Carbonation and listwanite formation in the Leka ophiolite, north-central Norway

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Abstract

The majority of previous reports on listwanitation since the term was first introduced by Rose (1837), have focused on its association with Au deposits. In recent years, environmental-threat in the form of an elevated atmospheric CO₂ level has led to an interest in listwanitation as it may represent an analog to a large-scale industrial process of CO₂ storage. This rising interest has led to a more thorough investigation of listwanites all over the world. As the first report, the thesis focuses on describing and providing information regarding listwanitation of the rocks in the Leka Ophiolite Complex (LOC).

The association of listwanitation with highly altered and deformed rocks situated in ophiolites, which is also the case for the LOC, complicates the study of listwanites as it often occurs concurrent with/ or post other processes. This association further complicates the interpretation of age relations, fluid sources and element migration.

To classify a rock as listwanite, a few features needs to be fulfilled according to Halls and Zhao (1995). Field observations, petrographically and geochemical data presented here, show that several of the samples analyzed classifies as listwanite.

Listwanitation affected rocks of both pyroxenitic and gabbroic protoliths. During listwanitation of the plagioclase-bearing gabbro, a number of Al-rich phases including tourmaline and kaolinite was formed. T-XCO₂ diagram of one of the gabbro-listwanites display that temperatures during formation of the observed mineral assemblage range between 90 °C and 220 °C at XCO₂ = 0. Formation of smectite-illite, Cr-muscovite and fuchsite indicate that K⁺ metasomatism has occurred which is commonly associated with seawater derived fluids. Boron needed to form tourmaline in the gabbro-listwanites also indicate seawater involvement probably associated with ocean-floor serpentinization and smectite-formation. Based on these findings, listwanitation is regarded to be associated with ocean-floor metamorphism with a fluid source partly composed of seawater.

Listwanitation occur along large NE-SW trending faults or shear zones. It is suggested that if not concurrent, listwanitation precedes the faulting and that the carbonation and hydration may have played a vital role in the deformation of the LOC.
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Chapter 1: Introduction

Ophiolites are formed as a part of the oceanic lithosphere in spreading ridge environments, and later emplaced onto continental margins (Dewey, 1976; Prestvik, 1980; Furnes et al., 1988; Searle and Cox, 1999; Titus et al., 2002; Berger et al., 2005; Burg, 2013). The study of ophiolites is of great importance as they may preserve information regarding active modern ocean-floor processes. Ocean-floor metamorphism includes amongst many processes; amphibolitization, rodingitization (Austrheim and Prestvik, 2008), serpentinization (Iyer et al., 2008) and carbonation (Plissart et al., 2009) that are highly fluid dependent reactions with the two latter ones being sensitive to changes in \( f(\text{CO}_2) \). Products from these alterations are recognized in numerous ophiolites. However, in which settings and under what conditions they occurred are topics of dispute.

The term “listwanite” was first introduced by Rose (1837) and is today mainly used for rocks composed of magnesite and quartz with minor fuchsite (Cr-muscovite) (Halls and Zhao, 1995). Listwanites are associated with rocks that contain carbonates, serpentine, talc, pyrite, quartz and variations of micas (Halls and Zhao, 1995), that are also formed as a result of carbonation which is often associated with serpentinization (Halls and Zhao, 1995).

