within the Årsheimneset websterite matrix, although considerably less than in the associated eclogite, and as inclusions in garnet and pyroxene which show multi- and poly-crystalline textures (Fig. 3.5d). Carbonate only forms a major phase in particular patches in the garnet websterites at Årsheimneset and Remøysunde and Grytting. At Grytting it includes clinopyroxene, orthopyroxene, biotite and rutile. At Årsheimneset it hosts biotite, rutile, clinopyroxene, amphibole, garnet and zircon inclusions and has irregular contacts with matrix phases (Fig. 3.5e,f). Amphibole-carbonate contacts can be marked by reaction textures and carbonate appears to be resorbing garnet at some contacts. It is a siderite-magnesite phase but rims are often marked by ankerite and calcite coronas. Almost completely exsolved clinohumite is observed in the garnet websterite at Eiksunddal.

Apatite is found in the matrix, sometimes in concentrations, and in garnets and often contains aligned, short black needles (*Fig. 3.5g*). Rutile occurs as small, irregular-ovoid, interstitial grains in the matrix and as inclusions within garnet, clinopyroxene and orthopyroxene. Large (>1cm-diameter) and elongate rutile grains are also observed (*Fig. A3.3e*). In garnet some rutile inclusions are very small and occur in clusters forming small grains and unaligned needles. Zircons are mostly ovoid and consistently ~10µm in diameter. They are hosted in garnet, biotite and clinopyroxene. Opaque phases include Fe-Ti and Ti-Fe oxides and Fe-Ti-Cr oxides and various Fe, Ni, Fe-Ni, Ni-Fe and Fe-Cu sulphides. They typically have interstitial textures and are associated with biotite-amphibole aggregates and rutile (*Fig. 3.5h*). Some at Nybø form disseminated trails along the lineation. Some larger crystals have tiny, dark-green Al-Fe(-Cr) hercynite spinel within them or at their edges.



Figure 3.5: Images of features in garnet websterites. a) photograph showing biotite included in garnet [Årsheimneset, A347-1C] PPL, x2.5 mag, FOV 4.2mm; b) photograph showing biotite included in orthopyroxene [Svartberget, SV1A-1A] PPL, x2.5 mag, FOV 4.2mm; c) BSE SEM image showing exsolution lamellae in amphibole [Årsheimneset, QC26G-B3-A]; d) photograph showing polycrystalline quartz inclusion in garnet in garnet websterite [Årsheimneset, A347-10C] XPL, x6.3 mag, FOV 1.64mm; e-f) photographs showing carbonate with inclusions of biotite, amphibole, clinopyroxene, garnet and rutile [Årsheimneset, QC26G-5F], PPL-XPL, x2.5 mag, FOV 4.2mm; g) BSE SEM image showing Fe oxide needles in apatite [Årsheimneset, A347-B6-A]; h) photograph showing opaque phases associated with biotite and rutile and with spinel [Årsheimneset, QC26G-6A], PPL, x2.5 mag, FOV 4.2mm].
# 3.4.5. Laterally-Zoned Veins

Petrographic observations of veining features focus on laterally-zoned veins from Årsheimneset, Svartberget and Myrbærneset. The similarly coarse grained, decussate-textured veins from Årsheimneset and Svartberget are described first.

Websterite selvages contain orthopyroxene, clinopyroxene and biotite with or without major garnet and amphibole and with accessory rutile, zircon, monazite and opaques. Clinopyroxene is typically the more abundant pyroxene and elongate laths can contain biotite, streaks and larger inclusions of orthopyroxene and garnet when present in the matrix. Orthopyroxene also forms elongate laths; they include clinopyroxene and biotite. Amphibole form poikiloblastic crystals hosting clinopyroxene, orthopyroxene and biotite (*Fig. 3.6a,b*). Garnets form small, occasional, inclusion-poor blasts in the veins at Svartberget and Årsheimneset and also large, clinopyroxene and amphibole inclusion-rich porphyroblasts in veins at Årsheimneset similar to those in the eclogite. Rutile forms rare, tiny grains. Zircons are ovoid with more euhedral cores and typically ~10-15µm in diameter although some have up to 50µm diameters. Opaque phases are Fe oxide, Th-U oxide and Fe-Cu sulphide.

Clinopyroxenite selvages contain clinopyroxene, biotite, amphibole and accessory apatite, rutile, zircon and opaques. Minor quartz and garnet are also sometimes present in those from Årsheimneset. The decussate textures across the contact with websterite selvages are continuous which is marked lastly by the loss of orthopyroxene. Some clinopyroxenite selvages at Svartberget have patches with finer grained, granoblastic textures. Biotite is seen directly replacing orthopyroxene in a vein at Svartberget (*Fig. 3.6c*) and is both well distributed in the clinopyroxenite matrix and hosted within clinopyroxene crystals (*Fig. 3.6d*). It shows orange-greencolourless pleochroism (Mg-rich) and hosts zircons. Amphibole forms poikiloblastic crystals and wide rims around clinopyroxenes and is intergrown with biotite. When present, quartz occurs as interstitial, multicrystalline grains and some inclusions show polycrystalline textures (*Fig. 3.6e*). The rare, small garnets within the clinopyroxenite selvage have inclusion-rich domains with tiny inclusions of biotite, apatite and amphibole. They are associated with matrix biotite which appears to resorb them. Rutile and zircons are less common within the clinopyroxenite selvages. Rutile grains



Figure 3.6: Images of features in laterally-zoned veins. a-b) photographs showing pale coloured, poikiloblastic amphibole hosting clinopyroxene, orthopyroxene and biotite [Svartberget, QC36GB-1A], PPL-XPL, x2.5 mag, FOV 4.2mm; c) photograph showing replacement of orthopyroxene by biotite in garnet clinopyroxenite selvage/core [Svartberget, QC36UA-5A] PPL, x6.3 mag, FOV 1.64mm; d) photograph showing biotite included in clinopyroxene in clinopyroxenite selvage [Svartberget, QC36GA-7A] PPL, x2.5 mag, FOV 4.2mm; e) polycrystalline quartz inclusion in clinopyroxene in vein selvage [Årsheimneset, A631B-3E] XPL, FOV 1.64mm; f) scan of thin section showing zircon-rich glimmerite selvage, particularly on the garnetite side and garnet blasts with interstitial matrix [Svartberget, QC36GA] PPL, FOV 2cm; g) scan of thin section showing subhedral, inclusion-poor garnets [Årsheimneset, QC26JA] PPL, FOV 2cm; h-i) scans of thin sections showing garnetite mass with inclusions [Svartberget, QC36UB] PPL-XPL, FOV 2cm;

are included in clinopyroxene and apatite and distributed in the matrix and concentrated in trails parallel to the zoning particularly adjacent to the vein core: elongate grains are aligned along the transition. Grains increasingly contain ilmenite exsolutions towards the vein core. Apatite contains aligned, short black Fe-rich needles, rutile and fluid inclusions. Opaque phases are Fe-Cu sulphides: one grain occurs with quartz in a vein selvage from Svartberget.

Glimmerite selvages (pure biotite) are only observed in veins at Svartberget. They contain numerous inclusions of zircon and monazite with sizes up to 150µm in diameter although most are ~50µm. These accessory phases are most concentrated in a band on the garnetite side of the selvage (*Fig. 3.6f*). Garnet clinopyroxenite selvages contain accessory rutile, apatite and zircon. Garnets form granular aggregates with finer, interstitial, granoblastic clinopyroxene or larger porphyroblasts with few inclusions of clinopyroxene, amphibole, needles and grains of rutile, biotite, apatite and multiphase inclusions. Garnets also contain spinel and opaques and trails of tiny, aligned needles and blades of a highly birefringent phase (rutile?). Rutile grains are often concentrated towards the centre and the margins of the zone and some grains contain opaque ilmenite blades and needles.

Garnetite vein cores have a range of assemblages and forms individual, an-euhedral blasts with interstitial phases (Fig. 3.6f,g) or more solid, aggregate masses with large inclusions (Figs. 3.6h,i). Those at Årsheimneset contain major biotite and guartz, minor apatite, white mica and accessory rutile, zircon, monazite, opaques and allanite. Quartz and mica sometimes form central zones with garnet either side. Blasts have straight, more euhedral crystal faces when in contact with quartz (Fig. 3.6g). Matrix phases are typically included within the garnets. Some garnets are inclusion-poor such as some from Årsheimneset (Fig. 3.6g) although garnets with inclusions of biotite, clinopyroxene, quartz, apatite, amphibole, zircon and domains with lots of tiny inclusions are also present: biotite and rutile are dominantly found in garnet rims and some garnets contain only quartz. Quartz inclusions at Arsheimneset often show palisade and polycrystalline textures (Figs. 3.7a,b) and some quartz crystals in the matrix of the vein cores show multicrystalline textures (Figs. 3.7c,d). Aggregate masses of garnet at Svartberget (Figs. 3.6h,i) contain small inclusions of biotite, apatite, amphibole and multiphase inclusions and coarse, an-subhedral blasts occur within a matrix of interstitial biotite and/or clinopyroxene (Fig. 3.6f), minor



Figure 3.7: Images of features in laterally-zoned veins. a) palisade-textured quartz inclusion in garnet in vein core [Årsheimneset, ARS-1-1A] XPL, FOV 0.65mm; b) polycrystalline quartz inclusion in garnet in vein core [Årsheimneset, ARS-1-1E, XPL] FOV 0.65mm; c-d) multicrystalline quartz lying interstitially between garnets in a vein core [Årsheimneset, QC26JB] PPL-XPL, FOV 4.2m; e-f) scans of a thin section of a zoned vein showing the fine grained, granoblastic texture and Fe oxide opaque-rich nature of the core and selvage [QC34F] PPL-XPL, FOV 2cm; g) photograph showing opaque-spinel-rich garnetite [Myrbærneset, QC34F-8A] PPL, x2.5 mag, FOV 4.2mm.

amphibole, white mica, apatite and accessory zircon, rutile, monazite, apatite, allanite and opaques. Accessory guartz is also found included in garnet but is not present in the matrix. They range from inclusion-poor to inclusion-rich containing inclusions of Ti-rich biotite, clinopyroxene, rutile, apatite, amphibole, monazite, zircon and plagioclase. Garnets also typically contain multiphase inclusions: these are described in more detail in section 3.4.8.. Clinopyroxene crystals in the vein cores at Svartberget are also seen to include monocrystalline quartz. Biotite shows pale green-orange-colourless pleochroism and contains numerous, small and a few larger inclusions of zircon and allanite. It is generally coarser and more aligned than that in the vein selvage. Quartz in the Årsheimneset veins is interstitial between garnet blasts and shows multicrystalline, granoblastic textures (Figs. 3.7c, d). It hosts biotite, rutile and zircon and trails of fluid inclusions (see section 3.4.9.). Sub-grain boundaries range from straight and sharp to irregular and blurry and most exhibit undulose extinction. Quartz in the foliated white mica-quartz vein core observed at Årsheimneset contains small inclusions of white mica. The white mica in the matrix is partially altered to biotite-plagioclase-microperthitic feldspar-quartz intergrowths. Apatite crystals can contain aligned, tiny black needles, small rutile blebs and fluid inclusions. Rutile has a heterogeneous distribution and is present in disseminated bands towards the edges of some vein cores or in seams within the core. It is hosted in garnet and quartz. Zircon is present within all vein cores generally hosted within garnet, biotite, clinopyroxene and rutile: sizes range up to 150µm-diameters. Monazite is often associated with apatite. Opaque phases in Svartberget vein cores are Fe, Fe-Cr, Cr-Fe, Fe-Ti, Ti-Fe, Mo, Fe-Cu-Mo and Fe-Cu oxides and Fe-Cu and Cu-Fe sulphides: oxides are often exsolved, multiphase grains.

In contrast all the zones of the veins at Myrbærneset have medium-grained, granoblastic textures similar to the peridotite host (*Figs. 3.7e-g*). Garnetite cores consist of aggregates of smaller grains with intergranular clinopyroxene and accessory opaque phases (Fe and Ti-Fe oxides) with spinel rather than rutile (*Fig. 3.7g*). Clinopyroxenite selvages are very thin (<1cm) and biotite is absent. Garnet in the garnet websterite selvages form irregularly shaped sub-grain aggregates and orthopyroxene is the dominant pyroxene; biotite, again, is absent. The selvages also contain occasional larger, irregularly shaped orthopyroxene and clinopyroxene crystals which typically contain tiny, ovoid inclusions.



Figure 3.8: Images of retrogressive features. a) BSE SEM image showing symplectitic alteration of clinopyroxene in an eclogite [Myrbærneset, QC34A-B9-A]; b) photograph showing the coarsening of symplectite behind the reaction front [Svartberget, QC36HB-1B] XPL, x6.3 mag, FOV 1.64mm; c) photograph of exsolution needles in clinopyroxene [Svartberget, QC36EA-4A], PPL, x6.3 mag, FOV 1.64mm; d) BSE SEM image of exsolution needles in clinopyroxene comprised of quartz (darker) and amphibole (lighter) [QC36GA-B6-A]; e) photograph showing late amphibole corona between garnet and clinopyroxene [Årsheimneset, A605-1A] PPL, x2.5 mag, FOV 4.2mm; f) BSE SEM image of amphibole fringe between garnet and clinopyroxene [Myrbærneset, QC34A-B5-A]; g-h) photographs of more pervasive retrograde amphibole with quartz [Svartberget, QC36HB-2B] PPL-XPL, x2.5 mag, FOV 4.2mm.

## 3.4.6. <u>Retrogression</u>

The most common expression of retrogression is the symplectitic alteration of clinopyroxene to an increasingly coarse clinopyroxene-plagioclase intergrowth (*Figs. 3.8a,b*) which develops into an amphibole-plagioclase intergrowth and poikiloblastic plagioclase around coarse amphibole with continued alteration. It is observed in eclogites, garnet websterites, vein selvages and cores where clinopyroxene is present. A range of degree of symplectitic alteration is observed from small pockets to pervasive alteration and can be seen within one thin section but is most commonly along grain boundaries and fractures. Chemical zoning of clinopyroxene towards the symplectite is observed in the SEM picked out by changes in grey levels (*Fig. 3.8a*). Clinopyroxene crystals in eclogites, garnet websterites and vein zones are also characterised by numerous, aligned needles of quartz and amphibole (*Figs. 3.8c,d*).

Another common retrogressive feature is the partial to complete development of strongly coloured, turquoise-green pleochroic (Fe-rich) amphibole coronas between garnet and clinopyroxene crystals (*Fig. 3.8e*). This is observed in all thin sections where these two crystals meet, although not along every contact. Some contacts amphibole forms domal fringes matched by domes of symplectitic clinopyroxene (*Fig. 3.8f*). Chemical zoning of garnet rims is depicted in the SEM where present. An opaque Ti-Fe phase (ilmenite) is often present in well-developed coronas. The coronas merge with amphibole-myrmekitic quartz and amphibole after clinopyroxene symplectite in the matrix (*Figs. 3.8g,h*). Late amphibole also occurs after biotite although in some sections it appears to occur with it; on the whole amphibole-biotite relationships appear to be gradational between co-genetic to replacive. Amphibole has a browner colouration adjacent to rutile grains and some biotite crystals demonstrating diffusion of Ti. It also forms retrograde intergrowths with plagioclase ±biotite and an opaque phase after garnet in more altered portions.

Thick fractures in olivine and orthopyroxene crystals contain very fine-grained, orange-yellow iddingsite (*Fig. 3.9a*). Some small olivine grains are almost completely altered to iddingsite particularly near orthopyroxene and local domains of orthopyroxene are more pervasively altered to iddingsite. An opaque Fe oxide is present in some fractures (*Fig. 3.9a*). Many orthopyroxene crystals also contain fine exsolution lamellae (*Fig. 3.9b*).



Figure 3.9: Images of retrogressive features. a) photograph showing iddingsite-filled fractures in orthopyroxene and some with opaques [Sørøyane, N5-6A] PPL, x6.3 mag, FOV 1.64mm; b) photograph of fine exsolution lamellae in orthopyroxene [Grytting, Ø16-1G] XPL, x6.3 mag, FOV 1.64mm; c) photograph showing a kelyphite rim between garnet and olivine [Svartberget, QC36A] PPL, x2.5 mag, FOV 4.2mm; d) BSE SEM image of kelyphitic intergrowth containing orthopyroxene (dark background), spinel (lighter wiggles) and Fe oxide (white specks) with clinopyroxene inclusions (lighter ovoid blebs) [Svartberget, QC36A-B2-A]; e) BSE SEM image of retrograde intergrowth of orthopyroxene, Fe(-Ti) oxide, clinopyroxene and spinel between garnet and Fe oxide opaque [Myrbærneset, QC34B]; f) BSE SEM image of intergrowth of Ti-Fe oxide (ilmenite) with orthopyroxene between biotite and Ti-Fe-oxide [Årsheimneset, QC26G-B7-D]; g-h) photographs showing retrograde reaction between garnet and biotite forming a fibrous intergrowth of biotite and plagioclase [Årsheimneset, ARS-1-4A] PPL-XPL, x6.3 mag, FOV 1.64mm.

Most garnet-olivine contacts in olivine garnet websterites are lined by a fine kelyphite of orthopyroxene, hercynite spinel and an Fe oxide with clinopyroxene inclusions inherited from the altered garnet (*Figs 3.9c,d*). An orthopyroxene-rich band often lines the olivine contact. In addition, delicate intergrowths of an Fe oxide opaque phase with spinel with or without clinopyroxene and orthopyroxene occur between opaques, garnet and clinopyroxene (*Fig. 3.9e*). Fine intergrowths of ilmenite and orthopyroxene occur between ilmenite and biotite (*Fig. 3.9f*).

Rims of plagioclase occur between quartz and biotite, between biotite and amphibole and between biotite and garnet. Some garnet-biotite contacts in vein cores and biotite-rich garnet websterites are marked by a feathery-fibrous biotiteplagioclase, amphibole-feldspar-quartz or amphibole-plagioclase-biotite-Fe oxide intergrowths which encroach into the garnet crystal (*Figs. 3.9g,h*). Garnet rims are richer in Mn at these contacts and biotite is green and Fe-richer. White mica in gneisses and veins is seen altering to orange-green biotite in a fine intergrowth with plagioclase initially at crystal edges (*Figs. 3.10a,b*). Biotite-plagioclase intergrowths at biotite-garnet contacts merge with the filling of fractures cutting through garnet crystals in Årsheimneset vein cores. These contain a feathery, zoned assemblage of calcic plagioclase, green biotite, calcic amphibole and cubic opaques (exsolved Fe-Ti and Fe oxides and Fe, Fe-Ni and Fe-Cu sulphides) (*Figs. 3.10c,d*). This assemblage is also present between garnet and biotite particularly near the fractures in garnet. Where the fractures meet adjacent quartz radiating, fibrous needles of a pale brown Fe-Mg silicate (orthoamphibole) are formed (*Figs. 3.10e,f*).

Carbonates are often partially altered to a fine-grained, orange-brown-stained Fe silicate and Fe oxide initially at crystal edges but some is pervasive (*Fig. 3.10g*). Many rutile grains contain a few lamellae and blades of ilmenite. Some in vein cores also contain aligned needles and some in websterites and vein cores are almost totally replaced by ilmenite. These occur within altered domains such as clinopyroxene symplectites and the late fractures in garnet. Rutile is also seen altering to titanite which forms coronas around the grains particularly when hosted within symplectitic clinopyroxenes. Apatite grains within pervasively symplectised domains in eclogite have thin rims of plagioclase.



Figure 3.10: Photographs of retrogressive features. a-b) photographs showing biotite-plagioclase intergrowth along edges of white mica crystals [Årsheimneset, QC26F-2B] PPL-XPL, x2.5 mag, FOV 4.2mm; c-d) photographs showing late fractures cutting through garnet blasts containing biotite, plagioclase and Fe oxide [Årsheimneset, QC26HA-4B] PPL-XPL, x2.5 mag, FOV 4.2mm; e-f) photographs showing radial Fe-Mg silicate (orthoamphibole) protruding into quartz at garnet-quartz-fracture contact [Årsheimneset, QC26HA-4C] PPL-XPL, x2.5 mag, FOV 4.2mm; g) photograph showing alteration of carbonate to an orange-brown stained product [Årsheimneset, A634-1C] PPL, x6.3 mag, FOV 1.64mm; h) photograph showing alteration of clinopyroxene to calcite [Årsheimneset, QC26HB-8A] XPL, x6.3 mag, FOV 1.64mm.

Fine fractures which cut through samples are filled by amphibole with the degree of alteration varying in different phases. Clinopyroxene is also seen altering pervasively to a calcite near a late, amphibolitised fracture in a vein selvage [QC26HB] (*Fig. 3.10h*) and tiny calcite crystals are also observed up biotite cleavages in a garnet websterite from Årsheimneset [QC26G] and in late, cross-fibre veins in the biotite websterite in the Svartberget contact [QC36S].

Sub-amphibolite-facies retrogression in the host gneisses is seen through the alteration of biotite to chlorite and alteration of white mica and plagioclase to sericite. Chlorite is more common nearer to a quartz-tourmaline segregation [QC26D]. Myrmekitic quartz-plagioclase textures and exsolved alkali feldspars are also observed in the host gneisses. Lastly, a fine-grained, granitic assemblage of biotite, twinned plagioclase, microperthitic K feldspar and quartz is present between crystals of white mica and quartz in a vein core from Årsheimneset [A631].

# 3.4.7. Svartberget Contact

The biotite websterite contains clinopyroxene, orthopyroxene, biotite, minor amphibole and accessory opaques and zircon (*Figs. 3.11a,b*). Biotite and clinopyroxene are included within each other. Biotites show orange-colourless pleochroism and some have green edges (*Fig. 3.11a*). Amphibole forms euhedral, pale coloured crystals in the matrix. Opaques form clusters of euhedral crystals associated with amphibole.

The felsic lens consists of quartz and a pseudomorphic intergrowth of biotite, white mica, cross-hatched, microperthitic microcline and plagioclase feldspar. Mica is crystallographically orientated (*Fig 3.11c*) and feldspar forms poikiloblastic crystals (*Fig. 3.11d*). Quartz forms large, unstrained crystals separating mica-feldspar intergrowths and rarely contains small pockets of them (*Figs. 3.11f,g*).

The biotite clinopyroxenite is heavily retrogressed and also contains zircon and apatite as part of the peak paragenesis. Retrogressive phases include amphibole, plagioclase, alkali feldspar (microcline) and quartz. Clinopyroxene forms anhedral, resorbed laths which include biotite and biotite-feldspar inclusions. Crystals show some exsolution but are dominantly characterised by anhedral blebs of feldspar



Figure 3.11: Images of features of the Svartberget contact unit. a-b) photographs showing the granoblastic texture of the biotite websterite [QC36S-11H] PPL-XPL, x2.5 mag, FOV 4.2mm; c-d) photographs showing crystallographically orientated mica and poikiloblastic alkali feldspar in the felsic lens [QC36Q-3A] PPL-XPL, x2.5 mag, FOV 4.2mm; e) photograph showing microperthite exsolution of alkali feldspar [QC36Q-7A] XPL, x2.5 mag, FOV 4.2mm; f-g) scans of thin section showing very coarse quartz crystals and phengite pseudomorph [QC36Q] PPL-XPL, FOV 2cm; h-i) photographs showing sillimanite-alkali feldspar-rich leucogneiss [QC36L-9B] PPL-XPL, x2.5 mag, FOV 4.2mm.

throughout. Amphibole is a strongly coloured, turquoise-green pleochroic retrograde phase which forms rims around biotite and replaces clinopyroxene. Euhedral crystals with myrmekitic quartz are also observed. Alkali feldspar contains fine, wavy microperthitic albite exsolutions and cross-hatched twinning. Albite and alkali feldspar also contain short, graphic blades of quartz. In addition to biotite, feldspar is also coarsely intergrown with late amphibole and can form poikiloblastic/interstitial crystals enveloping clinopyroxene, amphibole and biotite. Biotite shows greenorange pleochroism. The retrogressed thin, discontinuous garnetite veins and lenses in the clinopyroxenite domain contain biotite and accessory zircon and monazite.

The leucocratic gneiss on the other side of the contact consists of quartz, microcline, biotite, plagioclase, white mica, sillimanite, minor alkali feldspar and accessory zircon, monazite, apatite, spinel and xenotime. Quartz is dominant and forms coarse crystals and coarse sub-grains with irregular boundaries. Coarse white mica is partially to pervasively altered to fine, fibrous, highly birefringent white mica (sericite), a fine-coarse intergrowth of dark brown (Fe-rich) biotite and plagioclase with or without elongate masses of intergrown fibrolite sillimanite (*Figs. 3.11h,i*) or more coarse, skeletal white mica (muscovite). Plagioclase is present only in the intergrowth with biotite. Optically continuous crystals form poikiloblasts around the biotite. It shows multiple albite twinning and rare pericline twins. Microcline and minor perthitic alkali feldspar are common (*Fig. 3.11i*) particularly with quartz in thicker, biotite-poor bands. Monazite forms clustered, disseminated specks or anhedral grains in the matrix. Spinel is a gahnite-hercynite spinel (Zn-Fe-Al) and forms skeletal-dendritic crystals and disseminated specks in the matrix. Biotite is less abundant that in the surrounding gneiss and shows minor alteration to chlorite.

### 3.4.8. <u>Multiphase Solid Inclusions</u>

MPSIs are inclusions which contain more than one solid phase and many types are thought to form through the crystallisation of trapped fluid phases therefore providing insight into fluid activity; it is for this reason they were studied and are presented in this chapter. Typically, water is the dominant volatile crystallising to form hydrous phases with subordinate amounts of phosphates, carbonates, sulphates, chlorides (Ferrando *et al.*, 2005). *Tables 3.1 and 3.2* detail the characteristics and mineral assemblages of multiphase inclusions observed in

Rock Type	Section/Image Reference	Host Phase*	Assemblage*	Shape	Approx. Size (µm)	Number Observed	Comments
gneiss	QC26D-B3-C	grt	pl-ms-ttn-bt-[Si-Al-Mg-Fe- Mn]	elongate, octagonal	140	one	similar to retrograde assemblage
gneiss	QC26D-B3-C	grt	kfs-ab-ms-[Si-Ca-Al-K]	straight-sided, equant heptagonal	45	one	similar to retrograde assemblage
quartz eclogite	A632-B8-C	grt	am-rt-ap-qtz-cpx	irregular, elongate 370		one	similar assemblage to matrix
quartz eclogite	A632-B8-B	grt	am-qtz-zrn	equant 140		two	matrix phases; both inclusions in one garnet crystal
garnet websterite	QC26P (8a)	grt	am-spl-op	hexagonal ~290		numerous	constant volumetric ratios
garnetite vein core	ARS-1-B3-C	grt	bt-qtz-wmca	elongate, pentagonal	550	one	radial fractures
garnetite vein core	ARS-1-B4-B	grt	[Si-Al-Mg-Fe]-am-ap-aln- rt-ilm-[Al-Si-Fe-Mg]	irregular, ovoid	350	one	connects to another inclusion via amphibole neck
websterite selvage	QC26HC-B7-A/C	grt	am±bt-[Fe-S]	straight-sided, irregular octagon	55	two	off-shoot structures in one
websterite selvage	QC26HC-B11-C	am	ap-mnz (low Th)-[Fe-O]	irregular, ovoid	220	one	Fe-O forms a broken rim around apatite
websterite selvage	QC26HC-B11-A	срх	zrn-qtz-am	tiny, irregular	55	one	offshoot structure filled by amphibole
clinopyroxe nite selvage	QC26JB-B8-A	срх	rt-qtz-bt-am	irregular	300	one	sharp contacts with host

Table 3.1: Description of multiphase solid inclusions observed in thin sections of samples from Årsheimneset

\* mineral abbreviations as in Table 0.11b; minerals in [] brackets are unidentified phases

Tuble 3.2. Description of multiplices solid inclusions observed in thin sections of sumples from switcherget									
Rock	Rock Section/Image		Accomblago*	Shana+	Approx	Number	Commonte		
Type^	Reference	Phase*	Assemblage	Snaper	Size (µm)	Observed	Comments		
gw	QC36D-B16-D	grt	rt-bt-qtz-am-an-ab-[Si-Mg-Al-Fe]- [Si-Mg-Fe]	irr	275	one	some fractures stop at inclusion boundary, some continue into it		
gw	QC36D-B16-E/F	grt	qtz-am-bt	irr	150	two	offshoot structures		
gw	QC36D-B16-G	grt	qtz-bt-an-ab	irr	180	one	offshoot structures, anorthite needles		
gw	SV1C (5b)	grt	am-spl-op (aln?)	hex	90	one			
gvc	QC36EA-B5-A	grt	am-qtz-rt	irr	500	a few	sharp contact with host; quartz grains and/or myrmekitic quartz with am		
gvc	QC36EA-B17-A	grt	qtz-bt-[Si-Mg-Al-Fe]-[Si-Mg-Fe]- [Fe-Cu-S]	irr	100	one	sharp contact with host; diffusion rim		
gvc	QC36FA	grt	am-spl-op	irr-hex		numerous	retrograde, step-daughter assemblage(?)		
gvc	QC36GA-B4-B	grt	zrn-rt-bt-qtz	irr	30	one			
gvc	QC36GA-B8-E	grt	bt-pl-am	ovoid	80	one	diffusion rim; phases fill fractures either side of inclusion		
gvc	QC36GA-B9-G	grt	ap-qtz-bt-rt-am	hex, eq	160	one	radial fractures, cuspate terminations		
gvc	QC36GA-B9-4	grt	qtz-am-[Si-Al-Fe-Mg]	elg, irr	205	one	interstitial qtz-am; straight-sided		
gvc	QC36GA-B9-5	grt	an-bt-zrn-rt	elg, irr	135	one	diffusion rim in garnet?; some straight sides; goes into fractures in garnet		
gvc	QC36GA-B16-2	grt	am-qtz-bt-zrn-rt	irr, eq	105	one	fractures both stop at and cross boundary		
gvc	QC36GA-B20-A	grt	qtz-[Si-Al-Ca-Mg-Fe]-[Si-Mg-Fe- Al]	hex	265	one	fractures both stop at and cross boundary, offshoots		
gvc	QC36GA-B20-B	grt	bt-pl	eq, irr	65	one	coarse intergrowth		
gvc	QC36GA-B20-D	grt	ap-[Si-Al-Mg-Fe]-bt-am-an-ab-[Si- Mg-Fe-Al]-mnz	irr	600	one			
gvc	QC36GA-B20-E	grt	am-qtz-bt-[Si-Al-Mg-Fe]-[Si-Mg- Fe-Al]-zrn	hex, eq	145	one	fractures stop at boundary		
gvc	QC36UA	grt	am-spl ±op	hex, pyg	190	common	constant volumetric ratios		
gvc	QC36UA-B4-A-3	grt	am-spl-[Si-Al-Mg-Fe] ±op	irr	270	common			
gvc	QC36UB-B5-A	grt	am-[Si-Fe-Mg-Al]-aln	pyg (hex)	190	one	straight-sided; sharp contacts with host		
gvc	QC36UB-B5-A	grt	am-[Si-Al-Mg-Fe]-[Si-Fe-Mg-Al]- spl-[Fe-O]	irr	190	one	fractures stop at boundary		

Table 3.2: Description of multiphase solid inclusions observed in thin sections of samples from Svartberget

^ gw = garnet websterite; gvc = garnetite vein core

\* mineral abbreviations as in *Table 0.11b*; minerals in [] brackets are unidentified phases

+ irr = irregular; hex = hexagonal; elg = elongate; eq = equant; pyg = polygonal

samples from Årsheimneset and Svartberget, respectively and *Figures 3.12-13* show a representative suite of those observed. They are found in gneiss, eclogite, vein zones and garnet websterites typically hosted in garnet crystals and have various assemblages, shapes and sizes. Many have straight, flat faces with angular edges and sharp contacts with the host rocks (e.g. *Figs. 3.12d, 3.13f*), although some are more irregular and poorly-formed (e.g. *Figs. 3.12a, 3.13g*). Shapes range from equant polygons (*Fig. 3.12d*) to elongate (*Fig. 3.12c*) to irregular with some having offshoot structures (*Figs. 3.13c,d*) and cuspate terminations (*Fig. 3.13e*). Sizes range from 30-600µm in diameter. Fractures both stop at the contact and continue into the inclusion (*Fig. 3.13h*).

The following phases are identified within the multiphase inclusions (in alphabetical order): allanite, amphibole (calcic), apatite, biotite, clinopyroxene, ilmenite, K-feldspar, monazite (both Th-rich and Th-poor varieties), white mica, opaques, plagioclase (albite and anorthite), quartz, rutile, spinel (hercynite), titanite and zircon. Opaque phases include Fe oxides and Fe and Fe-Cu sulphides. In addition, a variety of unidentified Al-Mg-Fe silicate phases ±Ca ±Mn are also present with compositions including: Si-Al-Fe-Mg, Si-Mg-Fe, Si-Mg-Al-Fe, Si-Al-Mg-Fe(-Mn), Al-Si-Fe-Mg, Si-Mg-Fe-Al, Si-Fe-Mg-Al and Si-Al-Ca-Mg-Fe. They may be various types of amphibole, chlorite and/or talc. Two often occur together in the same inclusion, some have 60° cleavages, some have basal cleavages and others have spongy textures. No carbon phase was identified in any inclusion from the key localities despite a considerable search, particularly at Svartberget where diamond has been identified (Vrijmoed *et al.*, 2006).

Two inclusions of different assemblages are observed within the same garnet crystals in the gneiss [QC26D] and eclogite [A632] from Årsheimneset. In a garnet websterite from Svartberget [QC36D] there are four MPSIs in one garnet crystal, three of which have differing assemblages. Single phase inclusions are also found in close proximity to MPSIs such as biotite and rutile inclusions in garnets of vein cores. Some inclusions contain fine intergrowths of two or more phases such as plagioclase and biotite although most do not. A common assemblage observed in garnets in garnetite vein cores and garnet websterites is amphibole-spinel-opaque in euhedral (hexagonal) inclusions (*Figs. 3.12g,h*). This is the only assemblage which occurs between different



Figure 3.12: Images of multiphase solid inclusions observed in thin sections of samples from Årsheimneset and Svartberget. Mineral abbreviations as listed in Table 0.11b. i - Fe-Mg-Al silicate (amphibole?); ii - Al-Si-Fe-Mg-Ti phase; iii - Si-Al-Mg-Fe-Mn phase; iv – Si-Ca-Al-K phase. a) in garnet in eclogite [A632-B8-C] BSE SEM; b) garnetite vein core [ARS-1-B4-B] BSE SEM; c-d) gneiss [QC26D-B3-C] BSE SEM; e) garnet websterite selvage [QC26HC-B11-C] BSE SEM; f) clinopyroxenite selvage [QC26JB-B8-A] BSE SEM; g) photograph of amphibole-spinel ± opaque inclusions in a garnetite vein core [Svartberget, QC36UB-6A] PPL, x6.3 mag, FOV 1.64mm; h) photograph of amphibole-spinel-opaque inclusions in a garnet websterite [Årsheimneset, QC26P-8A] PPL, x6.3 mag, FOV 1.64mm.



Figure 3.13: BSE SEM images of a selection of multiphase solid inclusions observed in thin sections of samples from Svartberget. Mineral abbreviations as listed in Table 0.11b. i - Si-Mg-Fe phase; ii – Si-Al-Mg-Fe phase; ii – Si-Al-Mg-Fe-Ca phase; iv – Si-Al-Fe-Mg phase; v – Si-Al-Ca-Mg-Fe phase; vi – Si-Mg-Fe-Al phase. a) garnet websterite [QC36D-B16-E]; b) garnetite vein core [QC36GA-B4-B]; c) garnet websterite [QC36G-B16-G]; d) clinopyroxene garnetite vein core [QC36EA-B17-A]; e) garnetite vein core [QC36GA-B9-G]; f) garnetite vein core [QC36GA-B9-4]; g) garnetite vein core [QC36GA-B9-5]; h) garnetite vein core [QC36GA-B20-A].

samples and different rock types and is observed repeatedly. Amphibole is pale coloured and forms the majority of the inclusion.

### 3.4.9. Fluid Inclusions

Tiny, non-crystalline fluid inclusions are present in all thin sections and like MPSIs provide insight into fluid activity. Fluid inclusions are most common in quartz crystals. The most common type are spaced, angular-rounded, elongate to spherical inclusions arranged in inclined, well-defined planar trails. These are seen in garnet and quartz in both vein cores and gneisses and mark annealed fractures. The trails cut across mineral inclusions in garnet and current fractures and have various orientations. The trails are not always continuous but can be quite short and taper off before crystal edges. They curve and intersect like current fractures. Fewer inclusions and trails are seen towards garnet rims and none are seen in inclusion-free rims and smaller blasts. Trails are also found in amphibole, carbonates and clinopyroxene in websterites. Clinopyroxene and amphibole contain trails of squarerectangular inclusions aligned to the cleavage but the trails cross-cut the cleavage. Fewer trails are seen in websterite selvages than other zones. Trails in olivine garnet websterite are rare and when present are short, irregular, in garnet and contain branched and networks of fluid inclusions.

Inclusions in eclogites are less common than in websterites and vein cores but those observed have round, triangular to irregular shapes or are branching tubes and occur in planar trails in garnets. Branching, irregular, wiggly tubes, often densely spaced in trails are also seen in garnets in websterites and vein cores with similarly sized crystalline inclusions.

Isolated spherical-ovoid fluid inclusions are rare but are seen in garnet. Clusters or aggregates are more common but still rare in comparison to the planar trails. Solitary inclusions are larger than those in annealed fractures. Rare, isolated and clusters of tiny, round and irregular inclusions in quartz in vein cores and plagioclase, amphibole, quartz and biotite in gneiss are also observed. The inclusions are angular, oval, irregular or faceted. Inclusions of similar size and shape but which may be crystalline include dustings and clusters of tiny, elongate to round inclusions in garnet. These have high interference colours and may be tiny rutiles. Randomly arranged tiny needles are also observed in quartz and aligned needles in rutile in a trail at an angle to alignment.

### 3.5. Data Analysis

#### 3.5.1. Country Rocks

The samples observed indicate the peak paragenesis consisted of garnet, clinopyroxene, quartz, phengite, kyanite and rutile which retrogressed to an assemblage of amphibole, biotite, plagioclase, quartz, muscovite, sillimanite, epidote, titanite, ilmenite and Fe oxide. The presence of exsolution features in alkali feldspars indicate a K-feldspar phase also occurred at higher temperature conditions. It may have formed part of the peak paragenesis or formed during decompression whilst still at high temperatures. No phases observed in the country rocks are considered to be part of the prograde paragenesis.

The presence of quartz neoblasts and granular, multicrystalline quartz indicate recrystallisation after prior strain (Medaris Jr, 1980) and the persistence of large grains with undulose extinction and irregular, lobate, sub-grain boundaries suggests the gneisses did not fully texturally re-equilibrate upon exhumation and coarse crystals were a feature of the eclogite-facies paragenesis.

The fine to medium intergrowth of biotite and plagioclase which forms the matrix of the host gneisses and schists is considered to represent a coarsened product of the very fine biotite-plagioclase intergrowth observed along the edges of white mica crystals in quartz-rich seams (and vein cores) formed by the fluid-producing breakdown reaction of phengite (e.g. Franz *et al.*, 1986; Zheng *et al.*, 2009 and references cited therein). The reaction to form plagioclase requires components from other sources such as Na-Ca from clinopyroxene suggesting the breakdown of phengite involved co-existing phases. The dominance of this intergrowth indicates phengite formed a dominant phase in the peak assemblages of the country rocks. The presence in the matrix of relatively coarse white mica with biotite-plagioclase rims, particularly when surrounded by quartz, indicates persistence of phengite. The

surrounding quartz may have protected it from retrogressive fluids. White mica hosted in garnet may also be phengitic.

The pseudomorphic intergrowth of white mica, biotite, alkali feldspar and plagioclase produced in the felsic lens at the Svartberget contact is also considered to be after phengite with the white mica produced having a lower celadonite component. The original presence of phengite in the lens demonstrates a HP-UHP formation unlike other biotite-feldspar-bearing pegmatites observed at the northwest Svartberget contact and is a similar assemblage to the quartz-white mica vein core at Årsheimneset. There is no evidence of another peak phase in addition to quartz in the pegmatitic lens suggesting the intergrown products are sourced only from the precursor phengite and quartz: the dominance of alkali feldspar over plagioclase in the pseudomorphic intergrowth is considered to reflect the absence of Ca components from clinopyroxene during breakdown suggesting a Ca-bearing phase remained in the source rocks and the fluid was Ca-poor. The assemblage of biotite, plagioclase, K feldspar and quartz is also observed between crystals of white mica and quartz in a garnetite vein core from Årsheimneset and may represent products of a similar retrograde reaction.

No mineralogical indicators of UHP conditions such as coesite or diamond are evident in the country rocks suggesting they are pervasively retrogressed or they were not formed. The multicrystalline inclusions and textures of quartz may reflect the recovery from the transition from coesite as subgrains coarsen but are not considered reliable indicators as they may result from recovery from stress during exhumation. The tiny, cubic-hexagonal inclusions in some garnets in gneiss from Årsheimneset and Remøysunde have an appearance similar to diamond inclusions in garnet from other UHP metamorphic rocks (e.g. Sobolev and Shatsky, 1990); however none were found to have carbon compositions. Their wide-ranging compositions suggest they are solid phase inclusions and not fluid inclusions as similar compositions would be expected considering their proximity and local nature.

#### 3.5.2. Mafic-Ultramafic Bodies

The olivine garnet websterite assemblages of Svartberget and Myrbærneset samples indicate the bodies have a less ultramafic character than that at Kolmannskog and

than was assumed in the field. They may represent transitional compositions between garnet websterites and olivine-rich peridotite horizons in the parent bodies. The bodies have not been systematically sampled however so more olivine-rich domains may occur. The compositional range of opaque oxide and sulphide phases in the ultramafic bodies indicates they crystallised at a horizon rich in transition metal elements (Fe, Ti, Cr and Cu). From their qualitative chemistries, phases may include ulvöspinel, ilmenite, magnetite/haematite, chrome spinel, chalcopyrite and bornite. The exsolution of Fe-Ti oxides from each other in the olivine garnet websterite from Svartberget demonstrates the opaques were subject to retrogressive exsolution from high temperature conditions. The presence of haemoilmenite and ilmenohaematite suggests a reaction between rutile and magnetite components forming an ilmenitehaematite solid solution (Pinet and Smith, 1985).

The Fe oxide opaque 'blebs' and blades in pyroxenes in the Svartberget body indicate an Fe-rich nature and supports the Fe-Ti classification by Vrijmoed *et al.* (2006). Their presence dominantly in the cores of crystals and the aligned, faceted form of the blades suggests these are exsolutions from the pyroxenes, however, the ovoid, irregularly-sized nature of the 'blebs', suggest they are poorly-formed inclusions. They have a composition similar to some grains in the matrix but do not vary in composition like those in the matrix. Dustings of opaque minerals are known to occur in igneous plutonic Fe-Ti rocks. Those observed may therefore represent metamorphosed relics of such components of the primary igneous (Ti-augite) clinopyroxene redistributed during re-equilibration and constrained by cleavage planes.

There are few indications to the prograde evolution of the ultramafic rocks. The spongy, inclusion-rich garnets in the Svartberget olivine garnet websterite and some from Myrbærneset include only phases found in the matrix. Their almost skeletal form suggests garnet-forming components such as AI were scarce. The rare inclusion of biotite, amphibole and a carbonate phase in garnet indicates they were not devoid of volatiles during their prograde evolution. The finer grained, granoblastic textures of the olivine garnet websterites at Svartberget, Myrbærneset, some garnet peridotite-garnet websterite domains from Kolmannskog and eclogite from Nybø demonstrate dynamic recrystallisation occurred during deformation at eclogite-facies conditions at some localities. The occasional coarser pyroxenes and coarser textures

represent relict crystals of the prior texture. The eclogite at Myrbærneset is coarser grained with a decussate texture showing a heterogeneous response of proximal lithologies to stress.

The inclusion-rich garnet porphyroblasts in some eclogites, such as from Årsheimneset, with abundant clinopyroxene and amphibole inclusions indicate an amphibole-clinopyroxene-rich prior assemblage. The tiny amphibole inclusions with the more turquoise colouration are considered to be primary inclusions of prograde amphibole not formed through alteration of clinopyroxene inclusions. Their comparative compositions and with amphiboles in the matrix are investigated in Chapter 6. The absence of plagioclase inclusions in garnet indicates eclogitisation was pervasive and efficient. Dehydration of abundant amphibole may have provided abundant catalytic fluids to facilitate this. The presence of various accessory phases in the eclogites (rutile, zircon, apatite, allanite and monazite) suggests relatively high contents of phosphate, HFSE and REE elements were present in the protolith. The abundance of rutile is thought to reflect the water-insoluble nature of Ti which precluded its loss during dehydration and eclogitisation.

The presence of palisade-textured and polycrystalline quartz in eclogites from Årsheimneset petrographically confirms that coesite was stable and therefore UHP eclogite-facies conditions were reached at this locality but the rocks are now equilibrated to HP eclogite-facies conditions. Its absence in other localities does not mean they did not experience similar conditions but that free silica was not part of the peak assemblage or they experienced more complete recovery upon retrogression.

Textural equilibration of amphibole, biotite and carbonate indicated by sharp contacts, granoblastic junctions and the lack of retrogressive reaction between them and anhydrous phases and inclusions in garnet suggest they form part of the prograde-peak paragenesis of the ultramafic bodies. However interstitial relationships in the matrix, association with each other and contact shapes such as the embayed nature of some garnet-carbonate contacts suggest they may be later phases than the garnet-pyroxene-olivine assemblage forming at the expense of the anhydrous phases. Clinohumite is typically found in metamorphic or metasomatic carbonate rocks or in association with kimberlites and carbonatites but is known

from garnet peridotites (e.g. Yang, 2003). The exsolution of Ti-rich ilmenite suggests it is titanian clinohumite and may have formed through the hydration and addition of Ti to olivine. Its subsequent exsolution and alteration to olivine and ilmenite suggests it formed on the prograde path and is part of the peak paragenesis as it is stable to considerable mantle depths (Yang, 2003 and references cited therein). Its local nature within the body suggests its formation is related to the local presence of fluids or concentrations of minor elements (e.g. Ti, F). The coexistence of olivine and titanian clinohumite is seen in metamorphosed serpentinised peridotites (Scambelluri *et al.*, 1991) and therefore may owe its origins to an episode of hydration during earlier subduction and prograde metamorphism. The presence of almost completely exsolved clinohumite in the websterite from Eiksunddal may represent a more ultramafic portion of the body at this locality.

The pale colour and weak pleochroism of amphibole in the matrix of the maficultramafic rocks (and garnet websterites and vein zones) suggest it is an Mg-rich endmember such as tremolite or magnesiohornblende. The oblique nature of the exsolution lamellae are diagnostic of cummingtonite and hornblende (Deer et al., 1992) and demonstrate a high temperature formation exsolving upon cooling. The stronger green-turquoise colours of the retrogressive amphibole in coronas and after symplectitic clinopyroxene indicate a more Fe-rich hornblende formed later on the retrogressive path. The formation of amphibole coronas indicates partial hydration along grain boundaries. The formation of clinopyroxene-plagioclase symplectites in eclogites (and garnet websterites and vein zones) indicates the peak clinopyroxene is omphacitic and therefore an eclogite-facies paragenesis. The fine, delicate nature of the intergrowth and generally only partial replacement of clinopyroxene crystals indicates static, fluid-poor conditions with little deformation (e.g. Wain et al., 2001; Zheng, 2009; Hacker et al., 2010). Fluid-poor conditions are also indicated by the general preservation of eclogite-facies assemblages. The more heavily retrogressed portions are associated with fractures demonstrating a channelised flow of retrogressive fluids.

## 3.5.3. Garnet Websterites

The small, inclusion-poor nature of the garnets in the garnet websterite at Svartberget, compared to the inclusion-rich crystals of the olivine garnet websterite,

suggest they are newly grown and that the host body garnets are completely replaced during metasomatism. In comparison, the similar forms of porphyroblastic garnets in eclogite and garnet websterite at Årsheimneset, Eiksunddal and Remøysunde suggest they may be equivalent and had the same evolution. The inclusion-poor garnets and rims may represent metasomatic growth of garnet similar to that seen at Svartberget. The inclusion of biotite only in garnet rims at Årsheimneset suggests it formed during the later growth of the garnets and was not part of the prograde paragenesis. The inclusion of clinopyroxene and amphibole in garnet porphyroblast cores, however, suggest the prograde paragenesis of the garnet websterites were also amphibole-rich. The addition of biotite to the inclusion assemblage demonstrates a change in the matrix assemblage: a metasomatic, local introduction of fluids similar to that at Svartberget may have occurred.

