

**METAMORPHIC EVOLUTION OF THE
ELEKDAG ECLOGITES (CENTRAL PONTIDES)**

M.Sc. THESIS

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Department of Solid Earth Sciences

Geodynamics Programme

NOVEMBER 2019

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Thesis Advisor: Prof. Dr. Gultekin TOPUZ

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**ELEKDAĞ OFİYOLİTLERİNİN (ORTA PONTİDLER)
METAMORFİK EVRİMİ**

YÜKSEK LİSANS TEZİ

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To my spouse,

FOREWORD

First of all I would like to thank Professor Dr. Gultekin Topuz, my advisor for giving me the opportunity to work on this project with him as well as for his help and patience with the preparation of this thesis. I would like to also thank the committee members, Prof. Dr. Aral Okay for accompanying me during field work in the fall of 2014 as well as for his consideration of this thesis, Prof. Dr. Namik Aysal for his time and consideration in reviewing this Thesis. I am grateful for the assistance of Mahir Altinbaga in preparing thin sections for analysis, Dr. Mary Leech for the use of the SFSU mineralogical facilities as well as Bob Jones at the Stanford University Microprobe and nano-characterization laboratory.

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November 2019

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(Geoscientist)

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ABBREVIATIONS

Ab	: Albite
Al	: Aluminum
Alm	: Almandine Garnet
Amph	: Amphibole
Ar	: Argon
ASI	: Aluminum Saturation Index
BS	: Blueschist
C	: Centigrade/ Celsius
Ca	: Calcium
Cal	: Calcite
Chl	: Chlorite
CM	: Centimeter
cpfu	: cations per formula unit
Cpx	: Clinopyroxene
Di	: Diopside
EC	: Eclogite
EBS	: Electron Backscatter Spectroscopy
EDS	: Energy Dispersive Spectroscopy
Ep	: Epidote
Fe	: Iron
Gln	: Glaucophane
Grs	: Grossular Garnet
H₂O	: Water
HP	: High Pressure
HP-LT	: High Pressure-Low Temperature
HREE	: Heavy Rare Earth Elements

HT-LP	: High Temperature- Low Pressure
IAES	: Izmir-Ankara-Erzincan Suture
Jd	: Jadite
K	: Potassium
km	: kilometer
LOI	: Loss on ignition
LREE	: Light Rare Earth Elements
Lws	: Lawsonite
Ma	: Million years ago
Mg	: Magnesium
min	: minute
mm	: millimeter
Na	: Sodium
nm	: nanometer
O	: Oxygen
OH	: Hydroxite
Omph	: Omphacitic clinopyroxene
Phg	: Phengitic Mica
PPL	: Plain Polarized Light
ppm	: parts per million
PT	: Pressure-Temperature
Pump	: Pumpellyite
Pyp	: Pyrope Garnet
Rt	: Rutile
SEM	: Scanning Electron Microscope
Si	: Silicon
Sps	: Spessartine Garnet
Ttn	: Titanite/Sphene
µg	: microgram
X_{Mg}	: Magnesium number

XPL : Cross Polarized Light
WDS : Wavelength Dispersive Spectroscopy
Wt : Weight

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METAMORPHIC EVOLUTION OF THE ELEKDAG ECLOGITES

SUMMARY

The topic of the thesis is focused on the metamorphic and tectonic evolution of high-pressure metamorphic eclogites and by extension, the surrounding meta-lherzolite and serpentinite of the Elekdag Ophiolite within the Central Pontide Mountains of Turkey. The goal of this study is to continue the work of previous authors in better understanding the timing and conditions of metamorphism and their associated tectonic events through field mapping, detailed petrographic and microstructural analysis, investigation of stable equilibrium assemblages with wavelength-dispersive x-ray, energy-dispersive, angle-selective backscatter spectroscopies coupled with isochemical phase equilibria diagrams and conventional cation exchange geothermobarometry analysis.

The Elekdag Ophiolite and other related ophiolites in the Central Pontide Mountains of Northern Anatolia are interpreted as being mantle and crust accretionary prisms which formed along the southern edge of the Laurasian Supercontinent prior to the closure of the Paleotethys Ocean and the Formation of Pangea together with the Gondwana Supercontinent. Elekdag is closely associated with the larger Cangaldag Island Arc and the larger Domuzdag HP mélange. During subduction in the Cretaceous the Elekdag Complex experienced up to eclogite-facies metamorphic conditions.

Metamorphic rocks present in the study area include typical ophiolite assemblages with the addition of HP-LT blueschists, greenschist facies mineral assemblages, serpentinite, mica schists and HP-MT eclogites. The HP-LT units are contained within lenses along the boundary of the serpentinite body and are also in contact with the neighboring Domuzdag Complex. Typical mineral assemblages of the eclogites are garnet + omphacite + glaucophane + clinozoisite + white mica $\pm$ lawsonite $\pm$ tourmaline $\pm$ rutile. Blueschist mineral assemblages consist of garnet + glaucophane + clinozoisite + white mica + chlorite $\pm$ quartz $\pm$ lawsonite $\pm$ rutile. Most primary mineral assemblages have been heavily overprinted with lower grade hydrous mineral assemblages. Retrogression occurred due to lower pressures and temperatures during exhumation and the extensive infiltration of hydrothermal fluids within the rocks characterized by chlorite. Typical retrograde phases are glaucophane (for eclogites), as well as the common greenschist facies assemblage of chlorite + albite + clinozoisite $\pm$ stilpnomelane. The metabasite samples investigated show mid-ocean ridge basalt affinities (MORB) with at least one protolith being a cumulate. Constraints on maximum pressures and temperatures of metamorphism, and by extent depth of subduction and subsequent exhumation were inferred based upon geochemical and petrographic analysis.

Two distinct samples were modeled in isochemical phase diagrams. One, a cumulate metabasite yielded maximum PT conditions of 21 ± 2 kbar at 360 ± 50 °C based upon garnet and phengite compositional isopleths as well as a maximum temperature of

metamorphism of 400 °C according to Fe/Mg cation exchange thermometry between garnet and clinopyroxene. A second modeled sample yielded lower pressure ranges of 10 ± 2 kbar at a higher temperature of 500 °C which is in contrast to similarly thermometry results which give maximum temperatures of ~ 450 °C.

ELEKDAĞ EKLOJİTLERİNİN METAMORFİK EVRİMİ

ÖZET

Orta Pontidler’de yer alan yüksek basınç altında oluşan Elekdağ Ofiyoliti eklojitleri ile metalezolit ve serpantinitlerin metamorfik ve tektonik evrimi, bu tezin konusunu oluşturmaktadır. Çalışmanın amacı, önceki çalışmacıların araştırmalarına devam ederek, haritalama, detaylı petrografik ve mikrotektonik analiz yöntemlerine ilave olarak, dalga boyu dağınımlı x ışınları, enerji dağınımı ve geri saçınımlı spektroskopi metotlarını kullanarak izokimyasal faz denge diyagramları oluşturmak ve konvansiyonel katyon değişimi jeotermobarometri analizi yapmak suretiyle sabit mineral topluluklarını araştırarak, metamorfizma zamanı ve şartları ile tektonik olayları daha iyi ortaya koyabilmektir.

Elekdağ Ofiyoliti ve Orta Pontidler ilgili diğer ofiyolitler, Paleotetis Denizi’nin kapanmasına bağlı olarak Lavrasya’nın güney kısımları ile Pangea kıtası boyunca meydana gelen manto ve kabuk yığışım prizması şeklinde yorumlanmışlardır. Elekdağ, daha büyük olan Çangaldağ Ada Yayısı ve Domuzdağ yüksek basınç melanji ile ilişkilidir. Elekdağ Kompleksi, Kretase’deki dalma süresince eklojit fasiyesi metamorfik koşullarına maruz kalmıştır.

Çalışma alanındaki metamorfitler, tipik ofiyolit minerallerinin yanı sıra yüksek basınç-düşük sıcaklık mavişist ve yeşilşist mineralleri, serpantinit, mikaşist ve yüksek basınç-orta sıcaklık eklojitlerinden meydana gelirler. YB-DS birimleri, serpantinit sınırları boyunca mercekler içinde ve Domuzdağ Kompleksi ile ilişkili halde bulunmaktadır. Eklojitlerin tipik mineralleri granat + omfasit + glokofan + klinozoisit + beyaz mika ± lavsonit ± turmalin ± rutildir. Mavişist mineralleri ise granat + glokofan + klinozoisit + beyaz mika ± klorit ± kuvars ± lavsonit ± rutildir. Birçok temel mineral topluluğu, düşük dereceli su içeren mineral toplulukları tarafından şiddetli bir biçimde üzerlenmiştir. Retrogresyon, yüzeyleme ve kloritin varlığı ile karakterize edilen hidrotermal suların kayalar içerisine sızması ile daha düşük basınç ve sıcaklıklara bağlı olarak meydana gelmiştir. Karakteristik retrograd fazları, glokofanlar (eklojitler için) ve buna ilaveten yeşilşist fasiyesi mineralleri olan klorit ± albit ± klinozoisit ± stilpnomelandir. Çalışılan metabazit örnekleri, en az biri kümülat olan protolit içeren okyanus ortası sırtı bazaltlarını (MORB) işaret etmektedir. Azami metamorfizma basınç ve sıcaklıkları ile dalma derinliği ve daha sonraki yüzeyleme, jeokimyasal ve petrografik analizler ile belirlenmiştir. Ayrıcı nitelikteki iki numune, izokimyasal faz diyagramları ile modellenmiştir. Bunlardan biri olan kümülat metabazit numunesinin, granat ve fenjit bileşim izoplethleri baz alınarak, azami 21 ± 2 kbar ve 360 ± 50 °C basınç ve sıcaklık koşulları altında olduğu ve granat ile klinopiroksen arasındaki Fe/Mg katyon değişim termometrisine göre azami metamorfizma derecesinin 400 °C olduğu görülmüştür. Modellenen ikinci numunenin ise 10 ± 2 kbar gibi daha düşük basınç ve azami ~450 °C sıcaklık değerini veren termometri sonuçlarının aksine 500 °C gibi daha yüksek sıcaklık koşulları altında metamorfizmaya uğradığı sonucuna varılmıştır.

1. INTRODUCTION

The following chapter will provide a brief overview of the importance of ophiolitic rocks. The goal of the study will be introduced and the general geologic setting of the study area within the greater framework of Turkey will be described.

Ophiolites are sections of oceanic crust and underlying upper mantle which have been obducted and emplaced upon the margins of continents during continental-oceanic plate collisions. They are commonly found as incomplete units along active continental margins or former intracontinental suture zones (Robin Gill, 2010). Commonly accepted stratigraphy of the ophiolites from bottom to top are mantle peridotite, cumulate gabbro/peridotite, isotropic gabbro, sheeted dike, pillow basalts and ocean floor sedimentary rocks. Also common along subduction zone margins are metamorphic units which have been subducted to depth and subsequently exhumed to the surface. They are characterized by the presence of high-pressure low to medium-temperature mineral assemblages of blueschist and eclogite facies. Careful analysis of ophiolitic rocks and their associated metamorphic terranes can reveal much about the history of subduction and collision that occurred during their respective orogenies.

1.1 Purpose and Scope

This study deals with the metamorphic pressure and temperature evolution of the Elekdağ ophiolite and associated eclogite-facies rocks. The Elekdağ area is located within the Central Pontides of northern Turkey (Figure 1.1). The objective of this study is to continue the work of previous authors (Ustaömer and Robertson 1999; Alther et al, 2004; Okay et al, 2006; Tuysuz, 1990; Dönmez et al, 2014; Günay et al, 2016) in better understanding the timing and conditions of metamorphism and their associated tectonic events through detailed petrographic and microstructural analysis, investigation of stable equilibrium assemblages with wavelength-dispersive x-ray, energy-dispersive, angle-selective backscatter spectroscopies coupled with isochemical phase equilibria diagrams and conventional cation exchange geothermobarometry analysis. Before describing the geological characteristics of the

Elekdağ area in Central Pontides, the tectonic framework of Turkey will be summarized below, so that the importance of the Elekdağ area can be understood from the tectonic view of point.

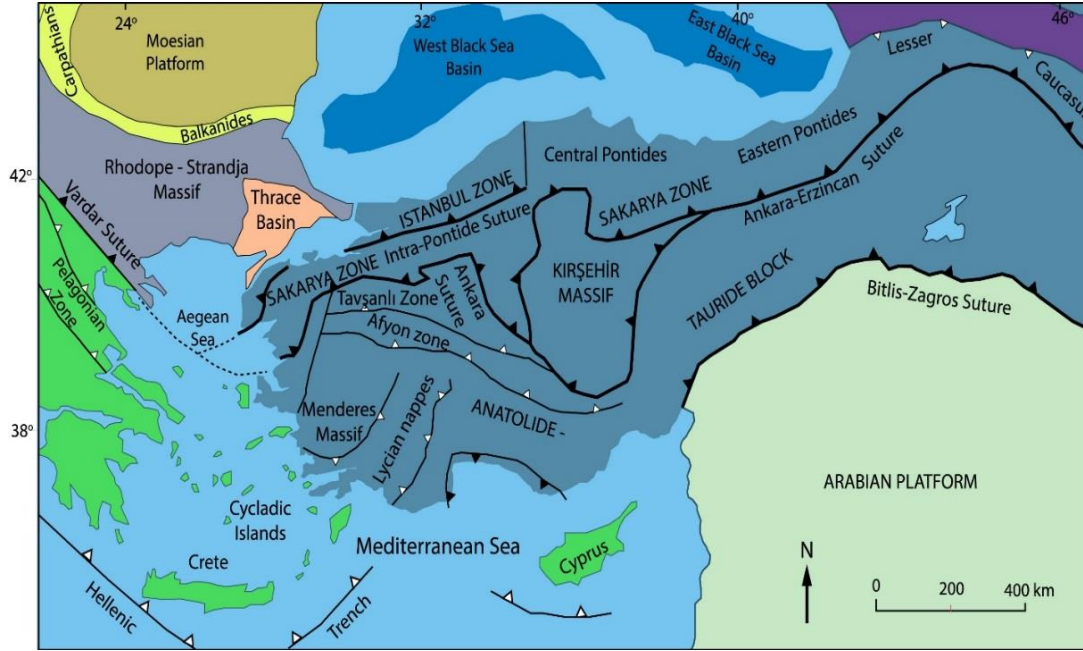


Figure 1.1 : General geologic map of the major tectonic units of Turkey (after Okay and Tuysuz, 1999).

1.2 General Geology of Turkey

Turkey is an amalgamation of several continental slivers, fragments and magmatic arcs, which were detached from Gondwana and accreted to Baltica at different times (Şengör and Yılmaz, 1981; Okay and Tuysuz, 1999; Okay et al, 2006; Okay and Nikishin, 2015) (Figure 1.1). These continental fragments are the Anatolide-Tauride block, the Kırşehir Massif, the Sakarya Zone, the Istanbul Zone and the Strandja-Rhodope zone. They are separated by a large-scale suture zones (Figure 1.1). The Arabian platform represent the northern margin of the Gondwana, and the East European craton represents the core of Laurussia. These suture zones are Bitlis Suture, Izmir-Ankara-Erzincan suture, Intra-Tauride suture and Intra-Pontide suture zone. All these suture zones are related to the consumption of branches of the NeoTethyan oceans. These sutures all represent north directed subduction. The Sakarya, İstanbul and Strandja zones are collectively referred to as the Pontides.

As the study area is located within the Central Pontides, which is part of the Sakarya Zone, the geological features of each zone of the Pontides are described below.

1.3 The Istanbul Zone

The Istanbul Zone is located in northwestern Turkey from Istanbul in the west to just east of Cide in the east (Figure 1.1). It is bordered to the south and east by the Sakarya Zone with the contact being the Intrapontide Suture. The western limit of the Istanbul Zone is the West Black Sea Fault which separates it from the Strandja Massif in Thrace. The Istanbul Zone consists of a Neoproterozoic crystalline basement (Okay and Tüysüz, 1999; Ozgul, 2012; Okay, 2014). This basement lithologically, geochemically and chronologically resembles the Pan-African basement of the Gondwana super-continent (Okay, 2006) and as such is considered a continental fragment of Gondwana. Overlying this Precambrian basement is a well-developed, continuous sequence of Paleozoic sediments which progress from the Ordovician continental deposits of the Polonezkoy Group to Lower Carboniferous (Viséan) turbiditic flysch (Trakya Formation) and coal bearing clastic deposits (Görür et al, 1997; Okay, 2006). The Paleozoic units are deformed by a NE verging deformation of thrusting and recumbent folds from plate convergence during the Variscan orogeny.

The Paleozoic sequence is unconformably overlain by Triassic sedimentary rocks and locally intruded by the Sancaktepe Permian acidic granodiorite (c. 255 Ma, Aysal et al, 2018). The Triassic sedimentary rocks are composed of a transgressive sequence beginning with red sandstones which are overlain by shallow marine sediments, deep marine limestones and finally by Late Triassic pelagic shales and sandstones (Okay, 2008). In the western region of the Istanbul zone unconformably overlying the Triassic sequence are Late Cretaceous to early Cenozoic volcanic and carbonate units however the eastern portion of the Istanbul zone contains a thick column of Middle Jurassic to Eocene sedimentary succession (Okay, 2008).

The Istanbul Zone is commonly regarded part of the Avalonia (Okay et al, 2012). During the Paleozoic, there was only Permian magmatism in the Istanbul Zone. The exact nature of the Intrapontide suture, which separates the Sakarya and Istanbul zones is contentious in terms of the timing of accretion. Proposed ages vary from Carboniferous to Cretaceous (Okay and Topuz, 2017; Tüysüz, 1999).

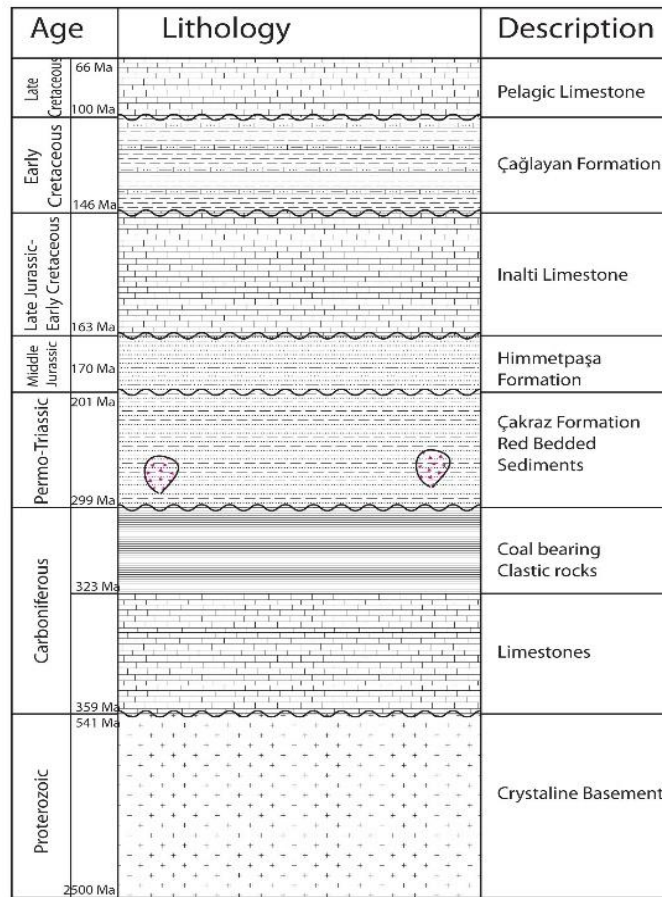


Figure 1.2 : Generalized stratigraphic column of the Istanbul Zone (after Okay and Topuz, 2017; Okay et al., 2013 and 2014) (Modify according to Okay, 2008 style).