The listwanite-forming carbonation, termed listwanititation, involves the replacement of silicates by carbonates in ultramafic rocks under low temperatures of approximately 270°C-390°C (Halls and Zhao, 1995). The replacement, caused by the silicates donating their metal cations; \( \text{Fe}^{2+}, \text{Mg}^{2+} \) and/or \( \text{Ca}^{2+} \) to carbonates, leads to a release of silica which forms quartz (Halls and Zhao, 1995). The donation usually follows a previous replacement e.g. serpentinization but in rare occasions the donation occurs directly from primary olivine and pyroxene (Halls and Zhao, 1995), e.g.: \( \text{Ol} + \text{CO}_2 \rightarrow 2\text{Mgs} + \text{Qtz} \). The direct replacement of Ol and Px with Qtz is together with gold association and large CO\(_2\) storage potential, features of interest regarding listwanititation (Halls and Zhao, 1995). Although listwanititation, is in most cases associated with serpentinization of peridotites (Halls and Zhao, 1995), gabbro and pyroxenite may also be carbonated to form listwanite-like rocks (Halls and Zhao, 1995; Plissart et al., 2009). Such rocks are of varies compositions, however, Cr-rich mica, quartz and carbonate are minerals commonly associated with gabbro and pyroxenite derived listwanites (Halls and Zhao, 1995; Plissart et al., 2009).
Amphibolitization, rodingitization (Gussone et al., In review; Austrheim and Prestvik, 2008), serpentinization (Iyer et al., 2008) and carbonation (Bjerga, 2014) of the rocks in the LOC have previously been described, however this is the first report on listwanitation. The rocks analyzed are of various protoliths, and present mineral assemblages and chemical compositions exhibit a large variation of alteration. Serpentinization are only found to have affected the peridotites, while carbonation have occurred to samples by all the investigated protoliths. Listwanitation, which include formation of quartz (Halls and Zhao, 1995; Plissart et al., 2009; Hinsken et al., 2017), is only found to have occurred to the protoliths of a pyroxenitic and gabbroic composition. While a stage prior to listwanitation are suggested to be represented by peridotite derived Soapstone (Fig. 3.2 C) which form by carbonation of peridotite and the studied soapstone contain talc and magnesite, but lack quartz.

According to Halls and Zhao (1995), formation of listwanite is associated with tectonism and plutonism generally localized near major faults or shear zones where hydrothermal fluids are channelized. The low temperature, high abundance of reactive ultramafic minerals and CO2 sufficiency needed to form listwanite, restricts the formation to unique settings where these conditions are met. The formation of listwanite in the LOC implies that these conditions have been met at some point. However, apart from the association with ocean-floor metamorphism in the thesis, when and in which tectonic setting listwanite formed in the LOC are hard to interpret.
Chapter 2: Regional geology

2.1 Geological setting

The Scandinavian Caledonides represent an Early Paleozoic collision belt with involving a range of lithologies, and structural properties reflecting the affects from numerous deformation events and intensive metamorphism (Roberts, 2003). The rock complexes composing the innumerable thrust sheets comprising The Scandinavian Caledonides are built up of a number thrust sheets emplaced as a result of the collision between the Baltica and Laurentia plates (Roberts, 2003; Corfu et al., 2007). The tectonic event involved; the rifting at the end of the Neoproterozoic, formation and the following closure of the Iapetus Ocean, and the collision between the two plates in Silurian to Early Devonian time (Roberts, 2003; Corfu et al., 2007). During the converging of the two, the Baltoscandian margin was subducted in a westward direction and overridden which led to the Precambrian Shield being thrusted by terranes derived from the Baltica margin, as well as exotic, oceanic and arc terranes from both the Iapetus Ocean and Laurentia (Stephens, 1988; Roberts, 2003; Corfu et al., 2007).

2.2 Allochthons

The collision between Laurentia and Baltica resulted in a tectonostratigraphy involving four recognizable tectonic units named according to their position in the nappe stack: Lower, Middle, Upper and Uppermost Allochthon can be recognized (Fig. 2.1) (Roberts, 2003). The autochthonous basement consists of Precambrian rocks of the Fennoscandian Shield, covered unconformably by Cambrian to Ordovician sediments (Stephens, 1988). The Lower Allochthon is composed by late Precambrian to Ordovician sediments with occasionally pieces of Precambrian crystalline rocks, and Silurian sediment covers appear at local scale (Stephens, 1988). The Middle Allochthon have close to similar composition as the Lower Allochthon (Stephens, 1988). However, it is also comprised of mylonitized Precambrian crystalline rocks and mainly late Precambrian sediments (Stephens, 1988). The lower part of the Upper allochthon consist of medium to high-grade metasedimentary rocks, while the upper part is composed of ophiolites, magmatic arcs and marginal basin associations suggested to have derived from the Laurentian side of the Iapetus ocean (Stephens, 1988; Stephens and Gee, 1989; Roberts, 2003; Corfu et al., 2007). The Uppermost Allochthon is composed of plutonic rocks, granitoids, sediments and ophiolites (Brueckner and van Roermund, 2004; Corfu et al., 2007),
and stable isotopic and radiometric dating strongly suggest that it is a part of the Laurentian affinity (Roberts et al., 2001, 2002; Melezhik et al., 2002; Yoshinobu et al., 2002; Roberts, 2003).