Like in the ultramafic rocks, the intergrown textures and clean, sharp faces of biotite and pale amphibole crystals suggest they are peak phases equilibrated to the same conditions as the anhydrous phases and their spatial association suggests their crystallisation was associated. Poikiloblastic and interstitial amphiboles, however, may have crystallised later growing at the expense of other phases (Jacob et al., 2009). The variable textures of biotite with amphibole suggests their relative timing may overlap but amphibole crystallisation continues later than that of biotite; perhaps as fluid compositions change with continued crystallisation of phases. The patchy distribution of carbonate in garnet websterites of eclogite bodies, its presence only in the matrix, inclusion of major anhydrous and hydrous phases and irregular contacts with them suggests it is a relatively later phase although still part of the peak assemblage and alteration of its rims during retrogression caused reaction textures with adjacent phases. The coexistence of quartz and Mg-rich orthopyroxene in some garnet websterites at Årsheimneset indicates high temperature conditions (>800°C) and/or low H<sub>2</sub>O activities and/or high Fe contents (otherwise talc would form) (e.g. Nakamura and Hirajima, 2000; Bucher and Frey, 2002).

The assemblage of accessory phases in the garnet websterite domains is similar to that in the eclogite. This suggests they contain a similar recipe of HFSEs and REEs. Opaques may include magnetite, ilmenite, ulvöspinel, cuprospinel, pyrite, chalcopyrite and bornite. The orientated needles, lamellae and blades of ilmenite in some rutile grains indicate they are exsolved from rutile. Ilmenite is known to be

stable at higher temperatures and lower pressures that rutile (Massonne, 2009) suggesting the rutiles formed prior to peak temperatures and exsolved upon increasing temperatures between  $P_{max}$  and  $T_{max}$ . The occurrence of ilmenite on its own or as considerable portions of rutile grains in garnet websterites and vein cores from Årsheimneset but not in the eclogite where it only occurs as fine lamellae within rutile and its association with biotite or retrograde assemblages suggests it is associated with metasomatism and retrogression.

The presence of both anhedral, 'spongy' garnets and inclusion-poor, granular garnets in both the olivine garnet websterite and garnet websterite at Myrbærneset suggests they both had a similar evolution to each other supporting the interpretation of a metamorphic origin of the garnet websterite body domains at this locality. The presence of coarser grain sizes of one garnet websterite from Myrbærneset may represent the prior coarser grained texture which was subsequently recrystallised to a granoblastic texture in other domains.

## 3.5.4. Laterally-Zoned Veins

The presence of garnet and rutile in the vein cores demonstrates they are a high pressure assemblage as plagioclase and ilmenite would form at lower pressure conditions. The inclusion of co-crystallising matrix phases in inclusion-rich vein core garnets suggests they are newly grown phases and not metamorphosed veins formed during earlier prograde metamorphism. The abundance of accessory phases such as apatite, monazite, rutile and zircon in the vein core garnets compared to the other zones indicates a greater content of HFSEs and phosphate. The presence of the zircon-monazite-rich band in the garnetite vein core side of the glimmerite selvage at Svartberget, the presence of trails of rutile at the edges of some vein cores and rarity in the clinopyroxenite and websterite selvages suggest this horizon concentrates HFSEs and REEs and perhaps represents the limit of their mobility from the fracture wall. The concentration of zircon nucleation and growth may also be related to the local Al and water contents which can affect the saturation of Zr in fluids (Zheng et al., 2009). A greater abundance of rutile in the garnet websterite domains at Svartberget may represent the inheritance of Ti-rich oxides from the peridotite. The observation of quartz with a Fe-Cu sulphide grain in a vein selvage at Svartberget demonstrates free silica does occur at Svartberget albeit a rare and small occurrence.

The inclusion of biotite, amphibole and carbonate in garnets and pyroxene laths in metasomatic garnet websterite domains and selvages suggests biotite-forming components were available at the initial stages of metasomatism and it formed part of the early peak metasomatic paragenesis. The inclusion of biotite in clinopyroxene, clinopyroxene in biotite and granoblastic matrix in the clinopyroxenite selvages demonstrates that biotite and clinopyroxene grew contemporaneously in this zone. The replacement of orthopyroxene and garnet by biotite observed in the clinopyroxenite selvage at Svartberget suggests biotite becomes the more stable Mgrich phase in this zone due to the local bulk chemistry. Sekine and Wyllie (1982) found that as the SiO<sub>2</sub> decreased phlogopite becomes dominant over orthopyroxene. It also suggests the vein zones may form successively with garnet websterite forming first.

The finer, granoblastic and sub-grain aggregates and occasional larger pyroxene crystals in the veins at Myrbærneset are indicate they too have undergone dynamic recrystallisation. This indicates deformation occurred post-vein formation at Myrbærneset in contrast with Svartberget where veins are not deformed. This suggests either a different timing of vein formation or deformation between the two locations. The foliated, granoblastic nature of the biotite websterite band at the Svartberget contact, however, suggests metasomatism of the margin may have been contemporaneous with deformation. The rare presence of rutile within the veins at Myrbærneset but abundance of opaque Fe-Ti oxides are further evidence that they formed in a different P-T-X system than those at Svartberget in addition to the reduction of volatile-bearing phases and clinopyroxene.

### 3.5.5. Other Retrogressive Features

The aligned nature, regular distribution, constant assemblage and similar proportions of quartz and amphibole in the needles in clinopyroxenes indicate they are exsolved phases from a homogeneous host (e.g. Ogasawara *et al.*, 2002; Proyer *et al.*, 2009). The presence of exsolved phases indicates re-equilibration of the rocks to lower temperatures (Berly *et al.*, 2006). The presence of quartz and amphibole together is indicative of a high pressure pyroxene with the Ca-Eskola (CaEs) component  $[Ca_{0.5}\Box_{0.5}AlSi_2O_6]$  (Terry *et al.*, 2000 and references cited therein; Hollocher *et al.*, 2007 and references cited therein). Exsolution of amphibole is indicative of high-Al

clinopyroxene unstable upon decreasing temperatures (Proyer *et al.*, 2009) and the hydration of clinopyroxene. Amphibole-forming OH<sup>-</sup> may be sourced from dissolved H<sub>2</sub>O and OH<sup>-</sup> in the clinopyroxenes: omphacite is known to be able to host considerable amounts of water under high pressure conditions (e.g. Ogasawara *et al.*, 2002 and references cited therein; Proyer *et al.*, 2009 and references cited therein) and this water rapidly exsolves upon decompression (e.g. Zheng, 2004). The typical absence of exsolution needles in the margins of the crystals suggests rims grew with a different composition or exsolved components diffused into the matrix (e.g. Proyer *et al.*, 2009). Precipitation of Si- and Al-rich phases in the cores reflects an efficient mechanism of expelling the components from the host lattice as they have slow diffusion rates (e.g. Proyer *et al.*, 2009). The characteristics of the fine exsolution lamellae in orthopyroxenes suggest they are clinopyroxene (augite) lamellae parallel to the (100) plane (Nesse, 2000) representing unmixing along the solvus and expulsion of Ca from the orthopyroxene lattice with decreasing temperatures.

The presence of calcite crystals along biotite cleavages, cross-fibre calcite in late fractures and the alteration of clinopyroxene to calcite is indicative of late CO<sub>2</sub>-rich/carbonic fluids. The local alteration of biotite to chlorite represents greenschist-facies retrogression and is suggestive of the local presence of late-stage aqueous fluids such as may be associated with tourmaline-forming veins where greater chloritisation occurs.

#### 3.5.6. Multiphase Solid Inclusions

Various explanations for MPSIs have been evoked: remnants of trapped supercritical liquids (e.g. Malaspina *et al.*, 2006a; Lang and Gilotti, 2007; Zhang *et al.* 2008); partial melt pockets of host phases formed during exhumation (e.g. Sheng *et al.*, 2013 and references cited therein); an *in situ* retrograde assemblage of a HP-UHP included phase (Liu *et al.*, 2013); or they are crystal aggregates enclosed during growth of the host phase. Hosts should not be considered inert containers (Stöckhert *et al.*, 2001) and step-daughter phases may also be produced through reaction of the fluid, daughter crystals and host (e.g. Svensen *et al.*, 1999; Stöckhert *et al.*, 2001).

Indications of being trapped former melt include consistent phase volumetric ratios in numerous inclusions of daughter phases (primary precipitates from the fluid)

(Ferrando *et al.*, 2005), negative host crystal shapes and off-shoot structures into the host crystal (e.g. Sheng *et al.*, 2013). Partial melting may occur as water is released during the dehydration of hydrous phase inclusions which decreases the solidus temperature of the inclusion-host micro-system (Perchuk *et al.*, 2005). Indications of being partial melts include radial cracks around the inclusions indicating volume expansion accompanying melting (Perchuk *et al.*, 2005) and new growth of the host phase around the pockets with a different composition (Perchuk *et al.*, 2005). Retrogression may be indicated by the tight intergrowth of retrograde phases (Liu *et al.*, 2013). Enclosure of solid, co-crystallising crystal aggregates would lack a reaction with the host phase which is indicated by an unzoned host around the inclusion and have sharp boundaries and irregular shapes (e.g. Zeng *et al.*, 2009).

Characteristics of the inclusions suggest some formed from a fluid, some are retrograde assemblages and some are crystal aggregates. The variation of assemblages and inconsistent volumetric ratios of the MPSIs suggest they were not all formed from a homogeneous trapped fluid; although, cross-sections through one part of the inclusion may not cut through all the phases present.

Offshoot structures, cuspate terminations and negative crystal shapes like Figures 3.12d,h suggest these inclusions crystallised from a fluid. Negative crystal shapes in garnet however may be due to the more dominant nature of the garnet structure over other crystal lattices. Assuming all inclusions in the vein cores are of the same generation, the presence of single phase inclusions of the peak paragenesis near the MPSIs suggests they were enclosed during peak conditions. Those MPSIs containing minerals that are mostly found in the matrix and as individual inclusions such as Figures 3.12a, b, e, f and 3.13b, e may be enclosed aggregates of small, co-crystallising crystals that became trapped during garnet growth. However, considering the vein cores formed in the presence of a fluid and bore the phases in the matrix, a trapped portion of the fluid would be expected to yield the same crystals. However, assemblages and volumetric ratios differ and accessory phases form a larger proportion of some inclusions. The presence of quartz in some inclusions at Svartberget indicates the inclusions crystallised from an evolved fluid as quartz was not a crystallising phase in the matrix. The inclusions which contain plagioclase feldspar represent sub-eclogite-facies assemblages formed by alteration of eclogitefacies phases. The delicate and intergrown nature of some inclusion phases (Figs.

*3.12d, 3.13c*) suggests these may be alteration assemblages. The two MPSIs found in the host gneisses at Årsheimneset (*Figs. 3.12c,d*) have euhedral shapes and sub-eclogite-facies assemblages containing feldspars, titanite and micas indicative of crystallisation from a felsic fluid. The intergrown nature suggests they may be alteration products of a prior phase or assemblage, which initially crystallised from a felsic fluid.

The common amphibole and spinel ±opaque inclusions may represent a retrograde, step-daughter assemblage formed through the reaction of a fluid inclusion with the host garnet. As they occur in garnet websterite garnets and garnetite vein cores, this suggests they are not a prograde feature as their hosts are interpreted to form at HP-UHP conditions. The existence of these inclusions in different rock types at different locations suggests they formed from the retrograde hydration of garnet.

## 3.5.7. Fluid Inclusions

The abundance of fluid inclusions in websterite and vein zones suggests these two lithologies are related to a fluid-rich environment. The irregular shapes of fluid inclusions and rare negative crystal shapes suggest they didn't form contemporaneously with garnet growth. The arrangement of most inclusions in planar trails suggests they are annealed fractures along which fluids flowed. This requires the fluids to be present once the garnet crystals have already formed and been fractured indicating the fluids are not primary, garnet-forming fluids, unless they are redistributed fluids from primary inclusions. However, as the trails are continuous and cross crystal boundaries it is more likely they have a later, external source or represent the movement of late fluids. As the fractures are annealed by garnet of the same composition (no linear compositional zoning is observed) and the veins are reasonably fresh the inclusions are probably of HP-UHP fluids.

### 3.6. Data Interpretation

## 3.6.1. Peak P-T Conditions and Metamorphic Evolution

A UHP nature is demonstrated petrographically at Årsheimneset by convincing textures (palisade and polycrystalline) for prior coesite of quartz inclusions in garnet

in eclogite, garnet websterite and vein zones. The rocks are now equilibrated to HP conditions as coesite is no longer present in the matrix. Convincing evidence of prior coesite was not observed at the other key localities despite quartz inclusions. Diamond was not observed despite diamond-facies conditions being obtained in thermobarometric studies at many of the key localities (e.g. Cuthbert *et al.*, 2000): their P-T conditions are investigated in Chapter 7. The lack of coesite, convincing textures of prior coesite or diamond does not preclude the experience of UHP conditions but that chemical conditions were not suitable for their formation or they are not within the few thin sections made of the rocks. For example, the absence of diamond (or graphite) in Svartberget thin sections, where it has previously been observed, is attributed to a heterogeneous distribution and local variation of the chemical system. The unzoned nature of fresh crystals and exsolution in clinopyroxene, orthopyroxene and amphibole indicate these are high temperature rocks.

The Fe-Ti oxide-rich nature of ultramafic rocks at Svartberget, Kolmannskog and Myrbærneset supports their classification as crustal ultramafic rocks as opposed to Mg-Cr-rich mantle peridotites. The low abundance of olivine at Svartberget and Myrbærneset also supports a more evolved nature to the protolith: this is expected to be reflected in their bulk composition in the following chapter. The minor biotite, amphibole, clinohumite and magnesite within olivine garnet websterites are interpreted to be metamorphic, prograde-peak, eclogite-facies phases considering their relatively low abundance, textures and grain sizes. Biotite at Myrbærneset is also interpreted to be a metamorphic phase despite its higher abundance. The dolomite mantles of magnesite crystals are interpreted to develop at peak conditions and calcitic rims of magnesite-dolomite crystals are interpreted to be late, retrograde features.

The prograde paragenesis of the eclogite-garnet websterite bodies are interpreted to be amphibole-rich considering the abundant turquoise-green amphibole inclusions within garnet porphyroblasts such as at Årsheimneset. Considering the dominantly anhydrous nature of the eclogite domains and pervasive transformation, they are thought to have been reasonably efficiently dehydrated during eclogitisation; a similarly efficient devolatisation is inferred for the adjacent garnet websterite domains with volatile-bearing phase-rich patches a result of metasomatism.

Dynamic recrystallisation and fabric development at Nybø, Svartberget, Båtneset and domains of Kolmannskog is considered to have occurred prior to the infiltration of fluids at UHP conditions. The absence of grain size reduction and deformation of eclogite domains (other than at Nybø) is considered to be due to strain partitioning and regional variation.

#### 3.6.2. Origins of Garnet Websterite

The contrasting forms of garnet and pyroxenes in the rocks at Svartberget to the host body and their inclusion of only co-genetic phases support the field interpretation of a metasomatic origin although inheriting components from the ultramafic rock. Biotite and amphibole within domains and vein selvages at Svartberget are interpreted to be metasomatic in origin and associated with the veinforming fluids. Biotite is interpreted to crystallise contemporaneously with the anhydrous phases with amphibole slightly later forming poikiloblastic and interstitial crystals. In comparison, the textural similarity of garnets, pyroxenes and opaque phases in the garnet websterite at Myrbærneset to those in the olivine garnet websterite (and to those in the olivine garnet websterite at Svartberget) support the field interpretation of a similar, metamorphic evolution. The presence of both inclusion-rich, porphyroblastic garnets similar to those in eclogite and small, inclusion-poor blasts including only co-genetic phases in the garnet websterite domain at Arsheimneset supports the interpretation of both a metamorphic and metasomatic origin. The lack of orthopyroxene inclusions in garnet or clinopyroxene in eclogite-associated garnet websterites and its association with biotite and amphibole in some domains is interpreted to indicate the domainal character of metamorphism forming in Ca-Al-poor domains due to heterogeneity of the protolith; rather than orthopyroxene forming only through metasomatism.

Biotite, pale amphibole and carbonate in garnet websterite domains of eclogite bodies are also interpreted to be newly grown peak phases associated with local metasomatism of the garnet websterites during eclogitisation. Biotite is interpreted to have formed first coinciding with the later stages of garnet and pyroxene growth (to form inclusions) and therefore during eclogitisation. Amphibole and carbonate are thought to have formed relatively later once garnet has stopped growing, although overlapping with biotite growth in order for amphibole to be intergrown with it. This interpretation requires the evolution of the garnet websterites to be associated with a K-H<sub>2</sub>O-rich component prior to peak conditions. Amphibole and carbonate may then have formed through hydration of clinopyroxene and carbonation of orthopyroxene by H<sub>2</sub>O-rich and CO<sub>2</sub>-rich fluid phases, respectively. However, they may also have formed stable phases during prograde metamorphism (e.g. Kiseeva et al., 2012) as in the Svartberget garnet peridotite. Multiphase solid inclusions in garnet websterites at Årsheimneset are interpreted to be peak phases, locally nucleated contemporaneously with garnet growth from a HFSE-REE-rich, solute-rich fluid due to their variable, accessory phase-rich assemblages; except for the am-spl-op inclusions which are interpreted to be a retrograde step-daughter assemblage forming during retrograde interaction of the host garnet with a trapped fluid. Those in garnet websterites from Svartberget are interpreted to be retrograde assemblages of altered peak phases which may have been enclosed with a fluid phase that has since crystallised. The quartz-feldsparamphibole-biotite-rich assemblages are interpreted to present a hydrous fluid of felsic composition.

There is no distinction of assemblage between textural types of garnet websterite (or eclogite) suggesting texture variation relates to texture of the protolith and deformational evolution.

## 3.6.3. Fluid-Rock Interaction

Fluid activity responsible for the laterally-zoned veins at Årsheimneset is confidently interpreted to have occurred at UHP conditions and not during retrogression due to convincing textures of prior coesite in vein zones and features indicative of eclogitefacies conditions such as symplectitic alteration and exsolution of clinopyroxene (and orthopyroxene and amphibole). They are not considered to be re-equilibrated, earlier formed veins because they lack obvious prograde characteristics such as colour (chemical) zoning or inclusion suites. The exsolution of ilmenite from rutile in vein cores also indicates a formation at high temperatures to enable cooling and exsolution. This interpretation is extended to the fluid activity at Svartberget due to the similarity in vein assemblages and characteristics despite the lack of convincing textures of prior coesite or microdiamonds in the thin sections observed. The similarity between the texture and pseudomorphic assemblage of the leucocratic

gneiss at the Svartberget contact and the pegmatitic leucocratic lens within the biotite websterite zone suggests they formed via retrogression of equivalent phases.

The characteristics and assemblages of the MPSIs in vein zones at Årsheimneset and Svartberget are interpreted to have formed as enclosed crystal aggregates and trapped  $H_2O$ -rich fluids which formed step-daughter assemblages. At Svartberget, some MPSIs in vein cores are also interpreted to have formed through retrogression of an eclogite-facies phase as they are intergrown and contain plagioclase feldspar.

The fluid responsible is interpreted to be rich in P, HFSEs (e.g. Ti, Zr) and REEs considering the increased abundance of phosphate-, HFSE- and REE-bearing accessory phases within them relative to garnet websterites, eclogites and olivine garnet websterites. However, Ti may be sourced from the host rock considering the abundance of rutile and ilmenite in the host bodies. The presence of nominally insoluble elements in H<sub>2</sub>O-rich fluids suggests they are supercritical in nature. Their overall compositional character is interpreted to be felsic and K-rich considering the abundance of quartz, feldspars and micas in the retrograde and step-daughter assemblages: the composition of the fluid phase is investigated in Chapter 5. The source of the fluid is considered to be associated with the breakdown of phengite in the country rocks which produces a K-rich fluid (e.g. Schmidt, 1996; Auzanneau *et al.*, 2006; Zheng *et al.*, 2009).

The textural characteristics of the veins at Myrbærneset are interpreted to represent dynamically recrystallised, lower grade veins formed from a K-poor but siliceous fluid, assuming deformation of the bodies occurred at similar times and P-T conditions.

### 3.7. Summary

This chapter presents observations of microscopic features not visible in the field such as major phase inclusion relationships, textural relationships, retrogressive reaction textures, multiphase solid inclusions and fluid inclusions and brings together features of lithologies from different localities.

Low abundances of olivine in Svartberget and Myrbærneset garnet peridotites are observed forming olivine garnet websterite domains. This suggests compositional heterogeneity was

present within the crustal, igneous parent bodies and retained by the metamorphosed rocks. The Fe-Ti classification of the bodies is supported by abundant, accessory Fe-Ti oxide and Fe-rich sulphide phases supporting the classification as crust-derived protoliths rather than being entrained portions of the mantle. No further mineralogical distinction between textural types of garnet websterites and eclogites is found. The metamorphic interpretation of garnet websterite body domains at Myrbærneset, Nybø, Båtneset and Kolmannskog are supported by the further petrographic observations of their textures. These bodies and the Svartberget garnet peridotite have undergone dynamic recrystallisation and varying degrees of fabric development during deformation at eclogite-facies conditions as evidenced by their pervasive, medium grained, granoblastic and variably lineated textures. The laterally-zoned veins at Myrbærneset are considered to have formed earlier than the metasomatic UHP features at Årsheimneset and Svartberget as they suffered dynamic recrystallisation along with the body and are characterised by different mineral assemblages.

Amphibole-rich garnet porphyroblasts in both garnet websterites and eclogites demonstrate they had an amphibole-rich, prograde paragenesis and implying eclogitisation involved considerable dehydration. Minor volatile-bearing phases such as biotite, amphibole and/or carbonate in both the olivine garnet websterites and eclogites demonstrate their protoliths were not fully devolatilised during eclogitisation. UHP conditions are petrographically confirmed for metamorphism and metasomatism only at Årsheimneset by convincing evidence of prior coesite in quartz inclusions in eclogite, garnet websterite and vein cores. The proposed metamorphic-metasomatic character of garnet websterite domains of the Årsheimneset, Remøysunde, Grytting and Eiksunddal metamafic eclogite bodies is supported by petrographic observations of inclusion and textural relationships with biotite forming as an early peak phase followed by amphibole and carbonate implying an evolution of the fluid-rock chemical system. Retrogressive features such as delicate and finely intergrown symplectites and kelyphites indicate initial retrogression was characterised by fluid-poor, static conditions and then followed by hydration accompanying deformation at lower pressure, amphibolite-facies conditions enabling amphibolitisation at a range of scales of pervasiveness.

The fluid source is considered to be associated with the partial breakdown and pseudomorphic retrogression of phengite in the country rocks as indicated by breakdown assemblages including biotite, white mica, plagioclase and K-feldspar in the country rocks, felsic lenses and veins. HFSE-rich accessory phases (such as zircon, monazite and apatite) in

LILE-LREE-rich biotite-bearing vein cores at Årsheimneset and Svartberget suggest the fluid had a supercritical nature as it was able to transport both silicate melt- and aqueous fluidcompatible elements.
### **Chapter 4 - Whole Rock Chemistry**

## 4.1. Introduction

This chapter presents major and trace element whole rock chemical compositions of the least retrogressed, equilibrated samples from eight of the key localities of mafic-ultramafic lithologies and metasomatic systems: no hand samples from Grytting could be collected. Geochemical variation in the country rock upon approach to the Svartberget body at the southeast contact is also investigated. Whole rock chemistry data was obtained by X-ray fluorescence (XRF) mass spectrometry.

## 4.2. Aims and Objectives

The main aims of acquiring the bulk geochemistry were: to classify the rocks chemically and determine potential protoliths of non-metasomatised rocks; to determine the overall mafic-ultramafic character of the bodies with respect to the crust-derived Fe-Ti-mantle-derived Mg-Cr classification; and to attribute a chemical signature to metasomatism to aid identification in other rock bodies. A final aim was to obtain data to carry out mass balance calculations to quantitatively estimate the composition of the metasomatic agent: such calculations are presented in the following chapter for the Svartberget vein system.

Objectives of chemical analysis include: comparing the different textural and structural types of garnet websterite and eclogite distinguished by field and petrographic observations (Chapters 2 and 3) to determine if they can be chemically distinguished so to discriminate between different origins and/or evolutions; comparing rock suites and vein zones and altered and unaltered rock so to determine the chemical change of metasomatism and the influence of the metasomatic fluid across the vein zone profile; comparing the chemical change associated with the veins from Myrbærneset to those from Svartberget and Årsheimneset to determine if they can be chemically distinguished; and analysing compositions as igneous lithologies to estimate potential protoliths.

#### 4.3. Data Acquisition

Thirty-five samples from the eight localities were selected for analysis. Each sample was cleaned, crushed with a hammer and first milled in a Siebtechnik TIMA tungsten carbidecobalt mill in the Rock Preparation Laboratory, SEAES, University of Manchester to produce a silky powder. 12g of powder and 3.0g of carboxymethylcellulose low density wax were mixed together using a Fritsch Pulverisette 6 agate mill in the Manchester Analytical Geochemistry Unit Facility, SEAES, University of Manchester and analysis pellets were pressed from this mixture: a total of fifty-five pellets were prepared, including duplicates of five random samples. XRF analyses were carried out with an AXIOS Sequential X-ray Fluorescence Spectrometer at the Manchester Analytical Geochemistry Unit Facility, SEAES, University of Manchester and ProTrace software programmes were used to analyse major and trace element concentrations, respectively, and correct the raw data. Four standard reference materials were analysed alongside each batch of pellets to determine instrument accuracy and the duplicate analyses were carried out to determine instrument precision; details are reported in the full methodology in the Appendix [B3].

The remaining, unmixed powder of each sample was used to determine the loss on ignition (LOI). 1.0-1.5g of the powders were held in ceramic crucibles and placed in a MINERALOGY/18/CLAD/OH oven at 105°C for one hour to lose adsorbed volatiles and a Carbolite furnace at 1000°C for one hour to lose lattice-hosted volatiles. Crucibles were allowed to cool to room temperature in a desiccator before weighing.

## 4.4. Data Presentation

#### 4.4.1. Major and Trace Element Concentrations

Whole rock data are presented in the Appendix [E1-E2]. All data points were compared against their individual detection limits: those below their limit are not reported or used. All data analyses are precise to within one standard deviation. Major element data are accurate to a maximum of ±0.9wt%. Trace element accuracy varies considerably between elements and between standards with a range from 0.0-282,042.9% and 0-618.2ppm. Accuracy and precision data are also presented in the Appendix [E3-E4]. Cobalt data may not be reliable due to potential contamination from the tungsten carbide-cobalt TIMA mill used to powder the samples: all samples showed an extreme enrichment in tungsten and tungsten data were discarded for this reason.

*Figures 4.1-4.12* display major and trace element compositions for all samples and compare them to igneous reference rock compositions which are detailed in *Table 4.1*. Marker formats are as described in *Table 0.13*. *Figures 4.13-4.16* display multi-element diagrams with normalised compositions. To aid reading, graphs displaying data for each locality are presented in the locality overviews in the Appendix [A1-9].

# 4.4.2. Igneous Classification

In an effort to determine potential protoliths, assuming isochemical metamorphism has taken place, the samples are presented in an igneous classification TAS diagram (*Fig. 4.17*). Normative compositions have been calculated to provide a basis for classification and are presented in *Figures 4.18-4.22*. Compositional field labels are detailed in *Tables 4.2* and *4.3*. CIPW (Cross, Iddings, Pirsson, Washington) normative compositions were calculated using the program of Hollocher (2013) in Microsoft Excel. The classification diagrams follow the British Geological Society (BGS) Rock Classification Scheme after Gillespie and Styles (1999). The CIPW normative compositions of the igneous reference rocks are also plotted to continue their comparison. These igneous classifications are appropriate for the host orthogneisses, eclogites, peridotites and metamorphic garnet websterites which show no evidence of metasomatism if they are isochemical, metamorphic equivalents of their igneous protoliths. For the rocks and domains that are interpreted to be metasomatic in origin, these calculations are for comparison with the unaltered rocks to determine the trend in chemical change.

#### 4.5. Data Analysis

### 4.5.1. Major Elements

There are two distinct groupings in the  $SiO_2$ -MgO plot (*Fig. 4.1a*) due to a bimodal distribution of  $SiO_2$ : 70-80% for gneisses and quartz-rich vein core domains and 40-55% for all other rock types and domains. The lowest  $SiO_2$  contents are shared by

Туре	Rock	Reference*	Graph Label
	Depleted MORB Mantle	W&H05	DMM
	Primitive Mantle	McD&S95	PM
	Normal MORB	W&H05	N-MORB
	Continental Crust	R&G04	CC
	Andean Andesite	W&T84	AA
	Adakite	C05	А
	Komatiite	C93	К
	Tonalite-Trondhjemite-	C93	TTG
Xenolites	Garnet Peridotite	M91	GP
Xenonces	Spinel Peridotite	M01	SD
Basic	Basalt	10131	3F 1
Dasic	Diabaso		1
	Delorito		2
	Cabbro		3
	Norito		4 5
	Tholoiite		5
	Howeiite		0
Intermediate	Hawalite		/
Aikaime	Latite		8
	Muracarita		9
			10
	Trachyandesite	LIVI76	11
		LIVI76	12
Oversaturated	Ademeilite	LM76	13
Acid	Andesite	LIM76	14
	Dacite	LIM76	15
	Diorite	LIM76	16
	Granite	LM76	17
	Granodiorite	LM76	18
	Rhyodacite	LM76	19
	Rhyolite	LM76	20
	Tonalite	LM76	21
Saturated Sialic	Syenite	LM76	22
	Trachyte	LM76	23
Ultrabasic	Anorthosite	LM76	24
	Dunite	LM76	25
	Harzburgite	LM76	26
	Lherzolite	LM76	27
	Peridotite	LM76	28
	Pyroxenite	LM76	29
	Websterite	LM76	30
Undersaturated	Basanite	LM76	31
Basic	Nepheline Leucite Basalt	LM76	32
	Nephelinite	LM76	33
	Tephrite	LM76	34
Undersaturated	Lujavrite	LM76	35
Sialic	Nepheline Syenite	LM76	36
	Phonolite	LM76	37
	Tinguaite	LM76	38

10018 4.1. 10118003 1818181118 1018 01001 100813	aneous reference rock arap	h labels.
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\* W&H05 – Workman and Hart (2005); McD&S95 – McDonough and Sun (1995); R&G04 – Rudnick and Gao (2004); W&T84 – Weaver and Tarney (1984); C05 – Condie (2005); C93 – Condie (1993); M91 – McDonough (1991); LM76 – le Maitre (1976).

quartz-absent garnetite vein cores, websterites and peridotites. MgO compositions vary between ~0 and 27% and so best reflects the overall compositional change between ultramafic, mafic and felsic compositions. The general trend follows that of the igneous rock reference suite demonstrating an origin associated with magmatic fractionation. Eclogites have similar SiO<sub>2</sub> (and lower than Svartberget peridotites) but a range of MgO contents: most plot close to N-MORB and other basaltic rocks (1,4,5) but those from Nybø have higher MgO indicating a more ultramafic character. All have lower MgO contents than associated peridotites reflecting a more basaltic nature. Garnet peridotites trend towards ultramafic, mantle rocks (27-45% MgO) but have lower MgO (21-27%). The compositional field of the websterites (12-22% MgO) overlaps that of the peridotites and eclogites (6-14% MgO) demonstrating an intermediary composition. The three rock types from Kolmannskog show a considerable range in MgO. Garnet peridotite from Kolmannskog is more MgO-rich than that at Svartberget and Myrbærneset supporting the more olivine-rich assemblages and the garnet peridotite and websterites from Svartberget are consistently richer in SiO<sub>2</sub> than those from Myrbærneset but have similar MgO concentrations. The biotite websterite at the contact at Svartberget has similar MgO and SiO<sub>2</sub> compositions as the peridotite suggesting these components were not greatly affected by metasomatism. There is little distinction between websterite selvages and websterite domains except for at Arsheimneset where the selvage (vein-adjacent) garnet websterite has considerably higher SiO<sub>2</sub>: this probably reflects the presence of abundant carbonate in the domain garnet websterite. Clinopyroxenite selvages lie between websterite selvages and garnetites producing a zoning pattern similar to that of the veins. Vein cores have the largest compositional  $SiO_2$ -MgO field ranging from compositions similar to websterite selvages towards the gneisses. Those from Svartberget and Myrbærneset are less SiO<sub>2</sub>-rich and more MgOrich than Årsheimneset and Remøysunde. The quartz-white mica vein core from Årsheimneset and quartz-rich garnetite vein core from Remøysunde have higher MgO concentrations than the country rocks but similar SiO<sub>2</sub> contents.

*Figure 4.1b* displays MgO-Al<sub>2</sub>O<sub>3</sub> compositions and similarly shows the samples strongly follow the trend of the igneous suite and re-produce the zoning pattern seen in the SiO<sub>2</sub>-MgO plot. The concentration of Al<sub>2</sub>O<sub>3</sub> against MgO demonstrates the purity of the garnetite veins: cores not 'diluted' by quartz or clinopyroxene plot above the trend. Peridotites, websterites and clinopyroxenite selvages with low



*Figure 4.1: Whole rock chemical data vs MgO of all samples. Marker formats as described in* Table 0.13. *Reference rocks as listed in* Table 4.1. *a) SiO*<sub>2</sub>-*MgO plot; b) Al*<sub>2</sub>*O*<sub>3</sub>-*MgO plot; c) FeO*-*MgO plot.* 

modal abundance of garnet contain the lowest Al<sub>2</sub>O<sub>3</sub> concentrations (<6%) demonstrating they are dominantly correlated with garnet contents (and feldspar in the gneisses). There is a positive correlation between FeO and MgO (*Fig. 4.1c*) but the lithological pattern is less well defined by FeO. The increased scatter in vein domains may reflect the heterogeneous presence of various Fe oxide and sulphide phases which form minor components in some rocks. Samples from Myrbærneset, Eiksunddal and Kolmannskog plot above the main trend showing they are enriched in Fe and Svartberget peridotites plot just below the trend demonstrating they are not enriched in Fe.

There is considerable scatter away from the igneous suite trend of CaO (Fig. 4.2a). Lowest CaO compositions are found in plagioclase-poor gneisses, clinopyroxene-poor garnetite vein cores at Årsheimneset, Remøysunde and Svartberget and clinopyroxene-poor peridotites and websterites at Myrbærneset and Kolmannskog. Highest CaO contents are found in eclogites and clinopyroxenite selvages reflecting a correlation with clinopyroxene (and plagioclase in the gneisses). Samples follow the MgO-Na<sub>2</sub>O trend of the igneous suite (*Fig. 4.2b*) although host gneisses show a range of lower concentrations for a given MgO suggesting they have lost components. Eclogites from Myrbærneset and Kolmannskog have the highest Na concentrations and peridotites have consistently low Na<sub>2</sub>O concentrations. Båtneset boasts the highest Na<sub>2</sub>O concentrations of the websterites. Vein cores have lower Na<sub>2</sub>O concentrations than other vein zones suggesting it was readily added during alteration. Highest K<sub>2</sub>O contents are found in the gneisses from Svartberget, micarich vein cores, glimmerite selvages, biotite clinopyroxenite selvages and biotite-rich websterite (*Fig. 4.2c*) showing a strong correlation with mica and the highest  $K_2O$ contents are found in biotite- and microcline-bearing leucogneiss [QC36L] demonstrating a concentration of K. Most metasomatic zones are enriched in K<sub>2</sub>O demonstrating its ubiquitous addition during alteration.

TiO<sub>2</sub> concentrations are highest in rutile-rich eclogites, vein cores and websterites (*Fig. 4.3a*). Except for vein cores, samples from Svartberget have relatively low TiO<sub>2</sub> concentrations more similar to mantle rocks. Many eclogites show elevated TiO<sub>2</sub> contents.  $P_2O_5$  concentrations are commonly less than ~0.3% (*Fig. 4.3b*). Two samples have high  $P_2O_5$  (>2.5%): an apatite-bearing websterite from Eiksunddal [QC30B] and an apatite-rich garnetite vein core from Svartberget [QC36U]. Other samples with



*Figure 4.2: Whole rock chemical data vs MgO of all samples. Marker formats as described in* Table 0.13. *Reference rocks as listed in* Table 4.1. *a) CaO-MgO plot; b)*  $Na_2O$ -MgO plot; *c)*  $K_2O$ -MgO plot.

elevated  $P_2O_5$  concentrations include eclogite, websterite and garnetite vein core from Årsheimneset and eclogite from Kolmannskog. There is no correlation of  $P_2O_5$ with CaO, Th, Ce or Y suggesting no one phase dominates  $P_2O_5$  concentrations: two groups are defined suggesting LREE-rich monazite and LREE-poor apatite are involved.

#### 4.5.2. Trace Elements

Mn is quite common in all mafic-ultramafic samples and shows no particular trend with MgO: most contain less than 2500ppm (*Fig. 4.3c*) however garnetites and glimmerite from Svartberget and Myrbærneset contain 4000-5500ppm forming a separate cluster. The gneiss from Årsheimneset contains considerably more Mn than that from Svartberget sitting alone at 3019ppm and reflects the Mn-rich garnets present. Overall Mn concentrations indicate garnet preferentially hosts Mn but its depletion in some garnet-rich rocks suggests it has another control.

Two trends are evident between MgO and Cr (Fig. 4.4a): a Cr-poor-MgO-rich trend and a Cr-rich-MgO-rich trend. The Cr-poor trend consists of gneisses, Båtneset and Eiksunddal websterites, Kolmannskog peridotite and all samples from Myrbærneset and Nybø. Those from Årsheimneset, Svartberget and Remøysunde and the websterite from Kolmannskog form the Cr-rich reach the concentrations of primitive mantle (~2700ppm). The eclogite from Årsheimneset and Remøysunde have lower Cr concentrations than those of their respective websterites. In both trends Cr is highest in the more ultramafic rock types suggesting it behaves as a compatible element in MgO-rich silicates and/or spinels. Ni concentrations do not extend as high as those of Cr (<1100ppm) nor do they reach concentrations similar to those in primitive mantle (Fig. 4.4b) suggesting lower olivine contents in which it is preferentially held. Two trends are present but less well-defined than those of MgO-Cr and samples from Svartberget join the lower concentration trend. The peridotites from Svartberget have similar Ni concentrations to those at Kolmannskog and Myrbærneset (400-550ppm). Most samples have less than 600ppm Ni but pyroxene-rich samples from Remøysunde and Årsheimneset have higher concentrations. It also shows no correlation with K<sub>2</sub>O (biotite) or Al<sub>2</sub>O<sub>3</sub> (garnet) suggesting it is held in pyroxenes in olivine-absent rocks. *Figure 4.6* shows a Ni-Cr plot where three groups are evident: Ni-rich, Cr-poor rocks from Myrbærneset, Kolmannskog (except websterite), Nybø,



Figure 4.3: Whole rock chemical data vs MgO of all samples. Marker formats as described in Table 0.13. Reference rocks as listed in Table 4.1. a)  $TiO_2$ -MgO plot; b)  $P_2O_5$ -MgO plot; c) Mn-MgO plot.



Figure 4.4: Whole rock chemical data of compatible elements vs MgO of all samples. Marker formats as described in Table 0.13. Reference rocks as listed in Table 4.1. a) Cr-MgO plot; b) Ni-MgO plot; c) Sc-MgO plot.

Eiksunddal and Båtneset; Ni-rich, Cr-rich rocks from Årsheimneset and Remøysunde extending towards primitive mantle; and Ni-'poor', Cr-rich rocks from Svartberget. Sc contents range from 0-45ppm (*Fig. 4.4c*): highest concentrations are found in the websterite from Kolmannskog and peridotite from Svartberget. The two trends picked out by Cr are repeated suggesting the behaviour of Sc is similar to Cr except it is the Sc-poor trend that extends towards the concentration of mantle rocks. There is no correlation of Sc with CaO, Al<sub>2</sub>O<sub>3</sub>, TiO<sub>2</sub> or Zr suggesting it is not dominantly controlled by any one phase (clinopyroxene, garnet, rutile or zircon) but is most compatible with MgO-rich silicate phases. There is no strong correlation of Zn or Cu with MgO although a positive, scattered correlation exists between Ni and Zn reflecting their similar behaviour. No strong correlations are seen with SO<sub>3</sub> to indicate an association particularly with accessory sulphide phases. Most samples contain less than 210ppm Zn and 150ppm Cu although the websterite from Kolmannskog [QC32A] has Cu concentrations of 1245ppm.

Ga forms the least scattered, least anomalous trend with MgO: it is negatively correlated (Fig. 4.5a) suggesting it behaves incompatibly in MgO-rich silicate phases and is concentrated into the fluid phase during fractionation processes such as crystallisation, metamorphism and metasomatism. Three outliers from the trend are garnetites from Remøysunde and Myrbærneset and the gneiss from Årsheimneset: all plot below the main trend. The gneisses from Svartberget have the highest Ga contents and the peridotites from Svartberget have the lowest, similar to primitive mantle concentrations. All garnetites plot on the lower side of the trend. Fig. 4.7 shows that Ga has a mostly positive correlation with Al<sub>2</sub>O<sub>3</sub> but some samples form a negatively correlated sub-trend. Ga typically substitutes for Al and their positive correlation suggests an Al-rich phase is responsible for the Ga concentration: the generally lower concentrations in garnet-rich rocks however suggest Ga is not only controlled by garnet. CaO-rich and Na<sub>2</sub>O-rich samples mostly have high Ga but CaO and Na<sub>2</sub>O are respectively very low and low in the gneisses where Ga is highest. This suggests Ga is held in an Al-rich, Ca-Na-poor phase in the gneisses (e.g. an aluminosilicate - Svartberget gneisses do contain sillimanite or alkali feldspar) or the correlations in the other rock types with CaO and Na<sub>2</sub>O are incidental.



Figure 4.5: Whole rock chemical data of incompatible elements vs MgO of all samples. Marker formats as described in Table 0.13. Reference rocks as listed in Table 4.1.a) Ga-MgO plot, Ga represents incompatible elements; b) Y-MgO plot, Y represents HREE; c) Nb-MgO plot, Nb represents HFSE.



*Figure 4.6: Whole rock Ni-Cr data of all samples. Marker formats as described in* Table 0.13. *Reference rocks as listed in* Table 4.1.



Figure 4.7: Whole rock  $Ga-Al_2O_3$  data of all samples. Marker formats as described in Table 0.13. Reference rocks as listed in Table 4.1.

Y, which behaves like HREE, is highest in garnetites and eclogites (~20-60ppm) and lowest in peridotites (<10ppm) (*Fig. 4.5b*): there is a crude negative correlation with MgO. Y is highest in an apatite-rich garnetite vein core [QC36U] and most vein cores from Svartberget plot above the main trend suggesting REE and Y were transported in the fluid and concentrated in garnet and apatite. In comparison the leucogneiss at the Svartberget contact has Y concentrations below the detection limit, as does the quartz-white mica vein core from Årsheimneset, unlike the gneisses suggesting some Y remains in the fluid source.

Nb concentrations are mostly below 20ppm but vein cores from Svartberget have 20-28ppm and Svartberget gneisses extend from 32-71ppm (*Fig. 4.5c*). There is no correlation of Nb with MgO, CaO, Zr or Y. Many K<sub>2</sub>O-rich and Al<sub>2</sub>O<sub>3</sub>-rich samples are Nb-rich but form no particular trend. Some samples form a positive trend between Nb and TiO<sub>2</sub> suggesting a correlation with rutile but the high Nb concentrations in TiO<sub>2</sub>-poor gneisses suggests another phase is also responsible for its content such as titanite or biotite. Like Y, the quartz-white mica vein core from Årsheimneset has very low Nb concentrations despite high concentrations in the gneiss. These high Nb contents are not correlated with Zr or Y and they occur in Ga-rich, K<sub>2</sub>O-Al<sub>2</sub>O<sub>3</sub>-rich, MgO-poor gneisses so perhaps feldspar or sillimanite also host Nb. The compositional variation of trace elements such as Nb and Y between the country rock samples may indicate evolution through geological processes such as partial melting or there is a heterogeneous distribution of accessory phases strongly controlling their abundance.

Sr shows the greatest correlation with CaO with concentrations gently increasing up to ~260ppm with increasing CaO (*Fig. 4.8*) suggesting Sr is hosted in clinopyroxene. Higher Sr concentrations (up to 416ppm) are present in eclogite from Myrbærneset and websterites from Eiksunddal and Båtneset. The K-feldspar-rich leucogneiss and gneiss from Årsheimneset have above-trend Sr contents in the CaO-poor samples suggesting K-rich feldspar hosts Sr in the country rocks. V is positively correlated with TiO<sub>2</sub> in general (*Fig. 4.9*) suggesting it is hosted in rutile. One particular garnetite vein core [QC36G] (with glimmerite) from Svartberget, websterites from Eiksunddal and Båtneset and samples from Myrbærneset are all enriched in TiO<sub>2</sub> relative to V suggesting other Ti-rich phases may be present such as ilmenite or magnetite.



*Figure 4.8: Whole rock Sr-CaO data of all samples. Marker formats as described in* Table 0.13. *Reference rocks as listed in* Table 4.1.



Figure 4.9: Whole rock V-TiO<sub>2</sub> chemical data of all samples. Marker formats as described in Table 0.13. Reference rocks as listed in Table 4.1.



Figure 4.10: Whole rock Ba- $K_2O$  data of all samples. Marker formats as described in Table 0.13. Reference rocks as listed in Table 4.1.



*Figure 4.11: Whole rock Ce-Th data of all samples. Marker formats as described in* Table 0.13. *Reference rocks as listed in* Table 4.1.

LILE, represented by Ba in Figure 4.10, are well-correlated with K<sub>2</sub>O although Svartberget leucogneisses [QC36K, QC36L] and biotite websterite [QC36S] plot below the trend and no host gneisses plot as high as the felsic reference rocks: this suggests LILE are dominantly hosted by mica (possibly preferring white mica over biotite) and the country rocks have lost components. Th contents range up to 30ppm but most are below 5ppm and are highest in vein cores and the gneiss from Svartberget (see *Fig. 4.11*). There is no correlation with  $P_2O_5$  suggesting a Th-bearing phosphate does not control their concentrations. Most U contents are below 4ppm although a concentration of over 7ppm is found in a leucogneiss from Svartberget [QC36K]. LILE concentrations are generally lowest in the mafic-ultramafic bodies and highest in vein cores and clinopyroxenite selvages from Svartberget and Årsheimneset. Ce, representing LREE, is not correlated with major elements such as MgO or K<sub>2</sub>O. It is highest in Th- and Ba-rich Svartberget vein cores and gneiss (Fig. 4.11) which are rich in monazite and apatite. Other samples contain less than 60ppm Ce. Of HFSEs, Hf is strongly correlated with Zr (Fig. 4.12) suggesting zircon dominantly controls Hf and Zr contents and both are enriched in biotite-rich Svartberget vein cores.



*Figure 4.12: Whole rock Hf-Zr data of all samples. Marker formats as described in* Table 0.13. *Reference rocks as listed in* Table 4.1.

### 4.5.3. <u>Trace Elements Patterns</u>

The various localities show a range of trace element enrichment and depletion relative to pyrolite (*Figs. 4.13-4.14*). Most normalised element variation is over an order of magnitude (e.g. U, REE, Zn) with some elements varying over three orders of magnitude (e.g. Rb, Zr, Cr). Vein related samples from Svartberget show the greatest enrichment in LILE relative to pyrolite. Sr shows the least variation from 0.43-20.92 times the concentration in pyrolite. Rb ranges from 0.83 to ~870 times that of pyrolite and Cs ranges up to over 3400 times that of pyrolite. The host mafic-ultramafic bodies and websterites, excluding Svartberget samples, show the smallest overall range between 1.17-36.67 (Rb) (*Fig. 4.13b*) and (0.83-64.67 Rb) (*Fig. 4.13c*), respectively.

HFSE are typically similar or enriched relative to pyrolite except for two samples [QC36S, QC36L] which show a similar depletion of Zr (0.20-0.25): these are the biotite websterite and leucogneiss at the Svartberget contact. Garnetite vein cores from Svartberget show the greatest enrichment relative to pyrolite (<110 times) although the Svartberget gneisses have the highest Nb contents (<107 times). Websterite domains and websterite and clinopyroxenite selvages have similar contents.