1.4 The Sakarya Zone

The Sakarya Zone is geographically, the largest of the three tectonic terranes of the Pontide Mountains. It is roughly 1500 km long and spans from northwestern Anatolia to the southeast coast of the Black Sea (Fig. 1.1). Its southern extent is the Izmir-Ankara-Erzincan Suture which delineates the boundaries with the Anatolide-Tauride Block in the east and west and the Kirsehir Massif in central Anatolia. The suture is defined by Early Jurassic ophiolite and accretionary complexes as well as Late Cretaceous accretionary complexes. Continental collision with the southern continental fragments have occurred during Early Eocene.

The basement of the Sakarya Zone is composed of a mix of pre-Jurassic units. These are (i) Early Carboniferous high grade metamorphic units (Topuz et al, 2004a; Topuz and Altherr, 2004; Topuz et al, 2007) , (ii) Devonian to Permian plutonic units (Okay

et al, 1996; Okay et al, 2006; Aysal et al, 2010; Sunal 2013), (iii) Permo-Triassic accretionary complexes (Okay and Göncüoğlu, 2004; Topuz et al, 2004, 2014, 2018). These basement rocks only outcrop in a few locations throughout the length of the Sakarya Zone (Figure 1.3). The relations between these units is unclear due to strong Alpine orogeny deformation. The oldest plutonic basement units are the Early Devonian granitoids from NW Anatolia with zircon ages of 399 ± 13 and $397.5 \pm$ Ma (Okay et al, 1996; Okay et al, 2006; Aysal et al, 2010; Sunal 2013). The Carboniferous to Permian acidic granitic rocks range in age from 340 to 270 Ma (Topuz et al, 2010; Dokuz, 2011; Kaygusuz et al, 2012 and 2014). The metamorphic basement is exposed in Pulur, Kurtoğlu and Sarıcakaya areas (Figure 1.3), and consists of high-temperature/middle to low pressure rocks comprising gneisses, mica schists, amphibolites (Okay et al, 1996, Topuz et al, 2004a; Topuz and Altherr, 2004; Topuz et al, 2007). U-Pb monazite ages from the Pulur Massif yielded ages of 331-327 Ma for the peak of the high-temperature metamorphism. The Permo-Triassic accretionary complexes occur in several isolated locations such as Pulur, Ağvanis, Tokat, Küre and Yenişehir areas (Figure 1.2). They comprise mainly greenschist-facies metabasite, phyllite, marble and subordinately metachert and serpentinite. In addition, there are local blocks of eclogite and blueschists (Okay et al, 1996; Okay and Monie, 1997; Okay et al, 2002; Topuz et al, 2004a; Topuz and Altherr, 2004; Topuz et al, 2007).



Figure 1.3 : Map showing main pre-Jurassic basement exposures of the Sakarya Zone (Modified from Okay and Topuz, 2017).

All of the pre-Jurassic basement is unconformably overlain by Lower to Middle Jurassic volcanic-clastic rocks in the eastern part of the Sakarya Zone and by near-shore sedimentary sequences in the western Sakarya Zone. This is in turn conformably overlain by Upper Jurassic-Lower Cretaceous carbonate units which are present through the entire Sakarya zone. These limestones are overlain by middle to late

Cretaceous sandstones and shales possibly due to uplift from the onset of the Alpine Orogeny. During Late Cretaceous, a second pulse of the arc magmatism resumed.

Age	Lithology	Description
66 Ma Cretaceous		Arc Magmatic Rocks
		Cretaceous Turbidites, Çağlayan Formation
145 Ma Jurassic		Neritic and Pelagic Carbonates
		
		Volcanoclastic and Felsic to Mafic Plutonic Rocks
201 Ma Permo-Triassic		Transgressive sedimentary sequences
		Low-grade accretionary complexes (260-200 Ma), Carboniferous Granitic rocks and HT-LP metamorphic rocks

Figure 1.4 : Generalized stratigraphic column of the Sakarya Zone (after Topuz et al, 2013; Okay et al, 2013).

As can be seen from the stratigraphic features above (Figure 1.4), the Sakarya Zone displays totally different Paleozoic stratigraphy from the Istanbul zone, and thus is regarded as the eastward extension of Armorica. The timing of detachment from Gondwana is poorly constrained. However, its accretion to the north is suggested to have occurred during Carboniferous (Okay and Topuz, 2017).

The Central Pontides, located within the Sakarya Zone, is of particular interest due to the presence of large areas of both high-temperature and high-pressure type metamorphism spanning all three periods of the Mesozoic. The Kure ophiolite and accretionary complex is Triassic in age. The Geme and Devrekani are both Jurassic HT-LP metamorphic complexes. Several Cretaceous accretionary wedges and

associated ophiolites are present as well. Additionally, the Central Pontides contains HP-LT eclogite facies metamorphic units.

One example of HP-LT metamorphism, the primary focus of this study is the Elekdag Ophiolite. The Elekdag Ophiolite is a Southwest-Northeast trending lherzolitic block approximately 35 km by 5 km (Alther et al, 2006). It is a tectonic unit within the HP-LT Domuzdag complex. The Domuzdag formation is northeast southwest trending and approximately 75 km in width and 11 km thick at its maximum (Okay et al, 2006). It is highly deformed and consists of primarily micaceous schists, significant quantities of metabasite along with minor amounts of marbles and metacherts. Additionally it contains large lenses of serpentinite. The metabasites and serpentinite are more common in the upper portion (Northern) of the unit while the blueschist facies mica-bearing schists are more common in the lower part of the complex (Okay, 2006). Much of the unit is highly retrogressed to greenschist facies assemblages. Elekdag consists of high-pressure serpentinite along with eclogite blocks. The serpentinite is primarily composed of the higher temperature serpentine mineral, antigorite. Eclogites are HP-LT and, while highly retrogressed, contain several examples of Lawsonite inclusions within garnet. Lawsonite is a temperature sensitive hydrous mineral which has a fairly consistent stability field in PT space and is a key indicator of HP-LT metamorphism. Commonly during exhumation lawsonite breaks down and is only evidenced by relict grains. Preservation of lawsonite is uncommon. Its presence is commonly thought to be a result of rapid cooling and continuously hydrous conditions.

Below, each of the different units included in the Sakarya Zone are described separately in detail (Figure 1.5).

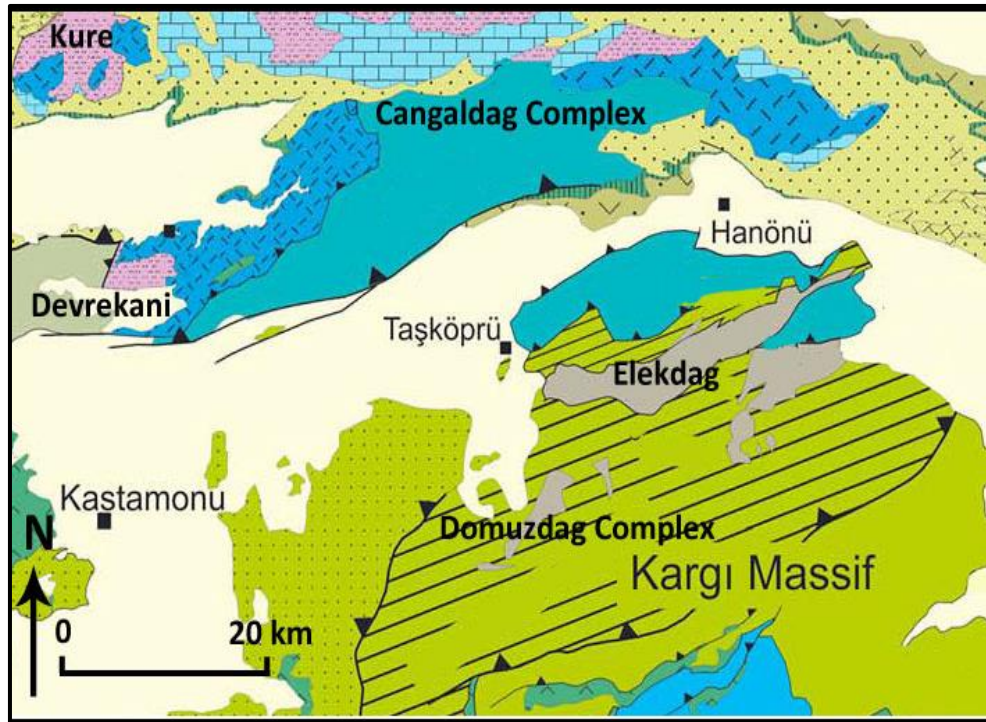


Figure 1.5 : Geological map of the Central Pontides (modified after Okay et al, 2013).

1.4.1 The Kure accretionary complex and ophiolite (Triassic)

The Kure Complex is an accretionary complex with an associated dismembered ophiolite. Thick sequences of pillow basalts and breccia, several small gabbro and ultramafic cumulate bodies (Cakir et al, 2006) and serpentinite form the base of the Kure Complex. The basalts are mostly MORB tholeiitic basalts (Ustaomer and Robertson, 1994). Overlying the basalts is a greater than 2 km thick sequence of strongly deformed, unmetamorphosed dark siltstone and sandstone, black shales and blocks of Upper Triassic limestone belonging to the Akgol Formation (Okay et al, 2015). The contact between the Akgol Formation and the underlying ophiolitic units is predominantly faults. The contact between the two units has seen significant mineralization (Bailey et al, 1967). The sedimentary units have been intruded by Middle Jurassic granitoids and acidic porphyries (Yilmaz and Boztug, 1986; Okay et al, 2015). Trace fossil dating places the Akgol Formation in the Late Triassic. The Akgol formation is unconformably overlain by undeformed Upper Jurassic units. Deformation was during the Early Jurassic or Late Triassic due to it not being present in either the intruding middle Jurassic rocks or the overlying Upper Jurassic sequences (Okay et al, 2015).

1.4.2 Elekdag ophiolite

One example of HP-LT metamorphism, the primary focus of this study is the Elekdag Ophiolite. The Elekdag Ophiolite is a Southwest-Northeast trending lherzolitic block approximately 35 km by 5 km (Alther et al, 2006). It is a tectonic unit within the HP-LT Domuzdag Complex. The Domuzdag formation is northeast southwest trending and approximately 75 km in width and 11 km thick at its maximum (Okay et al, 2006). It is highly deformed and consists of primarily micaceous schists, significant quantities of metabasite along with minor amounts of marbles and metacherts. Additionally it contains large lenses of serpentinite. The metabasites and serpentinite are more common in the upper portion (northern) of the unit while the blueschist facies mica-bearing schists are more common in the lower part of the complex (Okay, 2006). Much of the unit is highly retrogressed to greenschist facies assemblages. Elekdag consists of high-pressure serpentinite along with eclogite blocks. The serpentinite is primarily composed of the higher temperature serpentine mineral, antigorite. Eclogites are HP-LT and, while highly retrogressed, contain several examples of Lawsonite inclusions within garnet. Lawsonite is a temperature sensitive hydrous mineral which has a fairly consistent stability field in PT space and is a key indicator of HP-LT metamorphism. Commonly during exhumation lawsonite breaks down and is only evidenced by relict grains. Preservation of lawsonite is uncommon. Its presence is commonly thought to be a result of rapid cooling and continuously hydrous conditions.

This paper is primarily focused on the pressure and temperature conditions of metamorphism for the Elekdag Ophiolite and the implications for the greater tectonic assemblage of the Central Pontide region of northern Turkey.

1.4.3 Jurassic HT-LP metamorphism and associated magmatism

Metamorphism during the Jurassic period within the Central Pontides is of a high temperature low pressure nature. The Gemeni Complex and the Devrekani Complex are the large Jurassic metamorphic units in the Central Pontides. Magmatism during this period took the form of several shallow, chemically mostly intermediate intrusions. As a result heat flow and subsequent metamorphism was localized in contrast to the widespread recrystallization associated with subduction or regional metamorphic events. A lack of widespread evidence of deformation during the Middle-Jurassic in

the central pontides together with shallow level plutonic intrusions is evidence of an extensional setting (Okay et al, 2014).

The Geme Complex is an approximately 25 km by 5 km High-Temperature regional metamorphic unit. It is composed of mostly aluminum rich banded gneiss along with migmatite, amphibolite and cross cutting granitic veins in order of decreasing proportions (Okay et al, 2014). The Ca-amphibole + plagioclase + diopside gneiss is cut by the 163 Ma Dikmen Porphyry (Okay, 2014). Geothermobarometry and assemblage constraints on the gneiss indicate maximum pressures of 0.4 ± 0.1 GPa and the presence of migmatite suggests maximum temperature of metamorphism approached the hydrous granitic solidus of ~ 700 °C. Zircon grains from the granite have closure ages of 172 ± 3 Ma and Ar-Ar Biotite closure ages for the gneiss are 164 Ma. The Biotite closure temperature is 300 °C so this age likely represents the age of the intrusion. Inherited zircons within the Geme Complex show late Neoproterozoic ages with Carboniferous rims which is interpreted to mean the Geme Complex represents a piece of the central Pontides basement which has been remobilized (Okay, 2014).

Devrekani Massif considerably less studied. It is a similarly sized HP-LT metamorphic complex south of the Geme Complex within the Central Pontides.

1.4.4 Jurassic accretionary complexes

Jurassic accretionary complexes along with associated ophiolites are common in most of the countries located along the suture zones belonging to Laurasia-Gondwana and the intervening continental blocks. However they are rare in Turkey. The Refahiye Accretionary Complex and ophiolite in the Eastern Pontides and the Saka Complex in the Central Pontides are rare examples which show Early to Middle Jurassic ages (Topuz et al, 2013; Okay et al, 2013). The Refahiye metamorphic unit tectonically overlies Late Cretaceous accretionary rocks, while in the north it is truncated by the North Anatolian Fault which separates it from a low-grade metamorphosed Late Triassic accretionary complex. The Refahiye Complex is composed greenschists, marble, phyllite, amphibolite, eclogite and serpentinite. It is intruded by several gabbroic dikes. Radiogenic ages of white mica, calcic amphiboles and rutile gave closure temperature ages of 172 ± 4 Ma to 167 ± 4 Ma. The presence of a Jurassic accretionary complex in contact with both Late Triassic and Late Cretaceous units

without any intervening continental unit precludes the possibility of a Cimmeride continental ribbon separating the Paleo and Neo Tethyan oceans within Turkey.

The Saka Complex is a metamorphic complex within the Central Pontide Supercomplex. It is composed primarily of micaschists along with metabasites, serpentinite, marble and calcareous schists. Pieces of it are tectonically emplaced within other metamorphic terranes of the Supercomplex. PT estimates from the Saka Complex show $8-12 \pm 2$ kbar at 620 ± 30 °C. Ar ages from white mica give ages of between 170 and 162 Ma (Okay et al, 2013). It has been proposed that the Saka and Refahiye along with others may be portions of a larger accretionary complex that formed during the Jurassic period (Okay et al, 2013).

1.4.5 Cretaceous accretionary complexes

The accretionary complexes of the Cretaceous in Turkey are found in large areas along the Izmir-Ankara-Erzincan suture (IAES) within the Sakarya Zone. In some locations they are stacked below Jurassic aged accretionary wedges such as the Kure, however typically they are structurally below ophiolites and above the widespread carbonate platform of the Taurides. They have been subjected to various grades of metamorphism from high grade blueschist facies to greenschists facies conditions during the Late Cretaceous (Sherlock et al, 1999). They primarily are composed of oceanic sediments and basalts of all three periods of the Mesozoic Era (Okay et al, 2006). They represent the last of the Mesozoic amalgamations of Gondwana derived terranes to Laurasia.

The Kizilirmak Metamorphic Complex lies within the Kargi Massif of the Central Pontides, south of the Domuzdag Complex and is split by the Kizilirmak River. Ophiolitic rocks are comprised of metaperidotites both isotropic and cumulate gabbroic rocks along with basalts, serpentinite, and pelagic cherts.

2. FIELD RELATIONS AND PETROGRAPHY

In the following chapter the petrography and petrology of the rocks and minerals of the Elekdag Ophiolite will be described.

2.1 The Elekdag Ophiolite

The main body of Elekdag Ophiolite is approximately 35 km by 4 km (Figure 2.1). It consists primarily of lherzolite which has been greatly serpentized. The serpentinite is chiefly comprised of antigorite, the higher temperature member of the serpentine group of minerals. Along the boundaries of the metaophiolite proper are bands of serpentinite mélange which contain blocks of eclogites. These eclogites have been subject to varying degrees of retrograde metamorphism. Several blocks retain their peak metamorphism assemblages while most have been heavily retrogressed through blueschist to greenschist facies conditions. A notable feature of the Elekdag Ophiolites is the presence of primary lawsonite. Lawsonite is a hydrous calcium silicate which an index mineral along with glaucophane for high-pressure, low temperature subduction zone metamorphism. Primary lawsonite is somewhat uncommon and its presence is normally identified through the common pseudomorph assemblage of zoisite + white mica $\pm$ albite $\pm$ calcite $\pm$ pumpellyite $\pm$ chlorite in a rectangular relict.

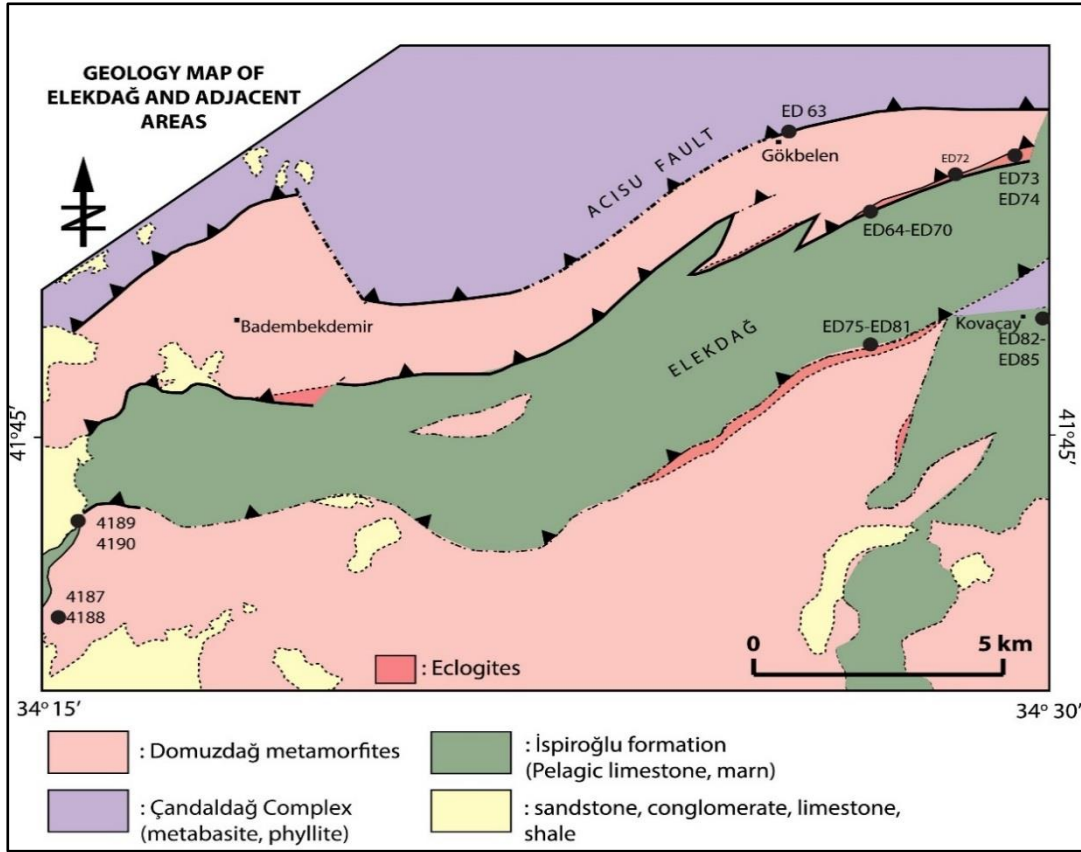


Figure 2.1 : Geologic map of the studied region, Elekdag Ophiolite with sample numbers and locations (modified from Okay et al, 2006).