2.3 The Scandinavian Caledonides

The Scandinavian Caledonides comprise the younger part of the succession concerning Scandinavia, while the older part, comprised of the autochthonous basement and numerous nappes, has a different origin (Stephens, 1988; Roberts, 2003; Corfu and Andersen, 2016).

The oldest part of the succession, the autochthonous basement, is comprised of Paleoproterozoic and Archean crystalline crust derived from the Fennoscandian Shield, with ubiquitous Anorthosite-Mangerite-Charnockite-Granite complexes with local differences caused by deformation, metamorphism, magmatism and intrusions (Corfu et al., 2014; Corfu and Andersen, 2016). These complexes reflects the formation of the Sveconorwegian orogeny, involving the related processes as convergence, subduction and/or post-collision between the Fennoscandian Shield and another plate, believed to be Amazonia during the Meso- and Neoproterozoic orogenies (Bingen et al., 2008; Corfu et al., 2014; Corfu and Andersen, 2016).

At the end of the Neoproterozoic the rifting of Baltica from Laurentia initiated, resolving in the formation of the Iapetus Ocean (Roberts, 2003; Brueckner and van Roermund, 2004; Corfu et al., 2007). The opening and the following closure of the Iapetus Ocean being the drive force behind the formation of the Scandinavian Caledonides by initiating four important tectonothermal events (Roberts, 2003; Brueckner and van Roermund, 2004). The four tectonothermal events recognized as compressive/transgressive events are; the Finnmarkian (Late Cambrian), Trondheim (Early Arenig), Taconian (Mid-Late Ordovician) and Scandian (Mid Silurian – Early Devonian) had an impact on different areas of the Scandinavian Caledonides (Roberts, 2003; Brueckner and van Roermund, 2004). The realization of that an event effecting northern Norway in Late Cambrian-earliest Ordovician time had occurred, came through Sturt et al. (1978) work with K-Ar and Rb-Sd dating (Roberts, 2003). A theory that was later supported by $^{40}$Ar-$^{39}$Ar and Sm-Nd dating (Mørk et al., 1988; Dallmeyer et al., 1991; Roberts, 2003). This event, the Finnmarkian event, affected the outermost Baltoscandian margin, parts of the Middle and the upper Allochthon and the Seve-Kalak super terrane (Andréasson et al., 1998; Roberts, 2003). Believed to be a result of a collision between the Baltoscandian margin and either a magmatic arc positioned in the Ægir sea or a partly developed
microcontinental block (Sturt and Roberts, 1991; Torsvik and Rehnstrom, 2001; Hartz and Torsvik, 2002; Roberts, 2003).

Later in the Early Arenig the Trondheim event occurred which led to deformation and metamorphism of central Norway, and it is recognized in the Köli Nappes of the upper Allochthon (Roberts, 2003). From analyzes of the basal parts of one of the fragmented ophiolites (Eide and Lardeaux, 2002) and U–Pb zircon dating of plagiogranite dykes and sheets in some of these ophiolites yielding ages of 493 – 482 Ma (Dunning and Pedersen, 1988; Melezhik et al., 2002; Roberts et al., 2002; Roberts, 2003), concludes that the event included both ophiolite obduction and blueschist metamorphic parageneses (Roberts, 2003). In the Hovin Group which lie unconformably upon the ophiolitic assemblages of the western parts of central Norway that are believed to have been obducted upon Gula Complex, the Early Ordovician faunas are mostly Laurentian and the oldest fossils are of Mid to Late Arenig age (Roberts, 2003). This both indicate the duration of the Trondheim event, and the interaction between the Laurentian and Baltic plate as a result of the Baltic plate’s anticlockwise rotation away from the Siberia plate (Torsvik and Rehnstrom, 2001; Roberts, 2003).