Period IV transition metals show the most dramatic variation. All samples show relative depletions in Cr and Ni relative to pyrolite and to the other transition metals as they are strongly compatible during mantle melting: this depletion is greatest in the host gneisses as expected and the quartz-white mica vein core from Årsheimneset [QC26F]. Sc, V, Cu and Zn show similar variation with most lying between 0.05 and 5 times pyrolite. The peridotite and biotite websterite from Svartberget show the least variation and are most similar to pyrolite. Eclogites and websterites show similar patterns and relative enrichments and depletions although eclogites extend to greater Cr and Ni depletions. The two groupings in Cr content of host bodies and websterites seen on the MgO-Cr plot are evident (*Figs. 4.13b,c*). All REE measured are enriched relative to chondrite although Yb analyses were rarely above the detection limit (*Figs. 4.15-4.16*): samples range from 2.4 times (Y) to just below 400 times (La). Most samples show gently decreasing patterns towards HREE; however one sample [QC36U], an apatite-rich garnetite vein core shows a slight enrichment of Y (HREE) over Sm (MREE) and some websterites and host bodies show



Figure 4.13: Multi-element spider diagram displaying pyrolite-normalised trace element data. Marker formats as described in Table 0.13. Pyrolite normalisation values from McDonough and Sun (1995). a) gneisses; b) host bodies; c) websterites; d) veins.



Figure 4.14: Multi-element spider diagram displaying pyrolite-normalised trace element data for all samples. Marker formats as described in Table 0.13. Pyrolite normalisation values from McDonough and Sun (1995).



*Figure 4.15: Multi-element diagram displaying chondrite-normalised rare earth element data. Marker formats as described in* Table 0.13. *Chondrite normalisation values from McDonough and Sun (1995). a) gneisses; b) host bodies; c) websterites; d) veins.* 

a slight enrichment of Sm (MREE) over Nd (LREE). Vein zones re-produce the zoning sequence with garnetite cores more enriched than clinopyroxenite selvages which are more enriched than websterite selvages. Gneisses have a similar range to garnetite vein cores although data are incomplete.



Figure 4.16: Multi-element diagram displaying chondrite-normalised rare earth element data for all samples. Marker formats as described in Table 0.13. Chondrite normalisation values from McDonough and Sun (1995).

### 4.5.4. Igneous Compositions

In the TAS diagram (*Fig. 4.17*), all eclogites fall in the basalt field. The Svartberget olivine garnet websterite falls in the basalt field demonstrating a more mafic character whilst that from Myrbærneset and the garnet peridotite Kolmannskog fall in the foidite and picrobasalt fields, respectively. Garnet websterites spread from the foidite to basaltic andesite fields. Those thought to have metamorphic origins, at least in part, plot up to the middle of the basalt field. The websterites from Båtneset have high alkali contents of ~3% and lie above the main igneous trend suggesting they have been enriched through a process other than magmatic differentiation. Gneisses cross the granite-granodiorite fields. All but the microcline-rich leucogneiss [QC36L] show lower alkali concentrations than the igneous reference suite more similar to andesite contents suggesting they may have lost components.



Figure 4.17: TAS diagram after le Bas et al. (1986) and Middlemost (1994). Marker formats as described in Table 0.13. Line intersections taken from Gillespie and Styles (1999). Compositional fields listed in Table 4.2. Reference rocks as listed in Table 4.1.

Table 4.2: TAS diagram compositional field labels. From	
Gillespie and Styles (1999).	
	-

Label	Rock (extrusive/intrusive)
1	foidite/foidolite
2	picrobasalt/peridotgabbro
2	tephrite (normative ol <10%)/basanite
5	(normative ol >10%)/foid gabbro
4	phonotephrite/foid monzodiorite
5	tephriphonolite/foid monzosyenite
6	phonolite/foid syenite
7	trachybasalt/alkali gabbro/monzogabbro
8	basaltic trachybasalt/monzodiorite
9	trachy-andesite/monzonite
10	tracyte (normative qtz <20%)/trachydacite
10	(normative qtz >20%)/ quartz monzonite
11	basalt/gabbro
12	basaltic andesite/gabbroic diorite
13	andesite/diorite
14	dacite/granodiorite
15	rhyolite/granite

The metasomatic websterites from Svartberget plot in the basaltic andesite field whilst the biotite websterite at the contact plots in the basalt field. The greater silica content may reflect the addition of Si during metasomatism or just greater silica content of the host body. The biotite websterite from Svartberget [QC36S] contains the highest alkali contents of any websterite of 3.66%. Biotite-rich vein cores from Svartberget fall in the alkali-rich tephrite/basanite field whilst others fall in the foidite and basalt fields. Garnetite from Myrbærneset falls in the picrobasalt field. Those from Remøysunde and Årsheimneset fall in the basalt and granodiorite fields. The clinopyroxenite selvages fall in the basalt and basaltic andesite fields. The quartzwhite mica vein core at Årsheimneset [QC26F] falls in the granite field showing a similar composition to the gneisses near the Svartberget contact.

All samples but two plot on the QAP and FAP diagrams (Figs. 4.18-4.19): the metamorphic websterite from Myrbærneset [QC34C] and metamorphic-metasomatic websterite from Årsheimneset [QC26G] have ultramafic CIPW normative compositions s.s. (QAPF <10%) falling in the olivine orthopyroxenite and orthopyroxenite fields, respectively (Fig. 4.22). Gneisses and a leucogneiss from Svartberget plot in the quartz granite field (1b) and the microcline-rich leucogneiss from Svartberget [QC36L] plots in the syenogranite field (3a) whilst the felsic reference rocks extend into the monzogranite field (3b). Most mafic-ultramafic metamorphic rocks fall in the diorite/gabbro/anorthosite field (10) and are expanded in Figures 4.20-4.21. Normative compositions of the garnet peridotites plot as monzonite (Kolmannskog), foid-bearing monzodiorite/monzogabbro (Myrbærneset) and mela-olivine-clinopyroxene norite (Svartberget); none have ultramafic s.s. normative compositions demonstrating an evolved nature and not peridotites s.s.. Normative compositions of eclogites plot as gabbro, foid-bearing gabbro and olivine orthopyroxene meso-gabbro/clinopyroxene norite near the tholeiite, N-MORB and norite reference rocks, respectively. Websterites plot in various fields in the normative composition diagrams. The biotite garnet websterite from Myrbærneset plots as monzogabbro. Other websterites plot as foid-bearing monzodiorite/monzogabbro (Båtneset), quartz monzodiorite/monzogabbro (Årsheimneset) and mela-olivine gabbro/norite (Nybø, Kolmannskog, Eiksunddal, Remøysunde). Eclogite normative compositions are all more plagioclase- and clinopyroxene-rich than their respective websterites.



Figure 4.18: QAP ternary diagram displaying CIPW normative compositions of appropriate samples. Q = quartz, A = alkali feldspar, P = plagioclase feldspar. Marker formats as described in Table 0.13. Composition fields as described in Table 4.3. Reference rocks as listed in Table 4.1.

abels. Froi	n Gillespie and Styles (1999).
Label	Rock
1a	quartzolite
1b	quartz granite
2	alkali feldspar granite
3a	syenogranite
3b	monzogranite
4	granodiorite
5	tonalite
6	alkali feldspar syenite
6*	quartz alkali feldspar syenite
6'	foid-bearing alkali feldspar syenite
7	syenite
7*	quartz syenite
7'	foid-bearing syenite
8	monzonite
8*	quartz monzonite
8'	foid-bearing monzonite
9	monzodiorite/monzogabbro
9*	quartz monzodiorite/monzogabbro
9'	foid-bearing monzodiorite/monzogabbro
10	diorite/gabbro/anorthosite
10*	quartz diorite/gabbro/anorthosite
10'	foid-bearing diorite/gabbro/anorthosite
11	foid-syenite
12	foid-monzosyenite
13	foid-monzosyenite/monzogabbro
14	foid-diorite/gabbro
15	foidolite

Table 4.3: QAP and FAP diagram compositional field
labels. From Gillespie and Styles (1999).



Figure 4.19: FAP ternary diagram displaying CIPW normative compositions of appropriate samples. F = feldspathoid, A = alkali feldspar, P = plagioclase feldspar. Marker formats as described in Table 0.13. Composition fields as described in Table 4.3. Reference rocks as listed in Table 4.1.



*Figure 4.20: PI-pyx-ol ternary diagram displaying CIPW normative compositions of appropriate samples. Marker formats as described in Table 0.13. Reference rocks as listed in the Table 4.1.* 



*Figure 4.21: PI-opx-cpx ternary diagram displaying CIPW normative compositions of appropriate samples. Marker formats as described in* Table 0.13. *Reference rocks as listed in* Table 4.1.



Figure 4.22: Ol-opx-cpx ternary diagram for ultramafic rocks (QAFP<10%) displaying CIPW normative compositions of appropriate samples. Marker formats as described in Table 0.13. Reference rocks as listed in Table 4.1.

Of the metasomatic rocks, for comparison, the garnetite vein core from Remøysunde falls in the quartz granite field (1b) but away from the gneisses with no alkali feldspar. The guartz-white mica vein core from Årsheimneset plots just in the alkali feldspar granite field (2). The syenite field (7) is characterised by biotite websterite, glimmerite selvage and a garnetite vein core from Svartberget reflecting the addition of alkalis but not silica. Clinopyroxenite selvages from Svartberget and Årsheimneset fall in the monzonite field (8) as does the apatite-rich garnetite vein core which plots near the peridotite from Kolmannskog. The quartz monzodiorite/monzogabbro field (9\*) is characterised by the websterite selvage and garnetite vein core from Årsheimneset which lies near the tonalite reference rock (21). The websterite domain and selvages and clinopyroxenite selvage from Svartberget fall in the monzodiorite/monzogabbro field (9) and the foidolite field (15) contains the garnetite from Myrbærneset with extremely silica-undersaturated normative composition trending towards the nephelinite reference rock (33). Of those that fall in the diorite/gabbro/anorthosite field (10), in the pl-px-ol plot (Fig. 4.20) the garnet clinopyroxenite from Remøysunde falls in the gabbro/norite field whilst the eclogitic vein selvage from Remøysunde and websterite and clinopyroxenite selvages from Myrbærneset fall in the olivine gabbro/norite composition field.

## 4.6. Data Interpretation

### 4.6.1 Host Mafic-Ultramafic Body Protoliths

Due to the low concentrations of Cr, Ni and Sc in the host gneisses and their compatible nature in common ultramafic phases, they are unlikely to be removed from the rocks during metamorphism or greatly introduced in a felsic partial melt during metasomatism; therefore they can be used to infer characteristics of the host body protoliths. The two trends depicted by Cr and Sc (*Figs. 4.4a,c*) can be interpreted to represent different protolith compositions for the various bodies. Those from Svartberget, Årsheimneset and Remøysunde formed from a Cr-Sc-rich magma and those from Eiksunddal, Båtneset, Nybø and Myrbærneset formed from a Cr-Sc-poor magma. The peridotite and websterite from Kolmannskog are found in separate groups demonstrating that both magma compositions can occur together, perhaps forming in large, crustal bodies. The lower MgO-Cr and higher FeO-TiO<sub>2</sub> contents of the Myrbærneset olivine garnet websterite and Kolmannskog garnet peridotite compared to mantle ultramafic rocks also confirms the status of the bodies as Fe-Ti crustal bodies as opposed to mantle Mg-Cr rocks as hypothesised. The Svartberget garnet peridotite, however, whilst not being MgO-rich like mantle peridotite, does not have enriched FeO and TiO<sub>2</sub> concentrations and is relatively more enriched in Cr than other peridotites with concentrations similar to primitive mantle. It is also richer in SiO<sub>2</sub> and the Ni-Cr plot (*Fig. 4.6*) indicates it is poor in Ni relative to the mantle trend. The lower MgO and Ni and apparent 'gain' of SiO<sub>2</sub> may represent an anomalous horizon in the crustal body which had less modal olivine, perhaps resulting from physical separation of the magma from an early cumulate, but was pyroxene and chromite-rich. On the TAS diagram (*Fig. 4.17*) Svartberget plots in the basaltic field reflecting its SiO<sub>2</sub>-enriched nature and if the analysed sample is representative of the whole body it should not be referred to as a peridotite.

The sequential but overlapping MgO-SiO<sub>2</sub> compositional fields of peridotites and websterites and eclogites suggests at an evolutionary link between the different rock types. The eclogites, which have 45-50% SiO<sub>2</sub>, are considered to be transitional between cumulate and separated, basaltic magmas as low silica eclogites with SiO<sub>2</sub> contents below 45% are considered to have protoliths associated with cumulates (Jahn et al., 2003). It is possible the eclogite domains have undergone little chemical change as they all plot within the basalt field on the TAS diagram (Fig. 4.17) and follow the major element compositional trends of the igneous reference rocks although relative K<sub>2</sub>O-loss and TiO<sub>2</sub>-enrichment is indicated: deeply subducted metabasaltic rocks are known to have equivalent compositions as their protoliths (e.g. Bebout, 2007) although eclogites are thought to have lower K than basalts (Forbes, 1965). Their normative compositions and variable Cr-Ni-Sc compositions reflect a range in the mafic-ultramafic nature of the protoliths. The Cr-Sc trends with MgO distinguish between the textural types of eclogites (and garnet websterites) (Årsheimneset and Remøysunde vs Nybø and Myrbærneset). Major element compositions are variable but do not distinguish between textural types as the coarser, decussate eclogites from Årsheimneset and Remøysunde lie between the finer, more pastel and granoblastic-lineated eclogites from Nybø, Myrbærneset and Kolmannskog. Regarding a structural distinction, those associated with peridotitebearing bodies, Myrbærneset and Kolmannskog, are the more evolved, Mg-poor, Sirich eclogites. They may therefore represent segregations of more evolved basaltic melt.

# 4.6.2. Origins of Websterites

There is no major chemical distinction between the different colour and textural types of garnet websterites. The coarse-pegmatitic, decussate, deeply coloured websterite at Remøysunde has very similar major element compositions to the finer grained, lineated, pastel-coloured websterite at Nybø. Furthermore, the selvage websterites are similar to the larger domain websterites indicating no great distinction in structural type either. Like the eclogites, however, a distinction is seen in the Cr-Sc contents between textural types. Ca, Na, Fe, K contents do vary but not according to texture or structure.

The concentrations of MgO, Cr and Ni (and to some extent Sc) of the Årsheimneset and Remøysunde garnet websterites are higher than their corresponding eclogites suggesting they are associated with a more ultramafic and olivine-bearing protolith and not through alteration of eclogite; although increases may arise through concentration effects with the preferential loss of other elements during metasomatic redistribution. The pattern is also seen in metamorphic garnet websterites from Nybø, Kolmannskog and Myrbærneset. The major element concentrations of the garnet websterites at Årsheimneset and Remøysunde sit along the igneous suite and show no great enrichment in trace elements. The samples analysed cannot, therefore, be geochemically distinguished as being metasomatic in origin; however, neither, can those from Svartberget. The signature of metasomatism is discussed further below.

The websterites with high CaO-Sr such as those from Kolmannskog, Nybø and Remøysunde trending towards and overlapping with the high CaO-Sr eclogite field suggests these had protoliths bearing anorthite plagioclase or clinopyroxene. Websterites from Båtneset with high Na<sub>2</sub>O, Al<sub>2</sub>O<sub>3</sub> and K<sub>2</sub>O and which plot above the main CaO-Sr trend suggests their protoliths were rich in alkali feldspar (e.g. Thrane, 2002). These websterites may reflect a horizon in their parent, crustal bodies where considerable feldspar was crystallising or accumulating. Websterites from Myrbærneset have low CaO-Sr contents suggesting these formed at a different horizon where plagioclase or clinopyroxene were not dominant phases: this is supported by their ultramafic, orthopyroxene-rich normative compositions. The only other main chemical distinction of samples from Myrbærneset is their strong enrichment of Fe.

#### 4.6.3. Geochemical Signature of Metasomatism

The difficulty in analysing this fluid-rock system is the signature which would reflect the addition of felsic materials from the host gneisses and Si-Al-K-H<sub>2</sub>O-rich metasomatic signature suggested by mineralogical changes (Chapters 2 and 3) is similar to the signature produced through the evolution of magma, such as a depletion of Cr, MgO and increased SiO<sub>2</sub>, K<sub>2</sub>O and Al<sub>2</sub>O<sub>3</sub> concentrations. Such change is observed between peridotite and websterite. It therefore cannot be determined immediately whether magmatic evolution or metasomatism is responsible but that if metasomatism is the cause, it produced chemical change indicative of a felsic fluid.

The re-production of the vein zonation pattern by MgO-Al<sub>2</sub>O<sub>3</sub> (*Fig. 4.1b*), -Ni (*Fig. 4.4b*) and -Ga (*Fig. 4.5a*) contents suggests that MgO, Al<sub>2</sub>O<sub>3</sub>, Ni and Ga are good reflectors of the extent of fractionation and/or metasomatism with Al and Ga behaving incompatibly, becoming concentrated in the fluid and MgO and Ni behaving compatibly and becoming depleted in the fluid. Within the ultramafic-felsic zoning pattern garnetites often plot off trend with lower Si and higher Al content and the zoning pattern is disrupted for Fe, Mn and Cr due to their preference for garnet. SiO<sub>2</sub>, FeO, Cr and Mn are therefore not good indicators of the overall character of rocks or metasomatism.

Vein domains from Svartberget and the biotite websterite at the contact show considerable enrichment in alkali and LILE contents demonstrating that increased alkalis and LILE are a good signature of the metasomatism: this increase can be considerable or subtle. HFSE, LREE and HREE contents are also increased, particularly in biotite-rich garnetite vein cores, suggesting both compatible and incompatible elements are mobile in the metasomatic agent and residual fluids are involved in alteration. This supports the hypothesis that the breakdown of phengite is involved in fluid production as phengite would be the dominant host for LILE in UHP felsic rocks: Ba is considered to reflect the presence of phengite in subduction zones (Sorensen *et al.*, 1997).

Relative to peridotite, garnet websterite selvages and domains at Svartberget contain higher silica and alkali concentrations suggesting initial metasomatism is associated with their addition from the metasomatic agent. Clinopyroxenite selvages at both Svartberget and Årsheimneset again have higher alkali concentrations than their websterite selvages but lower silica contents: this is well illustrated on the TAS diagram with trends projecting towards alkali-rich silica-undersaturated rocks such as nephelinite (33) and tephrite (34) and garnetite vein cores having silica contents similar to ultramafic rocks. This suggests silica causes initial enrichment at the alteration front and Si is mobile but metasomatism progresses through the interaction of other components such as K closer to the wall-rock. The similar SiO<sub>2</sub> composition of the biotite websterite at the Svartberget contact to the body peridotite suggests metasomatism may not involve considerable SiO<sub>2</sub> addition. However, vein cores from Årsheimneset and Remøysunde contain abundant guartz demonstrating considerable addition of  $SiO_2$ . The variable change suggests  $SiO_2$  is greatly redistributed during metasomatism. The major element composition of the metasomatic fluid at Svartberget is investigated in Chapter 5.

The different relative compositional change at the Svartberget contact compared to the interior veins may result from the timescale that chemical gradients exist, their steepness or their ability to evolve. For example, the relative lack of garnet in the contact zones may reflect the inability of fluid-rock interaction to reduce the silica content of the greater volume of fluid. Alternatively, it may result from a fluid of different composition.

The different normative compositions of the vein domains from Myrbærneset and the relative depletion in alkalis, LILE and LREE relative to their host rocks chemically distinguishes these from veining features at other localities, as hypothesised. Overall, the data suggest the vein in the peridotite domain at Myrbærneset formed through interaction with a fluid of different composition from those at Svartberget, Årsheimneset and Remøysunde which worked to remove mobile species.

The quartz-white mica vein core at Årsheimneset has a similar composition to the leucogneisses at Svartberget but with higher MgO. It has low CaO, Na<sub>2</sub>O and FeO and high K<sub>2</sub>O and SiO<sub>2</sub>. Its composition may be similar to that which produced the laterally-zoned veins in the garnet websterite domain with it representing residual metasomatic fluid.

#### 4.6.4. Garnetite Formation

Garnetite vein cores display the greatest compositional variation both between and within localities suggesting their resultant chemistry and assemblage depends on variable factors such as the extent of metasomatic interaction and the residual composition of any remaining fluid. Components added from the metasomatic agent such as LILE, HFSE and REE are most concentrated in the garnetite vein cores suggesting their compositions are more influenced by the incoming fluid than those of the zones closer to the peridotite. The presence of Cr, Ni, Sc and Mn in garnetite vein cores from Svartberget, whilst their host gneisses contain trace amounts, and the lower concentrations of Ga whilst the host gneisses contain greater amounts demonstrate the garnetites are still geochemically influenced by the host peridotite inheriting some of its trace element signature and the garnetites, therefore, should not be classified as restites *s.s.*.

## 4.6.5. <u>Country Rocks and Fluid Production</u>

It is possible the country rocks have also undergone little bulk chemical change as they plot within an expected field on the TAS diagram (*Fig. 4.17*) and they contain similar contents of MgO, SiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub>, FeO, K<sub>2</sub>O, TiO<sub>2</sub>, P<sub>2</sub>O<sub>5</sub>, Cr, Ni and Sc as the igneous reference rocks. However, Na<sub>2</sub>O and CaO contents are low (*Figs. 4.2a,b*) suggesting lower or a loss of plagioclase components. This is supported in the QAP diagram (*Fig. 4.18*) where all gneisses from Svartberget and Årsheimneset and the quartz-white mica vein core from Årsheimneset plot with higher quartz and lower plagioclase than the reference granitic and rhyolitic rocks. The gneiss from Årsheimneset also has a low alkali feldspar content indicating it may have also lost K through partial melting. Alternatively, these depletions may just reflect the protolith bulk composition. The relative enrichment of LREE over HREE, particularly considering the amount of garnet present in vein cores, suggests HREE were retained in the fluid source either through the contemporaneous growth of garnet during fluid production or because HREE-bearing phases were not involved in fluid production. The gneisses from Årsheimneset and Svartberget both contain garnet whilst the leucogneisses do not: this distinction is reflected in their HREE (Y) contents (*Fig. 4.5b*) suggesting the leucogneisses may be compositionally similar to the metasomatic fluid.

The metasomatic fluid carries LILE, HFSE, REE, lithophile, siderophile, chalcophile, atmophile, water soluble and water insoluble elements. The particular enrichment of LILE and HFSE in garnetite vein cores suggests such elements were concentrated in the metasomatic fluid and that both types of elements were mobile indicating a supercritical liquid is responsible. Relevant experiments indicate silicate melt-aqueous fluid miscibility occurs above 5.5GPa for pelitic rocks and above 4.5GPa for granite to quartz diorite (Massonne, 2009 and references cited therein) suggesting metasomatism occurred at such conditions: P-T conditions of metasomatism are investigated in Chapter 7.

#### 4.7. Summary

All host bodies show an evolved, crustal character relative to mantle rocks with respect to Mg although the Fe, Ti and Cr contents vary: Svartberget stands out as not being an Fe-Tienriched garnet peridotite body like those at Kolmannskog and Myrbærneset as previously classified (Vrijmoed *et al.*, 2006) and perhaps represents an olivine-depleted, pyroxene-rich, Cr-rich horizon reflecting compositional heterogeneity of the parent body. Furthermore, if the sample analysed is representative of the whole body, it should not be referred to as a peridotite: its CIPW-normalised composition is that of a mela-olivine-clinopyroxene norite, not an ultramafic rock.

Garnet peridotite, websterite and eclogites domains of the bodies may have undergone little chemical change during metamorphism as their CIPW-normalised compositions yield sensible igneous protoliths on TAS and QFAP diagrams for those which may be found in large, crustal, layered, magmatic bodies through magmatic fractionation and have similar compositions to igneous reference rocks. A wide variety of protoliths for the maficultramafic bodies are indicated implying considerable compositional heterogeneity occurred in parent bodies. No garnet peridotites have CIPW-normalised ultramafic s.s. compositions demonstrating an evolved nature and they are not peridotites s.s.; or their bulk geochemistries have been affected by metamorphic processes. The Svartberget garnet 'peridotite' plots as basalt on the TAS diagram and as mela-olivine-clinopyroxene norite on the QAPF diagram. The Myrbærneset garnet 'peridotite' plots as foidite on the TAS diagram and as foid-bearing monzodiorite/monzogabbro on the QAPF diagram. The Kolmannskog garnet 'peridotite' plots as picrobasalt on the TAS diagram and as monzonite on the QAPF diagram. Garnet websterites plot as foidite, basalt and basaltic andesite on the TAS diagram and all eclogites plot as basalts. Two garnet websterites have ultramafic CIPW-normalised compositions. A metamorphic garnet websterite from Myrbærneset plots as olivine orthopyroxenite. This sample has a low CaO content indicating it formed at a plagioclaseand/or clinopyroxene-poor horizon. The second is a carbonate-rich, metamorphicmetasomatic garnet websterite from Årsheimneset which plots as orthopyroxenite; a carbonate-poor sample plots as quartz monzodiorite/monzogabbro and may represent a non-metasomatised portion of the lithology. Garnet websterites from Nybø, Kolmannskog, Eiksunddal, Remøysunde plots as mela-olivine gabbro/norites with various compositions. The Myrbærneset biotite garnet websterite plots as monzogabbro. The Båtneset garnet websterites have high Na<sub>2</sub>O, Al<sub>2</sub>O<sub>3</sub> and  $K_2O$  contents indicating an alkali feldspar-rich protolith and plot as foid-bearing monzodiorite/monzogabbro. Eclogites plot as gabbro, foid-bearing gabbro and olivine orthopyroxene meso-gabbro/clinopyroxene norite on the QAPF diagram.

Two suites are distinguished by Cr-Sc-Ni-MgO contents thought to reflect different parent magmas and intrusion events or modal proportions of major phases in which they are compatible due to different crystallisation/cumulate horizons in their large parent bodies. Textural types of eclogite or garnet websterite are not geochemically distinctive indicating textures are not dependent on rock compositions and relate to other factors such as protolith textures, deformation history and metamorphic processes.

Garnet websterite body domains within eclogite bodies are interpreted as metamorphosed (and metasomatised), more ultramafic (olivine-bearing) portions of the protolith body; not metasomatised eclogite as they have higher concentrations of compatible elements such as Cr, Ni and Sc than their associated eclogite domains which would not be introduced through metasomatism by a felsic fluid. Geochemical distinction between metamorphosed and metasomatic garnet websterite is difficult as the compositional trend of igneous
evolution and metasomatism trends towards the same end-member (felsic compositions): it still relies heavily upon field and textural relations.

The dominant, major element compositional trend across the metasomatic zones of veining features is of increasing AI and decreasing Mg gradually changing in composition away from that of the host body towards, for the most part, that of the host gneisses. Other major elements are controlled by modal mineralogy spiking in extreme assemblages such as clinopyroxenite for Ca and garnetite as garnet has strong preferences for Fe, Cr and Mn. There is an increased geochemical influence of the metasomatic agent on zones towards the vein core, particularly of LILE and HFSE, however garnetites still represent a metasomatic zone, not a restite of the metasomatic agent, and the wide variation of their compositions demonstrate that metasomatic interaction is heterogeneous and dependent on local parameters such as the degree of metasomatic interaction, the local composition of the original rock and the compositional evolution of the fluid perhaps in relation to the distance from the fluid source. Different relative chemical change characterises the Svartberget contact zones relative to the vein system indicating a different fluid-rock chemical system implying the fluid had a different composition, a different evolutionary path and/or the fluid-rock ratio differed.

Analyses of the vein zones at Svartberget, Årsheimneset and Remøysunde compared to the unaltered host rocks suggest the metasomatic agent is siliceous and particularly rich in K and LILE adding these components into mafic-ultramafic bodies. No elements are seemingly incompatible as all types of elements (LILE, REE, HFSE) are contained within the veins suggesting it is a supercritical liquid whose composition is dominantly controlled by its source. Considering its composition, the source is inferred to be the adjacent UHP country rocks and its generation associated with the breakdown of phengite considering its K-LILErich nature.

### Chapter 5 – Mass Balance Calculations

### 5.1. Introduction

This chapter presents mass balance calculations using whole rock data of samples from Svartberget to calculate a composition of the metasomatic agent added to the body which produced the laterally-zoned veins. The Svartberget locality vein system was chosen for these calculations as rock assumed to closely approximate that which is altered is available for geochemical analysis, all vein domains are clearly metasomatic in origin, the vein system is well-developed and the zoning pattern is well-structured (in parts).

Three model parameters, each with two end-members, are defined producing eight possible models of addition. The first parameter addresses the transfer of material between metasomatic zones with a 'no transfer' end-member where each zone only allows the addition of the agent without transfer of material between zones and a 'transfer' endmember where the fluid-rock system acts completely openly. The second parameter addresses the amount of the metasomatic agent added to each zone with end-members where equal amounts are added and greatly different amounts are added. The third parameter addresses the nature of the garnetite zone with it representing either a metasomatic zone with a host body component or a restite of the metasomatic agent with no host body component.

## 5.2. Aims and Objectives

The main aim of carrying out mass balance calculations is to place constraints on the chemical composition of the metasomatic agent that infiltrated the Svartberget body. A second aim is to draw conclusions on the controls of the varying zoning patterns and mineral assemblages of the veins in relation to the calculated fluid composition.

Objectives are: to determine the origin and source of the agent by comparing the calculated composition to the compositions of other fluids produced in experiments or calculated compositions from other natural systems; to determine which model of vein development described above produces the most realistic results; and to test whether the metasomatic fluid is more similar to a felsic, partial melt or the composition of the quartz-white mica vein core from Årsheimneset.

# 5.3. Data Acquisition

Mass balance calculations are based on the following simple whole rock composition calculation:

*p*.unaltered host + *a*.metasomatic agent (additive) = *z*.metasomatic zone

where p, a and z are proportion coefficients. Calculations were carried out in Microsoft Excel with a spreadsheet template designed by G. Droop (see Appendix F1 for an example). The mass balance calculation is based on the following assumptions:

- 1. the equation balances when component masses are considered
- 2. there is no material lost from the fluid-rock system
- 3. there is a volume increase
- 4. density calculations for each metasomatic zone are representative
- 5. whole rock data for each metasomatic zone are representative
- 6. starting garnet peridotite whole rock data are representative
- 7. measured zone widths are representative
- the proportion of each element derived from the additive was the same in each zone (i.e. no fractionation of fluid composition with distance travelled from the fracture wall)

The calculated additive bulk compositions are based upon the following related requirements:

- 1. the mass of the metasomatic agent added is minimised
- 2. the resulting volume increase is minimised

As neither the composition of the metasomatic agent nor the amount of it added are known, there are an infinite number of solutions possible. The resulting compositions were therefore achieved by reducing one major component to zero as the proportion of additive was decreased. Oxides with proportions of less than 1% such as  $P_2O_5$  were not considered in this due to their relatively larger margin of error and lower scale of change. Whole rock data of sample QC36G were used for veins with glimmerite zones and the data of QC36E for those without. Vein zone width measurements are taken from the respective hand samples or close-up photographs of the veins. Density calculations are based upon modal estimations and mineral chemistries.

#### 5.3.1 <u>Uncertainty</u>

There is inherent error from the above assumptions and requirements in producing a result that accurately represents nature. Firstly that no material is lost from the system: this assumption is based on the hypotheses that fluid intruded the body along hydraulic pressure gradients and therefore did not flow out of it, up-gradient, and that fluids remained static once in the body. This assumption requires that the resulting rocks represent the entire compositional system as it is impossible to account for an absent component. It is uncertain, however, that all compositional zones are represented in the veins studied: Vrijmoed *et al.* (2013) report further inner zones within the vein cores not seen in the samples studied perhaps indicating the pooling of residual fluids elsewhere, leaving the local system and potentially taking leached or exchanged components. Furthermore, there may be cryptic or modally unobvious outer zones of garnet websterite or garnet peridotite not sampled and therefore not geochemically represented.

The assumptions that density calculations, whole rock data and zone width measurements are representative are required to carry out the calculations. Error is introduced with density calculations if modal proportions and mineral geochemistries are incorrect or resulting calculations are simply inaccurate. Whole rock data may not be representative of the vein zones if the portions of the samples used incorporated neighbouring zones repeating their geochemical contribution. Furthermore, the starting peridotite composition may not reflect that which is altered if there is considerable heterogeneity of the body. Zone width measurements are carried on a two-dimensional plane (particularly those from photographs) with no consideration for what happens in the third dimension: vein zones may thicken or thin and therefore calculated volumes and masses will be inaccurate.

The assumption that all elements in the additive and peridotite are involved in equal proportions is likely to be incorrect. Element fractionation may occur during

transport of the additive components towards the metasomatic front due to different element diffusivities and mobilities producing different, effective additive compositions. This assumption is particularly crucial for the models where no transfer of material can occur between the different zones. Fractionation also occurs due to different element partition coefficients in different minerals. If there is indeed residual fluid which escapes the local system, the more incompatible components will have been lost and will not be involved in the calculation.

Finally, the reduction of one major element oxide to zero is unrealistic because, as a common rock-forming element, it is unlikely that there is none involved in the fluid-rock interaction. The resulting composition therefore represents the actual additive composition relative to the zero'd element oxide; if there is some in the additive and other components are underestimated. Alternatively, the first element oxide returned to zero is perhaps actually lost from the system and so other components are overestimated.

#### 5.4. Data Presentation

Whole rock data can be found in the Appendix [E1-E2]. Vein width measurements are presented in *Tables 5.1-5.2. Table 5.1* displays the results of mass balance calculations where no transfer of material is allowed between the zones. *Table 5.2* displays the results of mass balance calculations where transfer of material is allowed between the zones. In the transfer model there is a negligible effect on the overall additive composition by varying the proportions of additive added to each zone as it treats the entire metasomatic system as one zone. Furthermore, including the garnetite as a metasomatic zone does not affect the composition of the additive or the mass added; for this reason the average composition and mass added from various calculations are presented for this model. Including the garnetite as a metasomatic zone by ~2% compared to if it were a restite (pure additive) but this excludes the volume of the garnetite which would need to be included to determine the total volume increase of the fluid-rock system.

The calculated additive composition results for the various models are plotted on a TAS diagram (*Fig. 5.1*) with igneous reference rocks and compositions of various experimental/calculated fluids from related literature for comparison. Details of these can

Whole-Rock Data	a	36G				36E				
Garnetite		m	m	r	r	m	m	m	r	r
Additive		=	Δ	=	Δ	 =	=	Δ	=	Δ
Zone Widths	Units									
Garnetite	cm	1	1	1	1	2	2	2	2	2
Glimmerite	cm	0.7	0.7	0.7	0.7	0	0	0	0	0
Clinopyroxenite	cm	2	2	2	2	4	4	4	4	4
Websterite	cm	4	4	4	4	8	8	8	8	8
Total	cm	7.7	7.7	7.7	7.7	14	14	14	14	14
Percentage Addit	tive to Z	ones								
Garnetite	%	77.49^	77.49^	100.00	100.00	54.50	43.26^	43.26*	100.00	100.00
Glimmerite	%	77.49	69.14^	69.13	69.13^					
Clinopyroxenite	%	77.49	57.5"	69.13	57.5"	54.5"	43.26	54.5"	54.5"	54.5"
Websterite	%	77.49	46.2"	69.13	46.2"	54.50	43.26	18.28^	54.50	18.28^
Average Composition of Additive										
SiO <sub>2</sub>	%	51.54	51.48	51.53	51.49	54.15	54.79	55.82	53.86	54.92
TiO <sub>2</sub>	%	0.64	0.82	0.66	0.79	0.40	0.46	0.55	0.37	0.47
$Al_2O_3$	%	11.14	13.69	11.47	13.18	11.21	12.91	15.67	10.44	13.25
FeO	%	7.65	6.96	7.56	7.10	8.47	8.23	7.85	8.58	8.18
MnO	%	0.20	0.18	0.20	0.19	0.34	0.37	0.41	0.33	0.37
MgO	%	15.74	13.37	15.43	13.85	12.17	9.68	5.62	13.30	9.17
CaO	%	6.84	5.36	6.65	5.66	7.58	6.80	5.52	7.94	6.64
Na <sub>2</sub> O	%	1.70	2.27	1.78	2.16	2.57	3.17	4.15	2.30	3.29
K <sub>2</sub> O	%	3.14	4.32	3.30	4.09	1.56	1.92	2.52	1.39	2.00
$P_2O_5$	%	0.01	0.01	0.01	0.01	1.62	2.03	2.71	1.43	2.12
H <sub>2</sub> O	%	0.15	0.16	0.15	0.16	-0.07	-0.12	-0.20	-0.05	-0.13
CO <sub>2</sub>	%	1.23	1.35	1.25	1.33	0.00	-0.24	-0.64	0.11	-0.29
SO <sub>3</sub>	%	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Total	%	100.0	100.0	100.0	100.0	 100.0	100.0	100.0	100.0	100.0
Mass Added	g/dm²	1952.8	1400.3	1857.2	1484.2	1262.8	1002.3	750.0	1432.5	961.6
Volume Increase	%	358.7	129.6	289.8	147.6	123.4	78.5	49.0	164.6	71.0

Table 5.1: Results of mass balance calculations for model where no transfer is permitted between zones.

= - equal percentage of additive added to each zone  $\Delta$  - different percentage of additive added to each zone

m - metasomatic garnetite

r - restitic garnetite

^ - zero CaO added

\* - zero MgO added

' - zero MnO added

Zone densities: garnetite = 3.73g/cm<sup>3</sup>; glimmerite = 2.9g/cm<sup>3</sup>; biotite clinopyroxenite = 3.22g/cm<sup>3</sup>; biotite websterite = 3.25g/cm<sup>3</sup>; garnet peridotite = 3.41g/cm<sup>3</sup>.

Detween zones.								
Whole-rock Data	36G	36G	36G		36E	36E	36E	
Width Source (sample/photo)	)	36G	4.7-1	4.7-5		36E	4.7-2	4.7-4
Zone Widths								
Garnetite	cm	1	2	3.1		2	1.2	1.7
Glimmerite	cm	0.7	1.2	0.2		0	0	0
Clinopyroxenite	cm	2	6.1	1.6		4	6.2	4
Websterite	cm	4	5.7	6.4		8	7.8	6
Total	cm	7.7	15	11.3		14	15.2	11.7
Proportion of Vein								
Garnetite	%	13.0	13.3	27.4		14.3	7.9	14.5
Glimmerite	%	9.1	8.0	1.8		0.0	0.0	0.0
Clinopyroxenite	%	26.0	40.7	14.2		28.6	40.8	34.2
Websterite	%	51.9	38.0	56.6		57.1	51.3	51.3
Clinopyroxenite/Websterite		0.50	1.07	0.25		0.50	0.79	0.67
Garnetite/Clinopyroxenite		0.50	0.33	1.94		0.50	0.19	0.43
Percentage Additive to Each Z	lone							
with metasomatic garnetite	%	27.44	23.96	35.78		24.00	23.67	24.78
with restitic garnetite	%	14.83	10.34	7.79		9.41	16.12	10.05
Average Composition of Addit	tive							
SiO <sub>2</sub>	%	51.25	50.51	48.33		57.26	59.33	56.56
TiO <sub>2</sub>	%	1.50	1.79	1.24		0.69	0.65	0.72
Al <sub>2</sub> O <sub>3</sub>	%	22.94	26.67	21.39		19.51	17.61	19.81
FeO	%	4.49	2.98	8.59		7.32	3.90	6.66
MnO	%	0.12	0.09	0.35		0.47	0.52	0.42
MgO	%	4.79	0.66	8.52		0.00	0.00	0.00
CaO	%	0.00	0.00	0.00		3.75	5.57	4.43
Na <sub>2</sub> O	%	4.31	5.06	3.13		5.49	6.40	5.45
K <sub>2</sub> O	%	8.61	10.19	6.72		3.35	3.97	3.50
$P_2O_5$	%	-0.01	-0.02	0.01		3.64	3.93	3.62
H <sub>2</sub> O	%	0.21	0.21	0.18		-0.31	-0.32	-0.30
CO <sub>2</sub>	%	1.79	1.85	1.54		-1.18	-0.66	-0.88
SO <sub>3</sub>	%	0.01	0.01	0.01		0.02	0.01	0.02
Total	%	100.00	100.00	100.00		100.00	100.00	100.00
Mass Added	g/dm <sup>2</sup>	691.4	1176.1	1362.9	• •	1112.3	884.8	959.9
Volume Increase (with metasomatic garnet)	%	39.2	32.6	55.4		32.2	32.1	33.7
Volume Increase (with	%	37.4	30.6	52.2		30.3	31.0	31.7

Table 5.2: Average results of mass balance calculations for model where element transfer is permitted between zones.

restitic garnet) % 57.4 50.6 52.2 50.5 51.0 51 Zone densities: garnetite = 3.73g/cm<sup>3</sup>; glimmerite = 2.9g/cm<sup>3</sup>; biotite clinopyroxenite = 3.22g/cm<sup>3</sup>; biotite websterite = 3.25g/cm<sup>3</sup>; garnet peridotite = 3.41g/cm<sup>3</sup>. be found in the figure notes (*Fig. 5.1*). The inner vein pegmatite of Vrijmoed *et al.* (2013) is also plotted as it possibly represents the residual fluid left following fluid-rock interaction between the Svartberget body and the metasomatic agent. CIPW normative compositions of the calculated additive, reference rocks and experimental/calculated fluid compositions have also been calculated and are presented in the Appendix [F2-F3]. The results are presented on normative compositional QAP and FAP diagrams (*Figs. 5.2-5.3*).

#### 5.5. Data Analysis

*Tables 5.1* and *5.2* show that the main differences in the overall additive composition between the two models are higher Al<sub>2</sub>O<sub>3</sub>, Na<sub>2</sub>O and K<sub>2</sub>O and lower MgO and CaO in the 'transfer' model. Furthermore, there is a considerably lower mass addition and volume increase in the 'transfer' model. Considering the field relationships, the high volume increases produced in the 'no-transfer' model calculations are probably unrealistic. The assumption that material cannot be transferred between the zones was made to provide an end-member to the model: fluid components would have had to reach the outermost websterite zone from the fracture wall via the other zones so transfer between zones must have existed. In the 'transfer' model the oxide calculated to be zero, as per the calculation requirements, changes between the whole rock data used or the presence of a glimmerite selvage. CaO is zero for 36G-glimmerite selvage calculations and MgO is zero for 36E-no glimmerite selvage calculations. Both CaO and MgO showed the greatest, pivotal change as the amounts of additive were amended. If the CaO and MgO contents of the additive are close to zero the relative concentrations of the other components should not be severely affected.

Including trace elements in the calculations doesn't vary the concentrations of the major elements greatly. Elements with the highest concentrations in the calculated additive are Zr, Ba, Rb, Cr, Ni and Sr. The inclusion of Cr and Ni in this list is unexpected but probably due to high Cr and Ni contents in the clinopyroxenite and websterite selvages. The higher concentrations of Zr, Ba and Rb reflect both HFSE and LILE were transported in the incoming fluid.

The TAS diagram (*Fig.5.1*) shows the 'no-transfer' model results (square markers) are well clustered, the additive has a consistently lower proportion of alkalis than the 'transfer' model results (diamond markers) and they plot in the trachybasalt, basaltic trachybasalt,



Fig. 5.1: TAS diagram after le Bas et al. (1986) and Middlemost (1994) displaying compositions for mass balance calculated fluids, compositions of experimental and calculated fluids and reference rocks. Diamonds mark additive compositions of the 'transfer' model. Squares mark additive compositions of 'no transfer' model. Triangles mark CIPW normative compositions of fluid compositions calculated or produced in experiments. Marker descriptions detailed in the table below. Reference materials listed in Table 4.1. Composition fields are listed in Table 4.2. Line intersections taken from Gillespie and Styles (1999).

Shape	Colour	Model	Shape	Colour	Source
diamond	red	36G		red (no	Sekine and Wyllie (1983); Table 3, Run #10, melt
diamond	black	36G/4.7-1	triangle	fill)	produced by reaction between granite-peridotite-H <sub>2</sub> O at 3.0GPa, 850°C
diamond	purple	36G/4.7-5			Hermann and Green (2001), Table 3, C-879,
diamond	blue	36E/4.7-4	triangle	green	composition reconstructed from multiphase inclusions in eclogite
diamond	green	36E	triangle	yellow	Hwang <i>et al.</i> (2001), section 3, composition reconstructed from multiphase inclusions in garnet in quartzofeldspathic rocks
diamond	orange	36E/4.7-2	triangle	blue	Wunder and Melzer (2003), Table 1, experimental melt composition in the KCMASH system at 3.5GPa, 850°C
square	purple	36G/m/=	triangle	light blue	Liu <i>et al.</i> (2009), Table 6, QP15, initial partial melt of phengite eclogite, 2.4GPa, 850°C
square square	red black	36G/m/∆ 36G/r/=	triangle	purple	Rapp <i>et al.</i> (2010), Table 3, CF-HP, melt produced in TTG-peridotite reaction experiment at 3.0GPa, 1100°C
square	light green	36G/r/∆	triangle	red	Zheng <i>et al.</i> (2011), Table 3, Melt-A, partial melt of
square	green	36E/m/=			
square	blue	36E/m/=	triangle	orange	Zheng <i>et al.</i> (2011), Table 3, Melt-A, partial melt of biotite gneiss at 950°C
square	light blue	36E/m/Δ	triangle	black	Gao <i>et al.</i> (2013); Table 7, mean multiphase inclusion in phengite eclogite
square	brown	36E/r/=	triangle	pink	Vrijmoed <i>et al.</i> (2013); Suppl_Table_5, Zone 8, SV07-1, pegmatite in centre of veins
square	orange	36E/r/∆	triangle	<b>black</b> (no fill)	Wang and Tang (2013); Table 3, 3.0GPa, 1250°C, spot 7, melt formed between olivine and tonalite reaction

basalt and basaltic andesite fields between the oversaturated acidic (14,16), intermediate alkaline (7,10,12) and basic (1-5) reference rocks. The most similar experimental fluid composition is found in an experimental melt composition in the KCMASH system formed at 3.5GPa, 850°C by Wunder and Melzer (2003) (blue triangle). The 'transfer' model results (diamond markers) are less well clustered and extend to very high alkali contents plotting in the trachy-andesite, trachyte (normative quartz <20%), phonotephrite, tephriphonolite and phonolite fields between the saturated sialic (22,23), undersaturated basic (34) and undersaturated sialic (35-38) reference rocks. The additives calculated for veins with glimmerite selvages have lower silica contents like basaltic magmas. The most similar experimental fluid compositions are found in a melt produced in a tonalite-trondhjemitegranodiorite (TTG)-peridotite reaction experiment at 3GPa, 1100°C by Rapp *et al.* (2010) (purple triangle). Other fluid compositions from multiphase inclusion reconstructions and partial melts follow a more typical tholeiitic-felsic compositional trend. The Vrijmoed *et al.* (2013) vein pegmatite plots close to the tonalite (21) and adakite (A) reference rocks suggesting residual metasomatic fluid evolves towards an andesitic composition.

In the QAP and FAP diagrams (Figs. 5.2-5.3) the 'transfer' model additives (diamond markers) from glimmerite-bearing veins plot in the syenite (7) and foid-bearing syenite (7') compositional fields away from any reference rock and those from the glimmerite-absent veins plot in the monzodiorite/monzogabbro (9) and foid-bearing monzodiorite/monzogabbro (9') fields between intermediate alkaline reference rocks (8,9,11). The glimmerite-absent (QC36E), transfer model calculations (orange, green and blue diamonds) plot off the quartz-saturation line suggesting the silica content of the chemical system may control the formation of biotite: the QC36E vein is more clinopyroxene-rich and biotite-poor than QC36G. Most of the 'no transfer' model additives (square markers) plot along the quartz saturation line in the monzodiorite/monzogabbro (9) and monzonite (8) fields although one has normative quartz in the monzonite field plotting close to the trachyandesite reference rock (11). The glimmerite-present (QC36G), 'transfer' model calculations (purple and red diamonds) plot near to where the biotite websterite, glimmerite selvage and a garnetite vein core samples plot (see Fig. 4.18) demonstrating they have similar CIPW normative compositions suggesting a greater influence over the composition of the vein cores.

CIPW normative compositions of most experimental/calculated fluid compositions plot in the granitic (1b-3b) and granodiorite (4) fields with most extending the reference rock trend



Figure 5.2: QAP ternary diagram displaying CIPW normative compositions of fluid compositions calculated or produced in experiments. Q = quartz, A = alkali feldspar, P = plagioclase feldspar. Reference materials listed in Table 4.1. Composition fields are listed in Table 4.3. Marker formats as described in Fig. 5.1.