In this study 25 samples were collected from Elekdag for further study. The samples collected consist of six eclogites in various states of retrogression, eleven blueschists also in various degrees of retrogression, two glaucophane-bearing white schists, four serpentinized peridotites and two greenschist facies metabasites. Of these samples, 17 thin sections were made. Of those thin sections six were selected for WDS microprobe analysis. Based upon microprobe results, three samples were selected for whole rock analysis in order to obtain the bulk chemical composition needed to generate isochemical phase equilibrium diagrams.

2.1.1 Mica schists

Mica schists are located along the margin of the lherzolite body of Elekdag. Their foliation is clearly visible in outcrops of several meters thick (Figure 2.7). They are well foliated with foliation defined by white mica (Figure 2.2 and 2.7) interlayered with granoblastic quartz and albite (Figure 2.5 and 2.6). Garnet porphyroblasts are set within the foliated matrix (Figure 2.6). The most common assemblage is white mica +

garnet + quartz + chlorite $\pm$ albite $\pm$ clinozoisite $\pm$ glaucophane (Figure 2.3 and 2.4). Glaucophane is rare and is the only HP-LT mineral preserved in the samples.

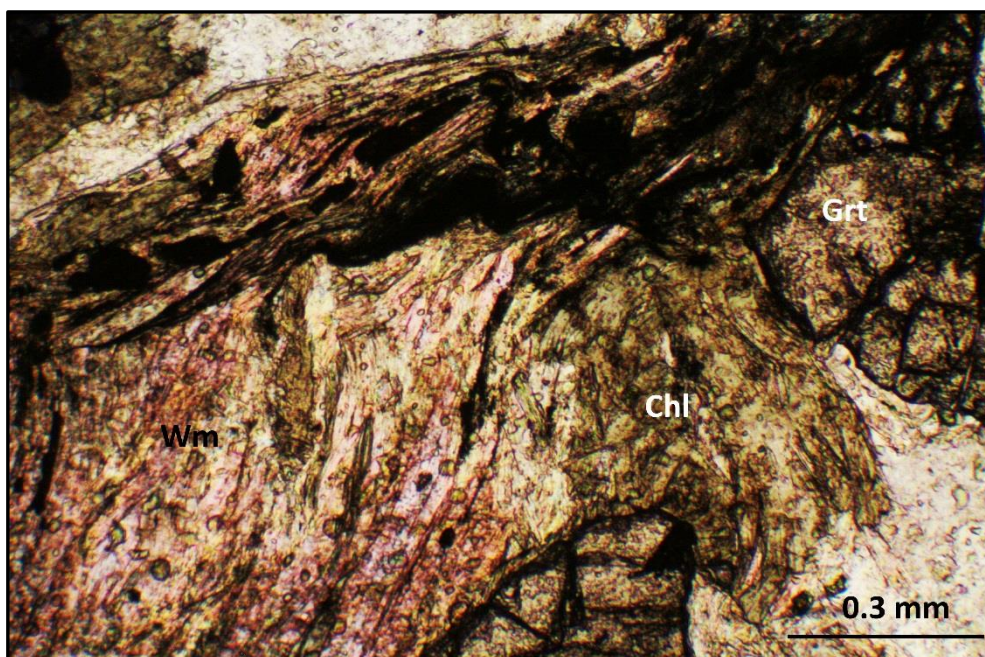


Figure 2.2 : Thin section photomicrograph of a white mica schist from NW of Aksaraykoy. Foliated textures of white mica with garnet and chlorite. The width of the micrograph is 1.5 mm. (Wm: White mica, Grt: Garnet, Chl: Chlorite).

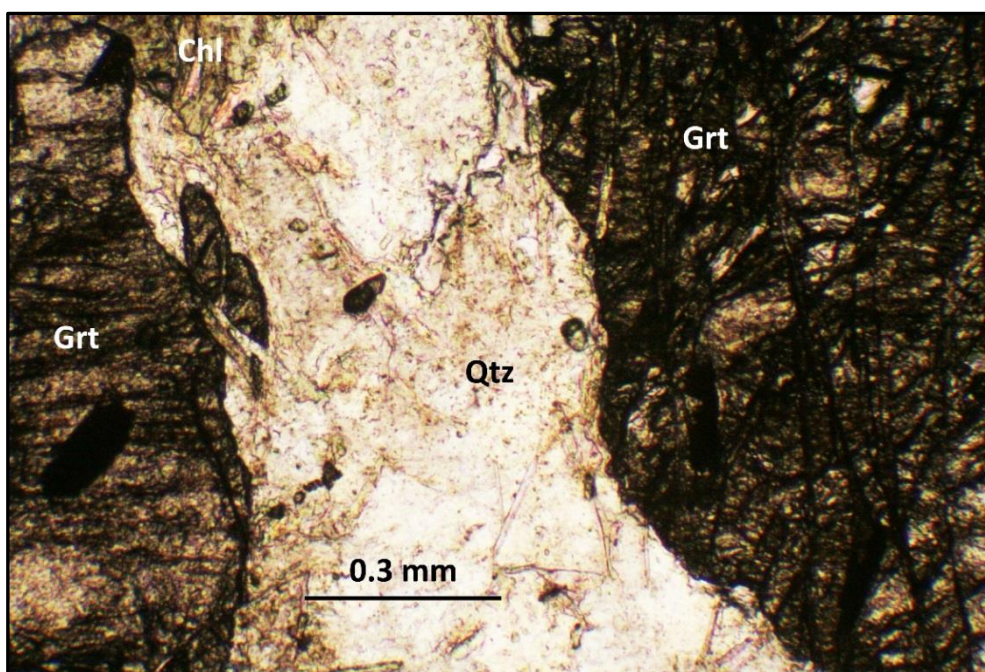


Figure 2.3 : Thin section of garnets set in a matrix of quartz, albite, minor glaucophane and chlorite (Grt: Garnet, Gln: Glaucophane, Ab: Albite, Qtz: Quartz, Chl: Chlorite).

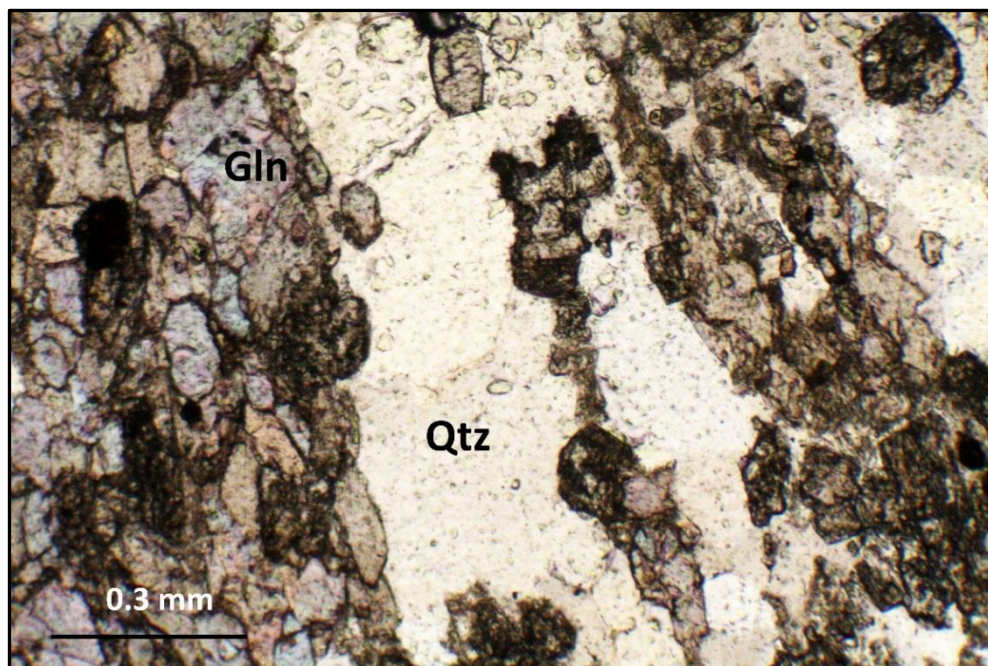


Figure 2.4 : Thin section of glaucophane and quartz banding (Gln: Glaucophane, Qtz: Quartz).

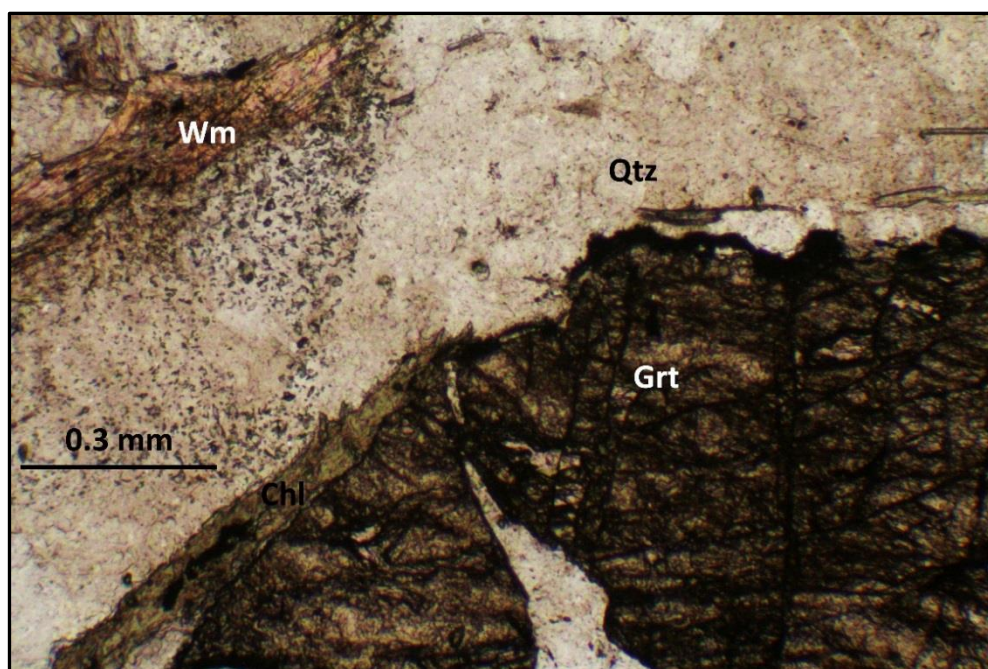


Figure 2.5 : Quartz band around garnet. Chlorite after garnet and white mica (Grt: Garnet, Chl: Chlorite, Qtz: Quartz, Ab: Albite, Wm: White mica).

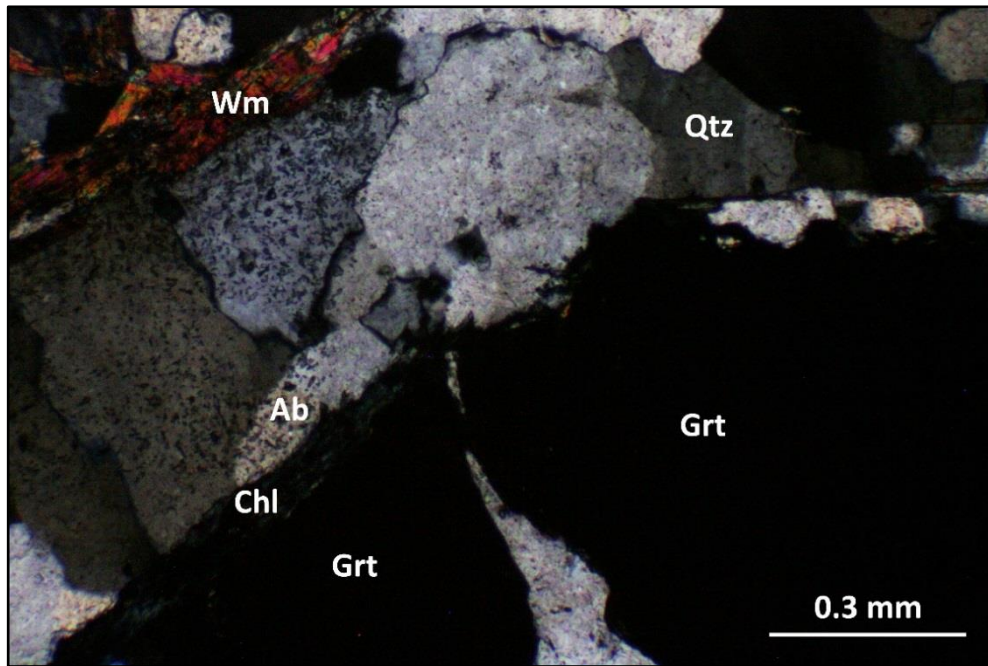


Figure 2.6 : Quartz band around garnet. Chlorite after garnet and white mica. In crossed polarized light (XPL) (Grt: Garnet, Chl: Chlorite, Qtz: Quartz, Ab: Albite, Wm: White mica).



Figure 2.7 : Outcrop of white mica schists in a quarry, NW of Aksaraykoy, Kastamonu.

2.1.2 Glaucophane garnet schists

Glaucophane garnet schists occur as boulders and as foliated layers in outcrop (Figure 2.8 and 2.9). Glaucophane rich layers are sometimes interlayered with eclogite facies assemblages (Figure 2.10). Garnets are present as porphyroblasts. The glaucophane and white mica matrix can be either foliated or massive. Layers of glaucophane are occasionally interlayered with quartzite or chlorite dominated foliations. The common assemblage of primary minerals is garnet + glaucophane + clinozoisite + white mica + chlorite $\pm$ quartz $\pm$ lawsonite $\pm$ rutile. In some examples lawsonite is broken down into pseudomorphs of clinozoisite + white mica $\pm$ albite $\pm$ chlorite. Most rocks have been hydrothermally altered during retrogression and chlorite is a common retrograde phase along with the phyllosilicate, stilpnomelane is present both in the matrix and within fractures of garnet (Figure 2.11).



Figure 2.8 : Metabasite knockers situated next to a road in Gokbelen, Kastamonu.



Figure 2.9 : Glaucophane interlayered with green schist facies assemblages in Gokbelen, Kastamonu.



Figure 2.10 : Interlayering of blue schist and green schist facies assemblages in outcrop in Gokbelen, Kastamonu.

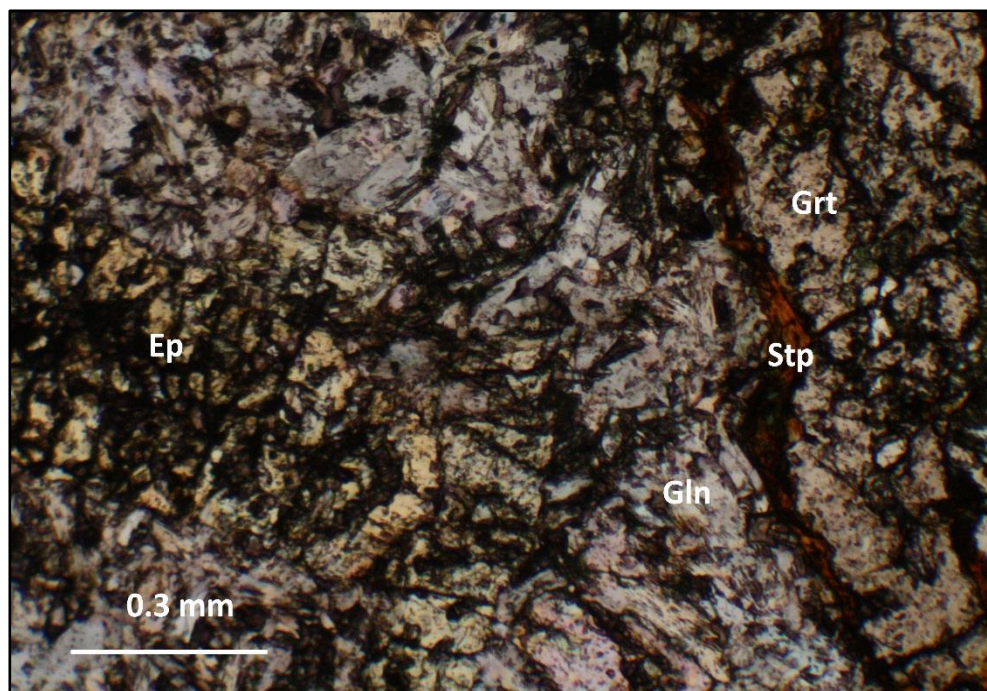


Figure 2.11 : Garnet bearing blueschist in thin section (Gln: Glaucophane, Grt: Garnet, Ep: Epidote, Stp: Stilpnomelane).

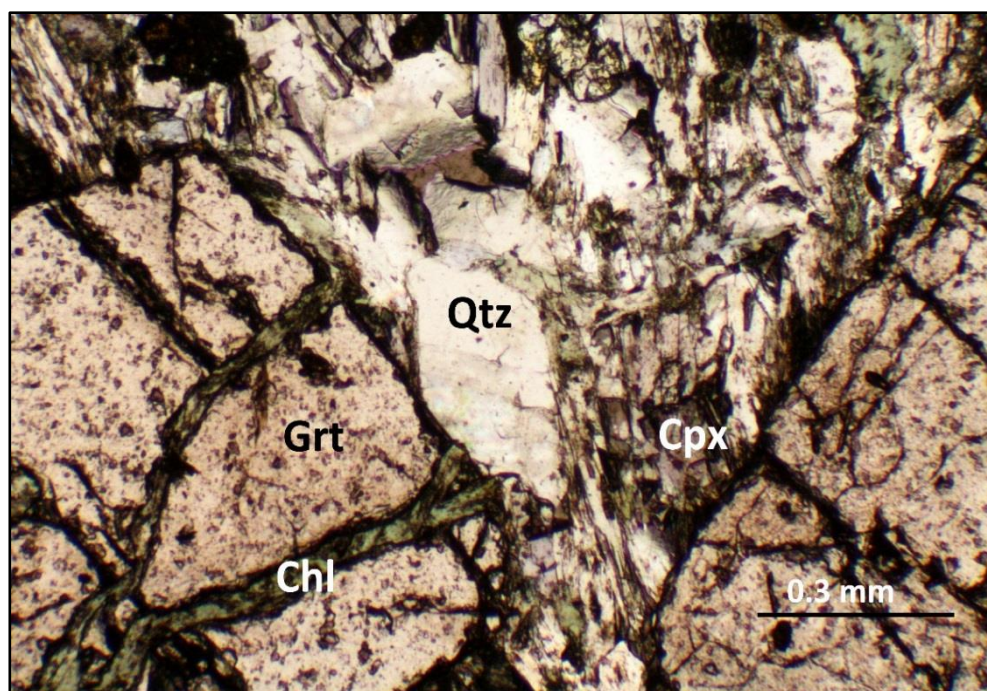


Figure 2.12 : Garnets and matrix. Garnets are heavily fractured. Chlorite veins along garnet fractures indicate the presence of hydrothermal activity during retrogression (Grt: Garnet, Gln: Glaucophane, Wm: White mica, Chl: Chlorite).