During the Mid to Late Ordovician, the following tectonothermal event with supporting evidence from the Upermost Allochthon and the very highest and outboard Köli terranes took place (Roberts, 2003). This event, the Taconian event, was first recognized in Arenig-Llanvirn sedimentary rocks, which were deformed and low grade metamorphosed prior to the intrusion of Smøla-Hitra batholith in Late Ordovician – Earliest Silurian (Roberts et al., 2001, 2007; Roberts, 2003). U-Pb dating of ophiolites and arc-related assemblages further south in Norway show that these complexes obducted onto the continental margin for around 470-465 Ma (Sturt et al., 1978; Dunning and Pedersen, 1988; Pedersen and Furnes, 1991; Pedersen et al., 1992; Yoshinobu et al., 2002; Roberts, 2003). In the following years, post the Taconian event, the complete closure of the Iapetus Ocean led to an oblique collision between the Baltic and Laurentian plate in Late Silurian to Early Devonian time (Roberts, 2003; Corfu et al., 2007). The collision, which also included the subduction of the Baltoscandian margin, is referred to as the Scandian event (Roberts, 2003; Corfu et al., 2007). During the Scandian event the Iapetan and Laurentian terranes, composing the Upermost Allochthon, were obducted onto the Scandinavian Caledonides nappes (Roberts, 2003). During the same event, all of the principal Allochthons were to a different extent, metamorphosized and exposed to magmatism (Roberts, 2003). The development of The Western Gneiss Region also occurred during the Scandian event, as subduction of the Baltoscandian margin led to UPH conditions which are only
obtainable at great depths in a subduction setting with very fast exhumation (Terry et al., 2000; Roberts, 2003).

**Post-Caledonian Extension:**

During the Scandian event the regional tectonic regime was of a compressional setting, this setting was later in the Early Devonian replaced by a tensional setting as a result of the convergence of the Baltica and Laurentia plate (Fossen, 1992; Roberts, 2003). The late- to post-orogenic extension, derived from the tensional setting limited to the western regions of the Scandinavian Caledonides, caused movement along detachment zones, formation of post-orogenic extensional basins and exhumation of high-pressure rocks (Andersen and Jamtveit, 1990; Fossen, 1992, 2000, 2010; Fossen and Dunlap, 1998; Eide et al., 2005). In southern Norway, two types of orogenic extension have been recognized by Fossen (Fossen, 1992). Backsliding of the orogenic wedge along already existing thrust sheets in the early stages of the collapse, followed by crustal collapse and formation of high angle listric faults (Fossen, 1992; Titus et al., 2002).

2.4 **The Helgeland Nappe complex (HNC)**

The Helgeland Nappe complex (HNC) is the structurally highest part of the Uppermost Allochthon in north-central Norway. It is comprised of five nappes mainly composed of medium-grade metamorphosed sedimentary rocks intruded by plutonic rocks (Fig. 2.1) (Heldal, 2001; Barnes et al., 2007; McArthur et al., 2014). The five nappes comprising the HNC, in a structurally ascending order; the Horta nappe, Sauren-Torghatten, Lower, Middle and the Upper nappe, are all recognized in the southern and central parts of the HNC and are divided into two group based on composition (Fig. 2.1) (Heldal, 2001; Barnes et al., 2007; McArthur et al., 2014). One of the groups, which has no exposed depositional basement, consists of the Horta, Lower and Upper nappes which are composed of migmatitic gneiss, calc-silicate rocks and marble (Heldal, 2001; Barnes et al., 2007; McArthur et al., 2014). The other group, deposited on exposed ophiolitic rocks, consists of the Middle and Sauren-Torghatten nappes which are composed of meta-conglomerates, marbles, pelitic schist and calc-silicate schist (Heldal, 2001; Yoshinobu et al., 2002; Barnes et al., 2007; McArthur et al., 2014). The former group is believed to have been deposited in continental-shelf environments while the latter in continental-margin basins (Heldal, 2001; Barnes et al., 2007). The group believed to have been deposited in continental-margin basins, is of great significance as they display the contact between the preserved Iapetan oceanic crust and first sedimentation (McArthur et al., 2014).
This contact is observed on three islands; Rødøy and Bolvær located offshore of the central parts of the HNC and Leka located in the South-Western parts (McArthur et al., 2014).