Figure 5.3: FAP ternary diagram displaying CIPW normative compositions of fluid compositions calculated or produced in experiments. F = feldspathoid, A = alkali feldspar, P = plagioclase feldspar. Reference materials listed in Table 4.1. Composition fields are listed in Table 4.3. Grey triangle – Frezzotti et al. (2007), average bulk compositions of inclusions. Black diamond – 36G/4.7-1 model.

and trending towards normative compositions of the felsic gneisses and quartz-rich veins but away from the calculated fluid compositions. Three plot with more plagioclase-rich normative compositions: the Rapp *et al.* (2010) TTG-peridotite melt which plots in the monzodiorite/monzogabbro (9) field, the Sekine and Wyllie (1983) melt produced by reaction between granite-peridotite-H<sub>2</sub>O which plots in the quartz monzodiorite/monzogabbro field (9\*) and the Vrijmoed *et al.* (2013) vein pegmatite composition which plots in the quartz diorite/gabbro/anorthosite (10\*) field demonstrating a reduction in alkalis and silica in the residual fluid.

### 5.6. Data Interpretation

Assuming the model with the smallest volume increase and mass added most closely approximates nature, the mass balance calculations indicate that a model where material is able to be transferred between the zones is the most realistic. Within this model, to determine the overall additive composition and mass, the distinction between a metasomatic or restitic garnetite and the proportions of additive added to each zone do not need to be made or known as they do not affect the outcome.

#### 5.6.1. Additive Composition

The additive compositions do vary between vein types (glimmerite-bearing or not) and between width measurements suggesting either the above assumptions do not hold and/or the additive varied locally in composition. Generally, the additive is a Si-Al-K-rich, Fe-Mg-Ca-poor with 48-60% SiO<sub>2</sub>, 17-27% Al<sub>2</sub>O<sub>3</sub>, 3-11% K<sub>2</sub>O, <10% MgO, CaO and FeO, 3-6% Na<sub>2</sub>O, <4% P<sub>2</sub>O<sub>5</sub> and <1% TiO<sub>2</sub> and MnO. This may not accurately represent the actual composition of the intruding fluid if element fractionation occurred, non-reactive components were present or residual fluids accumulated elsewhere, but it reflects the components added to the rock. Its overall character is undersaturated-saturated sialic, syenitic and not the felsic composition of the quartz-white mica vein core from Årsheimneset hypothesised. It contains similar components to the composition thought to characterise melts of felsic gneisses at high pressures being high in Si, K, LILE and LREE and low in Fe, Mg and Ca (and Cr and Ni) (Zheng *et al.*, 2009 and references cited therein) but these are more silica-rich and less alkaline than the composition calculated.

The QAP-FAP diagrams show the additives have silica undersaturated to silicaoversaturated normative compositions dominated by feldspars. The normative compositions of the 'transfer' model additives clearly do not have the similar quartzundersaturated nature as the reference rocks they plot close to on the TAS diagram. This suggests that whilst the additives may have low silica contents (maficintermediate compositions) they are not undersaturated with respect to quartz.

#### 5.6.2. Source of the Metasomatic Agent

According to the comparisons with experimental/calculated fluid compositions and reference rocks on the TAS diagram, the additive is not similar in composition to partial melts of felsic, mafic or ultramafic rocks or reconstructed compositions of multiphase inclusions. The composition is most similar to that of a melt produced through the reaction between peridotite and a TTG-like melt, similar to that of Rapp et al. (2010), having a composition between trachyandesite-syenite-tephritephonolite. Furthermore the QAPF plots suggest the fluid has a composition most similar to a fluid produced in a hydrated reaction between peridotite and granite, similar to that of Sekine and Wyllie (1983) having a less silica-rich, less alkalic, monzodioritic/monzogabbroic or trachyandesite-monzonite composition. This suggests the origin of the alkali-rich fluids is through the reaction between a felsic partial melt and the garnet peridotite body itself. Sanukitoid granitoids magmas, which are characterised by high Mg# (>0.6), high LILE and LREE and low HREE, are thought to have formed from the reaction of TTG liquids (derived from melting crustal rocks) and mantle peridotite and are in equilibrium with orthopyroxeneenriched peridotites/garnet websterites (Rapp et al., 2010). The metasomatised rock may therefore represent a similar system as the source region of such granitoids.

The metre-thick alteration zone at the body contacts may represent products of such a reaction. Fluid pressures at the contact may have induced hydrofracture of the body and influx of the hybrid fluid. The mineralogical variation between the contact's metasomatic zones (garnet-poor) and the zoned veins (garnet-rich) may reflect the contrasting composition between the infiltrated fluid and a partial melt of the country rocks.

## 5.6.3. Glimmerite Formation

The occurrence of glimmerite selvages is a localised event with the selvage structure and modal abundance of biotite varying on a dm-scale. This suggests the local geochemical environment has a dominant control on the structure and mineralogy of the vein. The cause of this variation may be due to protolith heterogeneity, change in the local fluid/rock ratio, compositional evolution of the fluid with interaction or change of more mechanical parameters such as fluid flow or fracture density.

Additive compositions calculated from veins without glimmerite selvages have Na<sub>2</sub>O>K<sub>2</sub>O and those with glimmerite selvages have K<sub>2</sub>O>Na<sub>2</sub>O. For veins with glimmerite selvages, the calculated additive has lower SiO<sub>2</sub> and Na<sub>2</sub>O and higher Al<sub>2</sub>O<sub>3</sub> and K<sub>2</sub>O than those without and CaO is the limiting component (reduced to zero) whereas MgO is the limiting component for veins without glimmerite selvages. These differences suggest the formation of glimmerite selvages is dependent on the relative major element composition and ratios such as Si/K, Si/Al, K/Na and Ca/Mg with glimmerite selvages forming where Ca, Na and Si contents are lower so biotite is formed over clinopyroxene: as seen, biotite-poor garnetite vein cores are typically more clinopyroxene-rich. The stability field of phlogopite is known to expand with decreasing SiO<sub>2</sub> in the KMASH granite-peridotite system (Sekine and Wyllie, 1982, 1983).

The formation of the biotite websterite at the Svartberget contact rather than a garnet websterite might also be due to major element ratios. Phlogopite is seen to form in reactions between alkali-rich fluid from phengite breakdown and peridotite (Marocchi *et al.*, 2010) and biotite websterite assemblages are known to form at high P-T conditions from Si-Al hydrous fluids (Becker *et al.*, 1999b and references cited therein).

### 5.7. Summary

The mass balance calculations suggest that the metasomatic agent which infiltrated the Svartberget body had an intermediate, highly alkaline-syenitic composition and not a felsic, granitic composition suggesting that other geological processes occurred to produce a fluid with a more unusual composition. Comparison to experimentally-determined and natural fluid compositions suggests the fluid composition is a hybrid fluid produced through the reaction between a felsic partial melt and peridotite such as what may have occurred at the contact of the Svartberget body with a partial melt of the surrounding gneisses at UHP conditions. Generally, the additive is a Si-Al-K-rich, Fe-Mg-Ca-poor with 48-60% SiO<sub>2</sub>, 17-27% Al<sub>2</sub>O<sub>3</sub>, 3-11% K<sub>2</sub>O, <10% MgO, CaO and FeO, 3-6% Na<sub>2</sub>O, <4% P<sub>2</sub>O<sub>5</sub> and <1% TiO<sub>2</sub> and MnO. Different additive compositions for different zoning patterns suggest the local fluid-rock system varies along the length of the vein network and is governed by local fluid-rock parameters such as Si/K and Si/Al ratios. The calculations also suggest that the volume increase associated with metasomatism of the Svartberget body is at least one third.

### **Chapter 6 - Mineral Chemistry**

### 6.1. Introduction

Major element compositions of garnet, clinopyroxene, orthopyroxene, biotite, amphibole, olivine and carbonate phases in samples of the mafic-ultramafic bodies and veining systems from Nybø, Grytting, Årsheimneset, Remøysunde, Kolmannskog, Myrbærneset and Svartberget were analysed to obtain mineral chemistries. Samples from Eiksunddal and Båtneset were not analysed due to sample and thin section suitability and restrictions of the project. Chemical data were obtained using electron microprobe analysis (EMPA).

EMPA was also used to obtain major and trace element compositions of accessory phases (zircons, monazites and rutiles) which were to be radiometrically dated; these data are presented and analysed in Chapter 8.

# 6.2. Aims and Objectives

The main aims of obtaining mineral chemistries are to quantitatively identify and nominally classify the major phases in the rocks, identify chemical trends and variability between and within rock types, obtain data for thermobarometric calculations (see Chapter 7) and classify clinopyroxene compositions to enable the classification of orthopyroxene-bearing rocks as garnet websterites or orthopyroxene eclogites. Orthopyroxene eclogites contain omphacitic clinopyroxenes with Na/(Ca+Na) values of 0.20-0.80 (Deer *et al.*, 1992): the addition of orthopyroxene to the eclogite assemblage should not affect the Na/(Ca+Na) ratio of clinopyroxene as neither enter the orthopyroxene structure in any great concentration.

Objectives were to: identify major element chemical zoning within crystals, particularly garnets, to help unravel the geological evolution of the rock and chemical system; compare the compositions of matrix and included phases to investigate the scale of chemical homogeneity and whether included phases retain compositions reflecting prograde or peak pressure conditions and therefore the prograde evolution of the rocks; define the limits of chemical variability of each rock type and distinguish between chemical variation due to metasomatism as the bulk composition of the fluid-rock system changes and retrogressive

exchange, such as Fe-Mg and Tschermakitic [M<sup>vi</sup>-1Si<sup>iv</sup>-1-Al<sup>vi</sup>Al<sup>iv</sup>] substitutions; and finally to quantitatively determine the nature of unidentified, microscopic phases.

## 6.3. Data Acquisition

EMPA was carried out on a multi-detector, wavelength-dispersive CAMECA SX100 electron microprobe (EMP) at the SEAES, University of Manchester and operated with an accelerating voltage of 15kV, beam current of 20nA and 2µm-diameter beam for most analyses: a wider beam of 20µm-diameter was used to analyse some exsolved phases such as clinopyroxene and amphibole. Raw data were corrected using the method of Pouchou and Pichoir (1991). Standards used to calibrate the electron microprobe for different elements are presented in the full methodology in the Appendix [B4]. The SEM was used, as in Chapter 3, to image the polished sections for purposes of crystal selection, location and recording EMP analysis sites: see the Appendix [B2] for full details.

Mössbauer spectroscopy was used to determine the Fe<sup>3+</sup> contents of a representative suite of clinopyroxenes: the content of ferric iron in clinopyroxene is particularly important for Fe-Mg thermometry as explained in the following chapter but the EMP is unable to distinguish between Fe<sup>2+</sup> and Fe<sup>3+</sup>. Clinopyroxenes from a range of rock types from Svartberget and Årsheimneset were selected for analysis. Crystal separates were handpicked and ground using a pestle and mortar to produce a powder. Mössbauer spectroscopy was carried out in the Mineral Analysis Facility at the Williamson Research Centre, SEAES, University of Manchester using a FAST ComTec Instrument with a <sup>57</sup>Co source. Samples were analysed at room temperature and in aerobic conditions for up to seven days until sufficient data points were obtained. The spectrometer and analysed samples were calibrated against 25µm-thick, pure Fe foil prior to sample analysis. Raw data were analysed using a FAST ComTec MCDWIN-MCA3 software programme. See the Appendix [B5] for a full methodology.

Data were prepared, analysed and plotted in graphs using Microsoft Office Excel 2007. Ternary plots were constructed using the TRI-PLOT programme by Graham and Midgley (2000) in Microsoft Office Excel 2007.

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#### 6.3.1. <u>Uncertainty</u>

To have confidence in the reliability of the data, each EMP element analysis was checked against its individual detection limit: average values are presented in the Appendix [G1]. To reduce further uncertainty associated with accuracy of the data, analyses with poor totals were discarded: analyses of anhydrous phases selected for presentation and further use have corrected totals of 99.00-100.99%, of 96.00-97.99% for amphiboles, of 94.00-95.99% for biotites and a wider range of 50.00-59.99% for carbonates. Due to these measures, different samples have different numbers of useable data points; therefore the full compositional range of each sample and the across-crystal chemical variation may not be accurately represented.

The data were corrected for ferric iron using various methods: clinopyroxene Fe<sup>3+</sup> was estimated using the results of the Mössbauer analyses; Fe<sup>3+</sup> in garnet and orthopyroxene was calculated using the charge balance method of Droop (Droop, 1987); all iron in biotite, olivine and carbonate is assumed to be ferrous (Fe<sub>2</sub>=Fe<sup>2+</sup>); and Fe<sup>3+</sup> in amphibole was calculated using the method of Holland and Blundy (1994) as other schemes did not produce useable data. The assumption for biotite that all iron is ferrous is based upon the knowledge that Fe<sup>3+</sup> in biotites increases with overall Fe (Deer *et al.*, 1992) and so biotites which have little total iron, as will be shown, have little ferric iron. Furthermore, biotites can contain vacant cation sites so calculating ferric iron contents using charge balance is inappropriate (e.g. Droop, 1987). The assumption of ferrous iron in olivine and carbonate is due to the inability of their structures to host Fe<sup>3+</sup>.

Correcting data for ferric iron using the charge balance method puts all analytical error onto the value of Fe<sup>3+</sup> as well as its actual value. The accuracy of concentrations of dominant elements like Si is therefore paramount as these have larger absolute errors if not relative errors. For this reason Si and Al analyses were counted for double the time. A further complication is the introduction of structural vacancies through substitutions such as the CaEs component in pyroxenes which affects the correctness of stoichiometric normalisation to a certain number of cations. It is therefore assumed in the normalisation of anhydrous mineral compositions that none host structural vacancies: this assumption may not hold well for UHP pyroxenes

as the CaEs component is thought to increase with increasing pressure (Patiño Douce, 2005; Smit *et al.*, 2008).

Fixing the Fe<sup>2+</sup>/Fe<sup>3+</sup> ratio to a constant value in clinopyroxenes following Mössbauer analysis means natural variation of the ratio is disregarded and any error of the Si concentration and/or the ratio is propagated through onto the Fe<sup>2+</sup> composition: a correlation can be seen on Fe-Si and crystal profile plots. The Si content therefore must be considered when analysing zoning profiles and Mg#. Mössbauer Fe<sup>3+</sup> measurements have been extrapolated to other clinopyroxene crystals from other samples and localities which may not be accurate but use of the charge balance method yields greater error.

Error in the accuracy of compositional data also results from the exsolution of peak phases, sub-microscopic included phases and sub-microscopic retrogressive alteration. Effort was made to select the freshest samples and crystals and some analyses were carried out with a 20µm beam diameter to compare to typical 2µm beam analyses. No consistent difference was seen: the main grouping of analyses was the same and the compositional range seen in the 20µm-diameter analyses was similar (and typically less) than that of the 2µm-diameter analyses.

# 6.4. Data Presentation

Results of the Mössbauer analysis and the proportions of ferric iron assumed for clinopyroxene in each sample are presented in *Table 6.1*: Mössbauer spectroscopy data are presented in the Appendix [H]. All corrected EMP data and stoichiometric calculations for each phase are presented in data tables in the Appendix [G2-G8] whilst representative analyses are presented below in *Table 6.2*. Average one sigma errors for each element in each phase are presented in the Appendix [G9].

All useable data are presented on classification and compositional graphs below. To aid reading, graphs with data from each locality are presented in the locality overviews in the Appendix [A1-9]. Abbreviations of phase components are explained in *Table 0.11b* and a key to the marker formats used in the graphs is presented in *Table 0.13* in the preliminary pages. The 'mid zone' refers to the middle zone between the core and outer rim of crystals

		Chargo Balanco	Mössbauer	Value Used in	
Sample	Rock Type†	Charge Balance	Spectroscopy	Calculations	
		average (%)	(%)	(%)	
A347	websterite (eclogite)	54 (37-78)		29	
Δ1308	websterite (eclogite)	63 (49-78)		29	
Ф16	websterite (eclogite)	37 (18-61)		29	
QC15B	eclogite	54 (22-86)		41	
QC15F	eclogite	63 (38-88)		41	
QC15J	websterite (eclogite)	55 (35-72)		29	
QC15N	websterite (eclogite)	72 (51-97)		29	
QC26E	eclogite	46 (28-85)	41	41	
QC26G	websterite (eclogite)	37 (23-49)	29	29	
QC26H	clinopyroxenite	61 (42-77)		29	
QC26H	websterite (eclogite)	63 (47-78)		29	
QC29A	websterite (eclogite)	42 (22-72)		29	
QC29L	eclogite	49 (35-64)		41	
QC32A	websterite (peridotite)	41 (20-60)		32	
QC34A	eclogite	73 (0-112)^		41	
QC34B	peridotite	47 (43-51)		23	
QC34E	websterite (peridotite)	52 (36-61)		23	
OC34F	clinopyroxenite selvage	38 (20-58)		23	
QCJH	(peridotite)	30 (20 30)		25	
QC36A	peridotite	94 (61-137)^	23	23	
QC36D	websterite (peridotite)	46	32	32	
QC36G	garnetite (peridotite)	57 (29-99)		35	
QC36G	clinopyroxenite (peridotite)	52 (9-111)^		35	
QC36G	websterite selvage (peridotite)	54 (14-91)	35	35	
SV1A	websterite (peridotite)	41 (16-80)		32	

Table 6.1: Ferric iron estimations and values used in clinopyroxene stoichiometry calculations

<sup>+</sup> rock in parentheses is the host body

\* numbers in parentheses are the range of calculated Fe<sup>3+</sup> using the charge balance method of Droop (1987) ^ - negative Fe<sup>2+</sup> calculated

Minoral*	art	art	CDY	CDY	0.01	0.01	ht	ht	2000	2000	al	crb	crb
wineral	gri	gr l Årshoi	Cpx	Cuart	Upx	Opx	DL Årshoi	DL	Årshoi	Bomdu	01 Swart	Crb	Cru
Locality	Nybø	Arshei-	Remøy-	SVdrt-	KOIMan-	SVdrt-	Arshei-	SVdrt-	Arshel-	Remøy-	Svart-	Nybø	Svart-
De als True a A		mneset	sunde	berget	пѕкод	berget	mneset	berget	mneset	sunde	berget		berget
коск Турел	gw	gw	be	ogw	gw	ogw	gvc	gw	De	gw	ogw	gw	ogw
Sample	D1308	QC26G	QC29L	QC36A	QC32A	QC36A	QC26HA	SVIA	QC26EB	QC29A	QC36A	QC15J	QC36A
Point	B2-26	B17-8	B5-C2	B10-9	B10-6	B4-2	B14-2	B6-5	B7-1	B1-1	<u> </u>	B5-1	B10-1
EMP data (wt%	)												
SiO <sub>2</sub>	39.62	39.52	52.81	53.31	54.46	55.10	40.28	38.52	42.65	55.11	38.87	-	-
TiO <sub>2</sub>	0.07	-	0.20	0.10	-	0.08	1.02	0.50	0.11	0.12	-	-	-
$AI_2O_3$	23.03	22.25	8.49	1.08	0.64	1.49	14.95	17.96	15.93	2.48	-	-	-
$Cr_2O_3$	0.02	0.06	0.14	0.12	0.02	0.04	0.05	0.48	0.07	0.14	0.014	-	-
Fe <sub>2</sub> O <sub>3</sub>	1.75	0.51	2.85	0.81	2.44	2.62	-	-	4.14	4.50			
FeO	15.95	21.64	3.69	2.45	12.82	7.78	10.11	7.27	8.65	0.96	15.02	9.84	1.77
MnO	0.30	0.58	0.05	0.07	0.21	0.24	0.04	-	0.07	0.04	0.30	0.05	0.19
MgO	13.95	11.02	10.44	17.12	28.92	32.31	20.13	20.60	12.17	21.86	44.98	42.94	22.53
CaO	4.93	4.21	16.67	24.20	0.28	0.27	0.09	-	9.98	10.89	-	0.35	30.12
Na₂O			4.34	0.39	0.04	-	0.79	0.81	3.39	1.03	-	-	-
K <sub>2</sub> O			0.02	-	-	-	7.55	8.27	0.05	0.49	-	-	-
Total	99.61	99.80	99.70	99.66	99.82	99.92	95.02	94.42	97.21	97.61	99.18	53.18	54.60
Stoichiometry (	pfu)												
Si	2.940	2.991	1.917	1.944	1.955	1.932	2.900	2.770	6.215	7.616	0.986		
Ті	0.004	-	0.005	0.003	-	0.002	0.055	0.027	0.012	0.012	-		
Al	2.014	1.985	0.363	0.047	0.027	0.061	1.269	1.522	2.735	0.404	-		
Cr	0.001	0.004	0.004	0.003	0.000	0.001	0.003	0.028	0.008	0.015	0.0003		
Fe <sup>3+</sup>	0.097	0.029	0.078	0.022	0.066	0.069	-	-	0.454	0.467			
Fe <sup>2+</sup>	0.990	1.370	0.112	0.075	0.385	0.228	0.609	0.437	1.054	0.111	0.319	0.113	0.044
Mn	0.019	0.037	0.002	0.002	0.006	0.007	0.003	-	0.009	0.004	0.0063	0.001	0.005
Mg	1.543	1.244	0.565	0.931	1.548	1.689	2.161	2.209	2.644	4.504	1.702	0.881	0.995
Ca	0.392	0.341	0.648	0.946	0.011	0.010	0.007	-	1.559	1.613	-	0.005	0.956
Na			0.305	0.028	0.003	-	0.111	0.113	0.959	0.275	-		
К			0.001	-	-	-	0.693	0.758	0.010	0.086	-		
Total	8.000	8.000	4.000	4.000	4.000	4.000	7.811	7.864	15.658	15.109	3.013	1.000	2.000
Mg#	0.609	0.476	0.834	0.926	0.801	0.881	0.780	0.835	0.715	0.976	0.842	0.886	0.958

Table 6.2: Representative compositional data.

\* phases as listed in Table 0.11b

^ be = bimineralic eclogite; qe = quartz eclogite; gw = garnet websterite; ogw = olivine garnet websterite; ws = websterite selvage; gvc = garnetite vein core; cs =

clinopyroxenite selvage

and is distinguished by a change in inclusion density or distance from the crystal boundary (typically >0.5mm).

Garnet end-member compositions are displayed on an alm-py-grs ternary diagram in *Figure 6.1* and representative garnet zoning profiles are displayed in *Figure 6.2*. *Figures 6.3-6.5* display various classification and compositional diagrams for pyroxenes and representative zoning profiles for clinopyroxene and orthopyroxene are presented in *Figures 6.6-6.7*. Orthopyroxene compositions are presented with clinopyroxenes in *Figure 6.4*. Compositional plots of biotite crystals are presented in *Figure 6.8*. Compositions of amphiboles are presented in the classification diagrams in *Figures 6.9-6.10*. Olivine compositions are presented in *Figure 6.11*. Olivine data from Myrbærneset are not suitable for use. Finally, compositions of carbonate phases are presented in *Figure 6.12* on a mgs-cal-sd ternary diagram.

### 6.5. Data Analysis

# 6.5.1. <u>Garnet</u>

The compositional range of garnet is  $Prp_{30.0-64.6}Alm_{22.3-56.4}Grs_{8.7-23.9}$  with the pyrope  $[Mg_3Al_2Si_3O_{12}]$ , almandine  $[Fe_3Al_2Si_3O_{12}]$  and grossular  $[Ca_3Al_2Si_3O_{12}]$  components forming 90-99% of the garnet composition. The dominant chemical variation is in almandine-pyrope contents both within and between rock types. There is little extension towards the grossular apex in websterite or vein garnets: most data lie along the 10% grossular line with a ±5% scatter. However, garnets in eclogites also vary in their grossular content such as those from Remøysunde and Myrbærneset, which have higher grossular contents and plot away from the main trend.

Garnets in the peridotite from Svartberget have the most pyrope-rich compositions of those analysed and those in eclogite from Årsheimneset are the most almandinerich whilst garnets in the eclogite from Myrbærneset are the most grossular-rich. The most pyrope-rich websterite garnets are in the sample from Grytting and the most almandine-rich websterite garnets are from Årsheimneset: both coarse body domains. The greatest Fe-Mg variation between rocks at a locality is between those from Svartberget and the least at Nybø. Those from Nybø also show the least variation in Ca between websterite and eclogite and garnet in a garnetite band from



*Figure 6.1: Alm-prp-grs ternary diagrams displaying garnet compositions. Marker formats as in* Table 0.13. *Phase components as in* Table 0.11b. *a) all samples; b) host bodies; c) websterites; d) veining features.* 



Figure 6.2: Major element-crystal profile plot displaying compositional variation across garnet crystals. Horizontal axis numbers refer to analysis sites. (e) – very edge, (r) – rim, (m) – mid, (c) – core. Element lines as described in Table 0.13. a) eclogite from Årsheimneset, core-rim traverse [QC26E, analysis box 5]; b) garnet websterite from Årsheimneset, rim-core-rim traverse [QC26G, analysis box 17]; c) vein core from Årsheimneset, rim-core-rim traverse [QC26H, analysis boxes 14-15].

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Nybø has the same composition and range as interstitial eclogite garnet. However, a mauve garnet porphyroblast in eclogite from Nybø has a slightly more almandinerich, pyrope-poor composition compared to the interstitial garnet. In comparison, the websterite garnets from Årsheimneset (*Fig. A3.10*), Remøysunde (*Fig. A4.8*) and Myrbærneset (*Fig. A8.8*) have more pyrope-rich, grossular-poor compositions than those in their respective eclogites. Compared to olivine websterite garnets, those in garnet websterites are less pyrope-rich (*Fig. A9.12*). Garnets in vein selvages at Myrbærneset plot with a slightly more almandine-rich composition than the websterite body domain and the garnetite garnets have a slightly more almandine-rich composition than the websterite selvage garnets (*Fig. A8.8*). Vein garnets at Svartberget also plot with a more almandine-rich composition than the websterite domains (*Fig. A9.12*). In contrast, vein garnets at Årsheimneset plot with a more pyrope-rich composition than their websterite domain (and eclogite) (*Fig. A3.10*).

The only consistent zoning seen in garnets is a drop in Mg and equivalent increase in Fe at the very edges of garnet crystals (*Fig. 6.2*): this is most pronounced in garnets with amphibole coronas (e.g.QC26HA). As this occurs in both metamorphic and metasomatic rocks, this reflects rims equilibrating to retrograde conditions rather than new growth in a system increasingly enriched in Fe. The minimal extent of the zoning and small distance over which it occurs also suggests it is related to retrograde diffusional effects. Cores and mid zones generally have flat zoning profiles although slightly higher Ca and Mg is seen in some crystal cores (e.g. QC15B garnetite band). Ca typically varies but no consistent zoning is observed. A slight increase in Mn concentrations at the very edges of some garnet rims is also noted: when absent Mn typically has the most restricted variation.

## 6.5.2. <u>Clinopyroxene</u>

Like garnet, clinopyroxenes in websterite plot between peridotite and eclogite compositions reproducing the mafic-ultramafic bulk variation pattern. The relative aegirine [NaFe<sup>3+</sup>Si<sub>2</sub>O<sub>6</sub>] content of clinopyroxene is constrained by Fe<sup>3+</sup> Mössbauer analyses and inferred estimations. Aegirine concentrations are below 8% for all clinopyroxene crystals in the Q-jd-aeg ternary diagram (*Fig. 6.3*). The dominant compositional trend is between jadeite and Q (Ca-Mg-Fe) components reflecting diopside-jadeite exchange [Ca<sub>-1</sub>Mg<sub>-1</sub>-NaAl]. The compositional range is Q<sub>42.8-99.1</sub>Jd<sub>0.6-</sub> <sup>53.4</sup>Aeg<sub>0.4-7.8</sub>. Compositions at some localities are well clustered such as at Remøysunde (*Fig. A4.10*) whilst others spread along the mixing line such as at Nybø (*Fig. A1.10*). Clinopyroxenes in vein zones overlap websteritic and eclogitic clinopyroxene compositions. They have a range of compositions at Svartberget but are more clustered at Årsheimneset where they are similar or more jadeite-rich than those in eclogite. The websterite selvages at both localities are more jadeite-rich than their corresponding websterite domains suggesting the introduction of Na.

Clinopyroxenes in the peridotite from Svartberget and in all analysed rock types from Myrbærneset except for the eclogite have strongly diopsidic clinopyroxenes with Na/(Na+Ca) values of 0.02-0.04 and 0.01-0.03, respectively. The eclogite from Myrbærneset has the highest jadeite component of all analysed with Na/(Na+Ca) up to 0.58. Most clinopyroxenes in eclogites analysed are omphacites falling in the omphacite compositional field of Morimoto *et al.* (1988) (*Fig. 6.3b*); although some from Nybø, Årsheimneset and Remøysunde have Na/(Na+Ca) below 0.20. Clinopyroxenes included within the cores of the mauve garnet porphyroblasts within the clinopyroxenite domains at Nybø (see *Fig. 2.11f*) have Na/(Na+Ca) up to 0.43. The typical bimineralic eclogite at this locality contains Na/(Na+Ca) of 0.20-0.30 making them omphacites: those within garnetite bands have the lowest Na contents of the samples from Nybø.

All clinopyroxenes in websterite domains have Na/(Na+Ca) <0.20 making them diopsides and fall in the calcic or 'quadrilateral' Ca-Fe-Mg compositional field according to the Q-J classification of Morimoto *et al.* (1988) (*Fig. 6.3b*). However, some websterite selvages, such as those at Svartberget and Årsheimneset contain omphacitic clinopyroxenes with Na/(Na+Ca) <0.27 and <0.29, respectively and fall in the omphacite field (*Fig. 6.3c*). Those in clinopyroxenite selvages also have Na/(Na+Ca) which cross the diopside-omphacite boundary e.g. 0.16-0.28 at Svartberget and cross the omphacite-quadrilateral compositional field boundary as do clinopyroxenes in garnetite vein cores at Svartberget (*Fig. 6.3c*). All clinopyroxenes in garnet websterites and vein domains from Myrbærneset are strongly quadrilateral clinopyroxenes/ diopsides with Na/(Na+Ca) below 0.04 distinguishing these from other websterites and veining features indicating an origin in a different, less Na-rich chemical system (as proposed in previous chapters).



Figure 6.3: Q-jd-aeg ternary diagrams displaying clinopyroxene compositions. Classification after Morimoto et al. (1988). Q = 'Quad' (quadrilateral) = Ca+Mg+Fe<sup>2+</sup>. Marker formats as in Table 0.13. Phase components as in Table 0.11b. a) all samples; b) host bodies; c) websterites; d) veining features.



Figure 6.4: Wo-fs-en ternary diagrams displaying pyroxene compositions. Classification after Morimoto et al. (1988). Marker formats as in Table 0.13. Phase components as in Table 0.11b. a) all samples; b) host bodies; c) websterites; d) veining features.



Figure 6.5: Q-J diagram after Morimoto et al. (1988) displaying clinopyroxene compositions.  $Q = Ca+Mg+Fe^{2+}$ ; J = 2Na. Marker formats as in Table 0.13. Phase components as in Table 0.11b. a) all samples; b) host bodies; c) websterites; d) veining features.

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Clinopyroxenes included in garnet in eclogite tend to have a similar upper Na/(Na+Ca) values as those in the matrix but do not extend to as low lower values such as at Årsheimneset where included crystals have Na/(Na+Ca) of 0.23-0.26 and those in the matrix have Na/(Na+Ca) of 0.18-0.27. In comparison, inclusions and matrix clinopyroxenes in garnet websterites have similar compositions such as at Kolmannskog (*Figs. A6.5, A6.7*) and clinopyroxene inclusions in clinopyroxenite selvages at Svartberget have similar Na/(Na+Ca) to those in the matrix of the websterite selvages.

Variation in Mg-Fe-Ca is displayed on the en-fs-wo ternary diagram (*Fig. 6.4*). The compositional range of clinopyroxenes is  $En_{40.4-51.4}Fs_{3.1-10.9}Wo_{40.3-51.5}$ . The majority plot in the diopside field but some (those from vein zones) extend into the augite field plotting along the  $\sim$ Fs<sub>10</sub> Ca-Mg mixing line towards greater enstatite contents. The most Fe-rich clinopyroxenes are in the matrix of eclogites from Remøysunde and Årsheimneset.

Major zoning of phase components is not observed in clinopyroxene crystals although Ca-Mg and Na-Al patterns do vary (*Fig. 6.6*). Na-Al patterns often follow each other (e.g. *Fig. 6.6b*) but Al sometimes mirrors Na (*Fig. 6.6d*) or mirrors Mg (*Fig. 6.6f*): the degree of variation typically increases towards crystal rims. The highest Na contents can sometimes be seen towards the rims of crystals and some rims are characterised by higher Ca and lower Mg.

### 6.5.3. Orthopyroxene

The compositional range of orthopyroxene crystals is En<sub>71.6-91.1</sub>Fs<sub>8.5-28.0</sub>Wo<sub>0.2-0.8</sub> with Fe-Mg exchange as the dominant substitution. Those in peridotites from Svartberget are the most enstatite-rich and have well-clustered compositions (Mg# 0.87-0.91). Those in websterites from Årsheimneset are the most ferrosilite-rich [Fe<sub>2</sub>Si<sub>2</sub>O<sub>6</sub>] and show the greatest compositional range. Those from Nybø, Grytting, Kolmannskog and Remøysunde have limited compositional variation. Orthopyroxenes in websterite selvages are less enstatite-rich than those in websterite domains.

Orthopyroxene crystals typically show no major zoning although discernible increases in Fe and mirroring decreases in Mg can be seen at crystal edges (*Fig. 6.7*). Al shows



Figure 6.6: Major element-crystal profile plots displaying compositional variation across clinopyroxene crystals. Horizontal axis numbers refer to analysis sites. (e) – very edge, (r) – rim, (m) – mid, (c) – core. Element lines as described in Table 0.13. a-b) bimineralic eclogite from Nybø, rim-core-mid traverse [QC15F, analysis box 3]; c-d) eclogite from Årsheimneset, rim-core-mid traverse [QC26E, analysis box 5]; e-f) eclogite from Remøysunde, rim-core-rim traverse [QC29L, analysis box 6].



Figure 6.7: Major element-crystal profile plot displaying compositional variation across orthopyroxene crystals. Horizontal axis numbers refer to analysis sites. (e) – very edge, (r) – rim, (m) – mid, (c) – core. Element lines as described in Table 0.13. a-b) garnet websterite from Nybø, rim-corerim traverse [QC15N, analysis box 1]; c-d) garnet websterite sample from Grytting, rim-core-rim traverse [Ø16, analysis box 6]; e-f) garnet websterite from Årsheimneset, rim-core traverse [A347, analysis box 5].

considerable relative variation with a range in concentrations of 0.006-0.089pfu. Lowest Al contents are not always found at the centre of crystals as they are shaped now (e.g. *Fig. 6.7f*). Al is typically much higher (twice the concentration) at rims but isn't always and mid zones of crystals can be high like most rims or low like most cores. Ca and Mn profiles across crystals are mostly reasonably flat although slight increases in Mn can also be seen in some rims. Ti, Cr, Na and K concentrations are typically below their detection limits.

## 6.5.4. <u>Biotite</u>

Biotites show a range in compositions with variation dominated by Mg-Tschermaks (MgTs) exchange [Si<sub>-1</sub>Mg<sub>-1</sub>-Al<sub>2</sub>] from phlogopite [KMg<sub>3</sub>AlSi<sub>3</sub>O<sub>10</sub>(OH)<sub>2</sub>] towards the 'eastonite' [K(Mg<sub>2</sub>Al)Al<sub>2</sub>Si<sub>2</sub>O<sub>10</sub>(OH)<sub>2</sub>] end-member. Both biotite compositional plots distinguish biotites from different rock types well; particularly the Al-Si plot (Fig. 6.8b) and particularly in samples from Svartberget (Fig. A9.23b). The increased scatter on the Al-Mg# plot (Fig. 6.8a) reflects the involvement of Mg in other substitutions such as Fe-Mg exchange which produces a extension towards the annite  $[KFe_3AISi_3O_{10}(OH)_2]$  and siderophyllite end-members  $[K(Fe^{2+}_2AI)AI_2Si_2O_{10}(OH)_2]$ . Biotites in samples from Arsheimneset and Remøysunde have the most phlogopitic compositions (highest Mg# (<0.92)) and those from the matrix of websterites from Kolmannskog have the lowest Mg# (>0.75) which is unsurprising considering the greater extent of amphibolite-facies retrogression at this locality. Biotites in websterites span the full range of Mg# observed. More 'eastonite'-dominant biotites are present in retrograde intergrowths reflecting greater amounts of MgTs substitution. A third exchange mechanism is evident in compositions of biotite included in garnetite vein cores from Svartberget as indicated by a negative correlation between Ti and total cations (reflecting a positive correlation between Ti and site vacancies) (see Fig. A9.24). A negative correlation also exists between Mg and Ti, Al and Ti and a positive correlation between total cations and Mg. These trends reflect 
-Ti substitution for octahedral Y site Mg (and/or AI) which is favoured at higher temperatures (Deer et al., 1992).

Overall, the compositional trend from ultramafic to more felsic bulk compositions (whether magmatic or metasomatic in origin) is reproduced in the biotite compositions: websteritic biotites have more phlogopitic compositions than those



*Figure 6.8: Compositional plots of biotites in all samples. Marker formats as in* Table 0.13. *Phase components as in* Table 0.11b. *a) Al-Mg# plot; b) Al-Si plot.* 

from associated eclogites and vein zones with higher Mg# and Si contents. An exception is Myrbærneset where the biotites from the olivine garnet websterite have a compositional range encompassing those in the garnet websterite.

There is no considerable distinction in upper Mg#, Si and lower Al compositions between matrix and included crystals from the same samples although matrix biotites extend to lower Mg# and Si contents and higher Al contents indicating continued Fe-Mg and MgTs exchange. Texturally late biotites (in reaction with garnet) extend to the highest Al contents measured such as those from Årsheimneset.

### 6.5.5. <u>Amphibole</u>

According to the classification of Leake *et al.* (1997) all amphiboles from all localities are calcic amphiboles having  $Na_B < 0.5pfu$  and belong to the hornblende *s.l.* and tremolite groups. Their subdivision into nominal classes is presented in *Fig. 6.9*. The various textural types of 'early' amphibole in the matrix of websterites, including granoblastic, decussate, poikiloblastic and interstitial, all appear to have similar compositions which are typically more tremolitic than those included in garnet or which form late coronas. Compositional zoning is observed in amphibole crystals particularly in late, retrograde, coronal amphibole which have higher Na contents in the rims immediately adjacent to the garnet around which it is forming.

The overall, dominant compositional trend of amphiboles is from tremolite to pargasite compositions reflecting a combination of both edenitic  $[Si_1 - 1 - NaAl]$  and MgTs substitution (Deer *et al.*, 1992) (*Fig. 6.10a*). The scatter about the trend line reflects the interplay of these two mechanisms and others such as Fe-Mg exchange which produces the greater scatter of plots vs Mg# (*Fig. 6.9*); the correlation between Mg with Fe is less well defined than that with Si, Al or Na suggesting it is not a dominant exchange mechanism.

Although not as well defined as that in biotite, there is a crude reproduction of maficultramafic bulk change. Websterites contain amphiboles in the matrix with the most tremolitic compositions and those in garnetite vein cores have more pargasitic compositions although there is considerable overlap. 'Early' amphiboles in the matrix



Figure 6.9: Mg#-Si plots of amphiboles in all samples. Phase classification after Leake et al. (1997). Marker formats as in Table 0.13. Phase components as in Table 0.11b. a) for amphiboles with  $(Na+K)_A > 0.5$ ; b) for amphiboles with  $(Na+K)_A > 0.5$ .


Figure 6.10: Compositional plots of amphiboles in all samples. Marker formats as in Table 0.13. Phase components as in Table 0.11b. a)  $(Na+K)_A$  plot; b) Al-Mg plot.

of eclogites and clinopyroxenite selvages plot in the middle of this trend. The more Na-rich nature of the amphiboles in the garnetite vein cores may reflect their clinopyroxene-poor nature.

Both inclusions within eclogitic and some websteritic garnets, such as from Årsheimneset and Remøysunde, inclusions within garnet in vein cores at Svartberget and late, coronal amphibole, such as those from Myrbærneset, have the greatest pargasitic compositions. In some instances, such as in vein zones from Svartberget, there is no considerable difference in composition between matrix and included phases. Some inclusions in garnet in vein cores from Svartberget plot below the main tremolite-pargasite trend showing less edenitic substitution for its tschermakitic substitution: this is also seen in inclusions in the mauve garnet porphyroblasts in eclogite from Nybø.

## 6.5.6. Olivine

There is a small compositional range of Fo<sub>83.8-84.9</sub>Fa<sub>14.9-15.9</sub>Te<sub>0.2-0.3</sub> (*Fig. 6.11*). This suggests there is little compositional variation or dominant substitution mechanism but that which took place was Fe-Mg exchange. Mn contents vary between 0.04-0.06pfu (~0.18-0.30wt%) and Ca and Cr contents are mostly below EMP detection limits.

# 6.5.7. <u>Carbonates</u>

There are three groupings of carbonate composition as defined on the mgs-cal-sd ternary plot (*Fig. 6.12*). Peak, 'early' carbonate crystals in the Svartberget olivine garnet websterite matrix and garnet websterites from Nybø and Årsheimneset consist of magnesite cores with dolomite mantles. Both magnesite and dolomite from garnet websterites have a slightly higher siderite [FeCO<sub>3</sub>] contents than those in olivine garnet websterite with Mg# of  $\geq$ 0.85 and Sd<sub>0.12-0.15</sub> compared to Sd<sub>0.07-0.08</sub> in magnesite and Sd<sub>0.04-0.08</sub> compared with Sd<sub>0.02</sub> in dolomite. Calcite is the late phase in fractures, up biotite cleavages and as outer, retrograde rims of altered matrix carbonate crystals.



*Figure 6.11: Fo-fa-te ternary plot of olivine compositions. Marker formats as described in* Table 0.13. *Phase components as listed in* Table 0.11b.



*Figure 6.12: Mgs-cal-sd ternary plot of carbonate compositions. Marker formats as described in* Table 0.13. *Phase components as listed in* Table 0.11b.

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### 6.6. Data Interpretation

# 6.6.1. Rock Classification

All clinopyroxenes analysed from garnet websterite body domains are diopsidic, not omphacitic in nature. The rocks are therefore garnet websterites *s.s.* and not orthopyroxene eclogites. However, most clinopyroxenes in vein websterite selvages from Årsheimneset and Svartberget and clinopyroxene-rich garnetite vein cores from Svartberget plot just into the omphacite field and so are orthopyroxene eclogites *s.s.* and eclogites *s.s.*. The increased jadeite component indicates a Na component was added during metasomatism. In comparison, the diopside-rich nature of clinopyroxenes in garnet websterites and vein zones at Myrbærneset suggests the body has a more depleted protolith than other garnet websterites and the metasomatising fluid at this locality was not rich in Na.

Clinopyroxenes, orthopyroxenes, olivines, biotites, amphiboles and carbonates are all close to Mg-rich end-members in terms of their Fe-Mg contents being dominantly diopside, enstatite, forsterite, phlogopite, tremolite-magnesiohornblende and magnesite-dolomite, respectively. The relative Fe-Mg character of orthopyroxenes and olivines reflect well the overall character of the meta-ultramafic rocks. Compared to olivines from a mantle 'Mg-Cr' garnet peridotite from the Almklovdalen body [QC20-21] in the WGR, which have Fo<sub>88-92</sub> (Medaris Jr, 1980), the olivines from Svartberget demonstrate a Mg-poorer nature (Fo<sub>84-85</sub>) and distinction from this classification of garnet peridotite potentially supporting the Fe-Ti classification of Vrijmoed et al. (2006). Olivine with Fo <90 are interpreted to be cumulate in origin rather than mantle rocks depleted by partial melting (Rehfeldt et al., 2008) supporting a crustal origin for the Svartberget peridotite. For comparison the olivines in the peridotite from Myrbærneset have (unreliable) compositions of approximately Fo<sub>80.0</sub>Fa<sub>19.8</sub>Te<sub>0.2</sub> demonstrating a more Fe-enriched nature similar to its whole rock composition (see Chapter 4) and intermediate nature of Svartberget.  $Cr_2O_3$  contents in the Svartberget olivine garnet websterite in garnets, clinopyroxenes and orthopyroxenes are average contents of 0.56, 0.10 and 0.02wt%, respectively. These are all considerably lower than those in the Almklovdalen garnet peridotite which have contents of 2.01-2.20, 1.13-1.15 and 0.10-0.14wt% in garnet, clinopyroxene and orthopyroxene, respectively (Medaris Jr, 1980) and higher than Myrbærneset

contents of 0.02, 0.02, 0.01wt%, respectively, demonstrating intermediate Cr values also.

# 6.6.2. <u>Chemical Zonation</u>

Chemical homogenisation and equilibration is thought to occur in major silicate phases at temperatures of over 700°C even in garnet in which diffusion is typically slower (e.g. Krogh, 1977; Krogh Ravna and Paquin, 2003). The flat chemical profiles of major elements in crystal interiors in both metamorphic and metasomatic rocks demonstrate either complete homogenisation and equilibration of variable prograde profiles or new growth at peak temperature conditions of over 700°C (Krogh Ravna and Paquin, 2003).

Major element variation is greatest in the rim zones and very edges of crystals and reflects re-equilibration to lower P-T conditions. Only the outer ~0.5mm of the crystals are affected demonstrating minimal diffusive re-equilibration occurred during retrogression. This indicates the duration of retrograde metamorphism was short or the rate of diffusive re-equilibration was slow. Na-Al-Ca-Mg patterns and contents in clinopyroxene differ most at crystal rims suggesting they are subject to exchange mechanisms other than jadeite substitution, such as MgTs exchange. The mirroring of Fe and Mg contents in orthopyroxenes demonstrates that Fe-Mg exchange is the dominant substitution mechanism at all stages although variation in Al demonstrates MgTs substitution also occurred. The increase of Mn at garnet (and orthopyroxene) crystal edges suggest either the introduction of Mn-rich fluids, release of Mn from another phase or more likely resorption of garnet (and orthopyroxene) with Mn back-diffusing due to incompatibility in the new phase.

Minor zoning of Na-Al in clinopyroxene and Al in orthopyroxene implies slight chemical zonation of pressure-sensitive components still persists. Assuming peak pressure conditions were recorded in crystal cores, the fact that Na-Al contents (jadeite) in clinopyroxene cores are not always highest and Al contents in orthopyroxene cores (MgTs) are not always lowest suggests the peak pressures crystal cores are no longer situated in the cores of the present crystals. This suggests asymmetrical crystal resorption via recrystallisation, asymmetric or non-concentric growth or heterogeneous diffusional exchange has taken place.

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## 6.6.3. <u>Compositional Variation</u>

Chemical variation of individual minerals in metamorphic rocks may be due to different bulk compositions of their protoliths or exchange reactions due to metamorphic equilibration at different P-T conditions: mafic-ultramafic protolith compositions have the potential to be greatly variable considering the cm-scale and larger compositional layering possible in magmatic crustal bodies. For example, increased Mg and Ca in garnet may reflect increased equilibration temperatures in the presence of clinopyroxene or a more ultramafic protolith. Eclogite garnets from each locality form separate clusters suggesting considerable difference in their protolith bulk compositions and the overall Fe-Mg compositional trend of garnets from olivine websterite to websterite to eclogite reproduces that of the whole rock compositional variation reflecting the ultramafic-mafic character of the host rocks; this is particularly evident in samples from Årsheimneset and Svartberget. However, higher Ca and lower Mg compositions in garnet like those in eclogites at Nybø and Myrbærneset relative to their websterites are also indicative of higher pressure conditions.

The variation in relative compositions of eclogite and websterite garnets at the same locality where the websterites are interpreted to be metamorphic in origin, for example at Nybø where garnet compositions are similar and at Myrbærneset where they are greatly different, suggests the cause of variation is due bulk composition. Furthermore, the slightly different compositions of the mauve porphyroblasts and the orange-pink interstitial garnet in eclogites from Nybø indicate the presence and persistence of local chemical heterogeneity; such small-scale, isolated variation precludes differing P-T conditions as the cause particularly as the blasts have flat zoning profiles. The more almandine-rich garnet and jadeite-rich included clinopyroxene compositions of the mauve porphyroblasts suggest a local Mg-Ca-poorer, Na-Al richer area perhaps originating from an albite-rich domain. The persistence of this chemical difference is thought to reflect incomplete homogeneity due to a lower temperature of equilibration at this locality. Samples from Nybø were expected to show the greatest degree of zonation and disequilibrium as the locality lies close to the estimated regional 700°C isotherm (*Figs. 1.9, 2.2*).