2.1.3 Eclogites

Eclogites, just as the blueschists occur as both blocks or in massive outcrop (Figure 2.14 and 2.15). Eclogites are dominated by the characteristic assemblage of garnet + omphacite + glaucophane + clinozoisite + white mica $\pm$ lawsonite $\pm$ tourmaline $\pm$ rutile. Samples are heavily overprinted with lower grade assemblages including chlorite, albite and some iron oxide. Clinozoisite is both a primary and retrograde mineral. Garnets are inclusion rich. Common inclusions within garnet include lawsonite, glaucophane, white mica and omphacite (Figure 2.13). Lawsonite inclusions are of particular note as they provide a temperature constraint on metamorphism, identifying the Elekdag samples as low temperature eclogites.

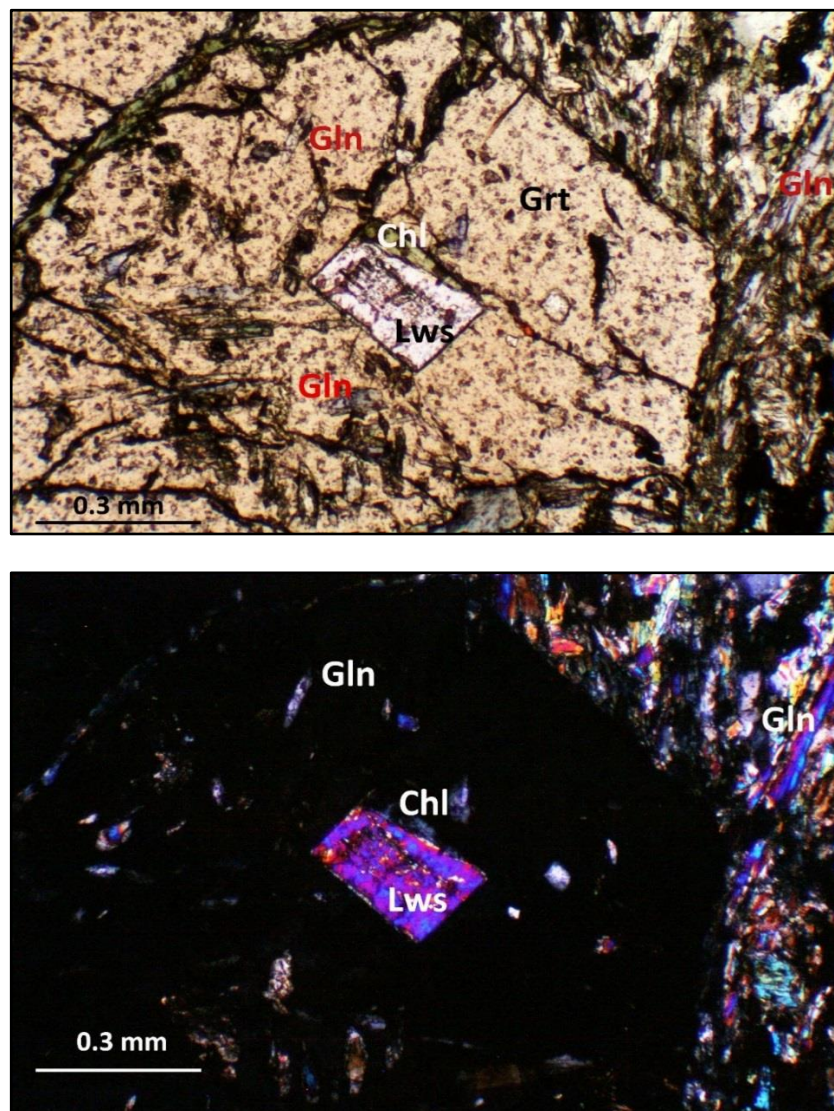


Figure 2.13 : Lawsonite and glaucophane inclusions within garnet (Lws: Lawsonite, Gln: Glaucophane, Grt: Garnet).



Figure 2.14 : Eclogite block, SE of Gokbelen village, Kastamonu.



Figure 2.15 : Eclogite facies outcrops near Kovacayir village, Kastamonu.

2.2 Mineral Chemistries

2.2.1 Garnet

Garnets are large euhedral porphyroblasts which have been fractured and penetrated by aqueous fluids during exhumation (Figure 2.12 and 2.13). They are inclusion rich, containing preserved lawsonite (Figure 2.13) as well as pseudomorphs of $\text{Czs}+\text{Wm}+/-\text{Ab}$. Other minerals present as inclusions within garnet are glaucophane, clinopyroxene (omphacitic), chlorite, plagioclase, clinozoisite, white mica and titanite. Glaucophane, omphacite, white mica are present both as prograde and retrograde phases. All fractures within garnets are populated by aqueous phases, most commonly chlorite, indicating samples were retrogressed in the presence of abundant water.

43 spot analyses of both garnet cores and rims were taken (Table 2.1, 2.2 and 2.3). Additionally, 3 diameter profiles of garnets were taken (Table 2.1, 2.2 and 2.3; Figure 2.18 and 2.19). Only 3 profiles were taken due to difficulties locating pure transects because of abundant inclusions, fracturing and water infiltration of the garnets making potential clean profiles hard to come by (Figure 2.16 and 2.17). The majority of garnets have $\text{Grs}(\text{Ca})_{21-26}\text{Pyp}(\text{Mg})_{03-20}\text{Sps}(\text{Mn})_{00-10}\text{Alm}(\text{Fe})_{40-70}$, however occasional garnets have spessartine cores as high as Sps_{53} which comes at the expense of Alm (Figure 2.19). Garnet analyses show small increases in Mg content at the expense of Ca and Fe from core to rim. This is a zoning pattern consistent with prograde (increasing pressure due to burial) garnet growth. Of the 4 components of the aluminous solid solution series $(\text{Ca},\text{Mg},\text{Mn},\text{Fe})_3\text{Al}_2(\text{SiO}_4)_3$ Pyrope (Mg), has the smallest cell parameter and the lowest entropy (S) and is the preferred composition under increasing pressure (P) making it a useful barometer.

Some garnets display unusually high spessartine cores of up to Sps_{54} at the expense of almandine (Figure 2.19).

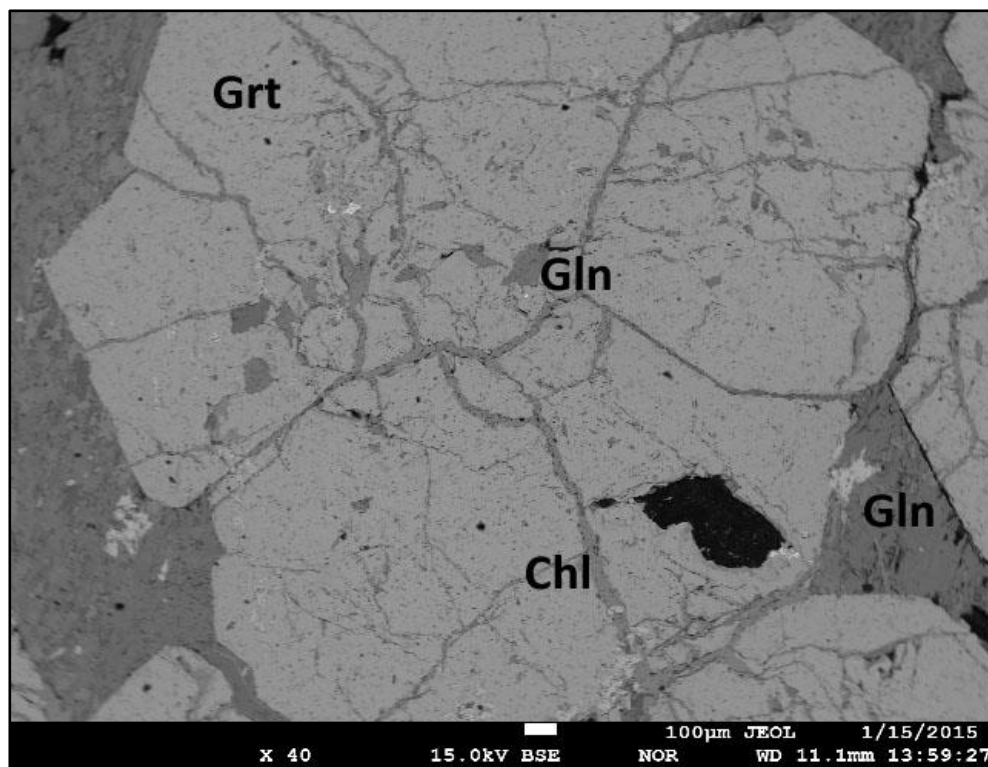


Figure 2.16 : SEM- BSE image of garnet with mineral replacement along fractures.

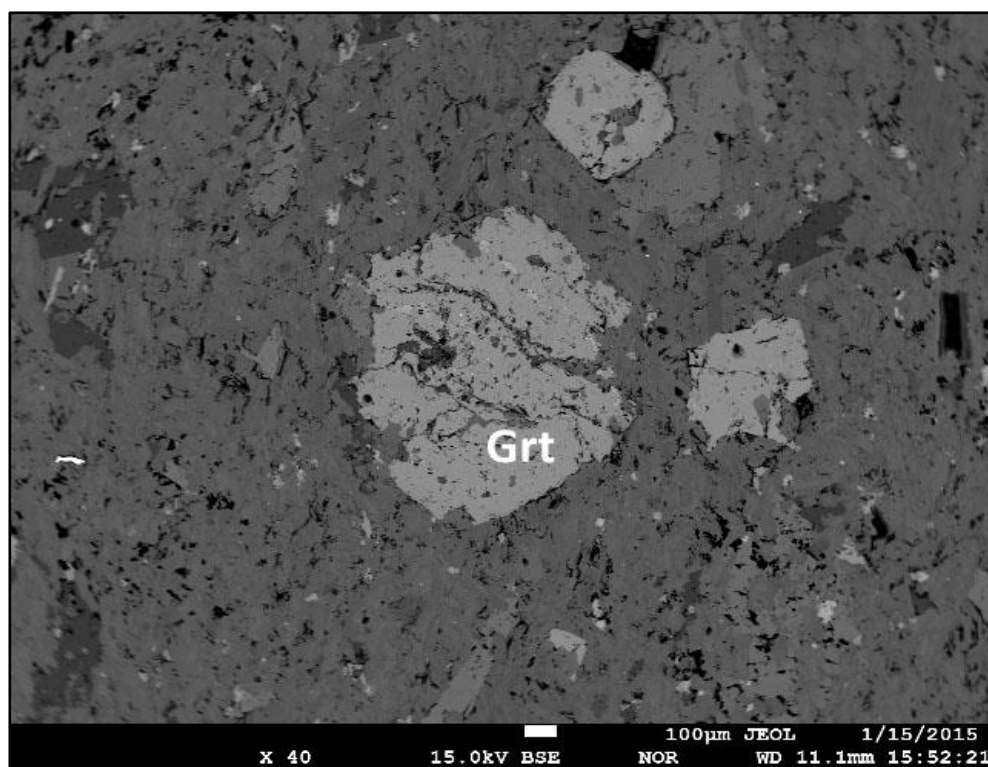


Figure 2.17 : SEM- BSE image of partially consumed garnets in matrix.

Table 2.1 : Electron microprobe (WDS) analyses of the cores of garnets from samples of HP-LT rocks of the Elekdag Ophiolite.

Sample	ED74	ED74	ED74	ED74	ED74	ED74	ED74	ED74	ED74	ED79	ED79	ED79	ED79	ED79	ED66	ED66	ED66	ED66	ED66	ED74
Mineral	Grt	Grt	Grt	Grt	Grt	Grt	Grt	Grt	Grt	Grt	Grt	Grt	Grt	Grt	Grt	Grt	Grt	Grt	Grt	Grt
Area	Core	Core	Core	Core	Core	Core	Core	Core	Core	Core	Core	Core	Core	Core	Core	Core	Core	Core	Core	Core
Analysis #	206	207	217	218	219	219	219	216	216	216	216	216	216	216	216	216	216	216	216	216
SiO ₂	37.37	37.21	35.76	34.79	37.19	37.43	37.38	37.29	36.69	37.16	38.08	37.11	37.30	37.50	37.27	36.86	37.26	37.80	40.98	38.84
TiO ₂	0.12	0.12	0.13	0.05	0.15	0.18	0.15	0.21	0.24	0.21	0.06	0.10	0.10	0.17	0.13	0.14	0.18	0.20	0.06	0.17
Al ₂ O ₃	21.34	21.16	20.66	20.32	21.13	21.21	20.93	20.60	20.14	20.43	20.75	20.63	20.42	20.57	20.64	20.27	20.48	20.70	22.94	19.71
Cr ₂ O ₃	0.02	0.04	-0.01	0.01	0.03	0.01	0.30	-0.01	-0.01	0.00	-0.01	-0.01	0.01	-0.01	-0.02	0.01	-0.01	0.01	0.02	0.01
FeO	30.55	31.00	29.21	27.90	29.11	29.72	29.19	22.39	22.84	21.30	31.49	31.31	31.36	24.08	25.36	28.13	24.05	23.67	29.99	28.66
MnO	1.36	1.06	3.22	3.18	3.84	2.62	3.66	11.71	11.10	13.67	0.39	0.41	0.39	8.78	7.47	4.28	8.33	8.06	0.34	2.12
MgO	1.76	1.75	1.28	1.10	1.16	1.33	1.18	0.50	0.46	0.44	1.75	1.66	1.65	0.71	0.77	0.92	0.66	0.86	3.34	1.28
CaO	8.30	8.47	7.97	9.06	8.25	8.51	8.32	8.36	8.67	7.91	8.23	8.62	8.88	8.67	8.48	8.90	9.09	9.29	8.26	8.59
Na ₂ O	0.01	0.01	0.01	-0.01	0.01	0.02	0.01	0.04	0.01	0.01	0.00	-0.02	0.02	0.01	0.01	0.01	0.02	0.01	0.01	0.08
K ₂ O	-0.14	-0.14	-0.13	-0.13	-0.12	-0.13	-0.14	-0.10	-0.12	-0.11	-0.16	-0.17	-0.18	-0.12	-0.14	-0.14	-0.14	-0.13	-0.17	-0.14
SUM	100.69	100.69	98.12	96.26	100.75	100.88	100.97	100.99	100.02	101.02	100.58	99.63	99.96	100.35	99.96	99.37	99.92	100.45	105.77	99.35
Analysis #	206	207	294	292	297	298	298	299	298	299	303	299	300	301	300	299	301	302	304	311
Si	2.97	2.97	2.94	1.95	2.17	2.18	2.19	1.64	1.65	1.66	3.14	3.15	3.16	3.27	3.28	3.29	3.43	3.44	3.45	3.50
Ti	0.01	0.01	0.01	0.00	0.01	0.01	0.01	0.01	0.02	0.01	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.00	0.01
Al	2.00	1.99	2.01	2.01	1.99	1.99	1.97	1.95	1.93	1.94	1.94	1.96	1.94	1.95	1.96	1.94	1.95	1.95	2.01	1.86
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe	2.03	2.07	2.01	1.96	1.95	1.98	1.95	1.50	1.55	1.43	2.09	2.11	2.11	1.62	1.71	1.91	1.62	1.58	1.86	1.92
Mn	0.09	0.07	0.23	0.23	0.26	0.18	0.25	0.80	0.76	0.93	0.03	0.03	0.03	0.60	0.51	0.30	0.57	0.55	0.02	0.14
Mg	0.21	0.21	0.16	0.14	0.14	0.16	0.14	0.06	0.06	0.05	0.21	0.20	0.20	0.09	0.09	0.11	0.08	0.10	0.37	0.15
Ca	0.71	0.72	0.70	0.82	0.71	0.73	0.71	0.72	0.75	0.68	0.70	0.74	0.77	0.75	0.73	0.77	0.79	0.80	0.66	0.74
Na	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01
K	-0.01	-0.02	-0.01	-0.01	-0.01	-0.01	-0.01	-0.01	-0.01	-0.01	-0.02	-0.02	-0.02	-0.01	-0.01	-0.02	-0.01	-0.01	-0.02	-0.02
SUM	20.01	20.02	20.04	20.06	20.02	20.01	20.01	20.02	20.04	20.03	19.99	20.02	20.02	20.00	20.00	20.02	20.01	19.99	19.95	19.94

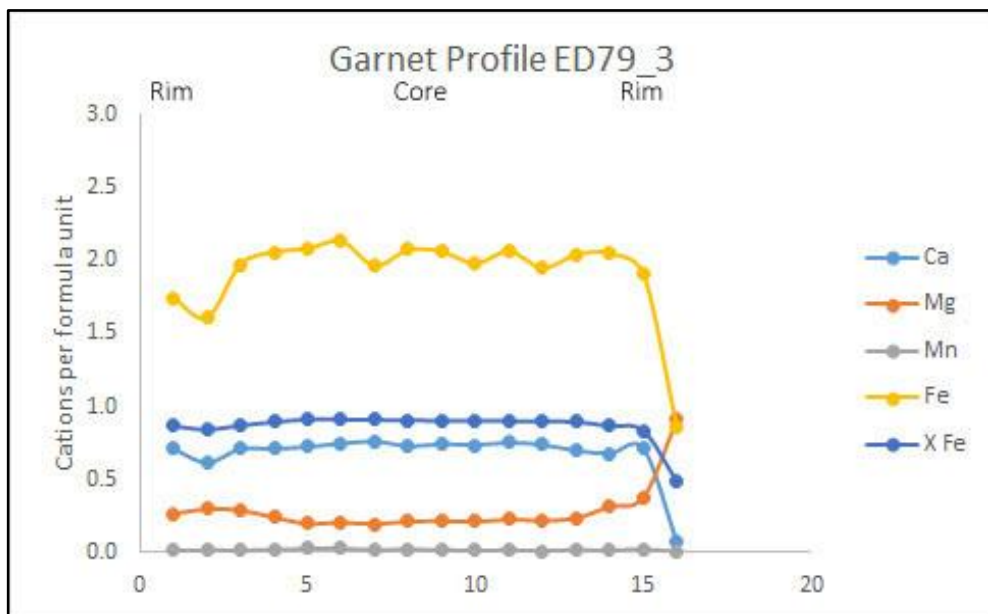
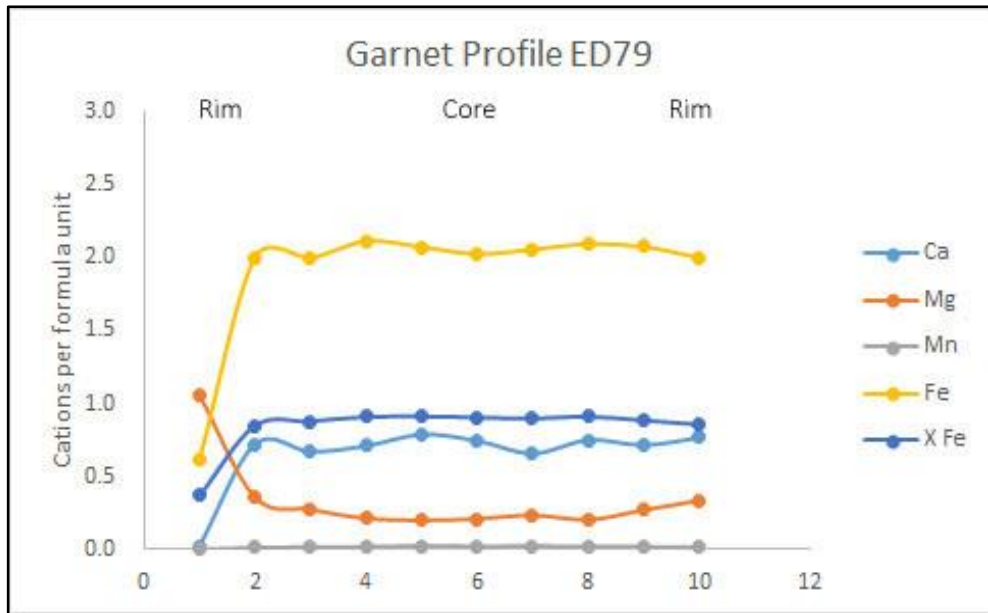


Figure 2.18 : Compositional profile of the cations for the aluminum series of garnets. Sample number ED-79 and ED-79_3. (Yellow: Almandine (Fe^{2+}), Blue: Grossular (Ca), Orange: Pyrope (Mg), Grey: Spessartine (Mn)).