Figure 2.1: Tectonostratigraphic map displaying the major units within the Helgeland Nappe Complex (HNC) and simplified geological cross section along the profile X – X’. From McArthur et al., (2014).
2.5 *Leka Ophiolite Complex (LOC)*

The Leka Ophiolite Complex (LOC) (Fig. 2.2) outcrops on the island of Leka (65°5′0″N 11°38′1″E) in Nord-Trøndelag, Norway. The LOC is considered to be one of the oldest and best preserved ophiolites within the Uppermost Allochthon in the Scandinavian Caledonides (Prestvik, 1972; Furnes *et al.*, 1988; Iyer *et al.*, 2008). U-Pb zircon dating of quartz-keratophyres in the LOC conducted by Dunning and Pedersen (1988) yields a crystallization age of 497±2 Ma, which correlates with the yielded age attained with the same methods of the Karmøy Ophiolite Complex (497±2 Ma) (Dunning and Pedersen, 1988). The LOC formed in the North Iapetus Ocean as a part of its oceanic lithosphere (Dunning and Pedersen, 1988; Titus *et al.*, 2002). Then obducted upon the Laurentia plate pre-the Taconian event, and later transferred onto Baltica during the Scandian orogeny where the mantle section formed an open synform (Dunning and Pedersen, 1988; Titus *et al.*, 2002; Roberts, 2003).

![Geological map of Leka, Storøya and Madsøya, displaying location of the various comprising units. Location of the studied area is marked with a black square. From Michels *et al.*, (2018).](image)
2.5.1 *General geology*

The well-preserved ophiolitic components are all exposed as several separated blocks in the LOC (Fig. 2.2) (Michels *et al.*, 2018). The mantle harzburgites, comprising the western region and central northern parts of the LOC, represents the mantle section while the layered crustal sequence, composed mainly of dunitic rocks with chromite layers, comprise the central parts of the LOC (Fig. 2.2) (Furnes *et al.*, 1988; Iyer *et al.*, 2008; Michels *et al.*, 2018). The third principle component, composed of metasediments of the Skei Group, lies unconfirmable on top of the most eastern parts of the metagabbros (Fig. 2.2) (Furnes *et al.*, 1988). The rocks present in the LOC have a metamorphic grade of upper greenschist to amphibolite facies (Prestvik, 1972; Dunning and Pedersen, 1988), and two distinct sets of faults (Dunning and Pedersen, 1988; Titus *et al.*, 2002). The two sets of faults formed due to brittle deformation, consist of the larger NE-SW trending faults and the smaller NW-SE trending ones (Furnes *et al.*, 1988; Titus *et al.*, 2002). At local scale deformation zones consisting of breccia and shear zones in different rock units are observed (Iyer *et al.*, 2008).

2.5.2 *Ophiolitic components*

2.5.2.1 *Mantle section*

The mantle section represents the structural lowest part of the lithostratigraphy exposed in the LOC (Fig. 2.3), and it is mainly composed of harzburgites with variable sized dunite bodies and pyroxenite veins (Furnes *et al.*, 1988; Albrektsen *et al.*, 1991). The eastern segment (Fig. 2.2) is 1.5 km thick, composed of all the components illustrated in Fig. 2.3, while the western segment only represents the top 500-600 m of the profile displayed in Fig. 2.3 (Furnes *et al.*, 1988).

The harzburgites in the mantle section display an imperfect foliation defined by 2-10mm thick bands, aligned and/or flattened aggregates or crystals of pyroxene, subsequently altered to serpentine (Furnes *et al.*, 1988; Iyer *et al.*, 2008). The foliation is a result of high-temperature plastic flow of the mantle sub an active spreading ridge (Nicolas *et al.*, 1973; Nicolas and Prinzhofer, 1983; Furnes *et al.*, 1988). Open folding of the lower parts of the mantle section led to a random oriented fabric, while the foliation in the upper part progressively obtain parallelism towards the layered cumulates (Furnes *et al.*, 1988). The changes in fabric correlates with abundance and type of dunite in the different parts of the section (Furnes *et al.*, 1988; Albrektsen *et al.*, 1991), with the abundance...
of dunite increasing towards the layered cumulates. Small dunite bodies and tabular dunites are present throughout the section, while branching dunite veins and large dunite bodies are only present in the upper parts (Furnes et al., 1988; Albrektsen et al., 1991). The progressively more dunitic composition towards the cumulate section is a feature associated with ophiolite complexes of a residual origin (Nicolas and Prinzhofer, 1983; Furnes et al., 1988).

The mantle section in the LOC is cut by various generations of veins and dikes composed of clinopyroxene, orthopyroxene and/or olivine websterite (Furnes et al., 1988; Maaløe, 2005). The veins and dikes occur either as cross-cutting or parallel to the fabric of the mantle harzburgite, with their variation in size and shape being a result of partial melting (Furnes et al., 1988; Maaløe, 2005).