Garnets within the vein cores at Årsheimneset are more Mg-rich than the websterite garnets. This opposes the compositional trend seen at Svartberget where vein garnets have a more Fe-rich character than those in the websterites. The compositions of vein garnets at both localities are actually similar suggesting a similar chemical system and they do not result from metasomatism or the chemical trend would be the same.

The dominant diopside-jadeite compositional variation of clinopyroxene both within and between rock types may reflect either equilibration at higher pressures enabling a greater jadeite component in the lattice structure or the equilibration in a less Narich environment/system at similar pressures. The reproduction of the ultramaficmafic trend suggests the variation is due to protolith heterogeneity. The wellclustered garnet and clinopyroxene compositions in both eclogite and websterite at Remøysunde and Årsheimneset suggest the lithologies are not transitional but are both well equilibrated and distinct. The trend of clinopyroxene compositions from Nybø (*Fig. A1.10*), in comparison to the clustering of associated garnet compositions suggests, clinopyroxene compositions are more sensitive to chemical change. Clinopyroxenes also show more retrogressive features than other minerals having exsolved phases and symplectitic intergrowths.

Compositional variation in biotite is due to MgTs substitution and possibly Fe-Mg exchange: variation in Mg# may be due to Fe-Mg exchange or MgTs substitution decreasing the relative amount of Mg but without an associated increase in Fe. Increased Fe in biotite in equilibrium with garnet such as seen in vein cores and glimmerite selvages from Svartberget usually suggests equilibration at higher temperatures (Perchuk, 1969). However, higher Fe in both garnet and biotite in vein garnets relative to websterite garnets at Svartberget suggest they grew in a relatively more Fe-rich environment rather than exchanging Fe-Mg due to changing P-T conditions. The higher Ti contents in biotites included in some garnets at Svartberget reflect a Ti-enriched environment. As noted, vein core garnets at Svartberget contain numerous, tiny rutile needles and larger grains: Ti becomes less compatible in the garnet lattice with increasing temperatures (Kiseeva *et al.*, 2012).

The presence of only one dominant trend in amphibole compositions compared to the range of types (prograde, 'early' and retrograde) suggests that the compositional

variation associated with metasomatism follows that associated with prograde and retrograde metamorphism towards more Na-K-Al-Fe-rich, Mg-Si-poor, pargasitic compositions. During retrogression, Na, Al and Fe are sourced from replaced omphacite and garnet whereas during metasomatism they are sourced from the incoming fluid.

### 6.6.4. Matrix vs Included Phases

Matrix and garnet-included phases mostly have similar compositions, particularly those which represent peak conditions; this suggests they formed at the same time under similar P-T-X conditions. Matrix phases do however extend to compositions representative of continued chemical exchange during retrogression: this relationship is seen in clinopyroxenes and biotites and suggests garnet sheltered the included phases from chemical communication with matrix phases. For example, some matrix clinopyroxenes in samples from Svartberget extend to lower Na/(Ca+Na) than included clinopyroxenes reflecting lower pressure conditions and matrix biotites extend to lower Mg# and higher Al contents than included biotites reflecting continued Fe-Mg and MgTs exchange which are dependent on P-T-X conditions (e.g. Harley, 1984a,b).

The pattern of amphibole compositional variation slightly differs from biotite and clinopyroxene in that both included and texturally late, retrogressive, matrix amphiboles in both eclogite and garnet websterite have greater pargasitic compositions than most texturally early matrix amphiboles. This suggests the included amphiboles are either retrograde, altered clinopyroxene inclusions or represent a prograde paragenesis formed under similar P-T-X conditions as the retrograde amphibole. The similar compositions of matrix and included amphiboles, in some instances however, such as in vein zones from Svartberget, suggests they grew contemporaneously with the garnet host in a well equilibrated system.

## 6.6.5. Fluid-Rock Interaction

The involvement of a metasomatic fluid adds another variable to the controlling factors of mineral compositions as P-T-X change no longer operates in a closed chemical system so variation may also reflect the loss or addition of elements

to/from the fluid phase which may itself vary in composition over time with continued influx. This makes the interpretation of mineral compositions in metasomatic rocks less straightforward. As the rocks are considered to be equilibrated to peak temperature conditions, chemical variation due to variable P-T conditions at each locality can be ruled out.

The compositional similarity of orthopyroxene eclogite selvage clinopyroxenes to clinopyroxenite selvage clinopyroxenes rather than to garnet websterite domain clinopyroxenes, both in the metasomatic websterite domains at Svartberget and the 'metamorphic' domains at Årsheimneset, suggests only the immediate vicinity of the veins is Na-Al-rich and considerable transport of these elements away from the vein does not occur, Ca-Mg contents of the fluid were low (supporting the calculated composition in Chapter 5) and the influence of the wall rock composition decreases towards the fracture wall. This supports the interpretation in Chapter 5 that zone compositions are influenced more by the composition of the rock rather than that of the fluid further away from the vein fracture. Na-Al in the fluids could be sourced from eclogite-facies phases which replaced albite plagioclase such as jadeite clinopyroxene.

The compositional trend re-producing the vein zonation pattern in biotite compositions at Svartberget and the ultramafic-felsic nature of the bulk rock suggests that biotite compositional change is able to reflect the overall compositional change of the fluid-rock system indicating it involved the addition of AI (and Fe). Mineral compositions in vein domains from Svartberget therefore suggest the metasomatic fluid is rich in Al-Na-K-Fe-Ti.

The fluids responsible for vein formation at Myrbærneset were not richer in Na-Al(-K-Ti) than the wall-rock as the vein clinopyroxenes retain an ultramafic diopsidic composition similar to the host garnet websterite and olivine garnet websterite and no other Na-bearing phase is present except for minor late amphibole.

### 6.7. <u>Summary</u>

Garnets are dominantly grospyral in composition with pyrope, almandine and grossular components forming 90-99% of the garnet composition. Compositional variation is mostly

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between pyrope and almandine although variable grossular contents are found. Orthopyroxenes, olivines, biotites, amphiboles and carbonates are all close to Mg-rich endmembers in terms of their Fe-Mg contents being dominantly enstatite, forsterite, phlogopite, tremolite-magnesiohornblende and magnesite-dolomite, respectively. Compositional variation of orthopyroxene, olivines and carbonates is dominantly towards Fe-rich end-members; although this is minimal for olivines and carbonates. Orthopyroxenes and carbonates in olivine garnet websterite are mostly more Mg-rich, Fe-poor than those in garnet websterites. The Fe-Mg composition of olivines (Fo <90) suggests the rocks were cumulates in origin rather than depleted mantle rocks. Eclogites contain the most grossularand almandine-rich garnets and olivine garnet websterites contain the most pyrope-rich garnets. Compositional variation of phlogopitic biotite is dominated by MgTs exchange trending towards the 'eastonite' end-member although Fe-Mg exchange towards annite compositions also occurs. Biotites in garnet websterites are the most phlogopitic and those in vein core garnetites and retrograde reactions are the most eastonitic. Compositional variation of amphibole is dominated by edenitic and MgTs substitution towards the pargasite end-member. Amphiboles in garnet websterites are the most tremolitic and those in vein core garnetites and retrograde reactions are generally the most pargasitic.

Clinopyroxenes are strongly Mg-rich in terms of their Fe-Mg content. Those in eclogites are confirmed as omphacitic (Na/(Na+Ca) = 0.20-0.80). Those in garnet websterite body domains, like olivine garnet websterites, are diopsidic so the rocks are thus confirmed as garnet websterites *s.s.*. Most clinopyroxenes in vein selvages websterites and vein cores from Årsheimneset and Svartberget are omphacitic demonstrating the rocks are orthopyroxene eclogites *s.s.* and eclogites *s.s.*, respectively. The variation is inferred to represent addition of Na-AI during metasomatism considering the occurrence omphacitic clinopyroxenes only in the immediate vicinity of the laterally-zoned veins. Websterites and vein selvages at Myrbærneset contain strongly Na-poor, diopsidic clinopyroxenes suggesting a more depleted, ultramafic protolith and indicating a Na-poor fluid was responsible for vein formation. Compositional variation in clinopyroxenes is dominantly between diopside and jadeite components.

Crystals in all rocks have flat chemical profiles except for minor retrogressive rims suggesting rocks were chemically equilibrated to high temperature conditions (>700°C). Most included crystals have the same compositions as crystals in the matrix except for more retrogressive compositions in some matrix crystals implying matrices continued to

equilibrate. Compositions of clinopyroxene inclusions in the mauve garnet porphyroblasts at Nybø differ from matrix compositions and probably represent the persistence of protolith-inherited chemical sub-domains isolated from the equilibrating matrix due to lower temperatures of equilibration and incomplete homogeneity.

The extent of chemical variation of a rock type between different localities and between rocks at the same locality is dominantly attributed to compositional variation of the protolith mafic-ultramafic rocks and the influence of metasomatism as opposed to varying P-T conditions as different compositions are found in rocks at the same locality which would have experienced the same P-T conditions and/or the chemical change cannot be explained by re-equilibration to different P-T conditions. Eclogites from the various localities contain phases with a range of compositions which reflect considerably different protolith bulk compositions.

### <u>Chapter 7 – Thermobarometry</u>

# 7.1. Introduction

Presented in this chapter are absolute thermobarometric estimates of peak P-T conditions using mineral equilibria and mineral chemistry data (presented in Chapter 6). Orthopyroxene-bearing rocks from seven localities were selected for analysis based on textural and mineralogical suitability and the quality of the thin sections. Those selected include garnet websterite body domains from Nybø, Grytting, Årsheimneset, Remøysunde, Kolmannskog and Myrbærneset, garnet websterite vein selvages from Myrbærneset and Svartberget, an orthopyroxene eclogite vein selvage from Årsheimneset and olivine garnet websterite from Svartberget.

All localities except Kolmannskog lie within the WGR UHP terrane. UHP conditions are therefore possible and expected from thermobarometric estimations. As no exsolution of pyroxene from garnet was observed in the samples, very ultrahigh pressures (>5GPa) are not expected (e.g. van Roermund and Drury, 1998; Ye *et al.*, 2000a). The major element chemical homogeneity in the interiors of garnet blasts in the samples suggests temperatures above 700°C were experienced long enough for equilibration to occur. However, the presence of fine rims with different compositions suggests minor reequilibration at lower P-T conditions occurred. Furthermore, the presence of quartz, not coesite, in the free silica-bearing rocks suggests the matrices of the samples re- equilibrated at HP conditions. The presence of plagioclase in the retrograde assemblages indicates partial re-equilibration to conditions outside the eclogite-facies field occurred. However, sub-eclogite-facies conditions are not expected from the samples chosen for analysis as these showed minimal retrogression and alteration to a plagioclase-bearing paragenesis.

## 7.2. Aims and Objectives

The main aims of thermobarometry are to obtain estimates of absolute P-T conditions of equilibration of the mafic-ultramafic bodies in order to determine whether metasomatic garnet websterites-orthopyroxene eclogites equilibrated at UHP conditions or whether they are lower pressure features associated with exhumation and to refine absolute P-T paths.

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Study of rocks from different localities allows any regional variation in the P-T characteristics of the WGR UHP terrane to be identified. A particular objective is therefore to obtain consistent, comparable, conservative estimates based on quality data and thermobarometers most suitable to the P-T-X conditions of the rocks analysed between different localities. This requires critical selection of calibrations and the use of reliable concentrations of particularly important elements. Thermobarometric estimates will be compared with the P-T conditions inferred from petrographic observations of this and other studies to test the accuracy and reliability of the estimates and then compared with the estimates of other studies with the further aim of integrating the results into the WGR dataset.

Reliability in the accuracy of P-T estimates will enable confidence in the estimates from the Kolmannskog locality with the objective of determining if it experienced UHP conditions: it has similar veining features to Svartberget but it currently resides outside of the defined eastern limit of the northernmost UHP domain (e.g. Vrijmoed *et al.*, 2008; Hacker *et al.*, 2010; Butler *et al.*, 2013). The Kolmannskog body does however show the greatest degree of retrogression and so may no longer preserve UHP conditions even if they were experienced.

A final objective is to corroborate or contest the 2.1GPa pressure gap found between the host body and the vein system at Svartberget by Vrijmoed *et al.* (2006).

# 7.3. Data Acquisition

Crystals selected for EMP analysis for thermobarometric calculations were preferably large and of similar size, part of the matrix, showed no major element zonation, had granoblastic or equilibrium textures and showed no reaction textures, exsolution or retrogressive alteration. Cores, mids and rims of crystals were analysed to obtain peak to retrogressive P-T estimations and construct a progression path.

SEM images were obtained as described in Appendix B2 and EMP data were collected and prepared as described in Chapter 6 and Appendix B4. The data points selected were those with the least erroneous wt% totals and which required the least stoichiometric normalisation. EMP data from the olivine garnet websterite from Myrbærneset were not suitable for use (as their wt% totals were outside the acceptable limits) and therefore P-T estimations have not been made. Microsoft Office Excel 2007 was used to carry out thermobarometric calculations and to plot P-T graphs.

*Table 7.1* displays the thermobarometers and calibrations chosen for use in P-T estimations. The rationales behind the choices are explained in the full methodology in the Appendix [B6]. Ferric iron contents were estimated using the various methods outlined in Chapter 6 and Al in M1 sites in orthopyroxene was calculated as half the total Al content. The rationale behind this latter decision is also explained in the methodology in the Appendix [B6] along with other sources of error and the considerable uncertainty in thermobarometric estimations.

Thermometer	Calibration	Code
orthopyroxene-clinopyroxene	Brey and Köhler (1990)	[BK90]
	Taylor (1998)	[T98]
enstatite-in-clinopyroxene	Nimis and Taylor (2000)	[NT00]
diopside-in-orthopyroxene	Nimis and Grütter (2010)	[NG10o]
garnet-clinopyroxene	Powell (1985)	[P85]
	Krogh Ravna (2000)	[KR00]
garnet-orthopyroxene	Harley (1984a)	[H84t]
	Nimis and Grütter (2010)	[NG10g]
garnet-olivine	O'Neill and Wood (1979)	[O'N90]
	corrected by O'Neill (1980)	
garnet-biotite	Indares and Martignole (1985a)	[IM85]
Barometer	Calibration	Code
garnet_orthonyroyene	Nickel and Green (1985)	[NG85]
gametormopyroxene	Harley (1984b)	[H84b]

Table 7.1: Chosen thermobarometers and calibrations to plot and their reference codes

# 7.4. Data Presentation

Data points and pairs used in thermobarometric calculations are presented in the Appendix [J1-J5] and P-T plots of calibration lines for each crystal zone for each sample are presented in the Appendix [J6]. *Table J6.1* details the line formats used in all the plots. Intersections of the Harley (1984b) garnet-orthopyroxene barometer calibration lines (orange) and Powell (1985) garnet-clinopyroxene thermometer calibration lines (dark green) were chosen as the most reliable estimations of P-T conditions (see the full methodology [B6] for rationale).

guiner or mop)			Crystal	P	T	Intersectio	on Errors	Max Errors*		
Locality	Rock <sup>^</sup>	Sample	Zone	(GPa)	(°C)	± P (GPa)	± T (°C)	± P (GPa)	± T (°C)	
Nybø	gw	Δ1308	core- mid	2.65	640	0.37	29	0.37	29	
Nybø	gw	Δ1308	rim	2.50	635	0.22	10	0.22	10	
Nybø	gw	QC15J/N	core	3.71	737	0.10	9	0.10	9	
Nybø	gw	QC15J/N	mid	3.73	736	0.14	20	0.14	20	
Nybø	gw	QC15J/N	rim	3.60	701	0.11	4	0.11	5	
Grytting	gw	Ø16	core	3.14	665	0.06	6	0.10	6	
Grytting	gw	Ø16	mid	2.71	646	0.40	19	0.40	19	
Grytting	gw	Ø16	rim	2.35	633	0.13	6	0.13	6	
Årsheimneset	gw	A347	core	3.53	696	0.03	6	0.10	6	
Årsheimneset	gw	A347	mid	2.89	662	0.09	6	0.10	6	
Årsheimneset	gw	A347	rim	2.23	638	0.23	9	0.23	9	
Årsheimneset	gw	QC26G	core	3.21	680	0.19	27	0.19	27	
Årsheimneset	oes	QC26HC	core	2.95	715	0.07	5	0.10	5	
Årsheimneset	oes	QC26HC	mid	3.05	709	0.16	16	0.16	16	
Remøysunde	gw	QC29A	core- mid	3.64	815	0.25	54	0.25	54	
Kolmannskog	gw	QC32A	core	2.87	750	0.26	5	0.26	5	
Kolmannskog	gw	QC32A	mid	3.02	754	0	0	0.10	5	
Myrbærneset	gw	QC34E	core	2.64	728	0.23	9	0.23	9	
Myrbærneset	gws	QC34F	core- mid	2.22	710	0.45	15	0.45	15	
Myrbærneset	gws	QC34F	rim	1.91	670	0	0	0.10	5	
Svartberget	ogw	QC36A	core	3.36	764	0.22	7	0.22	7	
Svartberget	ogw	QC36A	mid	2.68	737	0.06	4	0.10	5	
Svartberget	ogw	QC36A	rim	2.71	744	0.38	12	0.38	12	
Svartberget	gw	QC36D	rim	1.98	678	0	0	0.10	5	
Svartberget	gw	SV1A	core	3.85	792	0.27	9	0.27	9	
Svartberget	gw	SV1A	mid	3.14	769	0.28	14	0.28	14	
Svartberget	gw	SV1A	rim	2.39	698	0.38	12	0.38	12	

Table 7.2: P-T estimations and errors associated with the intersection space between Harley (1984b) garnet-orthopyroxene barometer and Powell (1985) garnet-clinopyroxene thermometer calibration lines

^ gw = garnet websterite; oes = orthopyroxene eclogite selvage; gws = garnet websterite selvage; ogw = olivine garnet websterite

\* EMP analytical errors of 0.1GPa and 5°C are used if intersection errors are below these values

Chosen P-T points and errors are presented in *Table 7.2*. Errors were calculated considering propagation of EMP data analytical error and the area of intersection of the calibration lines: see the full methodology [B6] for the method of calculation. Maximum errors are ±0.45GPa and ±54°C. *Figures 7.1-7.4* are composite plots of the final P-T estimations. Marker formats are as detailed in *Table 0.13*. *Figure 7.1* shows each crystal zone P-T point with its individual errors. *Figure 7.2* highlights the path taken by each sample according to the core to rim progression and *Figure 7.3* separates these into the three UHP domains. *Figure 7.4* compares the data to the P-T estimations of other studies in the WGR.

### 7.5. Data Analysis

All estimations plot within the eclogite-facies P-T field (*Fig. 7.2*). Cores of samples from Årsheimneset and Grytting both plot at pressures above the coesite-quartz boundary and some of those from Svartberget above the diamond-graphite boundary indicating the method of data acquisition and processing has produced sensible results in support of petrological observations even despite the use of conservative calibrations.

All crystal rims yield lower pressures and temperatures than their cores and/or mid zones supporting the interpretation in Chapter 6 of retrograde chemistries in rims. The overall trend of sample P-T paths registered from crystal cores to rims is a large decrease in pressure of up to ~1.5GPa and slight decreases in temperature of up to ~100°C. Retrograde paths are therefore not quite isothermal as proposed by many tectonic models of UHP terrane exhumation (e.g. Hacker, 2007; Kylander-Clark *et al.*, 2008; Hacker *et al.*, 2010). The fact that all core-rim traverses follow similar relative paths also brings confidence to the method of data acquisition and processing, particularly at Svartberget as the different lithologies will have experienced the same retrograde evolution. The paths suggest cores are equilibrated to peak temperatures as no increases in temperature are recorded from cores to mids to rims. Subduction-related P-T paths are typically clockwise loops with  $P_{max}$  before  $T_{max}$  (e.g. Peacock, 1992) and so peak temperatures may have occurred at lower pressures than peak pressure conditions; the pressures calculated are therefore considered to be minimum peak pressures.

The body domain websterite from Kolmannskog plots within the coesite-facies field along the P-T path of the Svartberget samples (*Figs. 7.2-7.3*) suggesting it experienced UHP



Figure 7.1: P-T plot showing results of thermobarometric calculations showing P-T estimations of each crystal zone in each sample with error margins. Marker formats as detailed in Table 0.13. Larger markers are cores, smaller markers are rims. Phase abbreviations as detailed in Table 0.11b. Dashed line = 5°C/km geotherm, lowest possible geotherm (Schreyer, 1988). Solid lines = polymorph phase boundaries (fields as labelled). Graphite-diamond boundary after Day (2012). Coesite-quartz boundary after Bose and Ganguly (1995). Albite = jadeite + quartz equilibria from Holland (1980).



Figure 7.2: P-T plot showing progression path of each sample. Arrows point towards rims. Marker formats and line colours as detailed in Table 0.13. Mineral abbreviations as detailed in Table 0.11b. Dashed line = 5°C/km geotherm, lowest possible geotherm (Schreyer, 1988). Solid lines = polymorph phase boundaries (fields as labelled). Graphite-diamond boundary after Day (2012). Coesite-quartz boundary after Bose and Ganguly (1995). Albite = jadeite + quartz equilibria from Holland (1980). Aluminosilicate boundary after Holdaway (1971). Dotted line = approximate eclogite-facies field from Robertson (1999).

conditions and followed a similar P-T path as the along-strike Svartberget body. Considering the confidence in the method of data acquisition and processing, this result can be relied upon. Other studies of the Kolmannskog Fe-Ti body produced estimates of 2.6GPa, 765°C (Cuthbert *et al.*, 2000 and references cited therein) demonstrating not all rocks, even within a single body, preserve peak conditions.

Crystal cores of the metamorphic, body domain garnet websterite from Myrbærneset plot just below the coesite-facies field at 2.64±0.23GPa, 728±9°C: a UHP nature of the Myrbærneset body is therefore not confirmed in this study although the error range does enter the coesite-facies field. The study of a body named Raknestangen along the north shore of Otrøy has yielded estimates of 3.2GPa, 750°C (Robinson *et al.*, 2003) demonstrating that this area has been subject to UHP conditions if not preserved by all rocks, like at Kolmannskog. Crystal cores of the vein selvage garnet websterite from Myrbærneset yields even lower pressures of 2.22±0.45GPa. The similar core-rim trend of the P-T path to the other samples suggests the veins last equilibrated on the retrograde path. Re-equilibration to retrograde conditions may have occurred during dynamic recrystallisation and formation of the cataclastic texture of the Myrbærneset body and veins or the smaller grain size enabled more complete re-equilibration.

*Figure 7.3* appears to show a reasonable distinction in temperature between the Stadlandet-Nordfjord (light blue) and Nordøyane (dark blue) domains except for the Årsheimneset vein selvage which lies between them. According to the current regional P-T gradient (Kylander-Clark *et al.*, 2008) (*Fig. 1.9*), the Nordfjord-Stadlandet UHP domain should have peak temperatures below 750°C and down to 700°C: all estimates of the samples from Årsheimneset, Grytting and Nybø obtained in this study support this gradient. However, the samples from the northernmost, Nordøyane UHP domain which, according to the map, should have peak temperatures near and above 850°C all yield temperatures below 800°C: the highest temperature conditions were obtained from Svartberget of 792°C.

Regional isobars (*Fig. 1.9*) (Hacker *et al.*, 2010) suggest portions of both the Sørøyane and Nordøyane UHP domains reached pressures of over 3.2GPa and portions of the Nordfjord-Stadlandet UHP domain reached pressures of over 3.0GPa and the discovery of diamond and majorite garnet only in the northernmost, Nordøyane UHP domain (*Fig. 1.9*) suggests it experienced the highest pressure conditions. The minimum pressure estimates presented in this study, however, indicate all three domains reached diamond-facies depths and



Figure 7.3: P-T plot showing progression path of each sample separated into UHP domains. Arrows point towards rims. Light blue = Nordfjord-Stadlandet (southernmost); mid blue = Sørøyane (middle); dark blue = Nordøyane (northernmost); red = Kolmannskog. Marker shape formats as described in Table 0.13. Mineral abbreviations as detailed in Table 0.11b. Dashed line = 5°C/km geotherm, lowest possible geotherm (Schreyer, 1988). Solid lines = polymorph phase boundaries (fields as labelled). Graphite-diamond boundary after Day (2012). Coesite-quartz boundary after Bose and Ganguly (1995). Albite = jadeite + quartz equilibria from Holland (1980). Aluminosilicate boundaries after Holdaway (1971). Dotted line = approximate eclogite-facies field from Robertson (1999).

pressures of over 3.5GPa; although agreeing that the highest pressure conditions were experienced by the Nordøyane UHP domain.

The body domains from Nybø, Årsheimneset and Grytting all follow similar absolute paths. Their similarity again gives confidence in the method of data acquisition and processing considering their geographical proximity and suggests they experienced the same evolution. Such a similarity is also seen in the trend of the Nordøyane domain sample P-T paths although the trend is more concave down than the Stadlandet-Nordfjord UHP domain trend.

The garnet websterite from Remøysunde from the middle Sørøyane UHP domain plots with the highest temperatures of 815°C. This does not follow the regional gradient, which indicates intermediate temperature of 750-800°C, suggesting either the temperatures experienced in the middle UHP domain were similar to those in the northernmost domain, the Nordøyane samples have preserved post-peak temperatures or the rock was not equilibrated leading to erroneous results. Similarly high temperatures have been estimated for the Remøysunde garnet websterite of 825°C but at considerably lower pressures of 2.5GPa (Robinson *et al.*, 2003). The more common migmatitic features observed at localities in the Sørøyane domain, such as Remøysunde and Eggesbø, support the higher temperature conditions produced.

The vein selvage orthopyroxene eclogite from Årsheimneset plots ~50°C above the garnet websterite domain. It also registers an atypical increase in pressure from cores to mid zones. The absence of a considerable temperature difference between the olivine garnet websterite and metasomatic garnet websterites at Svartberget suggests a temperature increase above ambient conditions is not related to vein formation and therefore not responsible for the off-trend location of this sample. The garnet-clinopyroxene thermometer for this sample mostly utilised data from clinopyroxenes included within inclusion-rich garnet porphyroblasts as opposed to large crystals in the matrix and therefore the crystals may not be chemically equilibrated with the matrix orthopyroxenes which were used in the garnet-orthopyroxene barometer. The estimates from this sample may therefore be erroneous and not reliable.

Crystal cores in a garnet websterite from Svartberget yields higher pressures than those of the olivine garnet websterite and plot in the diamond-facies field. This supports the

discovery of diamond in vein cores at this locality by Vrijmoed *et al.* (2006). The 5.5GPa pressures yielded by Vrijmoed *et al.* (2006) (above the 5°C/km geotherm (*Fig. 7.4*)) are not reproduced and only a ~0.5GPa difference with the host body is found as opposed to 2.1GPa. The higher pressures recorded by the garnet websterite relative to the olivine garnet websterite may reflect a real difference in peak equilibration conditions related to fluid activity or the preservation of different P-T conditions on the retrograde path: the smaller grain size of the olivine garnet websterite may enable cores to more easily reequilibrate to retrograde conditions. The higher pressure conditions of crystal rims in the olivine garnet websterite relative to the garnet websterites suggest it stopped reequilibrating to retrograde conditions earlier perhaps in relation to the presence of fluid. Greater insight into the equilibration relationship between host bodies and metasomatic features could be obtained with more comparisons of garnet websterites with vein cores and other mafic-ultramafic lithologies. Comparisons with non-orthopyroxene-bearing lithologies such as eclogites and vein cores are greatly hampered by the lack of a reliable garnet-clinopyroxene barometer.

Crystal cores from the Årsheimneset garnet websterite body domain also plot within the diamond-facies field as do some websterites from Nybø [QC15J/N]. In addition, the garnet websterite from Remøysunde sits on the graphite-diamond phase boundary. Diamond has not been reported at these localities indicating conditions may not have been suitable for the crystallisation of reduced carbon. The other garnet websterite analysed from Nybø [ $\Delta$ 1308] plots just in the coesite-facies field. The difference of over 1GPa with the other samples may be due to errors associated with the high Na contents of some domains at Nybø and incompatibility with the calibrations (due to their development based on experiments with chemical systems which did not consider Na). The similar paths and estimations with the Grytting and Årsheimneset samples however suggest they reflect accurate conditions considering their geographical proximity and lithological similarities. Alternatively, therefore, the difference between the websterites may reflect chemical heterogeneity shown to be preserved at Nybø and domains equilibrated to different P-T conditions. Other studies have produced estimates of similarly high pressures of 3.79GPa, 745°C and even up to 4.76GPa, 820°C at Nybø (Cuthbert et al., 2000 and references cited therein) suggesting the estimates are not erroneous. Studies of the Grytting eclogite and garnet websterite also have estimated P-T conditions similar to those presented here such as 3.07GPa, 688°C (Cuthbert et al., 2000 and references cited therein) although other



Figure 7.4: P-T plot showing data from this study compared against that of other studies. Sample marker formats and line colours as detailed in Table 0.13. Small, black filled circles = P-T estimates of HP-UHP 'orthopyroxene eclogites' in the WGR, data from Cuthbert et al. (2000) and references cited therein, Walsh and Hacker (2004). Small, black open circles = P-T estimates of HP-UHP Fe-Ti eclogites in the WGR, data of Cuthbert et al. (2000). Larger, black markers are locality estimates from Robinson et al. (2003), Carswell et al. (2006) and Vrijmoed et al. (2006, 2013). Small grey filled circles and grey shaded areas are P-T estimates of country rocks from Krabbendam and Wain (1997), Hacker (2007), Walsh and Hacker (2004), Kylander-Clark et al. (2008), Peterman et al. (2009). Dashed arrows are proposed P-T paths of the WGR. Coloured, dotted lines are metamorphic facies field boundaries, taken from Robertson (1999). Dashed line is the 5°C/km geotherm, lowest possible geotherm (Schreyer, 1988). Solid lines are polymorph phase boundaries (fields as labelled) and reaction equilibria. Mineral abbreviations as detailed in Table 0.11b. Coesite-quartz boundary after Bose and Ganguly (1995). Graphite-diamond boundary after Day (2012). Aluminosilicate boundaries after Holdaway (1971). Albite = jadeite + quartz equilibria from Holland (1980). Numbered grey dashed lines are melt reaction equilibria: 1) wet granite solidus, taken from Auzanneau et al. (2006) and references cited therein; 2) wet basalt solidus, taken from Gordon et al. (2013) and references cited therein; 3) phg-cpx-coe-fluid metapelite wet solidus melting reaction, taken from Schmidt et al. (2004); 4) greywacke melting, phg + cpx + grt + fluid ( $aH_2O<1$ ) = melt reaction solidus, taken from Auzanneau et al. (2006); 5) phengite melting, taken from Hermann and Green (2001); 6) KCMASH, after Hermann and Green (2001), taken from Liu et al. (2014); 7) ms + qtz = kfs + als + melt, taken from Auzanneau et al. (2006) and references cited therein.

estimates also extend up to higher P-T conditions such as 3.47GPa, 776°C (Cuthbert *et al.,* 2000 and references cited therein) (*Fig. 7.4*). However, it is not clear how directly comparable these P-T estimates are as they have been produced using different data analysis techniques.

# 7.6. Data Interpretation

### 7.6.1. <u>P-T Conditions</u>

The P-T estimations are in agreement with petrological observations of coesite at Grytting (e.g. Smith, 1984), quartz after coesite at Årsheimneset (this study) and diamond at Svartberget (Vrijmoed et al., 2008) affirming that the method of data acquisition and processing is accurate and can be relied upon when extrapolating to localities without positive mineralogical identification of UHP conditions. The trend of the P-T paths demonstrate that the rocks equilibrated to peak or post-peak temperature conditions as no increases in temperature are produced. Diffusion is greatly more dependent on temperature than pressure and diffusion rates increase with increasing temperature (Spear, 1995) so equilibration typically continues to peak temperatures. Pressures recorded therefore must be considered as minimum peak pressures, assuming clockwise P-T loops. The rocks analysed all stopped reequilibrating to changing ambient conditions within the HP eclogite-facies field as recorded by crystal rims. Sub-eclogite-facies conditions may be recorded by more altered mafic-ultramafic rocks as those analysed were chosen to yield peak conditions. It is expected therefore that the full range of P-T conditions preserved by the bodies is not represented by this study. The variation in P-T estimates of cores from the same locality and rims from the same locality is interpreted to represent reequilibration stopping at different points along the retrograde P-T paths in association with the local availability of catalysing fluid as temperatures and rates of diffusion decreased.

Considering the reliability of the data it is proposed the eastern limit of the Nordøyane UHP domain should be extended to include the Kolmannskog locality as illustrated in *Figure 7.5*.



Figure 7.5: Map showing the proposed eastern extension of the Nordøyane UHP domain to include the Kolmannskog locality. Image modified from Walsh et al. (2013). Bracketed numbers – locality 'QC' reference codes. Dashed line – current UHP domain boundary.

# 7.6.2. Origins of Garnet Websterite

All garnet websterites except for those at Myrbærneset are shown to have experienced UHP conditions and equilibrated to peak temperature conditions, including those of metasomatic origin like those at Svartberget. This demonstrates metasomatism occurred at UHP conditions and prior to peak temperatures as hypothesised. The garnet websterite body domain and vein selvage at Myrbærneset are inferred to have experienced UHP conditions considering the body's location and the thermobarometric studies of others but UHP conditions are not yielded by the samples analysed in this study.

### 7.6.3. P-T Conditions of Metasomatism

Metasomatic garnet websterites record similar peak temperature conditions as metamorphic lithologies and all rocks appear to be equilibrated to peak or post-peak temperatures. Assuming equilibration continued to peak temperature conditions, this means metasomatic rocks formed prior to peak temperature conditions and that it is not possible to determine at which point on the prograde path metasomatism occurred nor the peak pressures of metasomatism. It can be confirmed, however, that fluid-rock interaction and vein formation at Svartberget occurred at UHP conditions yielding diamond-facies estimates and minimum peak pressures of 3.85±0.27GPa at 792±9°C.

The relationship between conditions of metasomatism and those preserved in the unaltered host body rocks is inconclusive. The Arsheimneset vein selvage suggests metasomatism may occur at temperatures ~50°C higher than the conditions experienced by the main garnet websterite domain body but the metasomatic websterite at Svartberget follows a similar path as the host body. Local temperature rise above ambient conditions of the bodies is not thought to occur due to advected heat conducted into the wall-rock during fluid-rock interaction as Brady (1988) found that little temperature rise occurs near channelised fluid flow unless the channels exceed 100m-wide or there is a high and transient fluid flux with a steep thermal gradient. A difference in P-T conditions may have been recorded if diffusional equilibration in the unaltered rocks stopped earlier on the P-T path than the fluidassociated veins and metasomatic garnet websterites due to the absence of free catalytic fluid. The lower pressure conditions of the olivine garnet websterite crystal cores at Svartberget relative to the garnet websterite and the non-UHP estimates of the samples from Myrbærneset are interpreted to be due to their smaller grain sizes with crystal cores more easily re-equilibrated through retrogressive diffusion. More data from more vein systems and localities are required to gain insight into this relationship.

# 7.6.4. Regional Gradient and P-T Paths of the WGR Terrane

The data suggest the whole UHP terrane follows the same relative P-T path but at ~100°C higher temperatures in the north. This is a smaller difference than indicated

by other studies (Kylander-Clark *et al.*, 2008; Hacker *et al.*, 2010). This divergence in gradient may be due to the use of conservative calibrations in this study, the use of different thermometric techniques, different minerals recording different closure temperatures or the samples analysed have re-equilibrated to post-peak temperature conditions. The P-T paths trends towards the lower temperature estimates of the amphibole-facies country rocks. A path towards the >700°C estimates would require an increase in temperature from the conditions recorded in the crystal rims. Such a P-T path is not expected for subduction zones during exhumation suggesting the P-T estimates of the country rocks have overestimated temperatures.

Regarding pressure, this study suggests higher pressures were reached in all three UHP domains than is indicated in the current regional gradient (Hacker *et al.*, 2010) (*Fig. 1.9*) and proposed P-T paths of other studies (*Fig. 7.4*) (Peterman *et al.*, 2009; Hacker *et al.*, 2010). Similarly high pressures have been obtained at the localities studied and others by other studies, particularly at Nybø (*Fig. 7.4*) (Cuthbert *et al.*, 2000 and references cited therein).

Assuming a minimum Laurentian crustal thickness of 70km (similar to that of the Himalayas (Zheng, 2012 and references cited therein)) and an average density of 2850kg/m<sup>3</sup> for continental crust and 3300kg/m<sup>3</sup> for the mantle wedge, the maximum depth of the Svartberget garnet websterite (cores equilibrated to 3.85GPa) is 129km. Other maximum depths in the Nordøyane UHP domain are 91km at Myrbærneset and 102.97km at Kolmannskog. Maximum depths recorded in the Nordfjord-Stadlandet UHP domain at Nybø are 125km, 107km at Grytting and 119km at Årsheimneset and 122km at Remøysunde in the Sørøyane domain. These equate to thermal peak geothermal gradients of 5.9°C/km, 6.2°C/km and 5.9°C/km for Nybø, Grytting and Årsheimneset, respectively; 6.7°C/km for Remøysunde; and 7.3°C/km, 8.0°C/km and 6.2°C/km for Kolmannskog, Myrbærneset and Svartberget, respectively, demonstrating a higher geothermal gradient in the Nordøyane domain.

Retrograde paths suggest exhumation was not strictly isothermal but temperatures decreased ~100°C over the first ~1.5GPa. The paths trend towards amphibolite-facies conditions of ~0.25-1.15GPa, ~500-700°C (*Fig. 7.4*) agreeing with the pervasive retrogressive overprinting in the country rocks by an amphibolite-facies paragenesis.

Comparisons with P-T data estimated for other mafic-ultramafic bodies in the HP-UHP domains (*Fig. 7.4*) suggest the prograde path approached more from blueschistfacies conditions and the P-T loop is quite closed never attaining >800°C conditions.

# 7.7. Summary

The peak thermobarometric estimations of 3.14GPa, 665°C at Grytting, 3.53GPa, 696°C at Årsheimneset and 3.85GPa, 792°C at Svartberget agree with positive petrological observations of coesite, convincing textures after coesite and diamond in this and other studies and can be relied upon as accurate, conservative P-T estimations. All samples record UHP conditions at least in their crystal cores other than those from Myrbærneset which record HP conditions of 2.64GPa, 729°C. It is proposed that the Nordøyane UHP domain should be extended eastwards to incorporate the Kolmannskog Fe-Ti peridotite body (*Fig. 7.5*) which recorded coesite-facies UHP conditions of 3.02GPa, 754°C. The 5.5GPa pressures proposed by Vrijmoed *et al.* (2006) for the Svartberget vein system are not reproduced by this study implying such high pressures were not experienced during vein formation.

Samples are equilibrated to peak temperature conditions as no increases in temperature are produced so pressure estimates should be treated as minimum peak pressures. A difference in temperature of ~100°C between the Stadlandet-Nordfjord and Nordøyane UHP domains is apparent: more calculations must be carried out on samples from the Sørøyane domain to determine its place within the trend. The temperature estimates of the Stadlandet-Nordfjord UHP domain agree with the regional trend produced by other studies of 700-750°C (Kylander-Clark *et al.*, 2008) however the ~850°C temperatures of the Nordøyane UHP domain is not reproduced by this study: temperature estimates do not reach above 800°C.

The relationship between P-T conditions of metasomatism with those preserved by the host bodies is not clear due to inconsistent relationships between vein and body estimates; however, it is suggested metasomatism occurred prior to peak temperatures as vein domains follow the same retrograde path as mafic-ultramafic body domains. Comparisons suffer from the lack of a reliable calibration equation for a garnet-clinopyroxene barometer which would enable eclogites to be analysed and the results of the garnet-orthopyroxene barometer barometer compared.

### <u>Chapter 8 – Geochronology</u>

## 8.1. Introduction

This chapter presents U-Pb ages determined using accessory zircons, monazites and rutiles in samples from Årsheimneset and Svartberget which have well-established vein systems and garnet websterite domains. Petrographic observations and EMP data of the accessory phases are also presented.

Radiometric isotopic systems are sensitive to isotopic resetting whilst cooling through a closure temperature dependent on the mineral and isotopic system. Above the closure temperature diffusional loss occurs preventing daughter isotopes from accumulating. Diffusion of radiogenic Pb is thought to close in zircon at minimum temperatures of 900°C (MacDonald *et al.*, 2013 and references cited therein), in monazite at over 600°C or as high as 950°C (Mezger and Krogstad, 1997; Cherniak *et al.*, 2004) and in rutile down to 400°C (Cherniak, 2000; Zack *et al.*, 2011). Therefore the ages of the grains may relate to a range of temperatures but rutile ages should be younger than those of zircon and monazite. It is thought U-Pb systematics in rutile, however, may be robust to the high temperatures of UHP metamorphism (Schärer and Labrousse, 2003).

Thermobarometric estimates presented in Chapter 7 suggest temperatures remained below ~720°C at Årsheimneset and ~800°C at Svartberget. As the thermobarometric temperatures are below the resetting temperatures of zircon and possibly monazite, any concordant, Scandian ages recorded are considered to represent crystallisation ages and therefore the ages of the metasomatism, host body re-equilibration or fluid-present conditions in the country rocks rather than the timing of cooling from these events. Zircons within the host mafic-ultramafic bodies are expected to yield ages representing either the cooling from crystallisation of their magmatic protoliths or metamorphic conditions if they re-equilibrated since crystallisation. Considering the lower closure temperatures of rutile, recorded ages relate to the exhumation of the subducted terrane rather than HP-UHP conditions. Rutiles in websterites, vein cores and selvages of the veins from Svartberget should all yield ages corresponding to cooling from peak Scandian conditions as rutile is not present in the host body and so are all newly grown in association with metasomatism. Rutiles from the Årsheimneset body should also yield ages corresponding to cooling from peak Scandian conditions assuming re-equilibrated occurred during Scandian subduction.

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## 8.2. Aims and Objectives

The objective of carrying out radiometric dating is to obtain absolute ages of rock evolutionary stages. In this study it is used to determine the timing of isotopic resetting following peak metamorphism and recrystallisation of the mafic-ultramafic bodies and primarily, the timing of metasomatism and vein formation. Previous studies established the general timing of Scandian UHP metamorphism in the WGR to be 425-400Ma and exhumation-related, amphibolite-facies equilibration of the country rocks is dated at 400-385Ma (e.g. Gordon *et al.*, 2013 and references cited therein). As inclusions of polycrystalline quartz after coesite and microdiamond have been discovered in the veins at Årsheimneset and Svartberget (e.g. Carswell and Cuthbert, 2003c; Vrijmoed *et al.*, 2006), ages associated with UHP metamorphism were expected for metasomatism.

Zircon grains in the host gneisses, which are expected to reflect the evolution of the country rocks, were also analysed with the objective of comparing growth zones with zircons in the vein cores and websterites. Due to the resistant nature of zircon, they may yield dates relating to the original crystallisation of the crust-forming magmatic bodies in the Proterozoic or the many collisional and subduction events thought to have affected the rocks of the WGR up to the Scandian Orogeny (e.g. Drury *et al.*, 2001; Bingen *et al.*, 2008; Lapen *et al.*, 2009). However, outer growth zones of the zircons in the gneisses may yield Scandian ages as zircon-dissolving and -crystallising fluids must have been present in order to contain components to form the grains in the vein cores. Growth zones may also relate to a later, retrogressive episode if zircon-forming fluids continued to be present during exhumation, such as decompression melts, and therefore yield younger ages than peak conditions.

Multiple minerals were dated with the aim of obtaining ages for a range of closure temperatures and to correlate them with the P-T paths of garnet websterites constrained by thermobarometric estimates in the previous chapter: these data are used to define P-T-t paths for the two locations and calculate cooling rates.

Other aims were: to draw conclusions on the evolution of the bodies and country rocks and the source of the metasomatic agent; to add to the dataset of radiometric dates of the WGR; and to refine the regional trend in the timing of peak metamorphism and exhumation.

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## 8.3. Data Acquisition

Prior to isotopic measurements, petrographic study and geochemical analysis was carried out in order to characterise the grains to provide context to the age determinations, to associate ages with specific growth events and to determine the nature of the fluid from which the phases formed considering their component solubilities. BSE images of zircon, monazite and rutile grains and cathodoluminescence (CL) images of zircons were obtained using SEM to depict fractures and chemical zoning, respectively. BSE images were captured as described in Chapter 3 (and Appendix B2). CL imaging was carried out using a Zeiss Supra 40VP Field Emission SEM with Gatan MonoCL box at the Dalton Research Institute of Analytical Microscopy, Manchester Metropolitan University with the assistance of Dr Vladimir Vishnyakov. SMARTSEM software was used to capture images and the instrument was operated with an extra-high tension (EHT) voltage of 20kV, a 10-15nA beam current (300µm-aperture) and a working distance of ~16mm: see Appendix B7 for a full methodology. Major and trace element geochemical data used to characterise the grains were obtained using the EMP as described in Chapter 6 (and Appendix B4).

Laser ablation multiple collector inductively coupled plasma mass spectrometry (LA-MC-ICP MS) was used to obtain U-Th-Pb isotopic compositions of *in situ* zircon, monazite and rutile grains from a range of rock types from Årsheimneset and Svartberget. The principles of radiometric dating and the analysis technique are presented in the Appendix with the full methodology [B8]. All zircon and monazite crystals with growth zones large enough to be ablated and the most suitable rutile grains (large, smooth surfaces, no ilmenite blades/needles or inclusions) were selected for analysis. Carbon coats on the polished thin sections were removed and isotopic analyses were carried out using a New Wave Research nu-193nm solid state, laser ablation system coupled to a MC-ICP-MS in the Natural Environment Research Council (NERC) Isotope Geosciences Laboratory (NIGL) at the BGS Headquarters at Keyworth, Nottingham, UK with Dr Nick Roberts. The instrument was operated using the Attolab NuVu Enhanced v2.0.4 software programme. Zircons were analysed with a spot size of 25µm-diameter, monazites with 15µm-diameter and rutiles with  $35\mu$ m-diameter. Data were analysed in Microsoft Office Excel 2007 and ages were calculated using the Isoplot v4.15 Excel Add In programmed by Kenneth R. Ludwig (BGC, 2011) and with which concordia diagrams were plotted. Standard reference materials are detailed in the Appendix [B8]: they were analysed before and after every sample and/or every ~10 analyses to monitor instrument drift and correct the data.

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### 8.4. Data Presentation

# 8.4.1. <u>Petrography</u>

Tables summarising petrographic characteristics of the accessory phases are presented in *Tables 8.1-8.5*. Images displaying the various features described below are presented in *Figures 8.1-8.6*.

## 8.4.1.1. Zircon

Zircons grains in the gneisses are found sporadically distributed in the matrix and hosted within garnets: when hosted in garnet, radial fractures can be seen around the grains. In the matrix, they are mostly isolated but small composite crystals with multiple cores surrounded by a shared rim or aggregates where two or three separate crystals have grown towards each other are observed (Fig. 8.1a). Grain shapes vary and sizes range from 20µm in diameter to 70-800µm long with aspect ratios up to 2:1. Rarely, zircons host tiny inclusions and small crystals of xenotime are occasionally found attached to zircon grains (Figs. 8.1b-c). A range of fracture types is seen: some have fractures cutting across their entire width; some have rims with fine radial cracks which taper towards the core (Figs. 8.1d, f); others have fractures in their interior zones which taper within outer zones (Figs. 8.1h, j, l, v, x) or have well defined terminations at the transition to the rim zones; and some are heavily fractured obscuring zoning (Fig. 8.1b). Those with adjacent xenotime are often more heavily fractured (Fig. 8.1b). Grains typically have three zones: fractured inner zones which are darker in BSE, lighter in CL; smooth outer zones which are lighter in BSE, darker in CL; and thin rims slightly darker in BSE. In addition, some are textured and made up of overlapping, elongate strips (Fig. 8.1aa). Many also have darker, angular, elongate, euhedral, often oscillatory zoned or sector zoned cores. The lighter, ovoid overgrowths often cut across or through the cores (Figs. 8.1ac-aq).