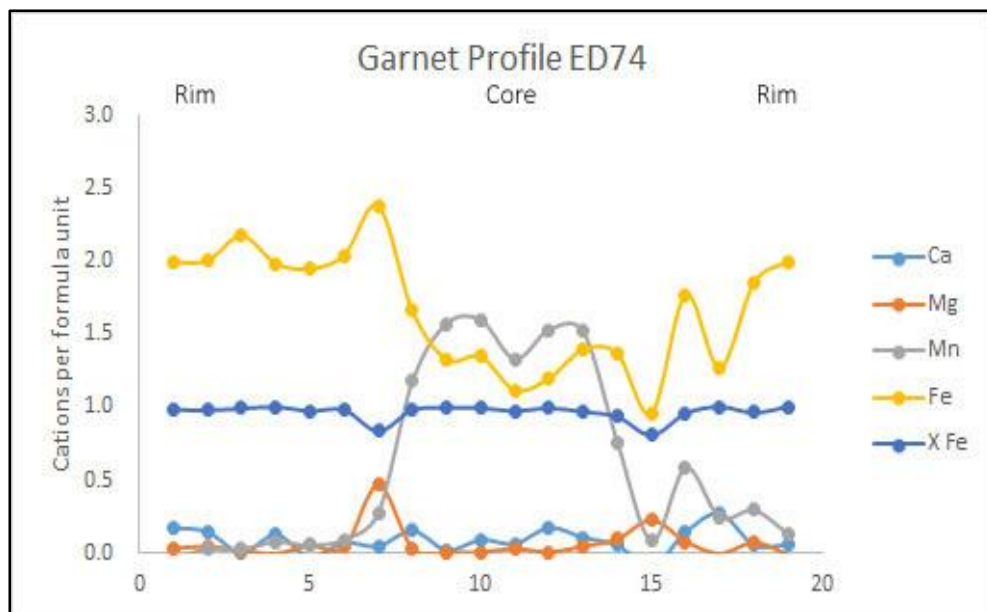


Figure 2.19: Compositional profile of the cations for the aluminum series of garnets. Sample number ED-74. (Yellow: Almandine (Fe^{2+}), Blue: Grossular (Ca), Orange: Pyrope (Mg), Grey: Spessartine (Mn)).

2.2.2 Clinopyroxene

Clinopyroxene falls within the compositional ranges of omphacite. Clinopyroxene is primarily magnesian with X_{Mg} content ranging from 0.75-0.88. Fe_2O_3 content is small (~ 0.064) and was determined from charge balancing on the basis of 4 cations and 6 oxygen. The sodic component of CPX is between 0.38-0.45 (Table 2.3). Of the sodium component, almost all of it is jadeite, aegerine or aegirine (Fe_2O_3 on the M1 site) values are between 0.15 and 0.130 with a mean of 0.06 with the remaining M1 site cation being Al. The calcium component is dominantly diopside Di_{75-88} (Figure 2.20).

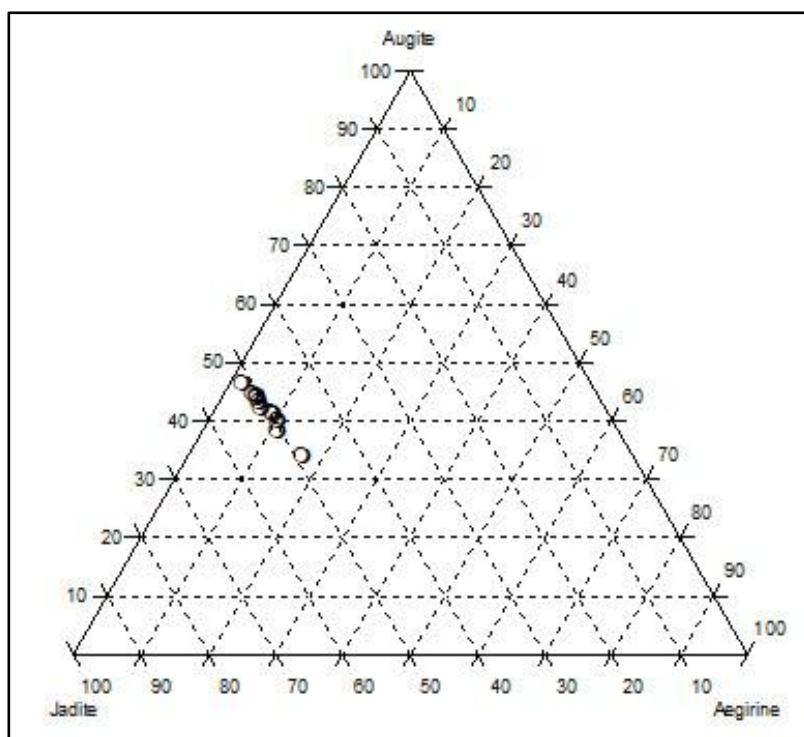


Figure 2.20 : Ternary Diagram of Na, Al and Fe^{3+} content of omphacite.

2.2.3 Amphibole

The predominant amphibole species in all samples is Na-rich (glaucophane) with minor amounts of actinolite-tremolite series in the more heavily retrogressed samples. Si contents are high, ranging from 7.87-8.00 cpfu (normalized to 15 cations, 23 O and 2H) (Table 2.4). Na_{M4} ranges from 1.60-1.95 cpfu (Figure 2.22). Sodic amphiboles are true glaucophanes with Al_{M2} contents (glaucophane – riebeckite) of $\text{Gln}_{67}\text{-Gln}_{93}$ and X_{Mg} of 0.50-0.69. Glaucophane occurs as inclusions within garnet and as a primary and secondary matrix mineral. Inclusions within garnet are small, less than 2mm (Figure 2.21). Primary matrix glaucophane is generally subhedral and larger than its secondary counterpart. Secondary grains can be massive or anhedral. In several blueschists glaucophane is parallel to the direction of foliation and forms layers with minor mica.

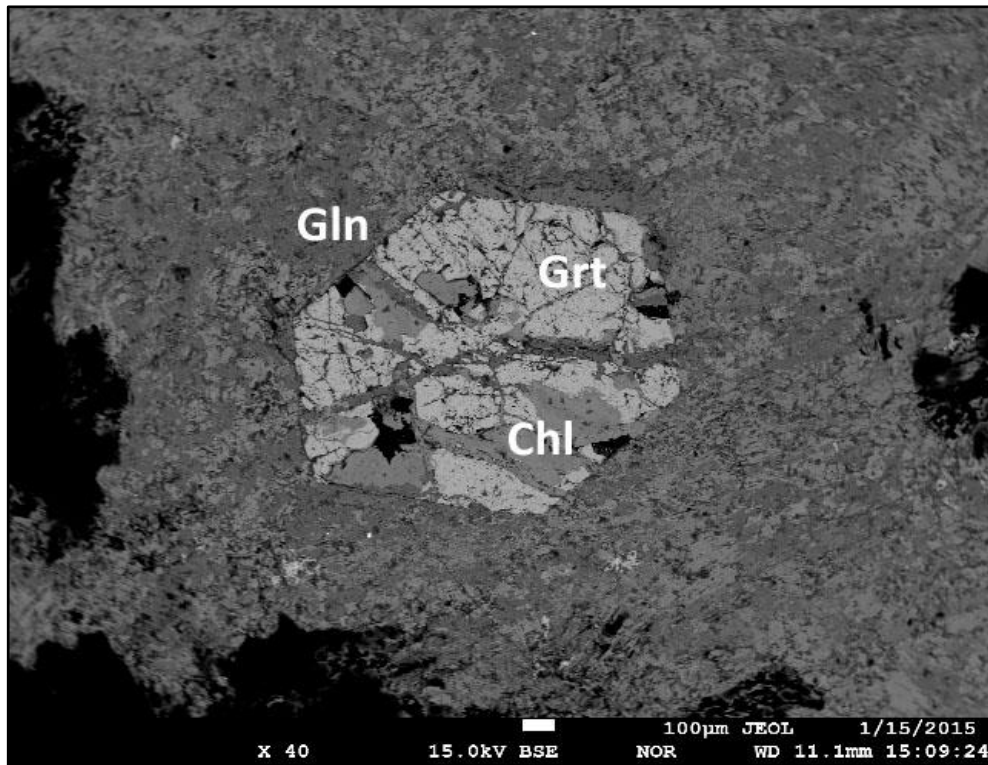


Figure 2.21 : SEM-BSE Image of fractured and partially retrogressed garnet sitting in a matrix of glaucophane, chlorite, phengite and clinozoisite.

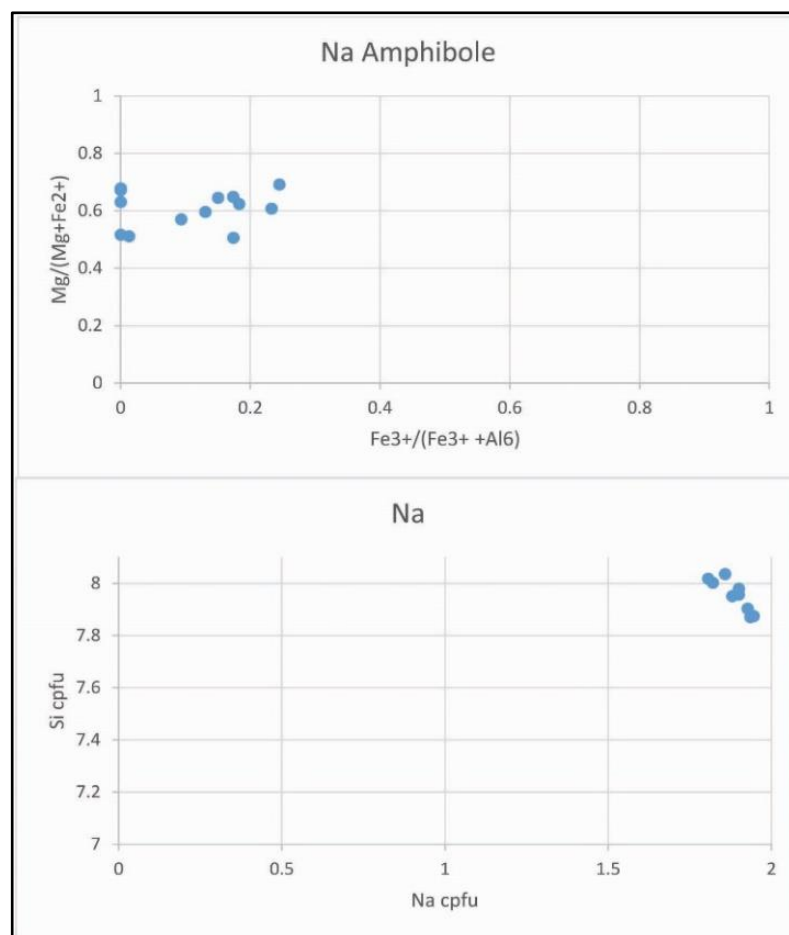


Figure 2.22 : Charts plotting sodic amphibole components. Top: X axis: glaucophane-riebeckite series, lower values correspond to glaucophane. Y axis: Magnesium content on the M5 site. Bottom chart plotting strength of the Na and Si components.

2.2.4 White mica

Microprobe analyses of dioctahedral whitemica were taken from several specimens. Si contents are 3.27-3.55 cpfu (Table 2.5). Samples display strong celadonite components (Mg+Mn+Fe) with values of 0.45-0.66 which correspond to the amount of Si in the mineral. K content on the A-site is high 0.95-0.98. Magnesium is the more common divalent cation and X_{Mg} values are 0.61-0.74. Micas are most common as matrix minerals although some inclusions within garnet exist. Si content in phengitic micas is a useful geobarometer for high pressure mineral assemblages.

2.2.5 Albite

Albite is present in nearly all samples as a late phase mineral. Chemical compositions are Ab₉₃₋₁₀₀ (Table 2.6). Albite in the metabasites is found as a break down product of some garnet inclusions or more commonly within the matrix, likely after omphacite or glaucophane. Albite is assumed to be low albite and not as relict magmatic plagioclase.

Table 2.6 : Electron microprobe analyses of plagioclase feldspar.

Sample	ED79	ED79	ED79	ED79	ED79	ED79	ED79	ED79	ED66
Mineral	Ab	Ab	Ab	Ab	Ab	Ab	Ab	Ab	Ab
Analysis #	167	168	282	283	284	186	187	306	331
SiO ₂	68.21	67.52	68.38	68.36	69.51	66.96	67.72	67.85	67.04
TiO ₂	0.01	0.01	0.01	0.00	-0.01	0.03	0.02	0.02	0.00
Al ₂ O ₃	20.05	19.90	18.72	18.59	18.76	18.95	19.49	19.97	21.03
Cr ₂ O ₃	0.01	0.01	0.01	-0.01	-0.01	0.00	0.00	-0.01	0.01
FeO	0.30	0.19	0.06	0.06	0.19	0.13	0.15	0.06	0.24
MnO	-0.01	0.04	0.01	0.01	0.01	0.03	0.02	-0.01	-0.02
MgO	-0.02	-0.01	-0.01	-0.01	-0.01	-0.01	0.00	-0.02	0.00
CaO	0.45	0.48	0.09	0.03	0.08	0.19	0.10	0.16	1.12
Na ₂ O	11.25	11.33	11.32	11.04	11.01	11.67	11.81	11.01	11.19
K ₂ O	0.02	0.02	0.02	0.01	-0.01	0.00	0.04	0.20	0.06
SUM	100.26	99.49	98.61	98.09	99.53	97.93	99.34	99.22	100.66
Si	2.97	2.97	3.02	3.03	3.04	2.99	2.98	2.98	2.92
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al	1.03	1.03	0.98	0.97	0.97	1.00	1.01	1.04	1.08
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe	0.01	0.01	0.00	0.00	0.01	0.01	0.01	0.00	0.01
Mn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ca	0.02	0.02	0.00	0.00	0.00	0.01	0.01	0.01	0.05
Na	0.95	0.97	0.97	0.95	0.93	1.01	1.01	0.94	0.95
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00
SUM	12.99	13.00	12.98	12.96	12.95	13.01	13.02	12.98	13.01
XAb	0.021	0.023	0.004	0.001	0.004	0.009	0.005	0.008	0.052
%Ab	95.1	96.6	97	94.9	93.3	101.1	100.8	93.8	94.5

2.2.6 Clinozoisite

Epidote-Clinozoisite solid solution minerals are highly aluminous with compositions of Czs₇₇₋₉₈ (Al/Al+ Fe³⁺) (Table 2.7). Clinozoisite is common within all metabasite samples due to heavy overprinting of peak assemblages during exhumation by greenschist facies assemblages. Grains are present as pseudomorphs after lawsonite, massive primary minerals and as replacement phases in the matrix and along garnet rims.

Table 2.7 : Electron microprobe analyses of clinozoisite-epidote minerals.

Sample	ED74	ED74	ED74	ED79	ED79	ED79	ED79	ED66	ED74	ED74	ED74
Mineral	Czo	Czo	Czo	Czo	Czo	Czo	Czo	Czo	Czo	Czo	Czo
Analysis #	209	210	211	303	304	305	322	330	353	354	355
SiO ₂	38.27	38.13	38.26	38.62	38.80	38.62	37.76	39.09	38.30	37.77	38.79
TiO ₂	0.08	0.07	0.08	0.13	0.14	0.09	0.08	0.13	0.12	0.13	0.07
Al ₂ O ₃	26.93	26.87	27.01	24.24	24.23	24.40	25.10	32.33	27.66	25.63	28.36
Cr ₂ O ₃	0.01	0.03	0.01	0.00	0.00	0.00	-0.01	0.04	0.01	0.08	0.05
FeO	7.93	8.02	8.02	10.68	10.73	9.09	9.25	0.95	6.37	8.22	5.23
MnO	0.10	0.14	0.04	0.05	0.01	0.03	0.04	0.01	0.05	0.12	0.05
MgO	0.01	0.00	0.01	0.07	0.09	0.04	0.04	-0.01	0.05	0.00	0.04
CaO	23.89	23.87	23.93	22.86	22.65	22.93	23.40	24.01	23.64	23.30	23.77
Na ₂ O	0.01	0.01	0.01	0.01	0.01	0.04	0.01	-0.01	-0.02	0.01	0.02
K ₂ O	-0.06	-0.03	-0.02	-0.06	-0.04	-0.06	-0.05	0.03	-0.03	-0.05	-0.04
SUM	97.17	97.13	97.34	96.60	96.62	95.17	95.63	96.56	96.14	95.21	96.33
Analysis #	209	210	211	303	304	305	322	330	353	354	355
Si	3.06	3.05	3.05	3.14	3.15	3.16	3.09	3.02	3.06	3.08	3.07
Ti	0.01	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.00
Al	2.53	2.53	2.54	2.32	2.32	2.35	2.42	2.94	2.61	2.47	2.65
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00
Fe	0.53	0.54	0.54	0.73	0.73	0.62	0.63	0.06	0.43	0.56	0.35
Mn	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00
Mg	0.00	0.00	0.00	0.01	0.01	0.01	0.01	0.00	0.01	0.00	0.01
Ca	2.04	2.05	2.04	1.99	1.97	2.01	2.05	1.98	2.02	2.04	2.02
Na	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00
K	-0.01	0.00	0.00	-0.01	-0.01	-0.01	-0.01	0.00	0.00	-0.01	0.00
SUM	20.67	20.68	20.68	20.69	20.68	20.66	20.70	20.51	20.63	20.67	20.60
Fe/(Al+Fe3+)	17.67	17.87	17.83	24.20	24.27	20.73	21.07	2.03	14.20	18.73	11.53

2.2.7 Lawsonite

Lawsonite occurs as well preserved inclusions within garnet where fracturing has not occurred. Pseudomorphs of lawsonite are common where fracturing and hydrothermal fluids have reached the inclusions. Pseudomorphs of lawsonite are easily recognized by their rectangular shape (Figure 2.23) and are comprised of clinozoisite+ white mica +/- albite. All lawsonite samples which were analyzed vary little from the empirical formula $\text{CaAl}_2\text{Si}_2\text{O}_7 \cdot \text{H}_2\text{O}$. Si values are 2 cpdf. Aluminum values range from 1.929 to 1.964 cpdf, while Ca values range from 0.946 to 0.953 cpdf. Fe is present in minimal amounts of no more than 0.081 cpdf (Table 2.8).

Table 2.8 : Electron microprobe analyses of lawsonite grains.