The mineral assemblage of the Lauvhatten mantle harzburgites (eastern segment) is composed of olivine, primary and secondary clinopyroxene, Cr-spinel, ferritchromite, magnetite and serpentine±brucite/clinochlore (Iyer et al., 2008). The forsterite composing the olivine vary within the range of Fo$_{88}$ and Fo$_{93}$, where the olivine in the peridotites have a Fo-content from 88 to 92 and from 89 to 93 within the dunites (Furnes et al., 1991). The difference in Fo-content of olivine in the mantle section compared to the olivine from the layered cumulates, together with the appearance of chromite layers indicate where the boundary between these two components are located (Furnes et al., 1988).
2.5.2.2 The layered crustal sequence

The sub-vertical layered ultramafic cumulates consists of layers on three different scales: the large scale layering, macro-rhythmic units and small-scale rhythmic layers (Furnes et al., 1988). The large scale layering, well-exposed within the Skråa block, has a thickness of up to several hundred meters and is comprised of dunite-dominated layers alternating with sequences dominated by wehrlites and pyroxenites (Fig. 2.3) (Furnes et al., 1988; Iyer et al., 2008). Within the large scale layering, macro-rhythmic units with a thickness of 10-50 m and variations in
composition both internally and in relation to each other are found (Furnes et al., 1988; Pedersen et al., 1993). Typical for these units in dunite subzones are olivine adcumulates at the base, chromitite, chromite and/or sulfide-enriched horizons occurring frequently 1-3 m above, and gradually enrichment of post-cumulus pyroxene upwards (Furnes et al., 1988; Pedersen et al., 1993). The gradual enrichment of pyroxene are first seen as scattered grains of clinopyroxene and locally orthopyroxene, and ends up with an increasing amount until cotectic proportions with olivine are reached (Pedersen et al., 1993). This is displayed by some units having olivine-clinopyroxene cumulates at the top, which indicate the crystallization sequence: olivine->chromite->clinopyroxene->orthopyroxene (Furnes et al., 1988). The same sequence of crystallization is displayed in some of the small-scale rhythmically layers, which are 10-30 cm and either uniform or modally graded (Furnes et al., 1988). Within the equal oriented rhythmic layering veins of wehrlitic and pyroxenitic segregation are common (Furnes et al., 1988; Iyer et al., 2008). These veins of a few to several tens of cm show variable orientations, with the most being aligned parallel to each other and the layering, but cross-cutting forming net-like patterns are not uncommon (Furnes et al., 1988; Iyer et al., 2008).

The sharp discordant boundary between the layered ultramafic cumulates and the overlaying gabbroic cumulates is well exposed in the south-western parts of the Steinstind block (Furnes et al., 1988; Iyer et al., 2008). The three different scales of layering observed in the ultramafic cumulates are less pronounced but still recognizable in the layered gabbros (Furnes et al., 1988). In this section the large scale layers are defined by thick ultramafic units interlayered with units of a composition including gabbroic rocks (Furnes et al., 1988). The first units with metagabbro composition are observed on top of the macro-rhythmic units from the ultramafic cumulates, extending the sequence of crystallization by adding plagioclase as an end member (Furnes et al., 1988). The addition of plagioclase led to alteration of dunite, giving wehrlite followed by gabbro, this being displayed in the decrease of dunite and the correlating increase of metagabbro upwards in the section (Furnes et al., 1988). In one part of the section, complete absence of dunite led to domination of metagabbro layers resulting in a several hundred meters thick metagabbroic section mainly composed of rhythmically layered mesocratic metagabbro with occasional layers of wehrlite, phroxenitic or melanocratic gabbro (Furnes et al., 1988). The transition from layered metagabbros to vari-textured metagabbros is exposed on Masøya and in the area SW of Skei, where rhythmic layering are shifting to fine lamination and finally vari-textured metagabbro (Furnes et al., 1988, 1991). The vari-textured metagabbro is massive with
grain size from fine to coarse and overlain by pseudo-stratigraphically dyke swarms and pillow lavas (Furnes et al., 1988, 1991).