Zircon grains found in both host eclogite and olivine garnet websterite bodies are typically rare and tiny (~10-20 $\mu$ m-diameter) (*Figs. 8.2c-e*) although one found with rutile in eclogite has a 115 $\mu$ m-diameter. In olivine garnet websterite from Svartberget they are found with rare biotite flakes and in garnet. Zircons in garnet websterites have a random, sporadic distribution and are often found with rutile,

Sample	Crystal	No.of Analyses	Rock Type*	Host	Shape	Size (µm)	No.of Zones	Euhedral Core	Zones Dated	Notes^
QC26EB	B1-A	2	qe	rt	ovoid	115	1	-	main	а
QC26EB	B20-A	1	qe	mtx"	triangular	60	1	-	main	
QC26G	В9-В	2	gw	grt	irregular	930	2	-	central	b
ARS-1	B1-A	3	gvc	bt	irregular	210	3	-	central/rim	c,d,e
ARS-1	B1-B	1	gvc	bt	irregular	120	3	-	central	c,e
ARS-1	B2-A	3	gvc	bt	irregular	200	4	х	central/rim	c,d,e,f,g
ARS-1	B5-B	1	gvc	bt	octagonal	45	3	-	central	e,h
ARS-1	B5-D	1	gvc	grt	rectangular	100	2	-	central	c,e
ARS-1	B11-D	1	gvc	qtz	circular	60	2	х	main	
ARS-1	B14-A	1	gvc	grt	ovoid	50	4	-	central/mid	i
QC26JB	B3-A	2	gvc	grt	hexagonal	105	1	-		b
QC26D	B1-A	4	g	mtx"	ovoid	165	6	?	central/mid/rim	c,e,f,g
QC26D	B3-B	2	g	mtx"	hexagon	125	3	?	central/mid	c,d,e

Table 8.1: Petrographic information of zircons analysed from samples from the Årsheimneset locality. Mineral abbreviations as listed in Table 0.11b.

\* qe - quartz eclogite; gw - garnet websterite; gvc - garnetite vein core; g - gneiss

" mtx - matrix

^ a – with rutile; b - in garnet rim; c - fractures in mid zones; d - fractures through centre; e - radial fractures in rim which stop at outer zone; f - tiny, euhedral core; g - intermittent rim; h - fractures through entire crystal; i - euhedral mid zone

Sample	Crystal	No.of Analyses	Rock Type"	Host	Shape	Size (µm)	No.of Zones	Euhedral Core	Zones Dated	Notes^
QC36GA	B1-A	5	gvc	bt	irregular	170	2	-	main	а
QC36GA	B2-A	1	gvc	grt	ovoid	70	3	-	main	а
QC36GA	B2-E	2	gvc	grt	irregular	90	5	х	outer/inner mid	b
QC36GA	B3-B	6	gvc	bt*	rectangular	110	3	х	core/mid	а
QC36GA	B7-A	1	gvc	grt	ovoid	80	4	Х	core	с, е
QC36GA	В7-В	2	gvc	grt	irregular	75	4	Х	main	а
QC36GA	B7-G	1	gvc	grt	circular	70	3	Х	mid/rim	а
QC36GA	B8-D	1	gvc	bt	hexagon	75	3	Х	central	
QC36GA	В9-В	4	gvc	grt	ovoid	110	5	х	central/mid/outer mid/rim	
QC36GA	B9-C	5	gvc	grt	circular	120	5	х	core/central/inner/ outer/rim	
QC36GA	В9-Е	1	gvc	grt	circular	50	4	Х	rim/outer mid	d
QC36GA	B9-F	1	gvc	grt	ovoid	90	4	Х	core	с
QC36GA	B10-A	3	gvc	grt	ovoid	110	3	Х	core/mid	с
QC36GA	B7-D	4	gs	bt	irregular	120	4	х	core/main (central)/outer	f
QC36GA	B7-F	1	gs	bt	ovoid	80	2	х	core	С
QC36GA	B2-B	2	gs	bt	ovoid	115	7	х	core/outer core	
QC36GA	B2-C	2	gs	bt	ovoid	150	2	х	core/outer core	с
SV1A	В7-В	2	gw	орх	rectangular	125	2	-	central	g
QC36J	B11-A	1	g	mtx'	hexagonal	100	4	Х	core	c, f, h
QC36J	B11-B	1	g	pl	hexagonal	40	3	Х	core/central	
QC36J	B12-C	1	g	mtx'	ovoid	100	4	Х	core/central	f, h
QC36J	B13-A	1	g	mtx'	ovoid	80	2	-	main (central)	i
QC36J	B13-B	1	g	mtx'	hexagonal	70	4	?	central/core	j
QC36J	B14-A	1	g	mtx'	ovoid	115	3	-	central	b, f, j
QC36J	B14-B	2	g	mtx'	ovoid	90	3	Х	core/mid	b
QC36J	B15-A	1	g	qtz	rectangular	65	4	Х	core	С
QC36J	B16-A	1	g	qtz	pear	70	3	Х	core/mid	
QC36J	B17-A	2	g	mtx'	ovoid	135	4	Х	core/rim	

Table 8.2: Petrographic information of zircons analysed from samples from the Svartberget locality. Mineral abbreviations as
listed in Table 0.11b.

" gvc - garnetite vein core; gs - glimmerite selvage; gw - garnet websterite; g - gneiss

\* inclusion in garnet

' mtx - matrix

^ a - feint zoning pattern; b - zoned core; c - oscillatory zoning in core zone; d - euhedral zones; e - fractures in outer zone; f - fractures through centre; g - fractures through entire crystal; h - fractures taper into rim; i - ablation pit hits second zone at depth; j - radial fractures in rim which stop at outer zone

Table 8.3: Petrographic information of monazites analysed from samples from the Svartberget locality. Mineral abbreviations as listed in Table 0.11b.

Sample	Crystal	No.of Analyses	Rock Type*	Host	Shape Size (µm)		No.of Zones	Core	Zones Dated	Notes^
QC36GA	B3-A	1	gvc	grt	irregular 90		4	х	central	
QC36GA	B8-C	1	gvc	bt	pear	85	3	-	central	а
QC36GA	B9-D	2	gvc	grt	irregular	70	3	-	central	b
QC36GA	B11-B	3	gvc	grt	rectangular	170	2	-	central	
QC36GA	B3-C	1	gs	bt	ovoid	90	2	-	central	
QC36GA	B7-C	1	gs	bt	irregular	120	3	-	central	c,d
QC36J	B8-D	2	g	pl	irregular	85	2	-	central	е
QC36J	B10-A	3	g	mtx"	irregular	180†	2		inner/rim	b
QC36J	B11-D	4	g	mtx"	irregular	135	2		rim	b
QC36J	B13-C	1	g	mtx"	irregular	40	4	-	mid	
QC36J	B14-D	2	g	qtz	rectangular	45	m'	-	central/outer	f
QC36L	B1-A	6	lg	mtx"	irregular	170	m'	-	various	f
QC36L	B1-B	1	lg	mtx"	irregular	110	m'	-	various	f
QC36L	B1-B	1	lg	mtx"	irregular	75	m'	-	various	f
QC36L	B1-B	2	lg	mtx"	irregular	105	m'	-	various	f
QC36L	B1-B	1	lg	mtx"	irregular	95	m'	-	various	f
QC36L	B2-A	4	lg	mtx"	elongate pentagon	330	2	-	inner/rim	f

\* gvc - garnetite vein core; gs - glimmerite selvage; g - gneiss; lg - leucogneiss " mtx - matrix ' m - multiple

† aggregate

^ a - fractures through entire crystal; b - feint zones; c - zircon inclusion; d - fractures in central zone; e - inclusions; f - patchy zoning

Arsheimneset loculity. Milleral abbreviations as listed in Table 0.11b.										
Sample	Crystal	No.of Analyses	Rock Type*	Host	Shape	Size (µm)	Notes^			
OC26FB	B1-B	2	de	mtx"	irregular	300				
OC26FB	B1-C	2	ae	срх	irregular	260				
OC26EB	B5-A	1	0e	cnx	ovoid	145	а			
OC26EB	B6-B	3	de de	срх	ovoid	195	u			
OC26EB	B6-C	2	de de	cpx	nentagonal	130				
OC26EB	B15-D	2	qe	cpx	ovoid	255	Э			
OC26EB	B15-D	2	qe	cpx	ovoid	1/6	a			
		2	qe	срх	irrogular	240				
	D10-D	2	qe	cµx mtv"	hovagonal	240				
	D2U	2	qe	mtv"	irrogular	290				
QCZOEB	BZZ-B	3	qe	mtv"	irregular	210				
A347	BT-R	2	gw	mtx	ovoid	210	_			
A347	B3-C	2	gw	am	ovoid	245	а			
A347	B4-B	3	gw	grt	irregular	270				
A347	B2-C	2	gw	grt	irregular	600				
A347	B8-A	2	gw	bt	ovoid	290				
A347	B9-A	2	gw	grt	irregular	265				
A347	B9-C	1	gw	grt	ovoid	370				
A347	B9-E	2	gw	grt	irregular	480				
A347	B9-F	2	gw	am	irregular	330				
A347	B10-A	3	gw	mtx"	hexagonal	400	а			
A347	B12-C	2	gw	срх	irregular	690				
QC26G	B11-B	1	gw	am	irregular	420				
QC26G	B27-D	2	gw	grt	irregular	235				
QC26G	B28-A	1	gw	орх	ovoid	215				
QC26G	B29-A	2	gw	grt	irregular	275				
QC26G	B30-A	2	gw	grt	irregular	315				
QC26P	B1-A	2	gw	am	ovoid	195				
QC26P	B3-A	2	gw	grt	square	125				
QC26P	B4-B	1	gw	mtx"	irregular	105				
QC26P	B5-B	2	gw	mtx"	octagonal	150				
QC26P	B7-B	2	gw	am'	irregular	145				
ARS-1	B4-C	3	gvc	am	irregular	375				
ARS-1	B10-B	3	gvc	grt	irregular	280	a,b			
ARS-1	B11-C	3	gvc	qtz	pear	560				
ARS-1	B15-A	3	gvc	mtx"	irregular	630				
ARS-1	B15-B	3	gvc	срх	irregular	660				
QC26HA	B19-D	1	gvc	grt	irregular	185	а			
QC26JA	B1-A	3	gvc	bt	irregular	535	а			
OC26JA	B3-A	1	gvc	grt	irregular	425	a.c			
OC26IB	B1-A	2	gvc	mtx"	triangular	275	а			
0C26HA	B23-Δ	1		cnx	ovoid	190	a			
0C26HA	B23 /	3	CS	срх	irregular	265	a			
	R4-R	2	C5	cpy	ovoid	205	u			
	B4°D B10₋C	<u>~</u> л	GMC	mtv"	elongate	51200	2			
	B10-C	+ 1	gws	CDV	ovoid	24000 160	a			
	D12-D	1 2	gws	cµx m+v"	irregular	200	h			
	010-0	<u>د</u>	8ws	t	inegulai	200	U			
	020	4	gvvo	gιι						

Table 8.4: Petrographic information of rutiles analysed from samples from the Årsheimneset locality. Mineral abbreviations as listed in Table 0.11b.

\* qe - quartz eclogite; gw - garnet websterite; gvc - garnetite vein core; cs -

clinopyroxenite selvage; gws - garnet websterite selvage

" mtx - matrix

' inclusion in garnet

^ a - exsolutions; b - in garnet rim; c - with quartz
Table 8.5: Petrographic information of rutiles analysed from samples f	from
the Svartberget locality. Mineral abbreviations as listed in Table 0.11k	э.
	-

Sample	Crystal	Analyses	коск Type*	Host	Shape	size (μm)	Notes^		
QC36D		1	gw	grt	triangular	145			
QC36D		1	gw	орх	ovoid	190			
QC36D		1	gw	орх	irregular	150			
QC36D		1	gw	орх	irregular	180			
QC36D		2	gw	срх	irregular	260			
QC36D		1	gw	срх	ovoid	170			
QC36D	B12-H	1	gw	срх	irregular	115			
QC36D	B16-C	1	gw grt		irregular	170			
QC36D	B17-A	1	gw	орх	rectangular	300			
QC36D	B17-B	1	gw	mtx"	irregular	290			
QC36EA	B4-C	2	gvc	grt	square	150			
QC36EA	B5-D	1	gvc	grt	ovoid	125	а		
QC36EA	B5-E	1	gvc	срх'	irregular	145			
QC36EA	B5-G	1	gvc	срх	ovoid	160			
QC36EA	B8-A	1	gvc	grt	ovoid	140	b		
QC36EA	B8-B	1	gvc	grt	ovoid	160	а		
QC36EA	B9-E	2	gvc	grt	irregular	350	С		
QC36EA	B10-A	3	gvc	срх'	rectangular	220			
QC36EA	B10-B	2	gvc	grt	ovoid	225	а		
QC36EA	B11-B	2	gvc	срх	ovoid	155			
QC36EA	B11-C	2	gvc	срх	ovoid	260	d		
QC36EA	B11-D	1	gvc	срх	triangular	135			
QC36GA	B7-H	3	gvc	grt	pear	200	а		
QC36GA	B7-I	3	gvc	grt	irregular	360	С		
QC36GA	B7-J	12	gvc	grt	rectangular	650			
QC36GA	В7-К	2	gvc	grt	ovoid	320			
QC36GA	B16-A	1	gvc	grt	ovoid	85	а		
QC36GA	B17-A	1	gvc	grt	ovoid	270	а		
QC36GA	B17-B	2	gvc	grt	ovoid	190	а		
QC36GA	B18-A	3	gvc	grt	circular	180			
QC36GA	B18-B	3	gvc	grt	pear	450	С		
QC36GA	B20-C	2	gvc	grt	pear	145			
QC36GA	B23-A	2	gvc	grt	ovoid	140	а		
QC36FA	B1-A	1	gcs	срх	rectangular	230			
QC36FA	B1-B	2	gcs	срх	ovoid	275			
QC36FA	B4-A	2	gcs	grt	rectangular	210	b		
QC36FA	B6-B	2	gcs	срх	irregular	265			
QC36FB	B2-A	2	gcs	am	rectangular	145			
QC36FB	B2-B	1	gcs	срх	pear	125	е		
QC36FB	B3-A	1	gcs	am	pear	245	е		
QC36FB	B5-A	1	gcs	am	rectangular	160	е		
QC36FB	B8-B	2	gcs	mtx"	irregular	310	е		
QC36GA	B29-A	2	CS	срх	ovoid	100	f		

\* gw - garnet websterite; gvc - garnetite vein core; gcs - garnet

clinopyroxenite selvage; cs - clinopyroxenite selvage

" mtx - matrix

' inclusion in garnet

^ a - in garnet rim; b - with clinopyroxene inclusion; c - with biotite inclusion;

d - near altered fracture; e - exsolutions; f - with quartz

ilmenite or hosted within clinopyroxene, garnet, orthopyroxene, biotite, amphibole and rutile with no notable preference. Grains are generally ovoid-spherical (*Figs. 8.2fj,m-o*) although some are more rectangular with straight faces and rounded corners (*Figs. 8.2k,I*). They typically have 10-20µm diameters but grains with 100µm diameters are rarely present. Aggregate grains are also seen (*Figs. 8.2i,m,o*). Most are featureless with no zoning although some have radial fractures around their rims (*Figs. 8.2j,n*), and central zones and small cores (*Figs. 8.2d,n*). Those hosted in amphibole are often more heavily fractured (*Fig. 8.2g*).

In vein cores, zircons are hosted in garnet, particularly in garnet rims, biotite, clinopyroxene, quartz and are often found adjacent to or within rutile or monazite grains. Grains are also found in late amphibole demonstrating their resistance to retrograde metamorphism. Grain diameters range from 10µm (*Fig. 8.3a*) to 150µm (*Figs. 8.3b,c*) but are typically 30-100µm: larger crystals are generally hosted in biotite or in biotite in garnet. Most are ovoid-spherical (*Figs. 8.3d,e*) although some are more euhedral (*Fig. 8.3f*) or subhedral with flat faces and rounded corners (*Fig. 8.3g*). Zoning varies from poorly- to well-developed and typically consist of a finely zoned, small core, wide inner and/or outer zones and a thin rim. Some have oscillatory-zoned, euhedral cores which can be the dominant zone of the grain (*Figs. 8.3s-v*) and some have more simple (*Figs. 8.3w-y*) or irregular zoning patterns (*Figs. 8.3z-ab*). Rims and outer zones often contain radial fractures and some inner zones have fractures which taper towards the rim. Zircons are also present in MPSIs (*Fig. 8.3a*) and some have radial fractures when hosted in garnet (*Fig. 8.3d*).

## 8.4.1.2. Monazite

Monazite grains in the gneisses occur as irregular, delicate, anhedral-interstitial aggregates (*Figs. 8.4a-e*) and individual, subhedral crystals within the matrix (*Figs. 8.4i*) and hosted in biotite. Crystals are typically strongly zoned (*Figs. 8.4c-e*) and some have patchwork patterns (*Fig. 8.4h*). Those in vein cores occur in zircon-rich patches hosted in garnet and biotite. These grains are more regularly shaped and simply zoned (*Figs. 8.4j-m*).



Figure 8.1: BSE and CL SEM images of zircons in gneisses from Årsheimneset and Svartberget. Red circles are LA-MC-ICP-MS analysis sites. a) composite zircon with three cores [QC26A-B2-C]; b-c) zircons with adjacent xenotime crystals [QC26A-B6-A, QC36K-B5-C]; d-g) BSE and CL images of zircons with radial fractures in their rims [QC26D-B1-A, QC36J-B6-A]; i-m) BSE and CL images of zircons with finite fractures within their inner zones [QC36J-B2-B, QC36J-B4-B, QC36J-B1-C]; n-z) BSE and CL images of zoned zircons with faceted cores [QC36J-B1-A, QC36J-B2-A, QC36J-B3-B, QC36J-B5-A, QC36J-B9-B, QC36J-B17-B, QC36K-B3-B]; aa-ab) BSE and CL images of an unzoned zircon [QC36J-B4-A]; ac-ag) BSE and CL images of zircons with oscillatory-zoned cores [QC36J-B15-A, QC36J-B6-C, QC36J-B8-A].



Figure 8.2: BSE SEM images of zircons in host bodies and websterites from Årsheimneset and Svartberget. Red circles are LA-MC-ICP-MS analysis sites. a) elongate zircon in eclogite [QC26EB-B18-A]; b) triangular zircon in eclogite with ablation pit [QC26EB-B20-A]; c-d) ovoid zircon in multiphase inclusion in garnet in eclogite with two growth zones [A632-B8-B]; e) small zircon in olivine garnet websterite [QC36A-B4-2]; f) zircon hosted in garnet in a garnet websterite from Årsheimneset [QC26G-B7-E]; g) zircon hosted in amphibole in a garnet websterite from Årsheimneset [QC26G-B7-E]; h) small zircons in garnet in garnet websterite from Svartberget [QC36D-B6-E]; i) composite zircon in garnet websterite from Svartberget [QC36D-B9-C]; j) small, zoned zircon in garnet websterite from Svartberget [SV1A-B7-B]; k-l) large, rectangular zircon in garnet websterite from Svartberget [SV1A-B7-B]; m) small, composite zircon in websterite selvage [QC36GB-B5-D]; n) spherical zircon with core and radial fractures in the rim and outer zones [QC36EB-B3-A]; o) composite, zoned and tiny zircons in websterite selvage [QC36EC-B4-B].



Figure 8.3: BSE and CL SEM images of zircons in vein core garnetites from Årsheimneset and Svartberget. Red circles are LA-MC-ICP-MS analysis sites. a) small zircon in multiphase inclusion in garnet [QC36GA-B4-B]; b-c) large, poorly-zoned zircon hosted in biotite [QC36GA-B1-A]; d-e) spherical zircon with a euhedral core and radial fractures in the host garnet [QC36GA-B5-B]; f) euhedral, octagonal zircon with radial fractures in the rim zone [ARS-1-B5-B]; g) triangular zircon with a euhedral core and radial fractures in the rim zone [ARS-1-B1-C]; h-r) BSE and CL images showing the typical zoning patterns and sizes of zircons [QC36GA-B7-G, QC36GA-B8-D, QC36GA-B9-B, QC36GA-B9-C, QC36GA-B9-F, QC36GA-B10-A]; s-v) zircons with dominant oscillatory-zoned cores [QC36GA-B7-F, QC36GA-B2-C]; w-y) more simply zoned zircons [QC36GA-B7-D, QC26JA-B2-A]; z-ab) irregularly-zoned zircons [QC36GA-B7-F, ARS-1-B1-A].



Figure 8.4: Photomicrographs and BSE SEM images of monazites in gneiss, leucogneiss and garnetite vein core samples from Svartberget. Red circles are LA-MC-ICP-MS analysis sites. a-b) photomicrographs of anhedral monazites with biotite in gneiss [QC36J] PPL-XPL, FOV 1.64mm; c) BSE SEM image of a patch-zoned monazite in leucogneiss [QC36L-B1-A]; d) BSE SEM image of monazite in gneiss [QC36J-B10-A]; e) BSE SEM image in monazite in leucogneiss [QC36L-B1-B]; f) BSE SEM image of more simply zoned monazite in gneiss [QC36J-B40-B]; g) BSE SEM image of a monazite in gneiss [QC36J-B40-B]; f) BSE SEM image of monazite in gneiss [QC36J-B40-B]; j) BSE SEM image of a monazite in the matrix of a leucogneiss showing feint, patchy zoning [QC36L-B2-A]; j) feint patch zoning in monazite in garnetite vein core [QC36GA-B9-D]; k) simply zoned monazite which includes zircons in garnetite vein core [QC36GA-B7-C]; l) simply zoned monazite in garnetite vein core [QC36GA-B8-C].

#### 8.4.1.3. <u>Rutile</u>

Rutile grains occur well-distributed throughout the Årsheimneset eclogite and garnet websterite and garnetite vein cores from both localities but are rare within clinopyroxenite vein domains. They are not present in the gneisses or olivine garnet websterite from Svartberget. In vein cores rutile is hosted in garnet and small grains are common in trails along the edge of the garnetite core. Rutiles hosted in garnets range from 660µm-diameter sizes (*Figs. 8.5a-d*) to tiny, clustered, randomly arranged, elongate grains (*Figs. 8.5e-f*). Grain shapes vary from rounded and ovoid (*Fig. 8.5g*) to interstitial and highly irregular (*Fig. 8.5h*) to granoblastic (*Fig. 8.5i*). Many crystals have blades of ilmenite cutting through them (*Figs. 8.6a-b*), some host small zircon grains and other inclusions (*Figs. 8.6b-d*) and some are featureless (*Fig. 8.6e*). Retrogressive features include replacement by ilmenite (*Fig. 8.6f*), exsolved ilmenite needles (*Fig. 8.6g*) and titanite rims (*Fig. 8.6h*); these are mostly found in those within symplectitic clinopyroxenes. Many have pitted surfaces precluding them being suitable for EMPA (*Fig. 8.5b*).

# 8.4.2. Mineral Chemistry

Representative EMP data are presented in *Table 8.6*: all data are presented in the Appendix [K1-K5]. Analysis points presented have wt% totals between 99.00 and 100.99wt% and are above their respective detection limits (see Appendix G1 for average values): all analyses of monazites from the leucogneiss from Svartberget [QC36L] are outside the acceptable wt% total range. Average one sigma errors of each element in each phase are presented in the Appendix [K6].

Unfortunately EMP analysis was largely unsuccessful as most trace element contents are below their respective detection limits or registered as a zero; for these reasons, few analyses produce useable trace element ratios with which to characterise the crystals. Pb, Th and U contents were also measured during LA-MC-ICP-MS and are presented with the radiometric data in the Appendix [L1-L5]. *Figures 8.7-8.10* display trace element analyses of the accessory phases from both analysis methods.



Figure 8.5: Photomicrographs and BSE SEM images of rutiles in samples from Årsheimneset and Svartberget showing the range in sizes and shapes. Red circles are LA-MC-ICP-MS analysis sites. a) PPL photomicrograph of a large rutile with zircon in a quartz eclogite [QC26EB-7A, FOV 1.64mm]; b) BSE SEM image of 'a' with zircon (with ablation pits) [QC26EB-B1-B]; c) PPL photomicrograph of a large, elongate rutile in a garnet websterite selvage [QC26HC-9A, FOV 4.2mm]; d) BSE SEM image of a large rutile with fine ilmenite lamellae, inclusion and four 35µmdiameter ablation pits [QC36GA-B7-J]; e-f) PPL photomicrographs of tiny rutile needles in garnet in garnetite vein cores [QC36EA-1A, FOV 4.2mm; QC36GA-5A, FOV 4.2mm]; g-i) BSE SEM images showing the range in grain shape from ovoid to interstitial to equant-granoblastic [QC36EA-B5-D; A347-B12-C; QC36EA-B4-C].



Figure 8.6: BSE SEM images of features of rutiles in samples from Årsheimneset and Svartberget. Red circles are LA-MC-ICP-MS analysis sites. a) rutile with ilmenite blades [A632-B1-D]; b) rutile with ilmenite blades, fine lamellae and a zircon inclusion [QC26HA-B19-D]; c) rutile with a zircon inclusion [A602-B1-D]; d) rutile with inclusions [QC26EB-B22-A]; e) featureless rutile [QC36GA-B23-A]; f) rutile being replaced by ilmenite [QC26G-B7-B]; g) rutile with orientated ilmenite exsolution needles [QC36FB-B3-C]; h) rutile with a titanite rim in clinopyroxene symplectite [QC26HA-B24-B].

Mineral		zircon	,	,,		monazite		rutile	
Locality"		Å	Å	S	S	S	S	Å	S
Rock Type	*	gvc	gw	g	gvc	g	g	gvc	bcs
Sample		ARS-1	QC26G	QC36J	QC36EA	QC36J	QC36J	QC26HA	QC36GA
Crystal		B1-A	B9-B	B2-A	B9-D	B10-A	B12-A	B19-D	B29-A
Analysis Po	oint	2	1'	6'	2'	3	1	3	2
Zone^		inner	inner	core	outer				
SiO <sub>2</sub>	wt%	30.80	30.097	31.84	29.74	0.20	0.71	-	-
TiO <sub>2</sub>		-	-	-	-	-	-	98.34	96.17
$Al_2O_3$		-		-	-	-	-	-	-
$Cr_2O_3$		-						0.46	0.68
FeO		-	0.32	0.39	0.60	-	-	0.23	0.45
MnO			-	-	-	-	-		
MgO			-	-	-	-	-		
CaO		-	-	-	0.06	1.27	0.05	-	0.07
Na <sub>2</sub> O			-	-	-	0.15	-		
K₂O			-	-	-	-	-		
P <sub>2</sub> O <sub>5</sub>			-	-	-	29.82	34.51		
ZrO <sub>2</sub>		65.81	68.63	65.59	66.43	-	-	-	-
HfO <sub>2</sub>		1.66	0.721	1.26	1.34	-	-	0.92	0.92
		-						0.13	0.86
		-					0.12	0.22	0.18
		-	-	- 0 17	-	-	0.12	-	-
		-	-	0.17	0.02	1.91	0.51	-	-
		0.29	-	0.25	-	12 02	1.70	-	-
		-		_	_	13.95	- 0.14	-	-
$Pr_2O_3$			_	_	_	27.47	-		
Nd <sub>2</sub> O <sub>2</sub>			-	-	-	11 27	0.73		
SmO		-	-	-	-	2.01	0.29	-	-
EuO				-		-	0.65		
Gd <sub>2</sub> O <sub>3</sub>				0.22		1.83	3.82		
Tb <sub>2</sub> O <sub>3</sub>				-		-	0.52		
Dy <sub>2</sub> O <sub>3</sub>				-		1.06	5.24		
Ho <sub>2</sub> O <sub>3</sub>				-		-	-		
$Er_2O_3$				-		-	4.19		
$Tm_2O_3$				-		-	-		
Yb <sub>2</sub> O <sub>3</sub>		0.30	-	-	0.09	-	3.56	-	-
$Y_2O_3$		0.39	-	0.27	0.76	3.53	41.62	-	-
Lu <sub>2</sub> O <sub>3</sub>		-	-	-	-	-	1.11	-	-
TULAI	- 	1226.2	99.77	1140.1	33.05	39.24	9009	100.29	33.33
U Th	ррпі	1320.3		1140.1	195	16740	8008		
III Dh		1510 7		1407.5	165	10749	4475		
PU V		1519.7		1074 9	2002	1201/	162991		
י בו				1074.0	2992	50303	103881		
Ce						117264	598		
Nd						48320	3108		
Sm						15251	2160		
Yb		1321.7			413		15641		
Lu					-		4898		
Nb								524.51	3413.27
Та								884.48	753.44
Th/U				1.29		2.23	0.56		
Ce/Y						8.43	0.004		
Nb/Ta								0.59	4.53

Table 8.6: Representative EMP data of accessory phases

" Å – Årsheimneset; S - Svartberget
\* gvc - garnetite vein core; g - gneiss; gw - garnet websterite; bcs - biotite clinopyroxenite selvage

'green numbers on printed images

^ zones - core, inner, outer, rim, u - unzoned

dash - below detection limit

blank - not analysed

## 8.4.2.1. Zircon

Many EMP analyses yielded Y values, which behaves similarly to HREE. Most are below 1500ppm although some have contents up to 3197ppm (*Fig. 8.7*). Y contents do not particularly differentiate between different growth zones: cores yield a wide range of contents and other growth zones all have a similar range and cluster between 300-900ppm. There are no strong or particular trends with Th or U contents and nothing can be discerned against Th/U ratios as only two points can be plotted. Ti contents are mostly below their detection limits (average 571ppm) but range from 360 to 1816ppm.

Only three EMP zircon analyses yielded Th/U ratios (*Figs. 8.8a,b*). The average EMP detection limit of U in zircon is 1861ppm and of Th is 683ppm (see Appendix D2.1). Zircons in the gneiss and leucogneiss have the highest concentrations of U and Th. The lower U and Th contents of gneiss zircons (from middle growth zones) are similar to the higher U and Th contents of vein core zircons. The highest U contents of up to 4478ppm are found in xenocrystic cores and middle growth zones of zircons from the leucogneiss at Svartberget (*Fig 8.8b*). Highest Th contents of up to 1468ppm are found in xenocrystic cores from a gneiss at Svartberget (*Fig 8.8b*) but some cores have lower Th contents like other growth zones. Cores, central zones, middle growth zones, rims and unzoned crystals have overlapping U contents.

LA-MC-ICP-MS analysis yielded lower concentrations of U and Th in zircon than EMP analysis (*Figs. 8.8c,d*). Highest U contents of up to 1816ppm are found again in cores from the Svartberget gneiss and middle growth zones from vein core zircons from Årsheimneset. Highest Th contents of up to 373ppm are found in xenocrystic cores from vein core and gneiss zircons from Svartberget. All analyses with Th contents over 50ppm plus one rim analysis from Årsheimneset are from xenocrystic cores from both the host gneisses and garnetite vein cores and have Th/U ratios above 0.1 (*Fig 8.8d*): three analyses have Th/U  $\ge$  0.4.

Unzoned zircons and growth zones of zoned zircons from all rock types, except for the Årsheimneset rim, all have low Th contents and Th/U ratios below 0.1. The zircon rim with the high Th content [ARS1-B1-A] is from a fractured rim (see *Table 8.1; Figure 8.3ab*). Th may accumulate in fractures during preferential expulsion from the



Figure 8.7: Y-Th-U plots showing EMP data of zircon crystals split into rock types (a,c) and growth zones (b,d). a-b) Y-U plots; c-d) Th-Y plots. Marker formats for a and c as detailed in Table 0.13. Marker formats for b and d: dark blue – central growth zone; mid blue – outer growth zones; light blue – rim zone; purple – xenocrystic cores.



Figure 8.8: Th-U plots showing mineral chemistry data of zircon crystals split into rock types (a,c) and growth zones (b,d). a-b) EMP data; c-d) LA-MC-ICP-MS data. Marker formats for a and c as detailed in Table 0.13. Marker formats for b and d: dark blue – central growth zone; mid blue – outer growth zones; light blue – rim zone; purple – xenocrystic cores. Th/U ratio lines as labelled.

lattice due to its larger size (Zack *et al.*, 2011); although care was taken to avoid fractures one may have been present in the analysed volume.

#### 8.4.2.2. Monazite

There are two distributions of Y concentrations in monazites in the gneiss from Svartberget; below 2,000ppm (0.2wt%) and ~160,000ppm (16.0wt%) demonstrating both Y-rich and Y-poor monazites are present. Y contents of monazites in the vein core and glimmerite selvage are below their detection limits (average 2089ppm). LREE (La, Ce) and MREE (Nd, Sm) are well clustered between the vein and the gneiss: the vein core and the glimmerite selvage have overlapping concentrations. Ce in monazites has similar compositions and range up to ~14.5wt% in the vein and ~12.2wt% in the gneiss. There is a slight negative correlation between Ce and Th and Ca and Th in the vein monazites. EMP U-Th data yields few data points on the Th-U plot as few have U contents over the detection limit. Th contents range up to 6.8wt% and U contents up to 1.24wt%. LA-MC-ICP-MS data yield considerably lower concentrations of up to 1.4wt% Th and 0.12wt% U. Vein monazites have much lower Th and U contents than EMP analysis with a range of 1172-3590ppm Th. There are three groupings of U-Th contents in the LA-MC-ICP-MS data: low U-low Th consisting of monazites from all four rock types analysed; mid U-high Th and high U-low Th which both consist of monazites from the gneiss and leucogneiss.

## 8.4.2.3. <u>Rutile</u>

Few EMP analyses of rutiles yielded trace element values over their detection limits except for Hf and Nb. Hf contents range between ~7100 to 8700ppm in most rock types from both localities. There are no trends with  $TiO_2$  or Nb. Ta concentrations above their detection limits range from ~700-1000ppm. Nb values range from ~250-5300ppm (*Fig. 8.10a*). Lowest Nb contents are found in eclogites, websterite domains and vein selvages from Årsheimneset. Highest Nb contents are found in rutiles in clinopyroxene-rich garnetite vein cores from Svartberget; these sit over 1000ppm above the Nb contents of rutiles in websterites from Svartberget which range up to 2185ppm. Nb contents of rutiles from samples from Årsheimneset are mostly below <1000ppm. There is no particular trend in Nb/Ta ratios in the analyses that yield both values. The analyses of vein samples which yield Nb/Ta values have strongly



Figure 8.9: Plots showing mineral chemistry data of monazite grains. Marker formats as detailed in Table 0.13. a) Ce-Th plot, EMP data; b) Ce-Y plot, EMP data; c) La-Sm plot, EMP data; d) CaO-Th plot, EMP data; e) Th-U plot, EMP data; f) Th-U plot, LA data.



Figure 8.10: Plots showing mineral chemistry data of rutile grains. Marker formats as detailed in Table 0.13. a) Nb-Ta plot, EMP data. Solid line = chondritic Nb/Ta (values from McDonough and Sun, 1995); b) Th-U plot, LA-MC-ICP-MS data.

sub-chondritic ratios (<17.14). Zr/Hf ratios are also strongly subchondritic (<37.09) ranging from 0.133 to 0.175.

Rutiles contain very little if any Th: the concentrations recorded by LA-MC-ICP-MS analysis are below 2.5ppm (*Fig. 8.10b*). U contents from LA-MC-ICP-MS analysis range from 35-849ppm. Svartberget rutiles are more U-rich than those in Årsheimneset. Highest U contents at each locality are in rutiles from garnetite vein cores.

## 8.4.3. <u>U-Pb Isotope Data</u>

Element and isotopic data obtained from the grains analysed are presented in the Appendix [L1-L5]. Ages of standard reference materials are presented in the Appendix [L6]. Plots of age data are presented in *Figures 8.11-8.14, 8.15* and *8.16-8.17* for zircons, monazites and rutiles, respectively. Ages quoted in the text are weighted mean  $^{206}$ Pb/ $^{238}$ U ages for zircons and weighted mean corrected  $^{238}$ U/ $^{206}$ Pb ages for monazites and rutiles (corrected for common lead): all ages are reported in *Table 8.7* with their mean square weighted deviations (MSWD), probabilities and the number of analyses upon which the ages are calculated. Quoted errors are two-sigma errors for zircons and 95%-confidence levels for monazite and rutile analyses. Discordant and anomalous analyses were discarded until the MSWD was reduced to below 2.0: those based on only a few analyses have larger errors. Plots showing the analyses used to obtain ages for each rock type and those discarded are presented in the Appendix [M1-M5].

# 8.4.3.1. *Zircon*

Many zircon analyses from both localities yield discordant ages. These are typically from grains with fractured zones (as detailed in the petrographic tables) although there is no stead-fast correlation. The discordant growth zones of zircons from Årsheimneset are from the central zones of vein core and websterite zircons (*Figs. 8.11a, 8.13a*) and those from Svartberget are identified xenocrystic cores in vein and gneiss zircons (*Figs. 8.12a, 8.13b*) many of which retain high Th/U ratios (>0.1) (*Fig. 8.14b*). Projected upper intercepts at both Årsheimneset and Svartberget lie near 1.5Ga and older.

Mineral		Zircon	<i>joi acces</i>		attaniet		Monazite			peanary		Rutile	- mada t			
Mean Age (Ma)		206 /238	2σ	MSWD	Prob	No.	Corrected	95%	MSWD	Prob	No.	Corrected	95%	MSWD	Prob	No.
Rock Type	Sample	Pb/ 0	(Myr)	*	^		<sup>238</sup> U/ <sup>206</sup> Pb	(Myr)	*	^	"	<sup>238</sup> U/ <sup>206</sup> Pb	(Myr)	*	^	
Årsheimneset																
gneiss	QC26D	417.0	10.0			1										
quartz eclogite	QC26E	411.4	5.3	0.28	0.60	2						408.1	2.8	0.78	0.71	18
garnet websterite																
	A347											410.7	3.9	1.90	0.013	18
	QC26G											392.7	5.0	0.90	0.46	5
	QC26P											419.3	7.3	1.60	0.13	7
garnetite vein core		414.0	5.6	0.39	0.68	3										
	ARS-1											396.5	4.1	0.78	0.64	10
	QC26H/J											401.5	6.6	1.70	0.12	5
clinopyroxenite selvage	9															
	QC26H											401.6	5.4	0.66	0.52	3
	QC26J											415.1	8.5	0.101	0.75	2
websterite selvage	QC26H											384.7	3.9	0.57	0.80	9
Svartberget																
gneiss	QC36J	401.1	2.7	0.40	0.92	9	389.1	2.9	1.2	0.30	10					
leucogneiss	QC36L						398.8	3.0	0.5	0.84	7					
garnet websterite	QC36D	411.3	8.9	1.30	0.26	2						419.0	13.0	1.70	0.16	4
garnetite vein core	QC36E											388.1	3.5	1.13	0.34	9
	QC36G	410.6	2.6	1.17	0.27	20	409.2	6.4	2.5	0.01	9	381.6	2.8	1.50	0.053	26
clinopyroxene garnetit	e vein core															
	QC36F											407.0	5.2	0.74	0.56	5
	QC36F											416.6	4.1	0.44	0.78	5
glimmerite selvage	QC36G	403.0	5.6	1.80	0.14	4										
clinopyroxenite selvage	QC36G											384.9	8.7	0.23	0.63	2

Table 8.7: Mean weighted radiometric U-Pb ages for accessory phases attained by using selected, concordant lower intercept analyses indicated by † symbol in data tables

\* Mean Square Weighted Deviation

^ Probability

" Number of analyses



Figure 8.11: Plots showing age data for all zircon analyses of samples from Årsheimneset. Error ellipse line colours and marker formats as described in Table 0.13. a) Wetherill concordia diagram. Ellipses are two-sigma errors. Age labels in Ma. Inset boxes expanded view of boxed area ~400Ma. b-c)<sup>206</sup>Pb/<sup>238</sup>U age plots grouped into rock type. Data points are ordered by age with each horizontal black line indicating the weighted mean age for the population of data points it intersects. Error bars are two-sigma errors.



Figure 8.12: Plots showing age data for all zircon analyses of samples from Svartberget. Error ellipse line colours and marker formats as described in Table 0.13. a) Wetherill concordia diagram. Ellipses are two-sigma errors. Age labels in Ma. Inset boxes expanded view of boxed area ~400Ma. b-c)  $^{206}Pb/^{238}U$  age plots grouped into rock type. Data points are ordered by age with each horizontal black line indicating the weighted mean age for the population of data points it intersects. Error bars are two-sigma errors.



Figure 8.13: Wetherill concordia diagrams showing growth zones of zircons. Error ellipse line colours: purple = xenocrystic core; dark blue = main (if unzoned)/central; mid blue = outer growth zones; light blue = rim. Ellipses are two-sigma errors. Age labels in Ma. Inset boxes expanded view of boxed area ~400Ma. a) samples from Årsheimneset; b) samples from Svartberget.



Figure 8.14: Wetherill concordia diagrams showing Th/U ratios of zircons. Error ellipse line colours: dark green = Th/U > 0.4; bright green = Th/U 0.1-0.4; light green = Th/U < 0.1. Ellipses are twosigma errors. Age labels in Ma. Inset boxes expanded view of boxed area ~400Ma. a) samples from Årsheimneset; b) samples from Svartberget.

A group of nine analyses from core, middle and rim zones in zircons in gneiss from Svartberget cluster ~415-395Ma (see *Fig. M2.1c*) and yield a concordant age of 401.1±2.7Ma. Zircons in the gneisses from Årsheimneset mostly yield discordant ages close to the lower intercept (*Fig. 8.11a*) with only one plotting on the concordia line at 417±10Ma.

The younger, concordant population of unzoned zircons and main zones from vein cores from Årsheimneset cluster around 425-405Ma (see *Fig. M1.1a*) and yield an age of 414.0±5.6Ma. They all have low Th/U ratios (<0.1) (*Fig. 8.14a*). The young, concordant population of analyses of vein core zircons from Svartberget are from all growth zones (*Fig. 8.13b*) and all of which have low Th/U ratios (<0.1) (*Fig. 8.14b*). They cluster around 430-390Ma (see *Fig. M2.1a*) with no distinction between those hosted in biotite and those in garnet and yield an age of 410.6±2.6Ma. The younger of those hosted in biotite in the glimmerite selvage cluster around 410-390Ma and yield an age of 403.0±5.6Ma: those which spread along the concordia to older ages are discarded (see *Fig. M2.1b*).

Zircons in the eclogite from Årsheimneset (*Fig. M1.1b*) yield an age of 411.4 $\pm$ 5.3Ma. Those in the garnet websterite from Årsheimneset do not yield a concordant age but they lie along a trend between an upper intercept age of ~1.8Ga and Scandian lower intercept (*Fig. 8.11a*). They are hosted within the inclusion-poor rim of an inclusionrich garnet porphyroblast. Two analyses of zircons in the garnet websterite from Svartberget yield a concordant age of 411.3 $\pm$ 8.9Ma similar to and within error of that in the vein core.

#### 8.4.3.2. Monazite

All monazite analyses are from grains in samples from Svartberget. Analyses of monazites in gneisses span 400-380Ma and yield an age of  $389.1\pm2.9Ma$  (see *Fig. M3.1b*). Monazite grains in the leucogneiss span 410-390Ma and yield an older age of 398.8±3Ma (see *Fig. M3.1c*). Monazites in the vein core and glimmerite selvage are discordant and plot along a trend which intercepts the Tera-Wasserberg concordia at 390±21Ma: an upper intercept is calculated at 3435±840Ma (see *Fig. M3.1a*). They yield a corrected <sup>238</sup>U/<sup>206</sup>Pb age of 409.2±6.4Ma.



Figure 8.15: Plots showing age data for all monazite analyses of samples from Svartberget. Error ellipse line colours and marker formats as described in Table 0.13. Age labels in Ma. a) Tera-Wasserberg concordia diagram. Ellipses are two-sigma errors; b) weighted mean corrected <sup>238</sup>U/<sup>206</sup>Pb age plot grouped into rock type. Data points are ordered by age with each horizontal black line indicating the weighted mean age for the population of data points it intersects. Error bars are two-sigma errors.

## 8.4.3.3. <u>Rutile</u>

All rutile analyses from Årsheimneset plot on or near the concordia line with Scandian ages with a strong, overall clustering about 400Ma but with a considerable range of ~430-370Ma (*Fig. 8.16a*): rock type groups typically span 30-40Myr. Some are discordant plotting just above and away from the concordia line. Average corrected ages vary and range between 419.3±7.3Ma (garnet websterite) to 384.7±3.9Ma (garnet websterite selvage) with seemingly no systematic correlation to rock type; for example, garnet websterites yield both older, intermediate and younger ages of 419.3±7.3Ma, 410.7±3.9Ma and 392.7±5.0Ma (see *Fig. M4.1*).

Rutiles from Årsheimneset vein zones yield a range of ages: those in clinopyroxenite selvages yield ages of 415.1±8.5Ma and 401.6±5.4Ma whilst those in a orthopyroxene eclogite selvage yield an age of 384.7±3.9 Ma (see *Fig. M4.3c*). Those from vein cores range from 401.5±6.6 to 396.5±4.1Ma (see *Fig. M4.2*) which are ~15Ma younger than the associated zircons. Rutile grains in the eclogite yield an age of 408.1±2.8Ma (see *Fig. M4.2a*) which is ~3Ma younger than the zircon ages for this lithology.

All rutiles from Svartberget are also concordant and young but show a considerable spread along the concordia with an expanded range of 450-360Ma (*Fig. 8.17a*). Analyses from different rock types plot in distinct but overlapping groups with the websterite; plotting on the concordia with the oldest ages and garnetite vein cores with the youngest. Similar to the rutiles at Årsheimneset, each lithology shows a spread of 30-40Myr. Weighted mean corrected ages range from 419.0±13Ma (garnet websterite) to 381.6±2.8Ma (garnetite vein core): a similar range to those from Årsheimneset. Rutiles from clinopyroxene garnetite vein cores yield ages of 416.6±4.1 and 407.0±5.2 (see *Fig. M5.2*) whilst those from garnetite vein cores yield younger ages of 388.1±3.5Ma and 381.6±2.8Ma (see *Fig. M5.1*). The clinopyroxenite selvage yields an age of 384.9±8.7Ma (see *Fig. M5.1c*).



Figure 8.16: Plots showing age data of all rutile analyses of samples from Årsheimneset. Error ellipse line colours and marker formats as described in Table 0.13. Age labels in Ma. a) Tera-Wasserberg concordia diagram. Ellipses are two-sigma errors. b) corrected <sup>238</sup>U/<sup>206</sup>Pb age plot grouped into rock type. Data points are ordered by age with each horizontal black line indicating the weighted mean age for the population of data points it intersects. Error bars are two-sigma errors.



Figure 8.17: Plots showing age data of all rutile analyses of samples from Svartberget. Error ellipse line colours and marker formats as described in Table 0.13. Age labels in Ma. a) Tera-Wasserberg concordia diagram. Ellipses are two-sigma errors. b) corrected <sup>238</sup>U/<sup>206</sup>Pb age plot grouped into rock type. Data points are ordered by age with each horizontal black line indicating the weighted mean age for the population of data points it intersects. Error bars are two-sigma errors.

#### 8.5. Data Analysis

# 8.5.1. <u>Petrography</u>

## 8.5.1.1. Zircon

The zoned and crystalline nature of zircons suggests they are not metamict grains. The overall ovoid-rounded shape of most outer growth zones and lack of oscillatory zoning suggests they are metamorphic in nature (e.g. Sheng *et al.*, 2013). In comparison, the prismatic, faceted, euhedral shapes and fine oscillatory zoning of some core zones suggest they have an igneous origin formed by crystallisation from a melt (e.g. Schaltegger *et al.*, 1999; Bingen *et al.*, 2001; Gordon *et al.*, 2013; Sheng *et al.*, 2013). Eclogite-facies metamorphic overgrowth around inherited cores from an igneous rock is recognised in other rocks (e.g. Mezger and Krogstad, 1997; Bingen *et al.*, 2001). Their often truncated nature suggests these are fragments of larger zircons broken upon transport from their source. Their faceted, euhedral forms of many grains suggest they did not undergo much dissolution/resorption during transport otherwise more irregular edges may have formed (e.g. Zheng, 2012; Chen *et al.*, 2013 and references cited therein). Such diagnostic igneous characteristics are not observed in the zircons found in the eclogite and olivine garnet websterite thin sections.

The similar zoning patterns, sizes and shapes of the xenocrystic cores in zircons from both the gneisses and the vein cores suggest they may be equivalent. The lack of igneous characteristics of zircons in the host bodies but presence of xenocrystic cores in the vein cores implies an external source and physical transport of zircon grains associated with the metasomatic fluid (e.g. Zheng, 2012; Sheng *et al.*, 2012, 2013). The higher number of zircons and monazites present in the garnetite vein core side of the glimmerite selvage in the vein core from Svartberget suggests this horizon was saturated in Zr, REEs and P and conditions were appropriate for high nucleation rates. The formation of aggregate grains reflects grains with nucleation sites close together coming into contact during growth (Bingen *et al.*, 2001). Zr may be derived from the breakdown of Zr-bearing minerals (e.g. biotite, hornblende or ilmenite) or dissolution of protolith zircon grains (Zheng, 2012 and references cited therein). As the abundance of zircon in the host bodies is low, particularly in the olivine garnet websterite at Svartberget, and the vein cores have higher Zr contents (as shown in Chapter 4), Zr is probably sourced from the incoming fluid. The paucity of zircons and their general smaller size in clinopyroxenite and websterite selvages suggests abundant zircon-forming components or fragments did not travel far into the wallrock and were not considerably sourced from the host bodies.