Sample	ED79	ED79	ED79	ED79	ED66
Mineral	Lws	Lws	Lws	Lws	Lws incl-grt
Analysis #	159	160	307	308	341
SiO ₂	38.08	38.10	38.67	37.83	27.66
TiO ₂	0.04	0.05	0.04	0.08	0.02
Al ₂ O ₃	31.37	30.78	31.64	30.43	19.33
Cr ₂ O ₃	0.01	0.01	0.00	0.01	0.01
FeO	1.20	0.84	1.20	1.79	23.29
MnO	0.01	0.00	-0.01	0.00	0.18
MgO	0.01	-0.02	-0.01	0.01	16.09
CaO	16.73	16.47	16.80	16.53	0.02
Na ₂ O	-0.03	0.00	-0.03	-0.02	0.13
K ₂ O	-0.06	-0.11	-0.09	-0.07	-0.12
SUM	87.36	86.12	88.22	86.59	86.61
Analysis #	159	160	307	308	341
Si	2.023	2.046	2.032	2.034	5.814
Ti	0.002	0.002	0.002	0.003	0.004
Al	1.964	1.948	1.96	1.929	4.788
Cr	0	0	0	0	0.002
Fe	0.053	0.038	0.053	0.081	4.093
Mn	0	0	0	0	0.031
Mg	0	-0.001	-0.001	0.001	5.041
Ca	0.953	0.948	0.946	0.953	0.005
Na	-0.003	0	-0.003	-0.002	0.053
K	-0.004	-0.007	-0.006	-0.005	-0.033
SUM	12.989	12.974	12.982	12.994	47.797

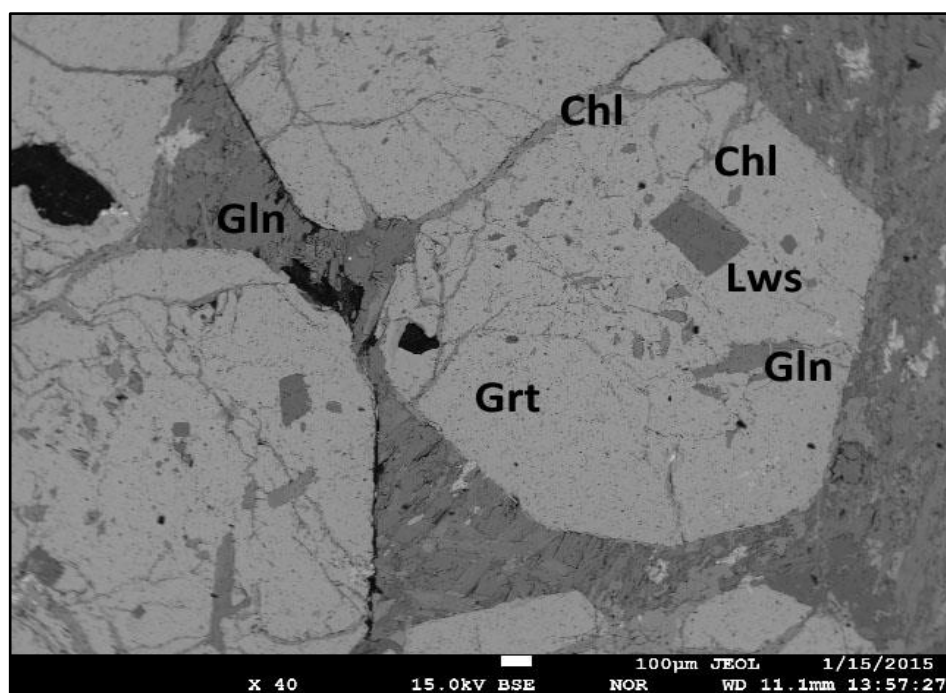


Figure 2.23 : SEM- BSE image of garnet cluster with clear rectangular lawsonite inclusions.

2.2.8 Chlorite

Chlorite is common in the metabasites sampled. X_{Mg} values are between 0.44 and 0.61 (Table 2.5) after being normalized to 28 atoms of Oxygen. It is the most common alteration product along fractures within garnets. It is also abundant within the matrix (Figure 2.24). Primary chlorite is difficult to identify with only a few small inclusions within garnet which could be alterations from fractures not evident due to the 2 dimensional character of the thin sections. No clear chemical indicator was observed within the few grains that were analyzed.

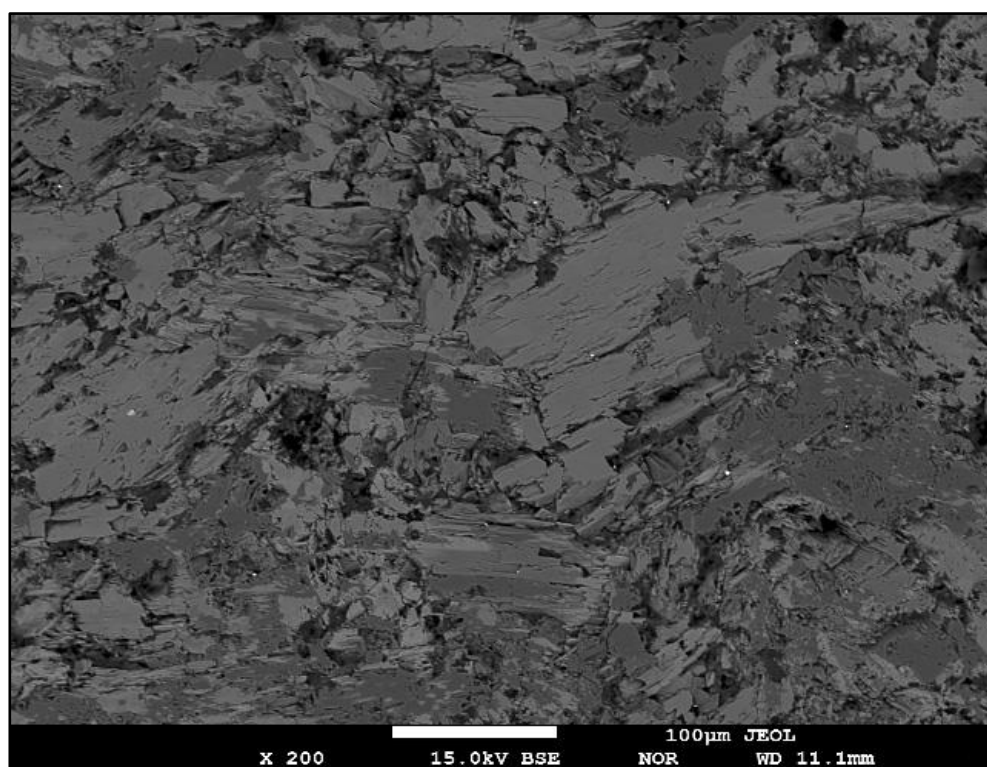


Figure 2.24 : SEM- BSE image of glaucophane and chlorite matrix.

2.3 Bulk Rock Composition

In order to better constrain the PT conditions of metamorphism, 3 metabasites were selected for whole rock analysis due to the at least partial presence of peak metamorphic assemblages. Their bulk composition of major oxides and minor elements as well as their geochemical characteristics are presented below (Table 2.9 and Figure 2.25 and 2.26).

The metabasite compositions are generally similar with some variation (SiO_2 = 46.27-49.37 wt%, TiO_2 = 1.37-1.83 wt%, Al_2O_3 = 13.73-16.64 wt%, FeO =10.77-15.83 wt%,

MnO= 0.19-0.24 wt%, MgO= 6.36-8.25 wt%, CaO= 5.76-12.05 wt%, K₂O= 0.18-0.68 wt%, Na₂O= 2.49-4.71 wt%, H₂O (from LOI)= 1.23-2.85 wt%). All samples plot within range of basalt on a TAS diagram, 2.5-5% K₂O+Na₂O, 46.27-49.37% SiO₂ (Cox et al 1979).

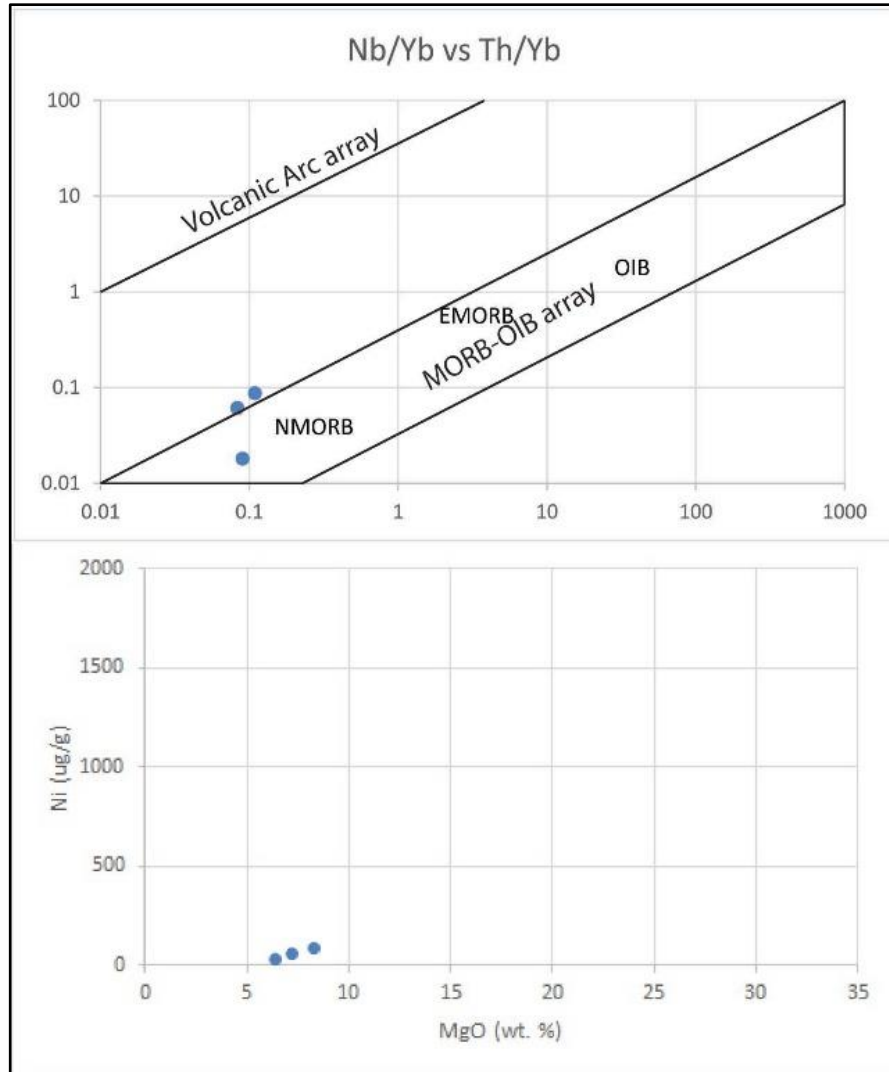


Figure 2.25 : a) Nb/Yb vs Th/Yb diagram. Shows that the 3 samples plot within the NMORB range of the MORB-OIB of Pearce 2008 (After Topuz et al., 2018 and Pearce 2008) b) Ni concentration vs MgO. Ni is a function of olivine fractionation.

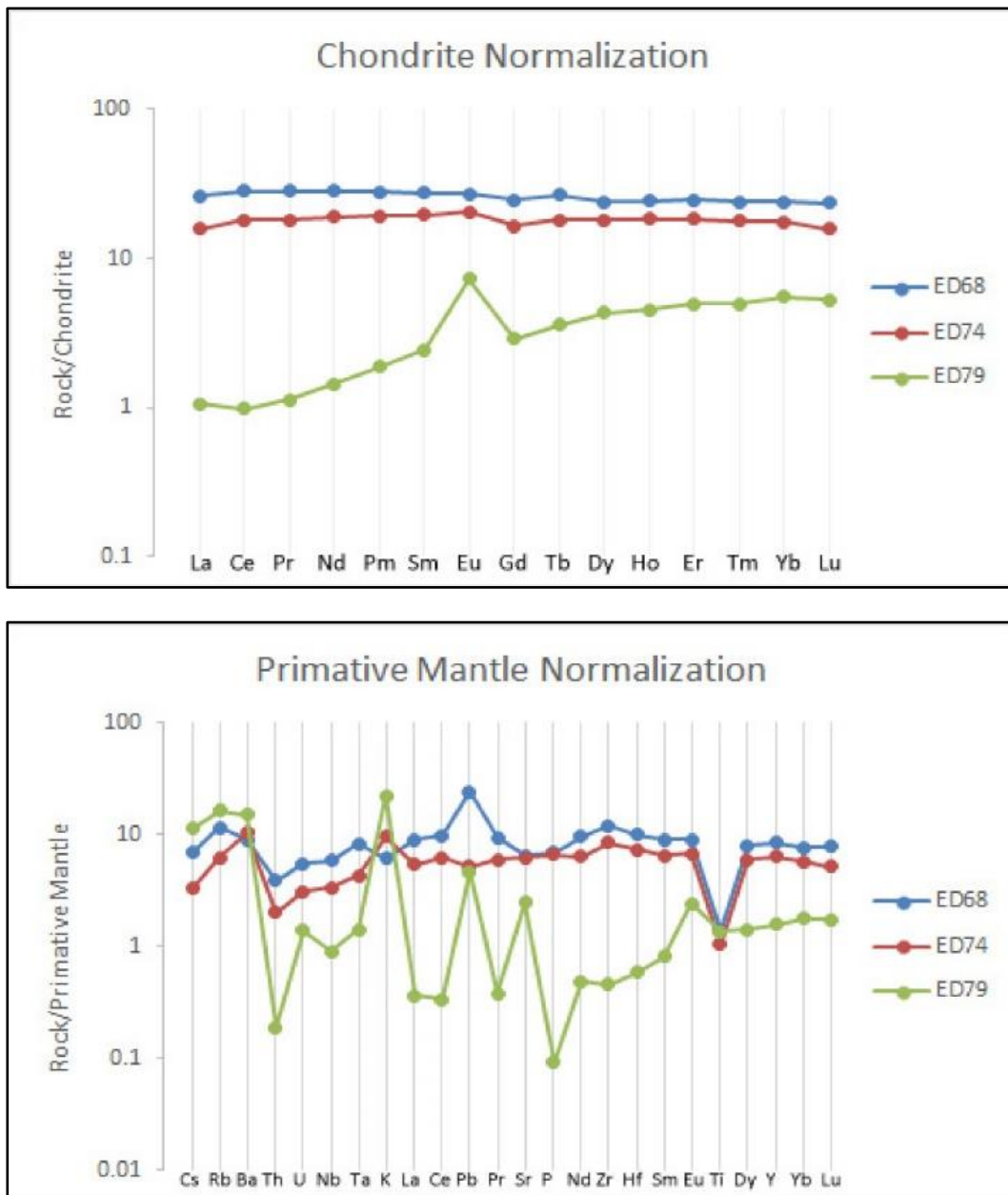


Figure 2.26 a) Primitive mantle normalized REE spider diagram. b) Chondrite normalized REE spider diagram.

Table 2.9 : Major Oxides and trace element compositions of three samples from the Elekdag Ophiolite.

Element	Units	ED-68	ED-74	ED-79	Element	Units	ED-68	ED-74	ED-79
SiO ₂	wt%	49.37	46.27	46.50	Ba	ug/g	63.16	73.25	105.49
TiO ₂	wt%	1.83	1.37	1.78	Y	ug/g	38.58	29.14	7.17
Al ₂ O ₃	wt%	13.73	15.41	16.64	Sn	ug/g	1.04	0.70	0.14
FeO	wt%	12.70	10.77	15.83	Sb	ug/g	0.41	0.08	0.38
MnO	wt%	0.24	0.19	0.23	Te	ug/g	<0.005	<0.005	<0.005
MgO	wt%	7.16	8.25	6.36	La	ug/g	6.23	3.76	0.25
CaO	wt%	9.94	12.05	5.76	Ce	ug/g	17.42	11.15	0.60
K ₂ O	wt%	0.18	0.29	0.68	Pr	ug/g	2.64	1.68	0.10
Na ₂ O	wt%	4.53	4.71	2.49	Nd	ug/g	13.04	8.74	0.67
LOI	wt%	1.23	1.62	2.85	Sm	ug/g	4.06	2.90	0.36
SUM	wt%	100.91	100.94	99.13	Eu	ug/g	1.53	1.15	0.41
P	ug/g	664.41	629.14	8.76	Gd	ug/g	4.89	3.32	0.58
S	ug/g	<0.01	<0.01	<0.01	Tb	ug/g	0.96	0.65	0.13
V	ug/g	337.22	261.30	270.34	Dy	ug/g	5.88	4.44	1.06
Co	ug/g	54.17	41.34	51.72	Ho	ug/g	1.32	1.01	0.25
Ni	ug/g	63.28	89.80	29.58	Er	ug/g	3.98	2.97	0.79
Cu	ug/g	39.87	51.76	25.99	Tm	ug/g	0.59	0.44	0.12
Zn	ug/g	83.12	84.24	109.01	Yb	ug/g	3.83	2.82	0.89
Ga	ug/g	17.01	16.69	16.72	Lu	ug/g	0.58	0.39	0.13
Se	ug/g	<0.01	<0.01	<0.01	Hf	ug/g	3.14	2.27	0.18
Rb	ug/g	7.40	4.00	10.52	Pb	ug/g	4.48	0.97	0.87
Sr	ug/g	137.23	131.09	52.33	U	ug/g	0.12	0.07	0.03
Zr	ug/g	136.48	96.15	5.14	Th	ug/g	0.33	0.17	0.02
Nb	ug/g	4.21	2.42	0.65	Ta	ug/g	0.34	0.18	0.06
Mo	ug/g	2.04	0.77	0.21	W	ug/g	116.76	20.13	35.45
Cs	ug/g	0.22	0.11	0.36					

3. ANALYTICAL METHODS AND ISOCHEMICAL PHASE MODELING

3.1 WDS Microprobe Analysis

Four samples; one white-mica schist and three eclogites with varying degrees of retrogression were analyzed using Wavelength Dispersive Spectroscopy by a JEOL JXA-8230 electron microprobe at the Stanford Mineral and Microchemical Analysis Facility at Stanford University (Figure 5) in order to better constrain the P-T history of the Elekdag lherzolite. Operational conditions were an accelerating voltage of 15 kV, a beam current of 20 nA and a beam width of 2 μm for all analyses. All results were then normalized according to the number of Oxygen atoms per base unit formula for each mineral, 12.00 for garnets and cpx, 22.00 for white micas. During the WDS analysis several BSE images were taken of the analyzed grains for reference.

3.2 Geothermobarometry

Within garnet-bearing metamorphic rocks the maximum temperature of metamorphism without melting can be estimated through the relative ratios of Fe^{2+} and Mg between coeval pairs of garnet and clinopyroxene. Coevolution is determined with the existence of straight mineral grain boundaries which do not show any signs of retrogression between them. There have been numerous garnet-clinopyroxene geothermobarometers calibrated over the years. Maximum pressure temperature boundaries were calibrated using 7 different geothermobarometers: (Krogh, 1988; Powell, 1985; Ellis and Green, 1979; Raheim and Green, 1974; Ganguly, 1979, cit. in Johnson et al, 1983; Ai, 1992; Krogh Ravna, 2000).

While variability between the different geothermobarometers can be minimal, in this study the most recent calibration of Krogh Ravna (2000) was plotted on the subsequent pseudosections.