The dunites present in the layered crustal sequence are composed of olivine, serpentine ± brucite, Cr-spinel and magnetite, and the wehrlites of primary and secondary clinopyroxene, olivine, serpentine ± brucite/clinochlore, Cr-Al spinel, ferritchromite and magnetite (Iyer et al., 2008). Clinopyroxene in the sequence is mainly composed by diopside and endiopside, with a SiO$_2$ content of 51 – 55 wt.% and 1 – 5.5 wt.% of Al$_2$O$_3$, and the composition of olivine vary between Fo$_{84}$ and Fo$_{93}$ (Furnes et al., 1991; Pedersen et al., 1993).

2.5.2.3 The overlaying sediments

Metasediments of the Skei Group lies unconformably on top of the layered sequence in the most eastern part of the LOC (Furnes et al., 1988). The sediments are composed of a lower sequence of alluvial fan and braided stream deposits and an upper succession of sandstones, mica-schists, conglomerates and limestones (Sturt et al., 1985; Furnes et al., 1988). With Sr and C analysis of marble and detrital zircon data the time of deposition are found to be of Ordovician (Barnes et al., 2007; McArthur et al., 2014). The sediments are all metamorphosed to lower greenschist facies, and elongated pebbles in the polymict conglomerate describes folding and strong deformation of the Skei Group (Sturt et al., 1985).

2.5.3 Alteration of the rocks in the LOC

2.5.3.1 Serpentinization

The ultramafic rocks present in the LOC originated as a part of the oceanic lithosphere, with their lithology displaying the effects of the ocean-floor metamorphism (Iyer et al., 2008). One of the effects caused by hydration processes over a wide range of decreasing temperatures is serpentinization, which is a process that changes the petrophysical and geochemical properties of a rock (Iyer et al., 2008). These changes are redistribution of major elements and formation of a less dense phase which leads to a serious increase in bulk volume (Iyer et al., 2008). The petrophysical changes of the ultramafic rocks does not occur at the same time, resulting in deformation of the surrounding rocks (Iyer et al., 2008). This is displayed in the LOC as fracturing of mm to dm thick altered orthopyroxenite dykes in serpentinized dunites (Iyer et al., 2008).

The geochemical changes caused by serpentinization in the LOC are locally redistribution of the major elements Mg, Si and Al within the rocks, and regional mobilization of grain sized Ca,
Na, Fe and Mn (Iyer et al., 2008). The mobilization of Ca is caused by alteration of primary clinopyroxenes to serpentine and clinochlore, which is a Ca-releasing reaction that enriches fluids with Ca (Austrheim and Prestvik, 2008; Iyer et al., 2008). The high concentration of Ca in the fluids results in a replacement of primary clinopyroxene, orthopyroxene and olivine by secondary diopside and formation of rodingites (Iyer et al., 2008). Rodingitization being a metasomatic process that alter ultramafic-mafic rocks with Ca$^{2+}$ and OH$^{-}$, are seen in the crustal layer coexisting with the serpentine (Austrheim and Prestvik, 2008; Iyer et al., 2008). Fe and Mn are mobilized simultaneously by dissolution and subsequent precipitation of various minerals during serpentinization (Iyer et al., 2008). This especially affects olivine, resulting in a diverse concentration of the two (Iyer et al., 2008).

2.5.3.2 Carbonation

Carbonation of the rocks in the LOC is a process highly associated with the previously described serpentinization (Iyer et al., 2008; Bjerga, 2014). Bjerga (2014) describes the carbonation of the rocks in the LOC as an alteration where carbonates (magnesite and/or dolomite) are formed in completely serpentinized peridotites. The carbonation, as with the serpentinization, is a highly fluid-dependent reaction, hence the association of carbonation with high-permeability zones (Bjerga, 2014). CO$_2$ added to the mineral assemblage by CO$_2$ enriched fluids under specific conditions is what resulted in the carbonation of the rocks in the LOC studied by Bjerga (2014).
Chapter 3: Field descriptions

3.1 Overview of area studied

The area studied (500m x 400m), is located at the transition zone between the layered crustal sequence and the metagabbro series (Fig. 2.2), is situated in the South-Western parts of the Steinstind block in the Leka Ophiolite Complex (Fig. 3.1) The morphology of the area is dominated by hills with well exposed rocks and vegetated lowland. The exposed rocks are of compositions associated with the ophiolitic components and are altered to variable extent (Fig. 2.2). Rocks associated with the layered crustal sequence are observed as less altered peridotite (pink), serpentinized and carbonatized peridotite (orange) and pyroxenite (purple). To what extent the peridotites have been altered are recognized in the color of the weathering skin, where the less altered peridotites have a darker weathering skin than the more altered ones (Fig. 3.2 A
and B). Pyroxenite is located at the top of the dome-like heap, in the sequence and in contact with the other listwanite (blue) assemblage north in Fig. 3.1.