The fine radial cracks often continue into radial fractures in the host garnet suggesting they may have formed at the same time and in relation to decompression during exhumation. Their abrupt termination at the xenocrystic core indicates the cores retained a greater structural strength and integrity than their metamorphic overgrowths which were more susceptible to brittle deformation. Fractures in the interior zones which taper towards the outer and rim zones may represent the extent of the annealing of older fractures during succeeding stages of growth, metamorphism and/or recrystallisation. The small crystals of xenotime (YPO<sub>4</sub>) occasionally found attached to zircon grains may have caused the increased fracturing observed in the associated zircons. The additional decay of minor U and Th in xenotime (Nesse, 2000) may cause a greater degree of radiation damage and metamictisation of the zircon crystal leading to a higher susceptibility to deformation and fracturing due to higher crystal defects. This process may also explain why zircons hosted in LILE-compatible amphibole are also more heavily fractured.

## 8.5.1.2. Monazite

As with zircon, the zoned and crystalline nature of monazites suggests they are not metamict grains. The irregular and delicate shapes of many monazite grains in the well-foliated host gneiss and leucogneiss and association with biotite suggest they grew during the retrogressive stage; perhaps later than the monazites in the metasomatic vein cores. The multiple chemical zones and patchwork patterns suggest the grains grew under variable, local chemical environments perhaps scavenging for suitable elements. The more simple zoning of monazites in vein cores, their more robust shapes and their association with small zircon grains suggest they grew in a more HFSE-HREE-enriched environment with fewer chemical changes or with a greater availability of components and under more physical stress.

#### 8.5.1.3. <u>Rutile</u>

Rutile grains show no geochemical or growth zoning suggesting they grew and/or fully re-equilibrated during metamorphism and the episode of metasomatism and they have few variations in impurity abundance or components. Their variable distribution, shapes and sizes suggests their forms were controlled by the local environment and availability of components. The common occurrence of trails of rutile grains along vein zones and between garnet and clinopyroxene crystals and often interstitial texture suggests they grew from late fluids held in pore spaces, microcracks and along grain boundaries.

The Ti-Fe blades and lamellae are considered to be ilmenite or haemoilmenite. Their origin is ambiguous although some are clearly exsolutions and alteration products of rutile. The tiny, orientated needles of ilmenite are more clearly exsolutions indicating increased temperatures (e.g. Bingen *et al.*, 2001; Massonne, 2009): Fe is a common impurity in rutile (Cherniak, 2000).

## 8.5.2. Mineral Chemistry

## 8.5.2.1. Zircon

The below-detection limit concentrations of trace elements in many zircon analyses indicate they contain low concentrations. Considering the predominantly below-detection limit Y contents of vein core zircons but often above-detection limit contents of Y in zircons from the gneiss probably results from the co-crystallisation of garnet in the vein core which would preferentially incorporate any HREE (e.g. Sheng *et al.*, 2012); alternatively it may reflect a low HREE content of the metasomatic agent due to residual garnet in the source or inability of the fluid to transport HREE.

The high Th/U of >0.4 in some zircon cores are suggestive of a magmatic origin and the low Th/U ratios of <0.1 of the outer growth zones are indicative of a metamorphic origin (e.g. Bingen *et al.*, 2004; Zheng, 2009; Berger *et al.*, 2010; Sheng *et al.*, 2013). A geochemical study of zircons from Svartberget carried out by Vrijmoed *et al.* (2013) also found they had low Th/U ratios of below 0.1. The cores with lower Th (<50ppm) and lower Th/U similar to the outer growth zones may have lost Th by diffusion during metamorphism-metasomatism or crystallised with epidote group minerals and/or monazite which preferentially incorporated Th (and REE) (e.g. Bingen *et al.*, 2001).

Grains from the garnetite vein cores are expected to have Scandian ages and low HREE contents as these are preferentially incorporated into garnet. Those in the host gneisses, if grown outside the stability field of garnet may have high HREE contents (e.g. Zheng, 2009; Sheng *et al.*, 2013).

## 8.5.2.2. Monazite

The clusters of REE contents in monazites separating the vein monazites from the gneiss monazites suggest different types formed under different chemical conditions. The considerable range of Th and range of U in monazites from the country rocks with respect to the LA-MC-ICP-MS data and vein with respect to EMP data suggests the availability of Th and U during monazite growth was variable perhaps effected by other accessory phases such as zircon. The considerable differences between the two analysis techniques suggest they are not equivalent, particularly considering the same grains and growth zones are analysed.

## 8.5.2.3. <u>Rutile</u>

The low U contents of rutiles in eclogites and websterites suggest considerable U inheritance does not occur during metasomatism and implies that U in rutiles in the vein zones is primarily sourced from the metasomatising fluid. The consistently low Th contents in rutile reflect Th not being readily accepted into the rutile lattice. The generally higher concentrations of Nb and U in rutiles (and of Y, U and Th in zircons) in vein cores from Svartberget over those from Årsheimneset suggest the metasomatic chemical environment at Svartberget was more enriched in trace elements. The strongly sub-chondritic Nb/Ta ratios of vein rutiles suggest they are sourced from released fluids and not residual mafic-ultramafic rocks (Zhang *et al.* 2008).

## 8.5.3. <u>U-Pb Isotope Data</u>

# 8.5.3.1. *Zircon*

The discordant nature of the central growth zones of zircons in vein cores from Årsheimneset suggests they are xenocrysts like those from Svartberget. Those from Svartberget plot along similar discordia to the analyses of zircon cores in the gneiss suggesting the cores in the vein zircons are xenocrysts of relict igneous zircons from the surrounding gneiss. At both localities discordant analyses plot along various discordia lines suggesting a range of crystallisation ages for the original, magmatic zircon: Vrijmoed et al. (2013) found similar a variation for Svartberget zircons. The upper intercept ages of 1.5Ga and older correspond to the Baltic Shield crust-forming events of the Gothian Orogeny (1.75-1.5Ga) during which extensive continental growth and magmatism occurred forming protoliths to the gneisses (Lapen et al., 2009). Most lines trend towards a lower intercept with Scandian ages (430-390Ma) suggesting they lost variable amounts of Pb during the Scandian event: no analyses plot at the upper intercept on the concordia line suggesting all xenocrystic cores experienced some degree of isotopic resetting. Alternatively, analyses may have ablated material from an outer, Scandian growth zone and therefore their isotopic concentrations represent mixtures of the two ages; although care was taken to avoid this.

The low Th/U ratios of concordant, lower intercept ages of identified xenocrystic cores in vein core zircons from Svartberget suggest these have completely reset and re-equilibrated to the metamorphic conditions of the Scandian event losing their prior accumulated radiogenic Pb and high Th contents: these are classified as 'metamorphosed' zircon. The low Th/U ratios of the concordant, lower intercept population of analyses of outer growth zones and unzoned zircons from both localities indicate they are 'metamorphic' zircons which grew during Scandian metamorphism. The two Svartberget vein core analyses with discordant ages which plot towards the lower intercept and identified as mid outer growth zones may be almost completely reset Proterozoic growth zones or have inherited components diffusing out of their xenocrystic cores into the surrounding metamorphic growth zone. No other analyses of zones identified as metamorphic yield discordant ages which trend towards an upper intercept age.

The discordant websterite zircons from Årsheimneset may represent either grains originally in the protolith prior to Scandian metasomatism with an upper intercept reflecting the magmatic age of the garnet websterite protolith or are xenocrysts of gneiss zircons like those in the vein cores with the upper intercept reflecting the magmatic age of the gneiss protolith. Their textural location is ambiguous as they lie within the inclusion-poor rim of a garnet porphyroblast which may be part of the prograde, metamorphic growth or a metasomatic overgrowth.

The persistence of igneous zircon isotopic components and ancient, discordant ages demonstrates the zircon xenocrysts did not undergo complete resetting and re-equilibration (e.g. Mezger and Krogstad, 1997) during the Scandian suggesting temperatures during this or other subsequent events did not reach above the closure temperature of zircon (~900°C) for any considerable length of time.

The discordant analyses which plot further away from the concordia line, such as those from the gneiss and vein cores from Årsheimneset, reflect the presence of common lead components. Those that plot below the concordant lower intercept, such as those from the gneiss from Årsheimneset, represent isotopic loss since recrystallisation/growth during the Scandian event perhaps related to metamictisation and enhanced diffusion of Pb, thermal disturbance or deformation/structural damage (Mezger and Krogstad, 1997). Concordant analyses which plot along or near the concordia with older ages, such as in the vein cores and glimmerite selvage from Svartberget (see App. M2.1), are thought to indicate inherited components: considering the presence within vein zones, which, overall give Scandian ages, an origin associated with earlier, pre-Scandian Caledonian events is not envisaged.

Vein core and garnet websterite zircon ages at Svartberget have similar ages *ca*. 411Ma suggesting this is a reliable age for metasomatism at this locality: the 410.6±2.6Ma age recorded by vein core zircons is considered to be a robust age being based upon 20 analyses. The younger age *ca*. 401Ma of the gneisses suggests the episode of zircon growth in the host rocks followed that of vein formation or continued for a longer period of time. At Årsheimneset, vein core formation occurred *ca*. 414Ma. The age of the gneiss is unreliable due to being based on only one analysis but indicates an episode of growth during the Scandian also. The age

difference between the vein core and glimmerite selvage at Svartberget is unexpected, although they are within error. The *ca*. 411Ma age of the eclogite zircons at Årsheimneset is similar to that of the vein core suggesting they equilibrated under similar conditions and metamorphic re-equilibration or growth of zircons in the host body continued alongside metasomatism and vein formation towards peak temperatures. This implies fluid production and metasomatism occurred under increasing temperatures.

Finally, the thin rim zones of zircon grains with radial fractures which yield either concordant ages similar to the other growth zones or discordant ages indicative of common Pb or Pb loss are considered to have grown at the same time as the other outer growth zones but have been modified by exhumation-related deformation and retrogression.

## 8.5.3.2. Monazite

The Scandian ages of monazite grains in the country rocks demonstrate they are all newly grown as indicated by their petrographic characteristics. The discordance of the analyses of monazites from the vein core and glimmerite selvage suggests they contain common lead rather than being discordant: the calculated upper intercept of 3435±840Ma is geologically meaningless. Their corrected age is similar to that of the zircons in the vein cores: this suggests they reflect ages of crystallisation if their effective closure temperatures were above the ambient temperature conditions. The later ages of the country rocks may reflect continued suitable conditions for monazite crystallisation in each rock type, variation in the local cooling temperatures of grains in each setting or they may relate to separate episodes of fluid activity and growth.

The monazites in the gneiss are ~12Ma younger than their associated zircons whilst those in the vein cores are more contemporaneous and within error. Assuming the zircons and monazites in each rock type crystallised at the same time, this difference suggests local environmental factors have affected the relative closure temperatures of the two phases otherwise a consistent difference would be expected. Assuming the leucogneiss formed in an environment where fluid was present for longer,

continued diffusion and exchange of components might have occurred to lower temperatures.

## 8.5.3.3. Rutile

The consistently young ages of rutile grains (419-381Ma) indicates they are all newly grown or equilibrated to Scandian conditions. The discordant analyses from Årsheimneset which plot away from the concordia line represent those which either contain common lead or have lost Pb components since the Scandian: the calculated age of the upper intercept of the discordia line (3852±850Ma) is geologically meaningless. The ca. 30Myr degree of scatter along the concordia of rutiles in each rock type may be related to: prolonged growth and resetting; inheritance of components from prograde phases such as titanite, amphibole and clinopyroxenes; post-crystallisation modification processes, such as deformation and retrogression; or variation in the closure temperature of grains to diffusion due to local environmental conditions and grain characteristics such as size and defect concentration. Furthermore, whilst the ca. 400Ma age of rutiles in vein cores from Årsheimneset are ~14Myr younger than their associated zircons and a similar difference of ~18Myr is seen in a garnetite vein core from Svartberget, suggesting rutiles were open to diffusion to lower closure temperatures than zircons in the vein cores, rutiles in the garnet websterite from Svartberget yield an age ~8Myr older than associated zircons. This suggests a selective process operated producing older ages in some rock types. The pattern and distinct clustering of the different rock types along the concordia at Svartberget suggests the cause of the scatter is related to lithology. Such a pattern may result from inheritance dependent upon the maficfelsic character of the mineral assemblage; however, the young age of the clinopyroxenite selvage contradicts the trend suggesting the variation may due to another process, at least in part.

An alternative explanation is variation in the closure temperatures of rutile grains. The younger ages may represent grains with lower closure temperatures, perhaps under fluid-present conditions, dating exhumation to crustal levels; whilst those with older ages closed to diffusion at higher temperatures and yield ages more representative of crystallisation and formation. The production of youngest ages closest to the vein centre might support this but the 14-17Myr age differences

between adjacent vein zones at Årsheimneset [QC26H, QC26J] and in grains of similar appearance suggest this may not be the process responsible.

The ages of rutiles in the eclogite and garnet websterite at Årsheimneset are considered to be robust considering the greater number of analyses upon which they are based. Their similar age ca. 409Ma suggests this is the age of cooling through the rutile closure temperature at this locality. This is just younger than the zircon age of 411Ma although this is not as reliable suggesting the rutile grains may have closed to diffusion at higher temperature than 400°C perhaps due to the absence of a free, catalysing fluid in the eclogite and garnet websterite relative to the vein zones which yield younger ages. Residual or retrogressive fluid in the veins may have enabled diffusion to continue until their lowest possible closure temperature. Vein cores at Arsheimneset show a greater degree of retrogression than the eclogites and websterites with garnets and clinopyroxenes cut by later, amphibole-rich fractures, pervasive symplectisation, reaction between garnet and biotite, dominant formation of amphibole coronas and pervasive replacement of garnet. This indicates fluids were present during retrogression which could have facilitated continued diffusion and reequilibration of rutiles in the vein zones to amphibolite-facies conditions. The range in ages yielded by garnet websterites supports a local presence of fluid enabling continued equilibration and diffusion in particular domains. Alternatively, the older ages similar to zircon ages may be coincidental and be due inheritance or a combination may be responsible for the rutile ages produced.

## 8.6. Data Interpretation

# 8.6.1. Mineral Growth

Zircon is considered inert in the absence of fluids (Hermann *et al.*, 2013) and under metamorphic conditions is thought to grow in association with prograde or retrograde metamorphic and melt reactions (McClelland and Lapen, 2013). Typically there is limited fluid availability for zircon growth during UHP conditions (McClelland and Lapen, 2013) and episodes of growth and recrystallisation occur in response to episodes of fluid availability and oversaturation of Zr (e.g. Zhao *et al.*, 2007 and references cited therein; Zheng, 2009; Zheng, 2012 and references cited therein Sheng *et al.*, 2012, 2013 and references cited therein; Hermann *et al.*, 2013). The
Scandian ages of zircons in UHP rocks (as indicated by their thermobarometry) are therefore interpreted to reflect the presence of fluids at UHP conditions.

Zr solubility in quartz-saturated aqueous fluids is typically low (Chen *et al.*, 2013 and references cited therein) and mafic melts, which are strongly undersaturated in Zr, would dissolve relict Zr very rapidly but would be unable to precipitate zircon (Zheng, 2012 and references cited therein). In felsic melts, however, Zr can become saturated (Zheng, 2012 and references cited therein; Chen *et al.*, 2013 and references cited therein) suggesting the incoming metasomatic fluid had some chemical characteristics and composition of a felsic, silicate melt.

Mafic-ultramafic rocks which contain zircon are thought to be either derived from partial melting of enriched mantle or underwent crustal assimilation during their emplacement (Zheng et al., 2009). There is evidence from Sr isotopic systematics to suggest the external mafic-ultramafic rocks in the WGR underwent pervasive metasomatism by crustal fluids at an early point in their history (e.g. Brueckner, 1977; Vrijmoed et al., 2013): addition of zircon-forming components may have occurred during this time. Scandian ages of zircon in the bodies, in the absence of a felsic melt, may result from new growth in metamorphic reactions, recrystallisation by solid-state/subsolidus transformation or in situ dissolution-re-precipitation of protolith zircon grains in the presence of a fluid (Zheng, 2009; Zheng, 2012; Chen et al., 2013; Sheng et al., 2013 and references cited therein). New zircon may grow during eclogitisation through the release of Zr during the breakdown of Zr-bearing prograde phases such as amphibole, titanite or ilmenite in the presence of an aqueous fluid (e.g. Bingen et al., 2001; Zheng, 2012). Considering the abundance of amphibole inclusions in garnet blasts, the abundance of rutile in the eclogite and its overall metabasaltic bulk composition, a prograde assemblage containing Zr-bearing amphibole and titanite (or ilmenite) is probable suggesting zircon originated during eclogite-facies metamorphism.

Zircons in the eclogite, garnet websterite and olivine garnet websterite samples are typically unzoned supporting a newly grown origin, however, some small zircon grains do have different growth zones suggesting some may contain older, recrystallised grains. Isotopic ratios and almost all geochemical features can be inherited during solid-state recrystallisation (Sheng *et al.*, 2012) suggesting it is not

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the responsible mechanism for Scandian ages; the xenocrystic cores in the gneisses and vein cores which have undergone some degree of component loss and isotopic resetting may have undergone solid-state recrystallisation. In comparison, *in situ* reprecipitation of zircon typically shows the highest degrees of reworking and element loss and can yield concordant ages for the metamorphic event (Thrane, 2002; Sheng *et al.*, 2012).

The Scandian ages and metamorphic petrographic and chemical characteristics of the outer growth zones of zircons in the gneiss demonstrate zircon grew under metamorphic not magmatic conditions and therefore that a metamorphic fluid capable of dissolving and precipitating Zr was present. The younger ages of zircons and monazites in the country rocks than the vein cores is interpreted to indicate prolonged fluid activity in the country rocks relative to the episode of metasomatism of the mafic-ultramafic bodies and more fluid was produced than infiltrated the bodies. Considering the sharp contacts and the chemical differences between the outer metamorphic growth zones and most cores, the outer zones are not considered to be recrystallised outer margins of the cores but newly grown zircon crystallised on the surface of magmatic zircons and record the timing of fluid availability. Considering the delicate form of the monazites within the deformed amphibolite-facies gneiss and their younger age, they are interpreted to have grown during retrograde metamorphism.

The zircon- and monazite-forming fluids may have been locally-derived, intermittent fluids or a more abundant, pervasive fluid associated with widespread partial melting. The geographical scale of Scandian zircon overgrowths has not been investigated but such fluids were present over at least the 1.25m from the contact with the peridotite body at Svartberget (to the location of sample QC36J). The persistence of magmatic zircons in the source rocks suggests extensive dissolution did not take place and that fluid activity was either short-lived, minimal or sporadic or the fluid did not have the capacity to consume all the zircon available.

Growth of zircon and monazite grains in vein zones is interpreted to have occurred during peak pressure conditions and from precipitation of components carried in the metasomatic fluid rather than redistributed zircon components sourced from the host bodies in which zircon and monazite are sparse, particularly the Svartberget

peridotite body. The UHP fluid must share similar characteristics to a magmatic, felsic melt in order for zircon components to be dissolved, become saturated and then precipitated from it (Zheng, 2012 and references cited therein). The presence of xenocrysts in the vein cores which plot along similar discordia as zircon cores in the gneisses indicates the fluid was viscous enough to physically transport solid zircon fragments from the country rocks into the mafic-ultramafic bodies during the Scandian metasomatic event. Rutiles within the veins are interpreted to have grown during metasomatism of the host bodies and record the timing of cooling from peak temperatures but more rutile-forming components may have been sourced from the host Ti-rich bodies.

The large, dateable zircons in garnet websterites at Svartberget are also considered to have formed from components from the metasomatic fluid considering the paucity of zircon and of larger zircons (>20µm-diameter) in the host peridotite. The smaller, unzoned zircons (<20µm-diameter) in the garnet websterites may have formed through prograde metamorphic reactions like those in the eclogite.

The Scandian growth/recrystallisation of accessory phases is considered to have occurred in association with the metamorphism of the mafic-ultramafic bodies, metasomatism during peak conditions and with fluid activity in the country rocks during post-peak temperature conditions.

## 8.6.2. <u>P-T-t Path</u>

The survival of zircon fragments in the host gneisses and vein cores indicates temperatures were below ~850°C as zircon can commonly survive anatexis below this (Chen *et al.*, 2013 and references cited therein) and the partial resetting of pre-Scandian ages indicate temperatures approached the closure temperature of zircon ~900°C (Chen *et al.*, 2013 and references cited therein).

To determine relative closure temperatures of the three minerals, the most reliable ages to compare are those in the QC36G vein core from Svartberget which yield ages of 410.6±2.6Ma for zircon, 409.2±6.4Ma for monazite and 381.6±2.8Ma for rutile. The similar ages of the zircons and monazites do not necessarily indicate similar closure temperatures just that they both formed together at temperatures below

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both their closure temperatures. They indicate however that the closure temperature for rutile is lower than that of zircon and monazite (as expected) and of the temperature of metasomatism being separated by *ca*. 30Myr of cooling. Assuming they formed at the same time, the closure temperatures of zircon and monazite may vary between different settings as a ~12Myr age difference is produced between those in the gneiss at Svartberget.

A P-T-t path for each locality is proposed in *Figure 8.18* using thermobarometric data presented in Chapter 6. At Svartberget, the timing of metasomatism and vein formation is considered to be *ca*. 410Ma (considering both zircon and monazite ages) under temperature conditions of up to ~800°C and pressures of at least 3.85GPa. Previous study of zircons from Svartberget yielded a lower intercept age of 397.2±1.2Ma (Vrijmoed *et al.*, 2013) younger than that found in this study. However, studies of zircons from other bodies in the northern Nordøyane UHP domain have yielded peak ages of 410Ma at Flemsøy (Krogh *et al.*, 2011) and *ca*. 410-400Ma at Finnøya and Otrøy (Gordon *et al.*, 2013). Monazites in garnets in diamondiferous country rocks at Fjørtoft yield an age of 415±6.8Ma which is considered to represent the timing of prograde metamorphism (Terry *et al.*, 2000) and 407±2.1Ma which is considered to represent the timing of peak pressure conditions (Terry *et al.*, 2000). These sit either side of the ages obtained from the vein cores in this study suggesting metasomatism occurred near peak pressure conditions.

Cooling to sub-eclogite-facies conditions is considered to have occurred by ca. 385±3Ma as indicated by the younger ages of rutiles in garnet websterite, vein cores and vein selvages at Svartberget. Vrijmoed *et al.* (2013) found a rutile from garnetite at Svartberget yielded a <sup>207</sup>Pb/<sup>206</sup>Pb age of 388±1.2Ma similar to that of this study.

A period of fluid activity in the host gneisses is dated at *ca*. 401-390Ma. Gordon *et al*. (2013) concluded that a partial melt may have been present in the country rocks in the Nordøyane UHP domain during peak UHP metamorphism and was present during initial decompression. The presence of zircons and monazites within the amphibolite-facies gneiss assemblage and younger ages are interpreted to indicate their growth occurred post-peak temperatures during retrogression: older ages of formation would be expected to be retained in zircon if growth during peak UHP conditions occurred. Vrijmoed *et al.* (2013) obtained a lower intercept age of 391.2±0.8Ma for a



Figure 8.18: Graph showing proposed P-T-t paths for the Årsheimneset and Svartberget localities. Marker formats and line colours as described in Table 0.13. Black dashed line is the 5°C/km geotherm, lowest possible geotherm (Schreyer, 1988). Solid black lines are polymorph phase boundaries (fields as labelled). Phase abbreviations as listed in Table 0.11b. Graphitediamond boundary after Day (2012). Coesite-quartz boundary after Bose and Ganguly (1995). Albite = jadeite + quartz equilibria from Holland (1980). Aluminosilicate boundaries after Holdaway (1971). Coloured, dotted lines are metamorphic facies fields from Robertson (1999).

zircon and <sup>207</sup>Pb/<sup>235</sup>U ages of 389.6±1.7Ma and 383.8±1.1Ma for monazites from a gneiss leucosome at Svartberget; slightly younger than those in this study but again demonstrating a post-peak-age formation. In studies of other Nordøyane localities, pegmatites and leucosomes thought to represent amphibolite-facies retrograde melts and to date the timing of exhumation to crustal depths are dated at 396±4Ma (Krogh *et al.*, 2011) and *ca*. 400-385Ma (Gordon *et al.*, 2013) similar to the cooling ages of rutiles from the Svartberget locality in this study. Furthermore, monazites in the gneiss matrix at Fjørtoft yield an age of 398±6Ma which is considered to represent exhumation to crustal levels (37km) (Terry *et al.*, 2000).

Ages of zircons at Årsheimneset in this study are not very reliable due to the low number of analyses upon which they are based. They indicate vein formation and metamorphic equilibration to peak conditions of ~3.5GPa, ~700°C (see Chapter 7) occurred at *ca*. 411-414Ma. The lack of distinction between metamorphic and metasomatic ages is interpreted to indicate metasomatism occurred prior to metamorphic equilibration at peak temperatures. Older rutile ages are within this range or within error of this range but these may be coincidental ages due to inherited components. Previous studies of zircons from Årsheimneset have been dated at *ca*. 400Ma (Carswell and Cuthbert, 2003c and references cited therein). Other studies of eclogites in the Nordfjord-Stadlandet UHP domain produced ages of 405–400Ma at Flatraket (Root *et al.*, 2004; Krogh *et al.*, 2011) and 406±4Ma and 402±2Ma in the Stadlandet region (Schärer and Labrousse, 2003) ~10Myr younger than those obtained in this study.

The younger ages of rutiles in garnet websterites and vein zones of *ca*. 396-384Ma are interpreted to represent the latest re-equilibration of U-Pb systematics to ambient conditions and reflect closure to diffusion at temperatures of  $\geq$ 400°C. This means that within 15-30Myr the Nordfjord-Stadlandet UHP domain had been exhumed to lower crustal levels. In other studies, rutile from an eclogite in the Stadlandet area produced a U-Pb age of 389±7 Ma similar to those obtained from Årsheimneset but is considered to date metamorphic equilibration at peak temperatures (770°C) during the early stages of decompression (Schärer and Labrousse, 2003). A similar U-Pb age range of 393–389Ma obtained from titanite grains in orthogneisses by Kylander-Clark *et al.* (2008) is considered to represent the timing of the widespread amphibolite-facies overprint of the country rocks showing

the rutile analyses to yield sensible ages and date the retrograde stage of the P-T-t path.

Knowledge of the absolute closure temperature of rutile, and therefore determination of accurate cooling rates, is difficult to predict. Using 400°C as the minimum rutile closure temperature and taking 800°C as the peak temperature of the Svartberget locality and 700°C for Årsheimneset, maximum average cooling rates are calculated as 13.8°C/Myr at Svartberget and 10-20°C/Myr at Årsheimneset considering the range of ages. If the effective closure temperature of rutile is over 400°C, cooling rates will be slower. As exhumation to crustal levels is considered to be near isothermal, cooling rates are likely to be variable and considerably faster than the average rates calculated during later stages of exhumation.

#### 8.6.3. <u>Regional Variation and Tectonic Implications</u>

Comparing the youngest ages for each mineral, zircons in gneisses from Svartberget are ~16Myr younger than those from Årsheimneset and those in Svartberget vein cores are ~4Myr younger than those from Årsheimneset; although the dates from Årsheimneset are not very reliable. As zircons are considered to be a crystallisation age and not a cooling age, peak metamorphism and metasomatism at Svartberget (northern UHP domain) appear to have occurred at a similar to slightly earlier time than at Årsheimneset (southern UHP domain). In contrast, other U-Pb studies of bodies from the two domains suggest peak metamorphism in the southern domain (405-400Ma) occurred at similar to slightly later times than that in the northern domain (410-400Ma) (Schärer and Labrousse, 2003; Root et al., 2004; Krogh et al., 2011; Gordon et al., 2013). Zircon ages from studies of bodies in the central Sørøyane UHP domain include 402±2Ma at Hareidlandet (Krogh et al., 2011) and 401±6Ma at Hareidlandet (Carswell et al., 2003a) similar to those for the Nordfjord-Stadlandet UHP domain. The difference between the domains (~5Myr) is similar to the typical error margins of the dates (<10Ma) so a real difference may not be resolvable and the age of peak conditions and metasomatism should be considered to have occurred at the same time in all domains.

The duration of Scandian residence at UHP conditions is hard to constrain. Both 'geologically brief' and prolonged (~30Ma) timescales are proposed (Brueckner and

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van Roermund, 2004; Spengler *et al.*, 2009). The average ages considered to represent UHP metamorphism in this study range from 414Ma to 411Ma at Årsheimneset and 411Ma to 409Ma at Svartberget indicating short intervals of UHP conditions at each locality. However, the ~20Ma span of zircon ages in each rock type may reflect a longer residence time rather than analytical scatter and suggests a range of ages rather than an average age should be used to represent timings of metamorphism, particularly considering metamorphism is a gradual, continuous process.

Regarding cooling ages, rutiles in vein zones from Svartberget yield ages younger than those from Årsheimneset by 15-20Myr. This suggests Svartberget was uplifted later than Årsheimneset. This may represent contemporaneous exhumation but from deeper depths at Svartberget and from higher temperatures (as indicated in Chapter 7) so cooling was naturally later or the northern UHP domain was physically uplifted later.

# 8.7. Summary

U-Pb ages of zircon, monazite and rutile grains from Årsheimneset and Svartberget were obtained using LA-MC-ICP-MS. Monazites and rutiles form newly grown or fully recrystallised, metamorphosed grains yielding Scandian ages. Zircons show a range of forms and zoning profiles depending upon their host rock: those in the country rocks and vein cores typically consist of a core with igneous petrological and geochemical characteristics, such as oscillatory zoning, angular shapes and high Th/U ratios (>0.1) which yield discordant ages, and outer growth zones with metamorphic petrological and geochemical characteristics being rounded, unzoned with low Th/U ratios (<0.1) and which yield Scandian ages. Zircons within metamorphic mafic-ultramafic domains are typically unzoned and tiny (10-20µm-diameters) being too small for *in situ* dating.

Discordant ages of xenocrystic zircon cores in metasomatic veins and cores of zircons in the country rocks point to protolith ages of over 1.5Ga which correspond to the Gothian Orogeny. The timing of Scandian metamorphic equilibration to peak temperature conditions at Årsheimneset (~700°C, 3.5GPa) is dated at 411.4±5.3Ma and that of vein formation at 414.0±5.6Ma from zircons in the quartz eclogite and garnetite vein cores, respectively. Cooling through rutile closure temperatures (≥400°C) is dated at *ca*. 400-

~385Ma. The timing of UHP metasomatism and vein formation at Svartberget is dated at *ca*. 410Ma and cooling through rutile closure temperatures is dated at *ca*. 390-380Ma. Fluid activity in the country rocks continued to *ca*. 400-390Ma in association with exhumation-related deformation and retrograde metamorphism as indicated by delicate monazites within the host gneisses and leucogneisses. Xenocrystic cores of zircons within vein cores are considered to demonstrate material transport from the surrounding country rocks into the mafic-ultramafic bodies during UHP conditions. The fluid responsible for UHP metasomatism is thought to have been felsic and viscous in order to precipitate zircon and transport xenocrysts, respectively.

#### Chapter 9 – Discussion

### 9.1. Origins and Evolution of Garnet Websterites and Orthopyroxene Eclogites

Garnet websterites are chemically transitional between eclogite and garnet peridotites (Aulbach *et al.*, 2002 and references cited therein) as is shown in Chapter 4. Identification of a metasomatic signature is difficult in a fluid-rock system with felsic and ultramafic components as metasomatic enrichment follows the trend of magmatic fractionation: structural and textural relationships are therefore vital in determining their origin.

Crustal garnet websterites are typically interpreted as arising from primary compositional variation of metamorphosed, igneous protoliths whether cumulate layers in large magmatic bodies within the continental crust, in ophiolites or as crystal segregates along magma conduits (e.g. Aulbach et al., 2002 and references cited therein) forming from gabbro-norite protoliths. The Hustad Igneous Complex on the coast to the north of the Svartberget and Kolmannskog localities with lithologies ranging from pyroxenite to granite is considered to be such a crustal protolith body (Austrheim et al., 2003; Kylander-Clark et al., 2007 and references cited therein). Garnet websterites in mantle-derived rocks are considered to be horizons depleted by partial melting or metasomatic in origin following interaction of silicaundersaturated phases in depleted peridotite with a siliceous fluid forming orthopyroxene from olivine (e.g. Malaspina et al., 2006; Rehfeldt et al., 2008 and references cited therein). They are not considered to be primary mantle rocks or to result from the evolution of known mantle partial melts due to their Si-rich composition (Rehfeldt et al., 2008). They are found in xenoliths suites of both old and young mantle (Rehfeldt et al., 2008) and in orogenic peridotite bodies as discrete dykes, veins and bands (Aulbach et al., 2002; Rehfeldt et al., 2008). Metasomatism by siliceous fluids from partially melted, subducted, eclogite-facies crust may result in a considerable abundance of garnet websterite in the lower mantle wedge (Aulbach et al., 2002; Rehfeldt et al., 2008 and references cited therein).

### 9.1.1. <u>UHP Metasomatic Origin</u>

Garnet websterites formed through metasomatism of crustal garnet peridotiteolivine garnet websterite bodies are observed at Svartberget, Kolmannskog and Myrbærneset where they are clearly related to fluid flow forming vein selvages to

zoned clinopyroxenite-garnetite veins. This is considered a crustal analogue to metasomatic garnet websterite formed in the mantle.

At Svartberget, garnet websterite forms up to ~3m-scale wide selvages at welldeveloped veins, irregular masses with other vein zones at some vein intersections and small blocks between veins and are diamond-bearing demonstrating a UHP formation (Vrijmoed et al., 2006). Clinopyroxenes in the larger domains are diopsides but those closest to the vein cores are omphacites reflecting the increased influence of the fluid composition towards the vein core and addition of Na-Al. This indicates that the geochemical signature of metasomatism is strongest close to the fluid pathway and so may not be easily distinguished or fully identified away from veins. The sodic character of metasomatic minerals in the mantle has been related to the sodic character of the metasomatising agent in other studies (e.g. Prouteau et al., 2001). The bulk compositions of the vein zones compared to those of their host garnet peridotite demonstrate metasomatism is also accompanied by an increase in K<sub>2</sub>O, SiO<sub>2</sub>, H<sub>2</sub>O, LILE, LREE, HFSE and P. Cr-Sc and Ga are seen to be good tracers of the influence of the host body composition at Svartberget as they decrease and increase, respectively, towards the vein core concentrations. However, such trends are also seen in igneous magmatic fractionation and so by themselves are not reliable indicators of the UHP metasomatism. The bulk composition of metasomatic garnet websterites depends strongly on the composition of the unaltered rock and so comparison of compositions between localities to determine their origin is not appropriate: comparison with the unaltered rock is required. For example, the SiO<sub>2</sub>rich nature and low FeO and TiO<sub>2</sub> contents of the Svartberget garnet websterite are thought to be, in part, inherited from the host body which has a higher silica content and lower FeO-TiO<sub>2</sub> abundance than other olivine-bearing host bodies.

Garnet websterite and biotite garnet websterite form the outermost metasomatic zones adjacent to the unaltered host. Their formation is thought to be governed by the following reactions demonstrating the addition of hydrous Si-Al-K fluids:

forsterite 
$$[Mg_2SiO_4] + SiO_2 = enstatite [Mg_2Si_2O_6]$$
 Eq. 9.1

forsterite  $[3Mg_2SiO_4] + 3SiO_2 + K_2O + Al_2O_3 + 2H_2O = phlogopite [2KMg_3AlSi_3O_{10}(OH)_2]$ 

Eq. 9.2

The veins are decussate-textured, undeformed and coarse-ultracoarse grained demonstrating brittle failure and fluid infiltration occurred after eclogite-facies dynamic recrystallisation of the host body and in the absence of directional stress. This type of metasomatism is invoked for the Kolmannskog locality also which shares similar characteristics. That at Myrbærneset has different characteristics showing no enrichment of Na<sub>2</sub>O, K<sub>2</sub>O, LILE or H<sub>2</sub>O and undergoing dynamic recrystallisation with the host body. Vein-forming metasomatism at this locality is therefore not equivalent to that at Svartberget and may not have occurred at UHP conditions. The infiltrating fluid is considered to have occurred at an earlier stage but still involving a siliceous fluid causing the consumption of olivine (*Eq. 9.1*) and phlogopite (Sekine and Wyllie, 1982). Two models of garnet websterite-forming metasomatism are therefore evident with that at Myrbærneset potentially expanding the mechanism of formation to prograde HP, eclogite-facies conditions.

# 9.1.2. Metamorphic Origin

The garnet websterites interpreted to have a metamorphic origin are envisaged to have developed their assemblage during prograde eclogitisation. Their pervasive eclogite-facies paragenesis demonstrates a fluid-present evolution otherwise transformation may not have gone to completion (e.g. Hermann, 2002; Bingen *et al.*, 2004). Localities assigned to such an origin include body domains at Nybø, Myrbærneset and Kolmannskog and the Båtneset pod. Body domains have similar textures with their associated garnet peridotites or eclogites demonstrating a shared structural and textural evolution. They are typically medium-coarse grained with granoblastic-interstitial textures and weak to strong lineations and spaced foliations in biotite-rich patches; and so are texturally distinct from the coarse-ultracoarse grained and decussate-textured, metasomatic garnet websterite at Svartberget. As the distribution of olivine in the Svartberget body is variable, even between the few thin sections taken, portions of it may form garnet websterite assemblages which would be metamorphic in origin.

The domains have undergone partial-complete dynamic recrystallisation from prior, coarser, eclogite-facies textures with other portions of the bodies. Recrystallisation is considered to have occurred during peak eclogite-facies metamorphic equilibration and not upon the later retrogressive path otherwise they would no longer be

equilibrated to UHP conditions such as the diamond-eclogite-facies conditions at Nybø. All the metamorphic domains are garnet websterites *s.s.* having diopsidic clinopyroxenes albeit more jadeite-rich than garnet peridotite clinopyroxenes. This intermediate composition between garnet peridotite and eclogite clinopyroxenes is similar to vein-distal metasomatic garnet websterites, however, and so is not a distinctive characteristic.

The domains are assigned a wholly metamorphic origin as they show no obvious evidence of developing through eclogite-facies metasomatism. At Nybø there are no UHP veining features and they lack patches rich in abundant, volatile-bearing phases. Minor phlogopite, tremolite-magnesiohornblende and carbonate are present but are considered to form part of the prograde-peak paragenesis with garnetclinopyroxene-orthopyroxene assemblages like the Svartberget body. Characteristics of metasomatic volatile-bearing phases are typically coarse-ultracoarse grain sizes, a patchy, heterogeneous distribution and random orientations. The biotite in biotiterich garnet websterites at Båtneset and Myrbærneset is medium-grained, foliated and more pervasively distributed; although so is the metasomatic biotite websterite at the southeast Svartberget contact. Such textures therefore may have occurred post-metasomatism forming in a metasomatised body margin like the biotite websterite at Svartberget with the foliation forming either during metasomatism due to stresses felt by the body margin or during later deformation and recrystallisation. The absence of analogous metasomatic features precludes the determination of the timing of deformation and textural recrystallisation relative to metasomatism at Svartberget. K-H<sub>2</sub>O components at Båtneset may have originated from an earlier episode of crustal metasomatism: although, the earlier episode of fluid activity invoked at nearby Myrbærneset is not characterised by enrichment of biotite. There are no features that confirm a metasomatic development and without relative structural information it is difficult to interpret, as with xenoliths. In the absence of strong evidence for a metasomatic origin, as with assignment of eclogite-facies rocks to the HP domain in the absence of evidence for UHPs, the Båtneset biotite garnet websterite is assigned a metamorphic origin forming due to alkali feldspar-rich protoliths and a hydrated prograde paragenesis: their CIPW normative compositions plot as monzogabbros.

Other various protoliths are invoked for the metamorphic garnet websterites studied which is in support of an origin in large, crustal magmatic bodies where different layers of cumulates and a range of bulk compositions can form. The Nybø garnet websterite samples' CIPW normalised compositions plot as meso/mela clinopyroxene norites between olivine-rich mantle rocks and olivine-poor basaltic rocks and they plot in the picrobasalt-basalt fields on a TAS diagram. They are considered to be an average bulk composition as they plot within the middle of the main garnet websterite grouping in the MgO plots (Chapter 4). They plot within the low Cr- and Sc-MgO trends with garnet websterites from Eiksunddal, Båtneset and Myrbærneset suggesting their protolith is not associated with those at other nearby, eclogitebearing localities. Although, the Kolmannskog garnet websterite plots in the high Cr and Sc trends with the Remøysunde, Årsheimneset and Svartberget garnet websterites unlike its associated peridotite and eclogite suggesting the trace element suites do not distinguish between parent bodies. The Kolmannskog garnet websterite reflects a more clinopyroxene-rich protolith as it plots on the Sr-CaO trend more similar to eclogites, it has high SiO<sub>2</sub> contents plotting in the basalt field on the TAS diagram and its CIPW normalised composition plots as an orthopyroxene melagabbro.

The overall character (MgO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub>) of the Myrbærneset garnet websterites are intermediate between the Svartberget olivine garnet websterites and garnet websterites of eclogite bodies and are enriched in Fe (having the highest Fe of any websterite) and Ca- and Na-poor. Biotite-rich domains occur and have higher MgO and K<sub>2</sub>O and lower SiO<sub>2</sub> than biotite-poor domains with a monzogabbro CIPW normative composition. The protolith is interpreted to have originated as a crustal magmatic product in a more ultramafic layer than other garnet websterites and with variable alkali feldspar contents.

### 9.1.3. Metamorphic-UHP Metasomatic Origin

The lithologies assigned a mixed metamorphic-metasomatic origin have both characteristics of a metamorphosed body domain and veining features indicative of the infiltration of external fluids and metasomatism at UHP conditions. Those assigned a mixed origin include Årsheimneset and Remøysunde, which are greatly similar, and a mixed origin is inferred for Grytting and Eiksunddal which are discussed

later. Metasomatism is not considered to be pervasive and so is not wholly responsible for the formation of the lithology as at Svartberget. It is thought the anhydrous garnet websterite assemblage would form without metasomatism and is enriched by fluid infiltration causing development of abundant volatile-bearing phases (phlogopite, tremolite, magnesite-dolomite) and laterally-zoned veins with similarities to those at Svartberget.

The garnet websterites at Årsheimneset and Remøysunde are coarse-ultracoarse grained, decussate-textured, deeply coloured and occur as distinct domain towards the edges of eclogite bodies. UHP fluid activity is invoked in formation of the ultracoarse grain sizes of the garnet websterites and their adjacent eclogite domains (e.g. Carswell *et al.*, 2003b); although such fluids may relate to the continued presence of internally-derived dehydration fluids during eclogitisation. The ultracoarse grain sizes and decussate textures demonstrate they suffered no dynamic crystallisation under stress at eclogite-facies conditions; unlike at Nybø.

A metamorphic origin is assigned due to the higher Cr, Sc and Ni contents of the garnet websterites relative to the adjacent eclogites indicating they had a more ultramafic protolith representing compositional heterogeneity of the parent body as with the metamorphic garnet websterite-garnet peridotite bodies; rather than forming through metasomatism of the eclogite by felsic fluids. Furthermore, the similar texture and inclusion-suite of garnet porphyroblasts in the garnet websterite at Årsheimneset to the eclogite suggests they underwent a similar prolonged, prograde evolution from an amphibole-rich precursor paragenesis. Considering the similarity of the lithologies at Årsheimneset and Remøysunde, their protoliths may be a common feature of the intrusive parent bodies. The Remøysunde garnet websterite samples have average major element bulk compositions for garnet websterites similar to those from Nybø, show no particular enrichment in  $K_2O$ , SiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub> or LILE and generally plot along the igneous fractionation trend. On the TAS diagram they plot in the basalt/gabbro field and have CIPW normalised compositions of melaclinopyroxene norites with other garnet websterites: this is considered to represent the non-metasomatised portion of the domain similar to that at Nybø. Metasomatised portions of the garnet websterites plot away from these average compositions such as the Arsheimneset garnet websterite domain which contains

abundant carbonate and plots as an ultramafic orthopyroxenite and in the foidite field on the TAS diagram.

The main metamorphic domains contain diopsidic clinopyroxenes but, as at Svartberget, the clinopyroxenes in the garnet websterite closest to the veins are omphacitic forming orthopyroxene eclogite assemblages *s.s.* again demonstrating the addition of Na-Al by the infiltrating fluid and its greater compositional influence closer to the vein core. The Årsheimneset vein selvage has a CIPW normalised composition of a quartz monzogabbro and plots in the basaltic andesite field on the TAS diagram near the Svartberget vein selvage suggesting an addition of SiO<sub>2</sub> compared to the average metamorphic garnet websterites at Remøysunde. However it shows no great enrichment of alkalis or LILE demonstrating only a partial signature of metasomatism.

The metasomatising, vein-forming fluid is not considered to be of the same composition as that which formed the laterally-zoned veins at Svartberget as the same hybrid composition would not develop in the absence of the olivine-bearing lithology. The absence of a vein network similar to that at Svartberget may relate to the contrasting rheology and texture of the Årsheimneset and Remøysunde bodies or to the absence of fluid accumulation and pressure increase at their margins. The absence of accumulation and metasomatism of the body margins and hybridisation of the fluid may explain the more felsic, quartz-white mica-rich nature of the vein cores at Årsheimneset. Distinct compositions of infiltrating fluids are proposed to develop at different localities through varying degrees of interaction with the body margins of different compositions.

A mixed origin is only inferred at Grytting as veining features are not observed but volatile-bearing phases are abundant and have a more patchy distribution, coarseultracoarse grain sizes and are texturally late similar to those at Årsheimneset and Remøysunde; although they may result from an inefficiently devolatilised protolith, as inferred for Båtneset. The garnet websterite is interbanded with eclogite (e.g. Carswell *et al.*, 2003b; *Fig. 2.6*) which is interpreted to represent primary, compositional layering of the protolith and reflecting a metamorphic origin to the domain. However, Carswell *et al.* (2003b) interpreted this to indicate a formation by metasomatism of the eclogite by crustal fluids (as did Carswell and Cuthbert (2003c)

for Årsheimneset) supported by the high, crustal <sup>87</sup>Sr/<sup>86</sup>Sr ratios in the garnet websterite found by Brueckner (1977) and Griffin and Brueckner (1985). The more ultramafic bulk chemistry of the garnet websterites at Årsheimneset and Remøysunde compared to their eclogites and inferred for Grytting, however, suggests this is not the case as metasomatism by crustal, felsic fluids would not produce more ultramafic compositions. High crustal <sup>87</sup>Sr/<sup>86</sup>Sr ratios may have developed during UHP partial modal metasomatism of the garnet websterite as is inferred in this study.

As structural observations were not carried out at Eiksunddal, interpretation of the type of garnet websterite is more difficult. The Eiksunddal body is thought to be a metamorphosed, layered, mafic-ultramafic complex (e.g. Jamtveit, 1987a,b). A similar origin and evolution to those at Årsheimneset and Remøysunde is suggested by its coarse-ultracoarse grain sizes, decussate texture, porphyroblastic and inclusion-rich garnets and being biotite-rich and associated with eclogite. However, no carbonate, abundant 'early' amphibole or laterally-zoned veins are observed. Biotite-orthopyroxene-rich veins are suggestive of Si-K-rich fluids and one sample is apatite-rich which may be considered evidence of fluid activity and metasomatism (Menzies and Hawkesworth, 1987; Marocchi et al., 2009). These vein samples plot on an igneous fractionation trend on the TAS diagram spanning the foidite-picrobasalt fields and their low K<sub>2</sub>O, Na<sub>2</sub>O and Ba contents suggest their protoliths were not alkali feldspar-rich. They have low MgO and SiO<sub>2</sub>, high Al<sub>2</sub>O<sub>3</sub> but average CaO contents for garnet websterites suggesting CaO in their protoliths was held in anorthosite plagioclase rather than clinopyroxene. This is supported by having CIPW normalised compositions of (clinopyroxene) mela-norites with one having higher anorthite and olivine contents being almost a meso-norite. As at Grytting, the garnet websterite is attributed a metamorphic origin but with an inferred, partial eclogitefacies metasomatic evolution although it is not clear whether metasomatism occurred at UHP conditions.