3.3 Isochemical Phase Modeling Parameters

Isochemical Phase Diagrams otherwise known as pseudosections were calculated using the Gibbs free energy minimization suite of software known as Perple_X. Perple_X uses total chemical system approximations for a given bulk composition and then identifies which mineral species represent the lowest free energy at any specific temperature and pressure conditions given. Through this modeling, mineral phase equilibria, mineral compositional variation and modal abundance can be modeled and used to estimate P-T conditions for a given set of mineral assemblages. Given the more comprehensive nature of the calculated system, results may be more precise than traditional mineral pairing equilibria. Bulk rock composition was determined using average chemical compositional values for representative minerals and their modal abundance within the sample. While this is generally considered less accurate than using XRF whole rock data, chemical and mineral variability within even a fist size sample can lead to large differences between the bulk chemistry and the elemental composition of the analyzed area which is on the order of millimeters.

Perple_X models used the internally-consistent end-member thermodynamic database hp04ver.data (ds55: Holland and Powell, 1998) with a total of 8 calibrated end-member solution models: omphacite (omph (GHP): Green et al, 2007), epidote (Ep(HP): holland and powell, 1998), binary feldspar (San: Waldbaum and Thompson, 1968), chlorite (Chl(HP): Holland and Powell, 1998), garnet (Gt(WPH): White et al., 2000), clinoamphibole (cAmph(DP2): Diener et al, JMG 2011), phengite (pheng(HP) Holland and Powell, 1998), binary plagioclase feldspar (pl(h): Holland and Powell, 1998).

Equilibrium assemblage diagrams of each sample were calculated in an 11 component system consisting of Na_2O - CaO - K_2O - FeO - MgO - MnO - Al_2O_3 - SiO_2 - H_2O - TiO_2 - O_2 (NCKFMMnASHTO). This component system was chosen so that most of the major rock forming minerals in metabasites would be considered for each diagram (Gaidies et al, 2006). The P-T range modeled was between 1-2.5 GPa and 623-923 Kelvin as is appropriate for eclogite facies subduction metamorphism and should be sufficient to include most of the prograde and retrograde assemblages of interest. Neither SiO_2 nor H_2O were considered as saturated components due to the analyzed rocks being metabasites rather than metapelites. Oilm (Orthoilmmanite), rieb (riebekite), gl

(glaucophane), parg (pargasite), ts (tschermakite), fanth (ferrous-anthophyllite), acti (actinolite), mrie (magnesio-rebekite), mrb (magnesio-rebekite), anth (anthophyllite) pure phase endmembers were all excluded because of conflicts with the various solution models listed previously. Each Isochemical Phase Diagram was generated using the preceding parameters.

4. GEOTHERMOBAROMETRY AND PHASE MODELING RESULTS

4.1 Results of Microprobe Analysis

The afore mentioned microprobe was used to analyze garnets cores and rims for four different thin sections (4187, ED-66, ED-74 and ED-79) with the intent on using the results for both conventional geothermobarometry (Fe^{2+} -Mg exchange) and isochemical phase modeling. Selected samples are presented in Table 1. The garnets are predominately Almandine (Fe^{2+}) but exhibit distinct variations in Grossular (Ca) and Pyrope (Mg) from core to rim. Spessartine (Mn) contents also varied from core to rim. Zoning is typical of prograde subduction metamorphism with grossular contents being higher in the core and being replaced by more Mg dominant compositions towards the rim indicating increasing pressure as the garnet grew. Garnet almandine content ranges from 1.58-2.11 in the cores to 1.86-2.37 moles per crystalline unit at the rim. The Ca content of the garnets varies between analyses but falls within the range of 0.66-0.82 atoms 12.00 Oxygen unit at the core and decreasing to between 0.64-0.74 at the rims. Manganese (Mn), while present in small amounts (with the exception one profiled garnet) is none the less important when determining the Gibbs free energy of a given composition and so it was included in all calculations. Core values ranged from 0.07-0.26 and 0.01-0.04 for rims. The decrease in Ca, and to a lesser extent Mn from rim to core corresponds to an equal increase in pyrope (Mg) content: 0.14-0.37 in the core to 0.23-0.66 at the rims. This is as is expected from an increase in pressure where the smaller cellular unit (~90% unit volume) of pyrope is more stable than its larger grossular and spessartine counterparts. In addition to garnet, clinopyroxene forms the second most important constituent mineral in eclogites from a thermodynamic modeling perspective. The clinopyroxenes present in the samples varied greatly in composition, most likely due to the extensive retrograde metamorphism all samples experienced. They were omphacitic, with varying diopside (Mg,Ca-Si) and the higher pressure jadite (Na-Al) components. Clinopyroxene rims tended to be slightly higher in jadite content than cores, 0.376-0.457 Na per 12.00 Oxygen. Also of interest during analysis were white micas. Many showed distinct

phengite compositions (Silicon contents higher than 3 per 22 Oxygens) indicating increased pressure conditions and a useful barometer when modeling HP-UHP metabasites. Si-Mg values in white micas range from 3.265-3.546 when analyses are normalized to 22.00 Oxygen.

4.2 Geothermobarometry

The maximum temperature of many metamorphic rocks can be determined by measuring the ratio of Fe^{2+} and Mg within two minerals whose chemical formulae contain both and who have grown contemporaneously. As minerals heat up, their crystal lattice expands and this allows the exchange for cations between neighboring minerals, the rate and amount of exchange is directly dependent upon the temperature of the mineral in question, once temperatures decrease, the lattice locks again and the exchange ceases, freezing the process and recording the maximum temperature. Garnet and biotite are commonly used for metapelites whereas garnet and clinopyroxene are used with more mafic metamorphic rocks such as eclogites. As was previously mentioned, several different geothermometers have been calibrated. In this study maximum temperatures of metamorphism were investigated using 7 different geothermometers for a couple of mineral pairs. For the sake of simplicity only the most recently publicized calibration of Krogh Ravna (2000) is included within pseudosections. There was significant differences between the different thermometer calibrations. The Krogh Ravna (2000) calibration yielded significantly lower temperatures than the older calibrations (Table 4.1).

Table 4.1 : Results of different Fe/Mg exchange geothermometers of garnet clinopyroxene pairs.

Analysis #		Krogh (1988) Powell (1985) Ellis and Green (1979) Raheim & Green (1974) Ganguly (1979) Ai (1992) Krogh Ravna (2000)						
		T (K)	T (°C)	T (K)	T (°C)	T (K)	T (°C)	T (°C)
333	P @ 5 kbar	592.1	628.7	652.4	610.7	762.3	635.4	546.9
	P @ 25 kbar	319.1	355.7	379.4	337.5	489.2	362.3	273.8
	P @ 5 kbar	629.6	664.0	689.6	701.2	794.5	565.0	612.6
	P @ 25 kbar	356.6	391.0	416.6	428.0	521.4	291.8	339.4
334	P @ 5 kbar	667.1	710.8	735.1	687.3	846.0	731.6	627.6
	P @ 25 kbar	394.1	437.8	462.1	414.1	572.9	458.5	354.4
	P @ 5 kbar	710.3	751.0	777.3	789.1	881.9	731.6	703.5
	P @ 25 kbar	437.3	478.0	504.3	515.9	608.8	458.5	430.4
339	P @ 5 kbar	649.2	685.9	710.0	661.9	819.7	701.8	602.8
	P @ 25 kbar	376.2	412.9	437.0	388.8	546.5	428.6	329.6
	P @ 5 kbar	690.5	724.5	750.5	760.0	854.3	623.8	675.2
	P @ 25 kbar	417.5	451.5	477.5	486.8	581.2	350.6	402.1
346	P @ 5 kbar	668.0	705.0	729.1	679.9	839.3	724.4	622.2
	P @ 25 kbar	395.0	432.0	456.1	406.8	566.2	451.3	349.1
	P @ 5 kbar	710.6	744.7	770.9	780.6	874.9	643.7	697.1
	P @ 25 kbar	437.6	471.7	497.9	507.5	601.7	370.6	424.0
202	P @ 5 kbar	633.1	669.9	693.8	647.5	803.1	682.9	585.8
	P @ 25 kbar	360.1	396.9	420.8	374.4	530.0	409.8	312.7
	P @ 5 kbar	673.3	707.5	733.4	743.5	837.1	607.1	656.3
	P @ 25 kbar	400.3	434.5	460.4	470.3	564.0	333.9	383.2

4.3 Isochemical Phase Equilibrium Diagrams

Of the four samples analyzed two (ED-74 and ED-79) had an easily analyzable prograde assemblage for the purposes of creating a model. The bulk assemblage was estimated using average chemical compositions and modal abundances. ED-79 was modeled (Figures 4.1, 4.2 and 4.3) with the following bulk composition in weight amounts; SiO₂ 46.50, TiO₂ 1.780, Al₂O₃ 16.64, FeO 15.83, MnO 0.23, MgO 6.360, CaO 5.760, Na₂O 2.490, K₂O 0.680, H₂O 2.850, O₂ 0.250. All models were considered to be undersaturated with H₂O. The O₂ value of 0.250 is used to convert a portion, in this case (~10%) of ferrous iron (Fe²⁺) in ferric iron (Fe³⁺) in accordance with observed values of Fe³⁺ (Augite) in Cpx analyses. Stable assemblages were calculated according to their lowest free energies. Once a stable, suitable model was generated isopleths of elemental variability within a given mineral were generated for the same P-T space. One set of isopleths was generated each for Ca and Mg content in garnet, Fe and Mn content in garnet and Si content in white mica (phengite). Ca varied from 0.210-0.280 with a step of 0.01, Mg varied from 0.010-0.120 with a step of 0.01, Mn varied from 0.020-0.1200: 0.01, Phengite varied from 3.10-3.55: 0.05. All of these ranges represent values of minerals from our analysis. Where the differing isopleths all overlap gives an indication of possible peak pressure conditions of metamorphism. In addition, the Geothermometer of Krogh Ravna (2000) is overlain in order to give a maximum temperature constraint. According to one pseudosection, approximate maximum pressure of sample ED-79 is around 21 kbar (minimum error of +/- 2kb) at a temperature of ~360 °C (Figure 4.1, 4.2 and 4.3). Pseudosections generated for sample ED-74 have much lower pressures of 10-12 kbar and temperatures of ~500±25 °C (Figure 4.4 and 4.5). Maximum temperature as determined with Grt-Cpx exchange thermometry for ED-79 at such pressures is ~ 400±25 °C. This is in relative agreement with Altherr et al., 2004 which using activities of endmembers determined a minimum pressure of 1.3 GPa at 300-400 °C. Okay et al. (2006) reported PT conditions based on endmember activity models of 17-20 kb at 500 °C.

Grossular Component contour interval: 0.1000E-01; range: 0.2100 $\Rightarrow$ 0.2800 Mole fraction
 Min/Max contours $\Rightarrow$ thick solid/dotted curves
 Pyrope Component contour interval: 0.1000E-01; range: 0.3000E-01 $\Rightarrow$ 0.1200 Mole fraction
 Min/Max contours $\Rightarrow$ thick solid/dotted curves

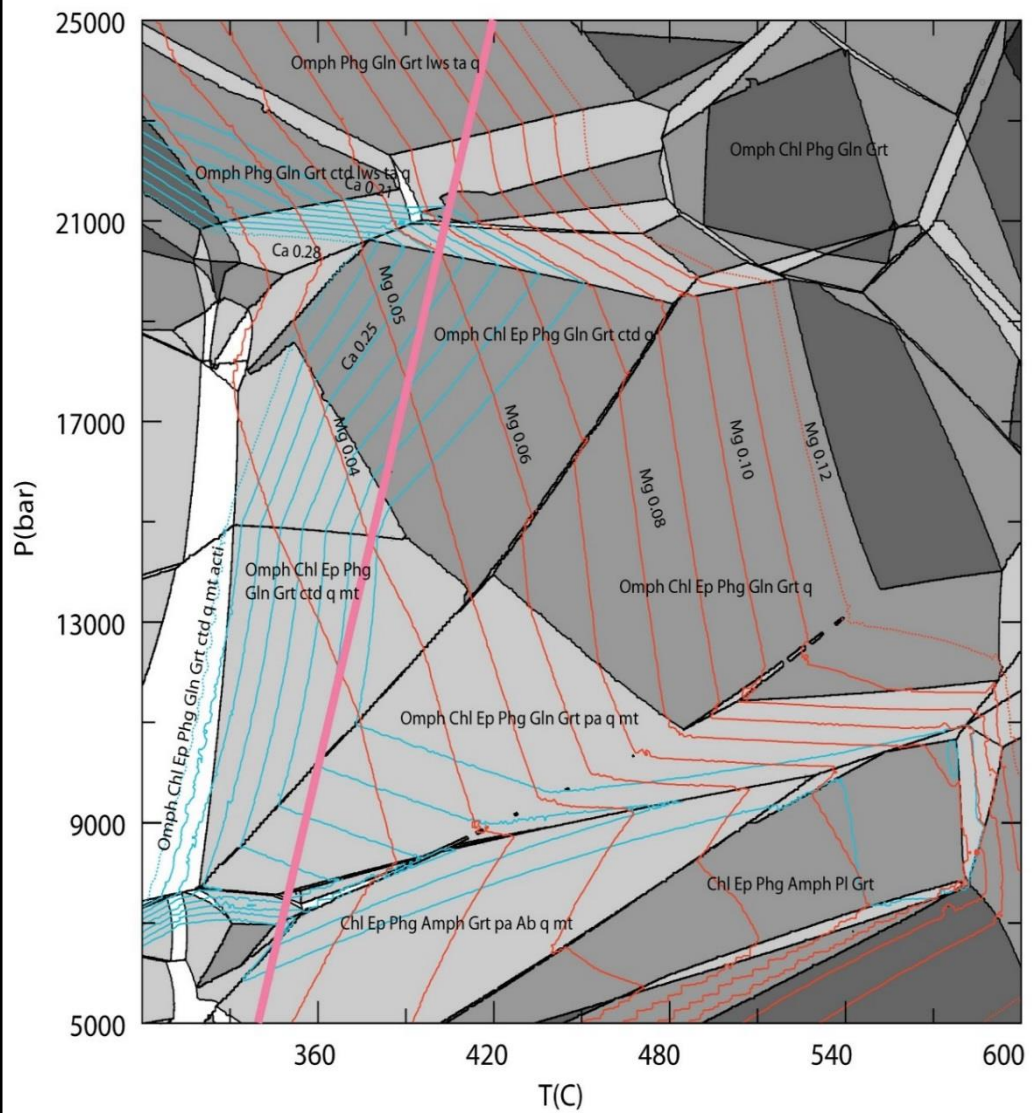


Figure 4.1 : Pseudosections of sample ED-79 a cumulate metabasite with colored garnet compositional isopleths and the geothermobarometer of Krogh Ravna (2000) plotted.

ed79f

Almandine Component contour interval: 0.2000E-01; range: 0.4800 => 0.7000 Mole fraction
variable range: 0.000 => 0.7880
Spessartine Component contour interval: 0.1000E-01; range: 0.2000E-01 => 0.1200 Mole fraction
variable range: 0.000 => 0.1508
Phengite Component contour interval: 0.5000E-01; range: 3.100 => 3.500 cpfu
variable range: 0.000 => 3.712

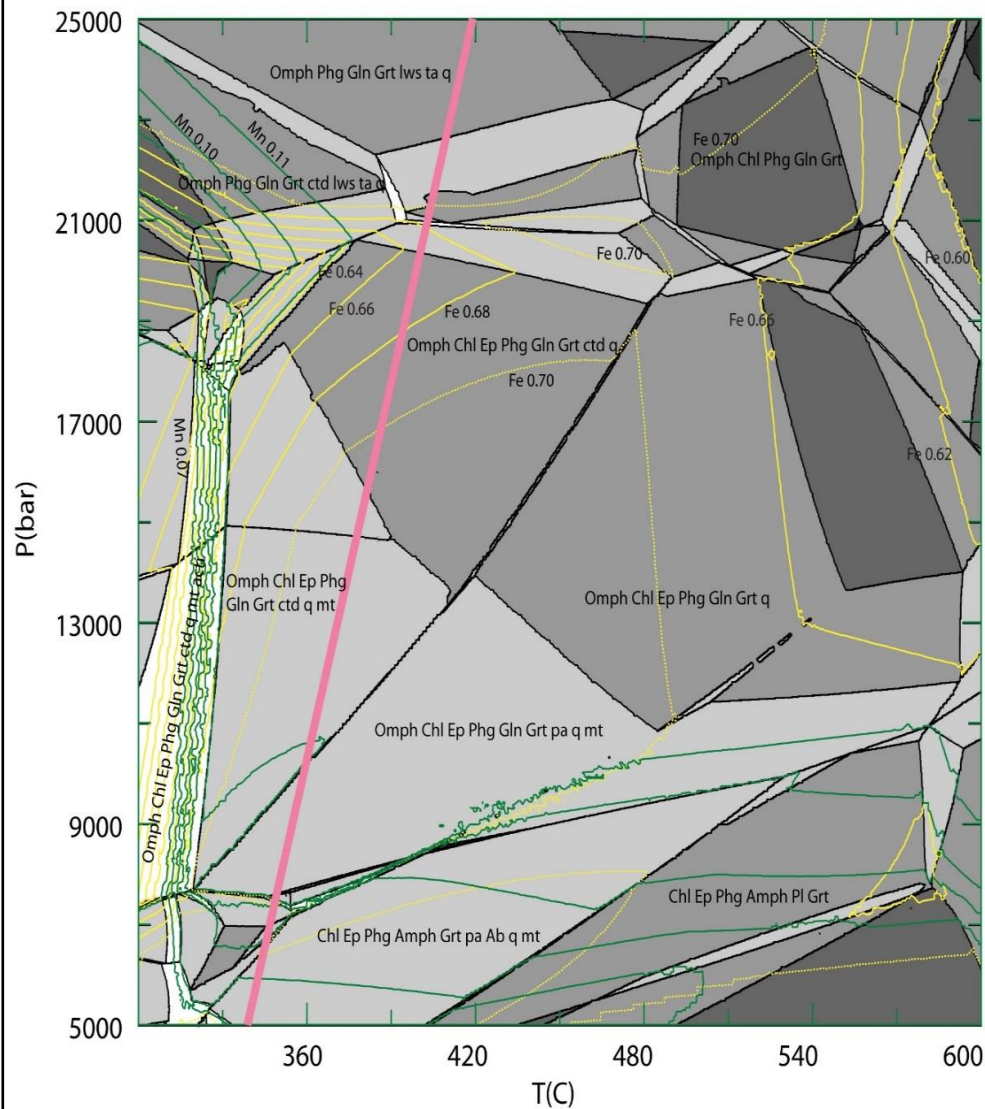


Figure 4.2 : Pseudosections of sample ED-79 a cumulate metabasite with colored garnet compositional isopleths and the geothermometer of Krogh Ravna (2000) plotted.

ed79f

C1Pheng(HP) ed79f_5.tab

contour interval: 0.5000E-01; range: 3.100 => 3.500

variable range: 0.000 => 3.712

Min/Max contours => thick solid/dotted curves

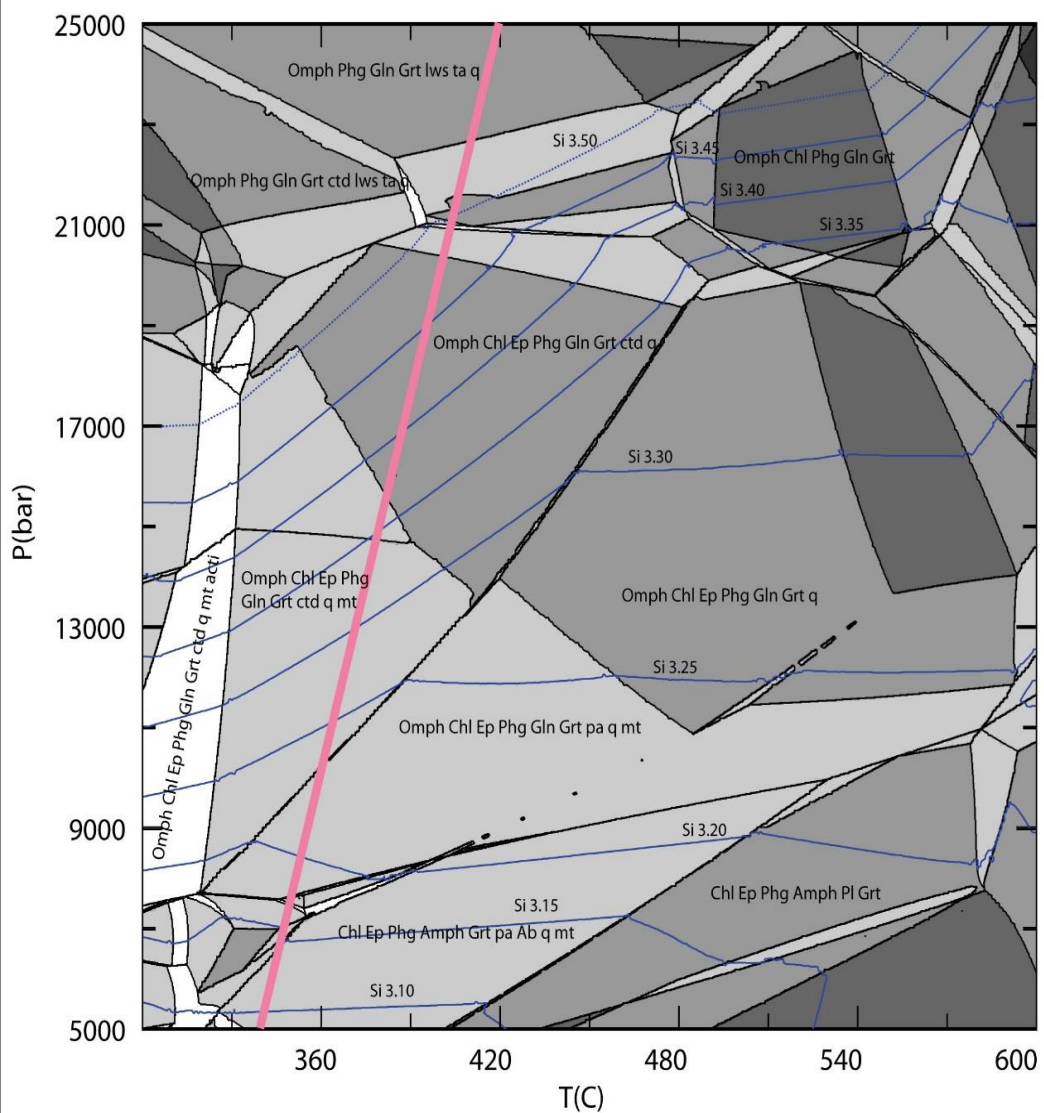


Figure 4.3 : Pseudosection of sample ED-79 a cumulate metabasite with the Si content of phengite isopleth and the geotherm of Krogh Ravna (2000) plotted.

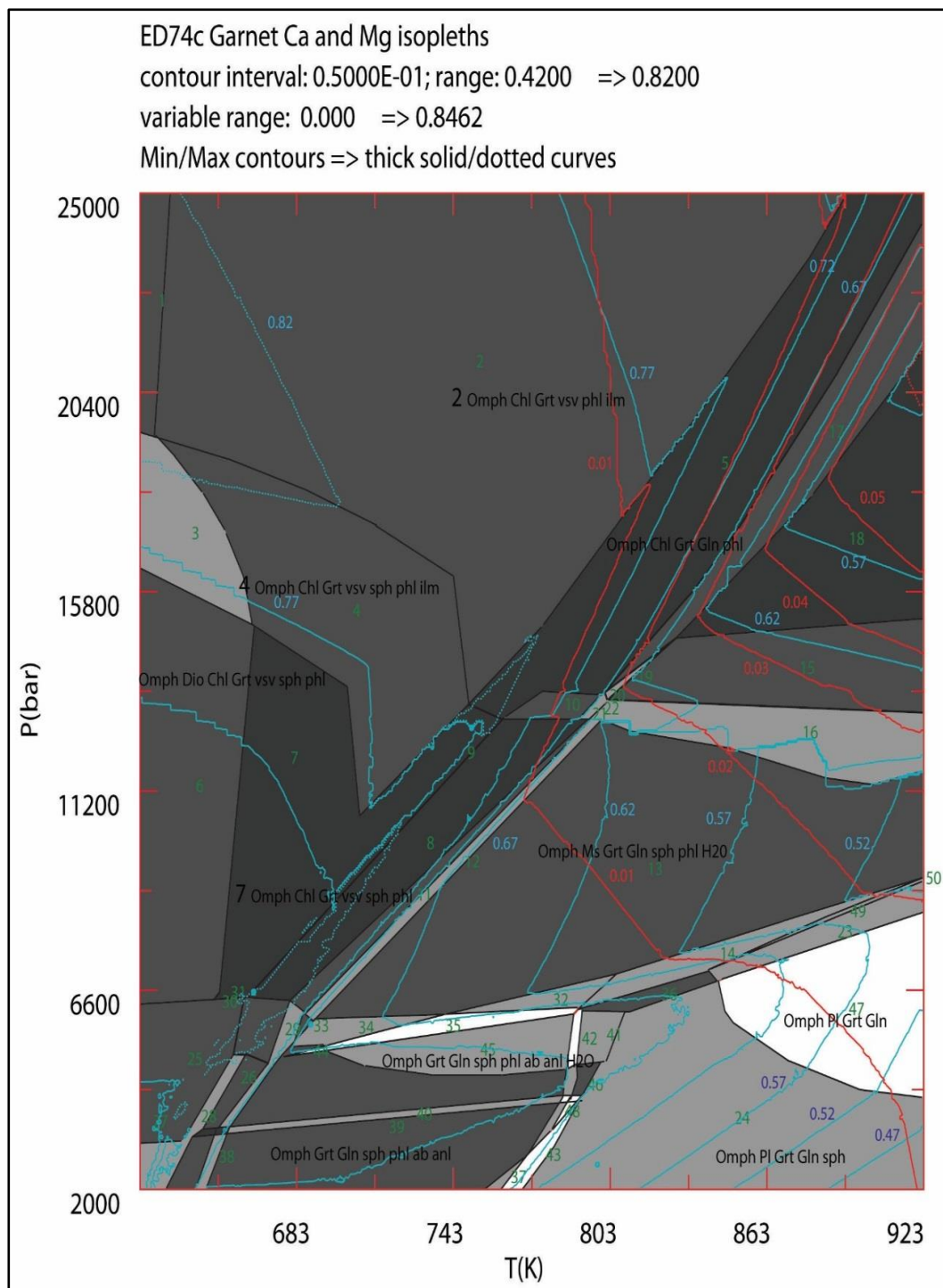


Figure 4.4 : Pseudosection of sample ED-74 with Grossular (Ca) and Pyrope (Mg) compositional isopleths plotted.

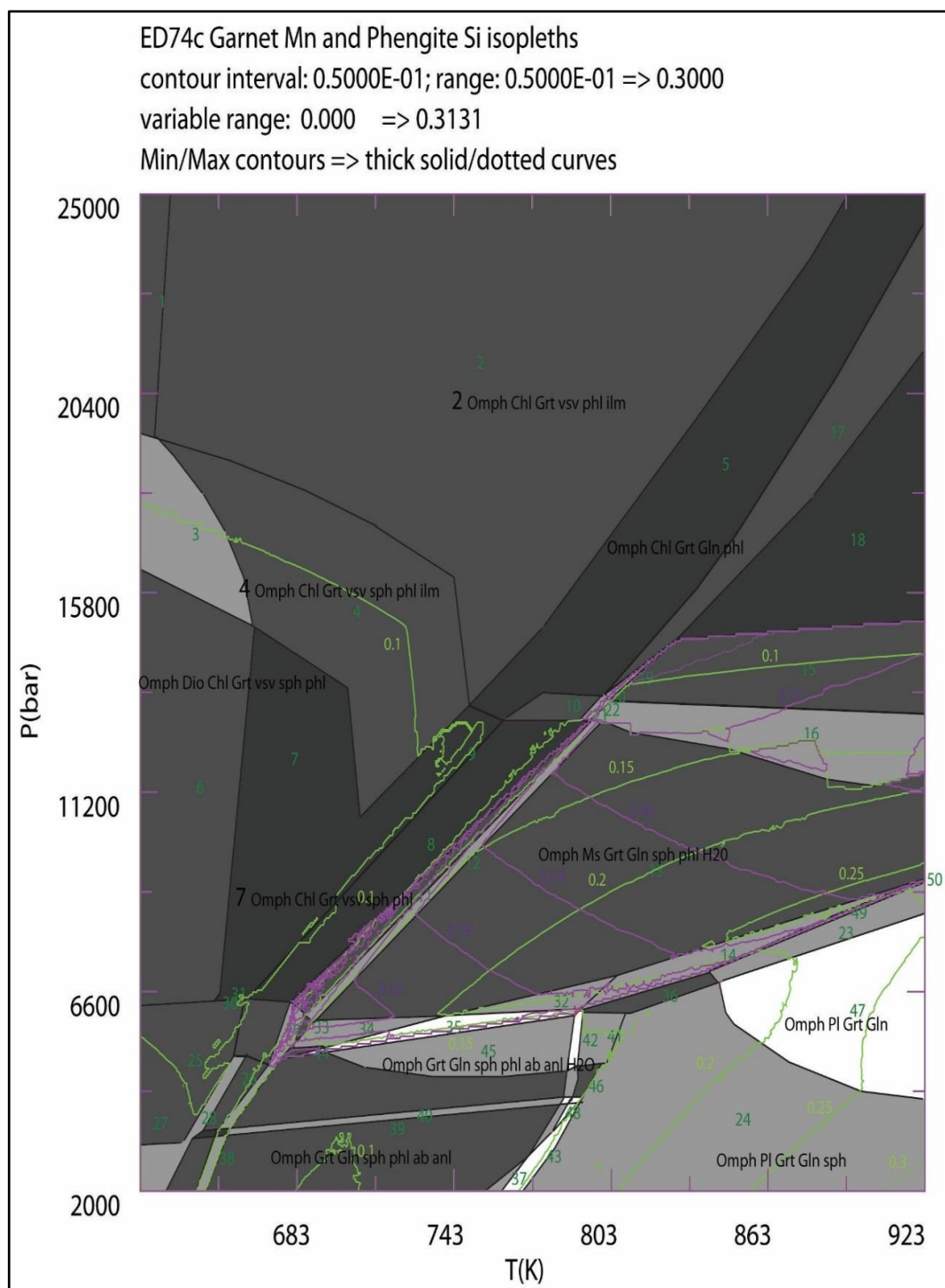


Figure 4.5 : Pseudosection of sample ED-74 with Spessartine (Mn) garnet composition and Phengitic mica Si content isopleths plotted.

5. DISCUSSION AND RECOMMENDATIONS

Two different explanations exist for the creation and evolution of the Elekdag Ophiolite. One is that the ophiolite itself is a supra-subduction zone ophiolite which formed above a subduction zone in front of the Çangaldağ volcanic arc (Ustaomer and Robertson, 1999) (Figure 5.1). The primary evidence for this is the presence of high ($> 0.60 \text{ Cr/Cr} + \text{Al}$) spinels as reported by Ustaomer (1993). The second model assumes an accretion to the southern margin of Eurasia similar to the accretion that took place in the cascades of northern California.

South of the passive margin of the Laurasian continent, the tectonic regime of the Tethyan oceanic basin reversed from a spreading center possibly due to slab roll back and became an active margin sometime prior to the Mesozoic. Tethyan oceanic crust began subducting below existing oceanic crust along southern margin of Laurasia. Dehydration of the down-going plate provided a fluid flux to the newly formed mantle wedge resulting in the formation of the Çangaldağ volcanic island arc. This arc is dated as pre-Jurassic due to the presence of granitic intrusions during the mid-Jurassic (Yilmaz, 1980; Yilmaz and Boztug, 1986). The current basal contact of the Çangaldağ Complex with the Domuzdağ is a late Cretaceous shear zone. The Elekdag Ophiolitic mélange under the supra-subduction zone model is assumed to have formed in the fore-arc region of the Çangal Island Arc because of the high Cr# of spinels found within (Donmez et al, 2014, Arai et al, 2006). In the second model the ophiolite is interpreted to have formed along the down going slab in a normal subduction environment (Ustaomer and Robertson 1999). Elekdag was viewed by (Ustaomer and Robertson 1999) as a discrete unit while others have considered it a part of the Domuzdağ which totally engulfs it. Further southward the Lower Cretaceous Domuzdağ accretionary mélange was assembled. Ages of metamorphism of the Domuzdağ are based on closure temperature ages for Phengitic mica and sodic amphiboles. Ar-Ar ages of 95.1 ± 0.6 to 116.5 ± 3.4 Ma were reported (Okay et al, 2006) and 83.7 ± 9.7 Ma (Ustaomer and Robertson 1999). Rb-Sr age dating of sodic amphiboles yielded ages of 106 ± 3 Ma, 107 ± 1 Ma, and 124 ± 9 Ma.

By the Cretaceous mantle wedge derived volcanism had begun ceasing and the Elekdag Ophiolite followed by the Domuzdağ mélangé and the Kargi microcontinent had amalgamated onto the Çangaldağ complex (Okay et al, 2006).

The metabasite samples in this study have NMORB affinities and PT estimates greater than would occur in a suprasubduction zone conditions and must have been part of the down-slab. Exhumation along reverse faults would have occurred during the amalgamation of the units along the margin of Laurasia. Chromitites found within the Elekdag ophiolite also show geochemical signatures of a mantle source in a forearc setting (Donmez et al, 2014).

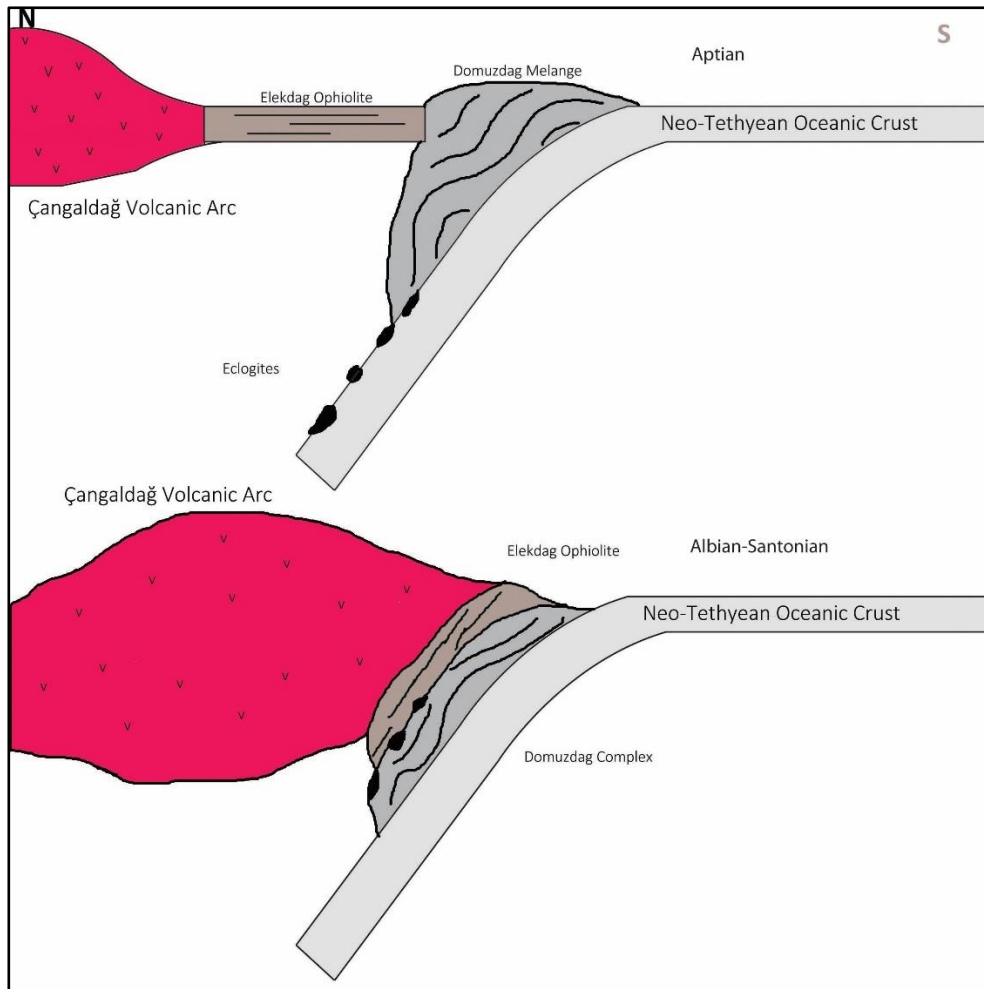


Figure 5.1 : Simplified schematic of a potential model for the geodynamic evolution of the Elekdag Ophiolite and exhumation of eclogite facies metabasites into a supra-subduction zone ophiolite.

6. CONCLUSIONS AND IMPLICATIONS

The metamorphic evolution of the Elekdag Ophiolite was investigated using petrographic and geochemical analysis of metamorphosed mafic rocks. Due to the highly retrograded nature of all samples, a quantitative prograde path is difficult to infer from the data. Isochemical phase diagrams indicate maximum pressures of metamorphism to be approximately 21 ± 2 kbar and peak temperatures of $\sim 400 \pm 25$ °C. The Krogh Ravna thermometer was plotted on the pseudosections however within the different geothermobarometers there was significant variation. The Krogh Ravna (2000) calibration gave lower maximum temperatures, lower than that of the eclogite facies, than the older calibrations of (Ellis and Green, 1979; Raheim and Green, 1974; Ganguly, 1979; Powell, 1985) which recorded temperatures of $\sim 450 \pm 25$ °C (Table 4.1). The error values given are the inherent minimum errors in the solution models and are likely larger due to uncertainties in the true bulk composition. To fully minimize the errors a 3D mineral chemical reconstruction would need to be done surrounding each pairing in order to account for variation in all dimensions not just the 2 dimensions represented by a thin section surface. Zoning within garnets does indicate a clear prograde evolution in their growth which is consistent with subduction zone metamorphism. Geochemical analysis of the metabasite samples show NMORB affinities. Other studies (Donmez et al, 2014; Gunay et al, 2016) report boninite affinities and theoretically supra-subduction zone ophiolites are characterized by their presence. While the samples analyzed in this study have the high magnesium content (oxide weight %) and high Mg# expected, the TiO₂ and SiO₂ content is lower than that of boninites as is the Ni content. Ti/Zr and La/Yb ratios also do not fall within the boninite range.

Given that one sample seems to have been subducted to approximately 21 ± 2 kbar, a key question is how can a supra-subduction zone serpentinite body contain blocks of subduction zone metamorphic rocks? It is possible that the metabasites were exhumed along reverse faults within the Domuzdağ complex which would have been positively buoyant relative to the mantle wedge.

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