# 9.1.4. <u>Retrogression</u>

All the garnet websterites share a similar partial retrograde evolution to sub-eclogitefacies P-T conditions evidenced by the presence of plagioclase feldspar in more altered samples, symplectisation of jadeite-bearing clinopyroxenes and production of

amphibole coronas. The partial P-T paths constructed from fresh samples suggest they progress into the amphibolite-facies field as sub-2GPa temperatures are below 800°C. More pervasive retrogression at body margins such as Årsheimneset form garnet amphibolite, orthopyroxene amphibolite and amphibolite zones and represent hydration by amphibolite-facies crustal fluids.

#### 9.2. Evolution of the Continental Crust at 3.5-4GPa During Deep Subduction-Exhumation

The diamond-eclogite-facies conditions of equilibration of metasomatic rocks at Svartberget and Årsheimneset demonstrate abundant fluid is present in the continental crust at ≥3.5-4GPa UHP conditions which approximate to depths of ~120-130km (e.g. Zheng, 2009). Partial melting of the felsic country rocks is inferred to have occurred by the presence of felsic, quartz-phengite-rich veins and lenses within the mafic-ultramafic bodies and Proterozoic xenocrystic zircon in garnetite vein cores. The country rocks in the vicinity of the bodies at UHP conditions are considered to be the source of the metasomatising fluids as long distance (hm-km-scale) transport of fluid at UHP conditions is not envisaged (e.g. Philippot, 1993; Oliver, 1996; Spandler and Hermann, 2006 and references cited therein; Zheng, 2012 and references cited therein). The 3.5-4GPa pressures obtained are considered minimum peak pressures and so fluid production may have occurred at higher pressures. The scale of partial melting is unconstrained as only sporadic bodies record interaction with crustal fluids.

The timing of the partial melting is inferred to have accompanied initial decompression and temperature increase between peak pressures and peak temperatures crossing steep dP/dT, melt-producing mineral equilibria such as phg+cpx+coe ±fluid = grt+ky+melt ±kfs (Hermann and Green, 2001), the wet granite solidus (Auzanneau *et al.*, 2006) and the wet metapelite (phg+cpx+coe+fluid) solidus (Schmidt *et al.*, 2004) (see *Figure 7.4*). This stage of the P-T path occurs with the initial stages of exhumation following detachment of the continental crust from the subducting oceanic slab and subcontinental lithospheric mantle and with asthenospheric upwelling (e.g. Massonne and O'Brien, 2003; Schmädicke *et al.*, 2010). Early fluid activity may have facilitated detachment and exhumation by reducing the strength of the crust and enabling strain to be strongly partitioned (e.g. Gordon *et al.*, 2013). Intense plastic deformation and strain partitioning in fluid-rich domains may facilitate fluid flow in the crust and fluid accumulation in lower stress environments such as at the pressure-shadowed contacts of large, competent mafic-ultramafic bodies. Fluid

production may have been triggered or facilitated by exsolution of H<sub>2</sub>O and OH<sup>-</sup> from NAMs upon initial decompression or phase reaction (e.g. Zheng, 2004; Zeng *et al.*, 2009). As UHP metasomatism is not seen in every body at every UHP locality, the production of the metasomatic fluid is not considered to be associated with the transformation of quartz to coesite and the transformation to the UHP paragenesis or it would be more ubiquitous; although, the known bodies in the WGR may hail from more central portions of the parent bodies where infiltrating fluid did not reach. The depth interval of fluid production depends strongly on the geotherm of the subduction zone and lithology of the continental crust. Those calculated in this study suggest partial melting occurs even at low geothermal gradients (<10°C/km).

There is no field evidence to suggest the metamafic-ultramafic lithologies of the large bodies underwent partial melting at 3-4GPa, 665-815°C although partial melting of small mafic pods and bands is seen to have occurred in association with retrograde P-T conditions during exhumation. The point of dismemberment of the large parent bodies on the P-T path is unclear but the presence of considerable metasomatism of the Svartberget body margin along only a portion of the contact suggests it occurred after UHP metasomatism; otherwise all contacts would show similar alteration. The P-T path of samples in the Nordøyane UHP domain lies just above the wet basalt solidus (*Fig. 7.4*) indicating temperatures in the WGR were high enough throughout the eclogite-facies during exhumation.

The calculated Si-Al-alkalis-rich composition of the metasomatic fluid, formation of hydrous phases and LILE-rich signature suggests it was produced in association with the breakdown of K-bearing, hydrous, LILE-rich phase phengite which is thought to be a dominant phase in the eclogite-facies paragenesis of the country rocks (Hermann and Green, 2001; Schmidt *et al.*, 2004; Patiño Douce, 2005; Auzanneau *et al.*, 2006; Scambelluri *et al.*, 2006 and references cited therein). No UHP reaction textures or restite assemblages are observed in the retrogressed country rocks so it is difficult to establish the reactant and produced phases involved in fluid generation (particularly as many fluid-producing, phengite breakdown reaction equations can be written). The production of abundant hydrous, LILE-rich phases and LREE-rich phases and yet having characteristics of a felsic, silicate melt transporting HFSE indicate the fluid was a supercritical liquid in which no element was particularly insoluble: experiments have suggested the SCEP of felsic compositions would be reached by the 3-4GPa pressures determined in this study (e.g. Kessel *et al.*, 2005). The

composition of the partial melt in the continental crust is therefore strongly dependent on the reactant, product and residual phases. The lack of HREE-enriched garnet in vein cores suggests HREE were retained in phases in the source rocks such as garnet. Reactant phases are inferred to have been phengite, coesite, and jadeite ±kyanite ±K feldspar to produce the Si-Al-K-Na-rich composition with accessory phases such as apatite, monazite and zircon dissolving into the supercritical liquid.

Transportation of the fluid towards mafic-ultramafic bodies is thought to reflect their greater competency and shadowing of stress. With minimal reaction and chemical exchange the fluid is thought to form quartz-phengite-dominated assemblages like the pegmatitic lens at the southeast Svartberget contact. However, interaction with the ultramafic lithologies is proposed to lead to a hybrid fluid composition relatively richer in alkalis and poorer in Si; as silica reacts with olivine to form orthopyroxene (*Eq. 9.1*), for example. Upon reaction with garnet peridotite it is proposed the fluid evolves an undersaturated-saturated sialic, syenitic composition but still retains a Si-K-Al-LILE-H<sub>2</sub>O-rich composition. This interaction produces biotite and pyroxene-rich zones such as biotite websterite and biotite clinopyroxenite as seen at the Svartberget contact. The lack of garnet websterite observed at the southeast contact may be due to steeper K gradients forming biotite websterite directly from garnet peridotite reflecting a more alkali-rich melt with a lower SiO<sub>2</sub> activity (Wang *et al.*, 2008):

pyrope  $[6Mg_3Al_2Si_3O_{12}] + olivine <math>[6Mg_2SiO_4] + 5K_2O + 6SiO_2 + 10H_2O = phlogopite$  $[10KMg_3AlSi_3O_{10}(OH)_2] + Al_2O_3$  Eq. 9.3

The formation of phlogopite from orthopyroxene and garnet is invoked for the formation of the biotite clinopyroxenite vein zones by *Eq. 9.4* and *Eq. 9.5*:

enstatite 
$$[3Mg_2Si_2O_6]$$
 + pyrope  $[2Mg_3Al_2Si_3O_{12}]$  +  $2K_2O$  +  $4H_2O$  = phlogopite
 [4KMg\_3AlSi\_3O\_{10}(OH)\_2]
 Eq. 9.4

 enstatite  $[3Mg_2Si_2O_6]$  +  $K_2O$  +  $Al_2O_3$  +  $2H_2O$  = phlogopite  $[2KMg_3AlSi_3O_{10}(OH)_2]$ 
 Eq. 9.5

A strong garnetite zone is thought not to form due to the abundance of fluid at the contact preventing considerable evolution of its composition and chemical gradients. Interaction of the hybrid fluid with fresh garnet peridotite in a closed system is proposed to produce biotite-rich, lateral zones of garnet websterite, clinopyroxenite and garnetite and eclogite vein cores such as seen along interior veins at Svartberget. Following garnetite and eclogite formation the residual fluid is thought to evolve towards a more andesitic-dacitic composition as represented by the retrogressed, vein-central, quartz-plagioclaseclinopyroxene-amphibole pegmatite seen by Vrijmoed *et al.* (2013) at Svartberget which is considered to be a fluid residuum.

Upon reaction with garnet websterite the partial melt of the country rocks is proposed to evolve to a less hybridised, more intermediate alkaline monzonitic, composition. Interaction leads to the formation of lateral zones of clinopyroxenite and quartz-micabearing garnetite and abundant formation of volatile-bearing phases such as biotite and amphibole through reactions such as *Eqs. 9.4-9.6*. Carbonate phases may also form as  $CO_2$ activities increase with removal of  $H_2O$  from the fluid through equations such as *Eq. 9.7* assuming there is an inherent C component in the fluid; as indicated by the presence of magnesite and quartz in the Årsheimneset garnet websterite.

diopside  $[2CaMgSi_2O_6]$  + enstatite  $[Mg_2Si_2O_6]$  +  $SiO_2$  +  $Al_2O_3$  +  $H_2O$  = magnesiohornblende  $[Ca_2Mg_4AlSi_7AlO_{22}(OH)_2]$  Eq. 9.6

enstatite 
$$[Mg_2Si_2O_6] + 2CO_2 = magnesite [2MgCO_3] + 2SiO_2$$
 Eq. 9.7

The geometries of the veins at Svartberget suggest they are related to brittle macrofracture possibly due to the build up of fluid pressure at body margins (e.g. Vrijmoed *et al.*, 2009). The diamond-eclogite-facies conditions of equilibration demonstrate brittle fracture of ultramafic lithologies is therefore possible at UHP conditions. Where no fractures occur, no garnet websterite development is found demonstrating the requirement of space-inducing mechanisms for material transfer and alteration of the interior of the body. In the absence of brittle deformation, therefore, fluid-rock interaction in the continental crust may be restricted to body margins. Assuming fluid production occurs at the initial stages of exhumation, mafic-ultramafic lithologies may only suffer a short duration of this type of metasomatism as exhumation is thought to be rapid. Supercritical liquid-production processes would cease with the decreasing P-T conditions during exhumation and the observed products of fluid-rock interaction would no longer be produced. UHP metasomatic interaction in exhumed continental crust is therefore fundamentally limited by its rapid exhumation. This may explain, in part, the rare nature of UHP metasomatic

features in these terranes. Processes during continued subduction are addressed in section 9.5.

### 9.3. Implications for Crust-Mantle Interactions at UHP Conditions

Bulk compositions of the lithologies of the studied mafic-ultramafic bodies hosted within the WGC suggest they represent metamorphic equivalents of igneous protoliths likely to be found in large, layered, crustal, magmatic, parent bodies and can be considered as partial analogues to mantle lithologies which are more Mg-Cr-rich and ultramafic in character. Assuming chemical gradients are similar, interaction between the crustal fluid and mantle wedge should follow a similar metasomatic reaction path and produce similar assemblages and zoning patterns as observed at Svartberget. Conclusions can be drawn from this study regarding fluid-mantle interaction over the 3.5-4GPa interval (120-130km depth) at which the bodies are equilibrated, assuming pressures of fluid production and metasomatism were not greatly above those recorded in the rocks.

Fluid, on the regional scale, is expected to travel upwards toward the crust-mantle interface down the pressure gradient due to buoyancy effects (Oliver, 1996 and references cited therein; Zheng, 2004). A low degree of partial melting may not produce enough fluid to reach this horizon, particularly if long distance flow is inhibited. However, with continued subduction and increased temperatures, a higher degree of melting may enable regional-scale fluid transportation. In addition, partial melting of the uppermost portions of the subducted continental crust would provide a quantity of fluid at the interface. Therefore, assuming partial melts reach the slab-mantle interface with their initial compositions intact, the metasomatic products observed at Svartberget may be produced at the base of the mantle wedge. The slab-mantle interface is thought to be characterised by a mélange zone of metasomatised mantle (Spandler and Picard, 2013) and metasomatism of UHP mantle wedge garnet peridotite entrained in the Sulu UHP terrane, China by crustal fluids is invoked to produce the enrichment of LILE, LREE, Zr and K with growth of zircon and volatile-bearing minerals at UHP conditions (Zhang *et al.*, 2011).

Geochemical gradients are likely to be greater considering the more ultramafic character of the mantle lithologies than crustal ultramafic lithologies resulting in more severe metasomatism of the immediately adjacent mantle. A continued fluid flux to the slabmantle interface is proposed to produce a metasomatic zonation more akin to the Svartberget contact than the interior veins with thick biotite-pyroxene-rich bands extending along the slab-mantle interface and without dominant development of garnetite or garnet websterite. (see Fig. 9.1). Hybridisation of the fluid at the interface with metasomatism toward a less extreme, felsic composition is proposed as inferred for the Svartberget contact. Fluid pathways penetrating into the mantle are proposed to be characterised by the zoning pattern of the interior veins through reaction with the hybrid fluid. An impermeable nature of the mantle wedge would promote accumulation of fluid at the slabmantle interface which may then induce fracturing: brittle fracture is shown to occur in mafic-ultramafic lithologies at such depths as demonstrated by the Svartberget body. Interaction with a continuous influx of fluid would lead to columns coated by metasomatic phases and a sequence of mineralogically distinct zones (e.g. Sekine and Wyllie, 1983; Vrijmoed et al., 2013). Where fluid accumulates enabling chemical gradients to develop and fluid compositions to evolve without the influx of fresh fluid, it is proposed garnet-rich vein core zones also develop with a biotite clinopyroxenite band followed by a glimmerite selvage and biotite garnetite vein core, an eclogite vein core or a well established garnetite band closest to the fluid-rock interface. Lesser garnetite is proposed to form where fluid can escape the local system.

Olivine-poor, garnet websterite (like portions of the Svartberget and Myrbærneset bodies) is expected to form furthest from the interface behind the fluid-mantle reaction front as silica is added as represented by *Eq. 9.1* (e.g. Kiseeva *et al.*, 2012) eventually forming garnet websterite assemblages upon complete consumption of olivine. Biotite garnet websterite may also form through the reaction of *Eq. 9.2* directly from garnet peridotite. Biotite garnet websterite lithologies have been produced in experimental reactions between hydrous granitic fluid and peridotite at 3GPa (Sekine and Wyllie, 1982). Abundant amphibole, such as seen in biotite websterite selvages, may also form through hydration of pyroxene as water activities increase (*Eq. 9.6*). The considerable production of phlogopite and amphibole-phlogopite 'blackwall' rocks, which are known from contacts between ultramafic mantle lithologies hosted in the continental crust, may also occur in relation to the more ultramafic character of the mantle and steeper chemical gradients (Rampone and Morten, 2001 and references cited therein; Tumiati *et al.*, 2007; Marocchi *et al.*, 2009) as they are not seen at Svartberget.

Towards the fluid-mantle interface a zone of biotite websterite is proposed to form represented by the reaction *Eq. 9.4*. This is followed by a zone of amphibole-rich biotite

clinopyroxenite. This zone represents a relative decrease in  $SiO_2$  in the original fluid with continued reaction causing biotite to become the stable Mg-bearing phase (Sekine and Wyllie, 1982): the removal of orthopyroxene is represented by *Eq. 9.5*. This sequence of reactions perhaps reflects the mobility of Si, K, Al from the fracture wall or variable concentration within the fluid and steepness of chemical gradients. The formation of biotite websterite from olivine garnet websterite at the Svartberget contact indicates steeper K chemical gradients across the interface with the original partial melt than the hybrid fluid and lower Si/K ratios.



Figure 9.1: Diagram showing proposed banding at the interface between continental crust and mantle wedge and fluid pathways based on the banding observed at the Svartberget locality. Thinner garnetite zones are expected as fluid is expected to escape the local system preventing residual fluid compositions to be involved in alteration except along discontinuous fractures. Lenses of hybridised fluid may persist in the biotite clinopyroxenite if fluid-rock interaction is incomplete.

Local accumulation of fluids may occur in discontinuous fractures at the periphery of fluid channels enabling the development of garnetite through reaction of evolved, Al-rich fluids (*Fig. 9.1*). Local concentration of minor/trace species such as phosphate, HFSE and REE in residual fluids in relation to their mobility and compatibility in the fluid may lead to horizons rich in accessory phases such as apatite, monazite, zircon, xenotime and rutile. The abundance of accessory phases in vein cores indicates they are mobilised from the continental crust but suggests they remain close to the fluid-rock interface and garnetite-rich zones. Therefore they may only characterise small-scale pathways penetrating into the mantle wedge and not the slab-mantle interface (*Fig. 9.1*) forming concentrated deposits where residual fluids accumulate and react such as at the ends of fractures or at impermeable barriers. Concentration of accessory phases such as apatite and zircon upon interaction with the continental crust are known in other studies (Marocchi *et al.*, 2009).

Quartz-rich vein zones are not envisaged to form along fluid pathways as any fluid not consumed during metasomatism would flow further into the mantle wedge or along the slab-mantle interface along pressure gradients and react with olivine. Quartz has been generated with phlogopite in granitic fluid-peridotite experiments at low temperatures (Sekine and Wyllie, 1982) but the higher mantle temperatures would cause instability.

One consideration which must be made is that the mantle wedge may already be metasomatised by fluids produced in the oceanic crust during preceding subduction causing peridotite lithologies to no longer exist at the slab-mantle interface. In this case, fluid-rock interaction seen at Årsheimneset and Remøysunde may be a more appropriate chemical system with alteration of a less ultramafic garnet websterite assemblage. Fluid release is thought to occur continuously throughout subduction (e.g. Poli and Schmidt, 1995; Becker et al., 2000; John et al., 2008) but as the major dehydration horizon of the oceanic slab is thought to occur at shallower depths during subduction in association with eclogitisation (e.g. Philippot, 1993; Katayama et al., 2006; Spandler and Hermann, 2006; John et al., 2008; Sheng et al., 2013) considerable dehydration-related fluid may not be produced by the oceanic slab at UHP depths as indicated by the dominantly anhydrous nature of the eclogite lithologies observed and presence of stable hydrous phases. Partial melting of eclogite is thought to produce a siliceous fluid (e.g. Rehfeldt et al., 2008 and references cited therein) which may metasomatise the mantle. However, the bodies hosted in the continental crust suggest partial melting of the mafic-ultramafic lithologies does not occur at 3-4GPa conditions in cold subduction zones. Therefore there may be an interval of mantle wedge

between horizons metasomatised by dehydration-related fluids and siliceous partial melts where the mantle wedge escaped considerable fluid accumulation, infiltration and alteration. Partial melting and fluid production within the continental crust within this interval, due to its felsic nature and lower partial melting temperatures, may therefore metasomatise reasonably fresh mantle.

Transport into the mantle wedge may only occur over short distances considering the reactivity of the fluid with ultramafic mantle lithologies (e.g. Kiseeva *et al.*, 2012 and references cited therein; Vrijmoed *et al.*, 2013) forming a relatively thin but intensely metasomatised unit at the base of the mantle wedge. The considerable alteration of domains of the Svartberget body indicates pervasive alteration can occur over geologically brief periods of time probably due to the reactivity of the fluid-rock system. Alteration is considered to occur for as long as geochemical gradients are present, fluid is available and components can diffuse through the wall rock to expand the reaction front. Any long-distance transport would require channelised, rapid flow with minimal diffusive interaction with the wall rock.

Permeability of the wall rock, fluid flux, the fluid/rock ratio and duration of fluid activity will determine the degree of alteration and width of the overall metasomatic zone (e.g. Kiseeva et al., 2012) but zone widths should remain proportional controlled by the component geochemical gradients assuming the fluid composition remains constant. However, fluids would chemically evolve along the pathways and over time upon transport and interaction forming different hybrid compositions resulting in different chemical systems, reactivity and the development of different metasomatic zone lithologies, relative widths or phases. This is seen on a small scale at Svartberget when vein core assemblages and zoning patterns vary. Furthermore, episodic influx and fluid escape would alter the chemical system away from the vein zone pattern proposed which is based on one influx of fluid into ultramafic rock and without fluid escape at Svartberget. Such a closed model in the mantle wedge may not be appropriate assuming larger-scale (hm-km-scale) and longer-term fluid production and transport. The system observed may only be applicable at the furthest reaches of short-distance (m-dm-scale) fluid transport in the mantle. The fluid may also ultimately evolve into a more CO<sub>2</sub>-rich fluid following abundant formation of phlogopite and amphibole and enable the formation of carbonate phases such as magnesite and dolomite.

## 9.4. Fate of Retained Subducted Continental Crust and Implications for Mantle Metasomatism

The extent of selective exhumation of crustal terranes from 3-5GPa depths due to relative buoyancy or such controls is not constrained and it is conceivable that a considerable proportion of continental crust is retained in the mantle attached to the preceding oceanic crust or lithospheric mantle or torn off due to frictional forces. Retained continental crust would eventually experience higher temperatures and undergo higher degrees of melting than experienced by the northwestern WGC and fluids produced would have a different composition to those inferred in the exhumed terrane such as lower concentrations of alkalis. Assemblages resulting from metasomatism outside the P-T conditions experienced therefore cannot be reliably inferred from the crustal analogues observed in this study.

With continued subduction attached to the oceanic slab, the continental crust would experience higher pressures and asthenospheric temperatures: oceanic crust has been geophysically imaged to the 660km discontinuity (Spandler and Picard, 2013). The interplay between pressure increase and temperature increase becomes important in metasomatism. Increased pressures would suppress fluid release and melting by decreasing porosity and permeability, preventing the release of H<sub>2</sub>O and OH<sup>-</sup> from NAMs and promoting the transformation of fluid-producing phases to higher pressure hydrous phases such as DHMSs (e.g. Philippot, 1993). The continental crust may therefore not release any fluid to induce metasomatism and become an impermeable block which transports volatiles into the deep upper mantle and transition zone. Increased temperatures however would induce higher degrees of partial melting, rheological weakening and ductile behaviour which may enable pervasive fluid production and escape into the surrounding mantle. Release of supercritical liquids along the extended subduction channel and above mantle would produce a track of metasomatic lithologies into the deep upper mantle, such as garnet websterite and biotite-rich pyroxenite or compositionally equivalent, denser assemblages containing majoritic garnet (e.g. Ye et al., 2000a; Drury et al, 2001), Ma and Ca silicate perovskites (after pyroxenes) at the mantle transition zone (e.g. Frost, 2008) and possibly DHMSs if mantle geotherms remain low (e.g. Komabayashi et al., 2005 and references cited therein). The residual crust would evolve a less felsic bulk composition and consequently become more dense, competent and with increased melting temperatures. High degrees of removal of its felsic components may produce a mafic restite with negative buoyancy which may sink to the mantle transition zone or even into the lower mantle (e.g. Hermann and Green, 2001).

In comparison, any continental crust separated from the preceding oceanic slab but retained in the mantle would still be buoyant. If not exhumed, or only partially exhumed, following the coherent continental slab, it may remain in the upper mantle in the subduction channel and eventually become incorporated into the subcontinental lithospheric mantle of the over-riding plate and collisional margin as the subduction zone evolves and diminishes following collision. At this level, partial melting may remain at a low degree, the crust may retain a dominantly felsic lithology for a longer period of time and steadily supply metasomatic partial melts to the subcontinental lithospheric mantle.

The mantle surrounding retained continental crust is expected to be metasomatised by fluids released from its upper surface or through thermal erosion of its margins producing hybrid fluids and induce further melting of the crust. The lithological contact may become gradational and zoned on the mantle side with continued interaction as seen at Svartberget. Alternatively, it may develop a thick garnetite band with the loss of the more reactive and mobile components as in the Ivrea Zone, the Alps (Rivalenti *et al.*, 1997). A garnetite band may act as a giant corona analogous to those in partially eclogitised gabbros and granulites between plagioclase and olivine or augite (e.g. Mørk, 1986) isolating reactive components and reducing chemical gradients preventing further alteration (e.g. Mørk, 1985, 1986).

Partial melts released from the continental crust may form part of subcontinental lithospheric mantle metasomatic system known from xenoliths suites (e.g. Rehfeldt *et al.*, 2008; Jacob *et al.*, 2009) and be a catalyst for compositional change and heterogeneity in the upper mantle. Mantle xenolith lithologies include dunite, Iherzolite, harzburgite, wehrlite, eclogite, pyroxenite and websterites (Taylor *et al.*, 2003). Whilst xenoliths provide no structural indications of their formation other than small-scale veining and fluid flow, similar lithologies and features may be associated with those observed in the continental terrane. Amphibole and phlogopite are characteristic metasomatic minerals in peridotitic mantle xenoliths (e.g. Bailey, 1982; Menzies and Hawkesworth, 1987; Prouteau *et al.*, 2001; Rampone and Morten, 2001) and indicate interaction of the mantle with alkaline, hydrous melts (Zanetti *et al.*, 1999 and references cited therein). Formation of Fe-richer orthopyroxene is also a common feature of metasomatised mantle peridotite xenoliths such as from African and Siberian cratons (Rampone and Morten, 2001; Wang *et al.*, 2008; Kiseeva *et al.*, 2012 and references cited therein) with orthopyroxene formation caused by enrichment by SiO<sub>2</sub>-rich fluids (Rampone and Morten, 2001; Lorand and Gregoire, 2010;

Kiseeva *et al.*, 2012). Similarly, group A (mantle-derived) eclogites ±orthopyroxene are thought to form by interaction of depleted peridotites with TTG and carbonatite melts from oceanic crust (Taylor *et al.*, 2003); their bulk chemistries are not compatible with known magmatic rocks and are not formed directly from crystallisation of a melt in the mantle (Taylor *et al.*, 2003 and references cited therein). Phlogopite is also produced in experiments between peridotite and siliceous fluids following crystallisation of orthopyroxene depending on the SiO<sub>2</sub> content and balance between KAlSi<sub>3</sub>O<sub>8</sub> and SiO<sub>2</sub> of the siliceous fluid (Sekine and Wyllie, 1983) and in experiments between phengite and forsterite-enstatite with enstatite and recrystallised phengite at the contact (Wunder and Melzer, 2003) with a residual liquid rich in alkalis, Al, Si and poor in Mg (Wunder and Melzer, 2003). Phlogopite is therefore proposed to form at the mantle wedge at over 100km depths at temperatures below 1200°C (Wunder and Melzer, 2003).

Another widespread feature of mantle metasomatism is the enrichment of alkalis, LILE, LREE, Ti, Al, incompatible elements and S (Harte *et al.*, 1987; Lloyd, 1987a; Menzies and Hawkesworth, 1987; Wang *et al.*, 2008) and apatite is considered evidence of fluids (Menzies and Hawkesworth, 1987): such enrichments cannot be generated by partial melting of depleted source regions (Menzies and Hawkesworth, 1987). The standard model of metasomatising, siliceous, TTG fluid production is formation from low degree melts of subducted oceanic crustal slabs or delaminated lower crust (e.g. Sekine and Wyllie, 1983; Wang *et al.*, 2008): it is hard to generate alkaline and trace element-rich melts by crystal fractionation or assimilation mechanisms (Lloyd, 1987a).

The metasomatic, LILE-LREE-enriched, biotite-amphibole-rich pyroxenite lithologies observed in this study demonstrate similar products to mantle metasomatism as seen in mantle xenoliths implying a similar mantle-fluid chemical system. Metasomatic mantle xenolith assemblages may therefore form in relation to supercritical liquids produced in the subducted continental crust in addition to/rather than melts of the subducted oceanic crust; particularly considering that the continental crust is a larger alkali, Al, LILE, LREE and HFSE reservoir than the oceanic crust. There is a ubiquitous occurrence of CO<sub>2</sub>-rich fluids in subcontinental lithospheric upper mantle xenoliths (Spera, 1987). CO<sub>2</sub> may be sourced from carbonate-rich horizons in the continental crust such as marbles in addition to pelagic oceanic sediments above the preceding oceanic crust.

## 9.5. Implications for Arc Magmatism and Element Recycling in Subduction Zones

The petrographic and geochemical data presented imply the metasomatism of the mantle wedge by partial melts from subducted continental crust at  $\geq$ 120-130km depths ( $\geq$ 3.5-4GPa) is a mechanism for transfer of a range of chemical species from the continental crust back to the mantle, particularly of Si, K, Al, Na, LILE (Ba, Rb), LREE (La, Ce), HFSE (Ti, Zr, Hf, Nb), P and H<sub>2</sub>O. The absence of arc magmatism above continental subduction zones suggest these elements are not recycled to the Earth's surface but are retained by the deep, sub continental lithospheric mantle and deeper mantle levels in continental crust subducted further with the oceanic crust. The recognition of continental UHP terranes worldwide suggests this may form a considerable component to the global budget and recycling of chemical species to the mantle. Stability of hydrous phases in deeper, cooler portions of the continental crust may recycle components to the deep mantle where they may be incorporated into mantle plumes and/or contribute to the instabilities which cause them.

The absence of arc magmatism above continental subduction margins may arise due to a variety of reasons. Firstly, major dehydration of the continental crust may not take place like the oceanic crust due to water conserving prograde metamorphic reactions and the prolonged stability of dominant hydrous phases such as phengite (e.g. Hermann, 2002; Zheng, 2009). Therefore, magma-inducing fluids may not be produced in any great quantity at the critical horizon to cause extensive magmatic activity in the mantle wedge: as mentioned arc-forming fluids may be released from the subducting slab below the forearc with a lateral transport component (Philippot, 1993; Poli and Schmidt, 1995; Sorensen *et al.*, 1997; Zack *et al.*, 2001). The mica-rich, amphibole-rich nature of crustal retrograde rocks in the WGC demonstrates the continental crust retains a considerable H<sub>2</sub>O content.

Secondly, fluid production may occur but not be able to reach or penetrate the slab-mantle interface, particularly if fluid flow only occurs over short distances at UHP conditions. Metasomatism in the continental crust suggests some fluid activity at >120-130km depths but the degree, scale and spread of production is unconstrained. The localities studied demonstrate fluid activity only in the vicinity of the bodies but evidence of widespread activity at UHP conditions is not seen due to retrogressive overprinting and relative movement.

Thirdly, the absence of arc magmatism may be due to melt-inducing fluids not reaching appropriate heights within the overlying subcontinental lithospheric mantle wedge where temperatures are high enough to partially melt the mantle and so magma is not produced. This may be due to reduced permeability or increased degrees of metasomatic interaction and efficient hydration of the mantle precluding the long-distance travel or long-term survival of any fluid in the mantle (e.g. Vrijmoed et al., 2013). Fluid infiltration and interaction may not occur if high permeability pathways do not exist within the mantle: the lack of metasomatic alteration of the Svartberget body where fractures do not occur demonstrates penetration and metasomatism at UHP conditions is dependent on brittle deformation. The localities studied demonstrate brittle fracture of mafic-ultramafic lithologies occurs at ~4GPa (at <700-800°C) conditions. Therefore, fluid at the slab-mantle interface should be able to penetrate someway into the mantle wedge. Fractures may be propagated by the intruding fluid (Spera, 1987) as is invoked for the wavy veins at Årsheimneset but used up quickly if the rate of fracture propagation is low and reactivity is high. The 2m-wide metasomatic unit at the Svartberget contact suggests fluid was abundant at body contacts but the absence of abundant, residual fluid in the interior of the body suggests it would be rapidly used up in the mantle. Back-arc basin magmatism may be related to the release of fluids from the breakdown of lawsonite and phengite at UHP conditions in the subducted oceanic crust (Spandler and Picard, 2013) demonstrating the considerable transport of fluids in the mantle wedge. The absence of magmatism above continental crust therefore may be more related to fluid reactivity rather than mechanisms of long-distance transport.

In contrast to extrusive arc magmatism, post-collisional, intrusive, magmatic, maficultramafic bodies are known from the UHP Dabie-Sulu orogenic belt (e.g. Dai *et al.*, 2012; Yang *et al.*, 2012a,b). Studies invoke an origin from partial melting of a lithochemically fertile mantle source generated by metasomatism of subcontinental lithospheric mantle by hydrous, felsic partial melts from the subducted continental crust causing enrichment in LILE and LREE and adding zircon xenocrysts (e.g. Dai *et al.*, 2012; Yang *et al.*, 2012a,b); such studies demonstrate potential consequences of the mantle metasomatism proposed.

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The structural, petrographic and chemical characteristics of orthopyroxene-bearing lithologies from nine localities across the three UHP domains of the WGR, western Norway were investigated in this study to gain insight into their origin and evolution. Many of the host mafic-ultramafic bodies show evidence of infiltration of external fluids and metasomatism at HP-UHP conditions: this is dominantly expressed through laterally-zoned, biotite-rich, websterite-clinopyroxenite-garnetite veins. The petrographic and chemical characteristics of these veins were also investigated to gain insight into fluid activity and parameters at upper mantle depths. The findings of the research carried out, in relation to the aims of the project, are summarised below. *Figure 10.1* displays the overall results of thermobarometric and geochronological studies and the proposed P-T-t path of the WGR.

- 1) Garnet websterite (garnet-orthopyroxene-clinopyroxene) lithologies are observed at the Nybø, Grytting, Årsheimneset, Remøysunde, Eiksunddal, Kolmannskog, Båtneset, Myrbærneset and Svartberget localities in the WGR. They form domains of eclogite bodies at Nybø, Grytting, Årsheimneset, Remøysunde and Eiksunddal, a domain in a garnet peridotite body at Kolmannskog, an entire small body at Båtneset, a domain within a garnet peridotite-eclogite body at Myrbærneset and vein-associated selvages and domains at Svartberget, Myrbærneset and Kolmannskog. They have varying textures, grain sizes and fabrics.
- 2) Assigning metamorphic or metasomatic origins to garnet websterites relies heavily upon structural field relations. It is difficult to distinguish a clear metasomatic signature, particularly of major element variation of Si, Al, Ca, Mg which follows that of magmatic differentiation but which therefore indicates the addition of felsic components with any metasomatism. The geochemical signature of UHP metasomatism is considerable enrichment of alkalis, LILE, LREE, HFSE and P.
- 3) Laterally-zoned vein selvage garnet websterites within the garnet peridotite/olivine garnet websterite body at Svartberget are confidently assigned a metasomatic origin due to their structural association with veins which require external components to produce the mineralogies observed. Vein-associated garnet websterites at Kolmannskog and Myrbærneset are also assigned a metasomatic origin considering their structural similarities with those at Svartberget.
- 4) Some veins at Svartberget demonstrate a metasomatic origin for eclogite being comprised of garnet and omphacitic clinopyroxene and for garnetite. Both are

associated with the interaction between ultramafic and more felsic components. Clinopyroxenes within the most vein-adjacent garnet websterites at Svartberget are omphacitic, therefore, also demonstrating a metasomatic origin for orthopyroxene eclogites.

- 5) Considering the features at Svartberget, a partially metasomatic origin is attributed to garnet websterite body domains in eclogite bodies at Årsheimneset and Remøysunde where laterally-zoned veins are hosted within volatile phase-rich garnet websterite: a metamorphic garnet websterite lithology is considered to have been metasomatised at these localities as veins are small and irregularly distributed within the garnet websterite domain. A similar origin is inferred for those at Grytting and Eiksunddal where similar lithologies occur but no laterally-zoned veins are observed. The immediate vein-adjacent garnet websterites at Årsheimneset also have omphacitic clinopyroxenes and therefore form orthopyroxene eclogites *s.s.*.
- 6) A wholly metamorphic origin is attributed to the garnet websterite body domains at Nybø, Kolmannskog and Myrbærneset and the Båtneset body which show no features strongly indicative of metasomatism; although early episodes of K-H<sub>2</sub>O-enrichment may have occurred at Båtneset and Myrbærneset which have biotite-rich lithologies. Clinopyroxenes in these garnet websterite body domains are not omphacitic suggesting orthopyroxene eclogites are metasomatic features related to the addition of Na-Al to garnet websterites/olivine garnet websterites at UHP conditions.
- 7) The four mafic-ultramafic bodies in the northern Nordøyane UHP domain (Kolmannskog, Båtneset, Myrbærneset and Svartberget) and Nybø in the southern Nordfjord-Stadlandet UHP domain underwent partial dynamic recrystallisation during prograde metamorphism under eclogite-facies conditions forming their finer grained, granoblastic textures. This occurred prior to metasomatism and vein formation at Svartberget and Kolmannskog and afterwards at Myrbærneset where the veins have also been recrystallised.
- 8) Diamond-eclogite facies UHP conditions were obtained from garnet websterites at Svartberget of 3.85±0.27GPa, 792±9°C, at Årsheimneset of 3.53±0.1GPa, 696±6°C and at Nybø of 3.71±0.10GPa, 737±9°C despite the use of conservative thermobarometric calibrations. These results are consistent with the identification of diamond at Svartberget but not the 5.5GPa pressures for the vein system of Vrijmoed *et al.* (2006).
- 9) Coesite-eclogite-facies UHP conditions were obtained from garnet websterites at Grytting of 3.14±0.1GPa, 665±6°C and at Remøysunde of 3.64±0.25GPa, 815±54°C.
   Coesite-eclogite-facies conditions of 3.02±0.1GPa, 754±5°C were also obtained from

the garnet websterite from the Kolmannskog locality within the HP domain. Therefore it is proposed the eastern limit of the Nordøyane UHP domain be extended to incorporate this locality (see *Fig. 7.5*).

- 10) HP P-T conditions were obtained a garnet websterite from at Myrbærneset of
  2.64±0.23GPa, 728±9°C; however, previous studies have obtained UHP conditions from
  this locality so it is assumed the rocks analysed re-equilibrated to retrograde HP
  conditions due to their finer grain size.
- 11) The P-T estimates suggest the Stadlandet-Nordfjord UHP domain also reached diamond-facies UHP conditions similar to those in the Nordøyane UHP domain (see *Fig. 10.1*) reducing the regional pressure gradient between the north and south indicated by mineralogical indicators.
- 12) Collated P-T paths (see Fig. 10.1) suggest the northern Nordøyane UHP domain experienced ~100°C higher temperatures than the southern Nordfjord-Stadlandet UHP domain and experienced a slightly higher geothermal gradient of 7-8°C/km compared to ca. 6°C/km. This is consistent with the established regional gradient although with a lower temperature gradient: temperatures do not reach the >850°C temperatures indicated by Kylander-Clarke et al. (2008).
- 13) Metasomatism of all bodies except for Myrbærneset is considered to have occurred at UHP conditions and prior to peak temperature conditions as there are no features indicative of a metamorphosed, prograde evolution and both metamorphic and metasomatic lithologies are equilibrated to peak temperatures (which occur at UHP conditions): no increases in temperature were recorded in their retrograde paths. The veins at Myrbærneset do not record UHP conditions and their textures suggest they formed earlier in the evolutionary path of the mafic-ultramafic bodies so a UHP formation is not evident.
- 14) Metasomatism of the Årsheimneset body is dated at 414±5.6Ma by U-Pb ages of zircons in vein cores and metamorphic equilibration to peak conditions (~700°C, 3.5GPa) is dated at 411.4±5.3Ma by U-Pb ages of zircons in the eclogite. This age is ~5Myr older than other studies in this area. Cooling through rutile closure temperatures (≥400°C) is dated at *ca*. 400-~385Ma; similar to the ages of the amphibolite-facies gneiss in the WGR (e.g. Gordon *et al.*, 2013).
- 15) Metasomatism of the Svartberget body is dated at  $410\pm2.6$ Ma and cooling through rutile closure temperatures ( $\geq400^{\circ}$ C) at *ca*. 390-380Ma.





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- 16) Fluid activity in the country rocks at Svartberget continued to *ca*. 400-390Ma in association with retrograde metamorphism as indicated by the U-Pb ages of zircon growth zones and monazites in gneiss and leucogneiss.
- 17) Laterally-zoned veins, except for those at Myrbærneset, are proposed to form from external fluids as they require components not in the host rocks to form the mineralogies observed. Mineralogies and whole rock geochemistries indicate they are related to an Si-Al-K-H<sub>2</sub>O-rich fluid thought to be generated in the adjacent country rocks in association with the breakdown of phengite. Vein development is thought to be characterised by open transfer of components between metasomatic zones as indicated by models of mass balance calculation and strongly influenced by the local chemical system as indicated by highly variable mineralogy along the length of veins.
- 18) The fluid responsible for UHP metasomatism is thought to be a supercritical liquid with felsic, silicate melt-like characteristics in order to precipitate zircon and transport zircon xenocrysts and HFSE and with a high water content to form abundant hydrous phases and transport LILE. The composition of the metasomatic fluid added to the garnet peridotite body at Svartberget is calculated to be 48-60% SiO<sub>2</sub>, 17-27% Al<sub>2</sub>O<sub>3</sub>, 3-11% K<sub>2</sub>O, <10% MgO, CaO and FeO, 3-6% Na<sub>2</sub>O, <4% P<sub>2</sub>O<sub>5</sub> and <1% TiO<sub>2</sub> and MnO with an overall undersaturated-saturated sialic, syenitic character which varied locally upon evolution through interaction with the wall rock. It is considered to be a hybrid fluid formed through interaction of the felsic, partial melt of the host country rocks with the margin of the body as its composition is similar to other hybrid fluids produced in reaction between granite and peridotite. Residual fluid following vein formation is thought to evolve towards a more intermediate alkaline to dacitic-andesitic composition. Infiltrating metasomatic fluids are thought to develop different hybridised compositions at different localities due to different, fluid-rock chemical systems.
- 19) The studied continental fluid-rock system demonstrates considerable fluid-rock interaction can occur at UHP conditions in subduction zones and implies a zoned metasomatic unit may form at the interface between the continental crust and mantle wedge at depths of ≥120-130km in continental subduction zones. As with vein development, its thickness is dependent on the abundance of fluid and duration of fluid availability. Interaction transfers considerable alkalis, trace elements and water to the deep mantle wedge by forming phases such as micas and amphiboles hydrating anhydrous peridotites. The zoning pattern is proposed to be garnet peridotite to olivine garnet websterite to garnet websterite to biotite
clinopyroxenite with lenses of garnetite with glimmerite selvages and concentrations of accessory phases including zircon, monazite, apatite and rutile. Outer zones may be thin depending upon the steepness of the chemical gradients.

- 20) Fluid pathways penetrating into the mantle will also be characterised by this zoning pattern although with different relative widths and possibly with more well-developed, central garnetite or eclogite zones if fluids accumulate and evolve. Penetrative fluids are considered to have a hybrid composition formed through reaction of the crustal partial melt with the margin of the mantle.
- 21) Pyroxene-rich, biotite-amphibole-rich mantle xenoliths may sample portions of banded, metasomatised mantle or >m-scale veining structures associated with fluid pathways. The metasomatic fluids which characterise the subcontinental lithospheric mantle as observed in xenoliths may relate to fluids sourced from the retained, subducted continental crust which is a larger reservoir of the element species enriched during mantle metasomatism compared to the oceanic crust.

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Firstly, samples of all rocks types and major phases at localities such as Grytting, Eiksunddal and Kolmannskog need to be chemically analysed in order to characterise the localities and to further establish the chemical relationships between the lithologies. More systematic sampling of all host mafic-ultramafic bodies and vein zones would help to determine the scale of lithological heterogeneity and subtle compositional domains in relation to metasomatic features, lithological boundaries and body contacts. Full REE whole rock analyses should be carried out to analyse depletion/enrichment patterns and particularly Eu anomalies to help determine body protoliths and phase involvement in fluid production. In addition, trace element concentrations of different minerals should be obtained and major phase crystal maps produced of major and trace elements to determine crystal-wide compositional variation that may not have been picked out by linear profiles and to find the 'chemical cores' of the crystals. This will enable more reliable thermobarometric estimates and the better determination of the chemical evolution and growth of the rocks. Furthermore, better trace element concentrations of accessory phases (zircon, monazite and rutile) should be obtained using a more sensitive analysis technique, such as LA-MC-ICP-MS, including a full REE analysis to investigate REE patterns to determine parameters of their crystallisation and to obtain reliable Ti and Zr contents to further estimate temperatures using the Ti-in-zircon and Zr-in-rutile thermometers (Ferry and Watson, 2007).

With regard to better interpreting the origin and evolution of garnet websterites and orthopyroxene eclogites, more observations need to be made from other localities in the WGR and other UHP terranes to better constrain the full range of variation in their occurrence, structure, mineralogy and texture. Samples from different mineralogical domains and positions in relation to lithological contacts and veining features should be studied and compared to determine distinguishing features and a geochemical signature which can be associated with metasomatism.

To progress the chemical characterisation of fluid-rock interaction, study should focus on the Svartberget body where the effects of metasomatism are most clear. The veins and websterites need to be mapped in detail with particular attention to the widths, mineralogies and zoning patterns of the veins and selvages and their interaction with the southeastern contact zone and other contacts around the body. The bulk geochemistry of

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the mineralogical and textural variation of the vein network needs to be systematically analysed to determine the lateral and temporal change of the fluid-rock chemical system. Such a study would also enable the full range of the zoning produced and chemistries to be defined and may help discern evolution of the metasomatic agent and any link between zone patterns and widths, garnetite mineralogies and the presence of restites. The immediately-adjacent host body rocks should be analysed with the veins as it is important to know the starting composition of unaltered rock and to determine any cryptic metasomatism or leaching that may have occurred next to the modal metasomatism of the veins. More mass balance calculations should be carried out on more vein systems to better constrain the average composition of the additive and multiple analyses should be carried out on the same veins and host body to be certain that whole rock data used in the calculations are representative. Following further study of the Svartberget locality, compositions of metasomatic additives of veins from Årsheimneset and Myrbærneset should be calculated to establish the variation in the chemical systems of other localities.

More detailed and systematic sampling and analysis of the zoned unit at the southeastern contact of Svartberget is required to gain further insight into the sequential chemical change that occurred at the body contact, to carry out mass balance calculations to compare the fluid-rock system to that to the interior veins and to compare the felsic, pegmatitic lens with mass balance-calculated fluid compositions to determine if it represents the metasomatic fluid. A detailed study of the immediately adjacent host gneisses and schists at UHP localities, particularly those in pressure shadows and country rock-body contacts, and body margins is required with a focus on finding features related to UHP fluid activity and metasomatism of body margins to associate production of the metasomatic fluids with the country rocks and their composition and to carry out more mass balance calculations.

Study of multiphase solid inclusions in more thin sections from the localities studied and from other UHP localities which show UHP metasomatism is required to identify inclusions which may have formed from trapped melt and may be used to re-calculate the fluid composition. Observations of such inclusions would also lead to the potential discovery of microdiamonds and coesite or convincing evidence of prior coesite which would help increase reliability of thermobarometric estimates and association of metasomatic features to UHP conditions.

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P-T estimates of metasomatism and peak metamorphism could be progressed by repeating EMPA and graphical analysis process on the samples analysed and analysing more samples to obtain more robust areas of intersection and reliable estimates. In particular, useable EMP data should be obtained from the olivine garnet websterite from Myrbærneset to compare P-T estimations with the garnet websterite and vein, from the vein selvage orthopyroxene eclogite at Arsheimneset and from the garnet websterite at Remøysunde which are considered unreliable estimations. Rocks from new localities should be analysed to further establish the regional gradient, particularly from the middle Sørøyane UHP for which only one P-T estimate is presented in this study. In addition, the methodology employed in this study should be used on the EMP data of other studies to obtain more comparable results. Finally, more altered, retrogressed rocks and reaction assemblages should be analysed to further define the retrograde P-T path to better constrain exhumation of the terrane. To aid thermobarometry, Mössbauer spectroscopy should be carried out on clinopyroxene separates from each rock in order to accurately determine Fe<sup>3+</sup> contents. A small study should also be carried out on separates of other minerals, such as garnet and orthopyroxene, to determine if the other methods of Fe<sup>3+</sup> calculation are accurate and reliable. Whole rock data should be used with programmes such as THERMOCALC to compare the results of the graphical analysis study and determine if they are equivalent techniques at such P-T-X conditions.

To build up a bigger picture of the regional timing of metasomatism, accessory phases in rocks at the other key localities should be analysed where possible and more reliable, robust ages for metamorphism and metasomatism at Årsheimneset should be obtained by analysing more zircons. Suitable grains included within the cores of garnet porphyroblasts should also be targeted in a bid to determine the timing of prograde and peak-pressure metamorphism. Analyses should also be carried out on the metasomatic zones of the southeast contact at Svartberget to determine the age of metasomatism at the body contact and to compare it with the age of the veins. Finally, U-Pb-bearing phases in more retrograde samples, including retrograde phases such as titanite, should be analysed along with retrograde thermobarometric estimates to better refine retrograde P-T paths.

Finally, stable isotope studies, such as oxygen and strontium isotopes, should be carried out on vein systems, initially at Svartberget and Årsheimneset where they are best established, to determine the scale of fluid-rock interaction and aid determining the source of the metasomatic agent.

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