Listwanite (blue) is found comprising the dome-like heap, a part of the sequence and an outcrop to the north (Fig. 3.5). The listwanites (blue) are commonly observed situated in heavily deformed zones which follows the NE-SW trending faults and shear zones described by Furnes et al. (1988) and Titus et al. (2002).

3.2 Peridotite

The majority of the rocks comprising the study area are of a peridotitic composition and are altered to variable extent (Fig. 3.1). Rocks associated with the variable extent of alteration are displayed in Fig. 3.2, where carbonatized peridotites are associated with the red-brown colored weathering skin from oxidation displayed in Fig. 3.2 A and D. The assemblages comprised of carbonatized peridotite are observed throughout the locality as up to 4 meters thick veins with lengths of up to tens of meters. The texture of the assemblages is massive with large fractures caused by brittle deformation, and they are commonly observed in sharp contact with surrounding rocks which are usually composed of a less altered peridotite (Fig. 3.2 A). The thick parallel bands of chromite displayed in Fig 3.2 D are an observation associated with the carbonatized peridotite.

The large horizontal vein displayed in Fig. 3.2 B is composed of altered peridotite (talc ± magnesite) and it is observed intruding a less altered peridotite. Serpentinite is formed along the main horizontal fracture situated in the center of the vein. The shorter perpendicular fractures are only observed in the vein and not in the surrounding less altered peridotite.

The observed soapstone (Fig. 3.2 C) consists of fine foliation of parallel micro rhythmic layers of talc and magnesite. The composition of the rhythmic layers indicate that these rocks are both serpentinized and carbonatized and may represent a stage prior to the listwanitination (Chapter 6).
Figure 3.2: A) Thick vein of carbonatized peridotite intruding a less altered peridotite. B) Horizontal vein of altered peridotite (magnesite + talc) with serpentine at the center intruding a less altered peridotite. The fractures are only observed in the vein and not in the surrounding less altered peridotite. C) Soapstone with parallel micro rhythmic layers of talc and magnesite. D) Thick chromite bands in the carbonatized peridotite.
3.3 Serpentinite

Fig. 3.3 shows formation of serpentinite in different settings. Fig. 3.3 B displays formation of thin serpentinite-veins along fractures in the partly altered peridotite (talc ± magnesite), while Fig. 3.3 C displays thin shaly green serpentinite formed on a less altered peridotite.

The north-eastern part of the study area is comprised of the stratigraphy seen in Fig. 3.3 A and D. A cross-section of the stratigraphy show that the overlaying altered peridotite are comprised of the same fine foliated parallel micro rhythmic layers of talc and magnesite associated with the soapstone (Fig. 3.2 C). The underlying layer is approximately 30-40cm thick and comprised of serpentinite, while the base layer in the stratigraphy is comprised of a less altered peridotite (Fig. 3.3 A).

Figure 3.3: A) Cross-section of the stratigraphy comprising the north-eastern part of the studied area. Layers from top to bottom consists of: altered peridotite, serpentinite and a less altered peridotite. B) Veins composed of serpentinite in an altered peridotite. C) Thin shaly green serpentinite formed on a less altered peridotite. D) A higher resolution image of the altered peridotite displayed in A).
3.4 Structural observations

In addition to the 5-10 cm thick chromite bands displayed in Fig. 3.2 A and Fig. 3.4 B, various veins of serpentine are observed. The dark veins of serpentine seen in Fig. 3.4 A, E and F varies in thickness, orientation and abundance, where the ones in Fig. 3.4 A have a random orientation that follows the fractures in the altered peridotite (talc ± magnesite) with a dynamic thickness of 1 – 3 cm. The serpentine veins in Fig. 3.4 E are >1cm and in high abundance, they have random orientation but seem to follow an internal deformation of the host rock. In addition, two horizontal white talc veins of approximately 1cm are observed. In Fig. 4 F, the serpentine veins appear parallel with a thickness of 1 – 2 cm and the higher elevation compared to the host rock can indicate volume increase during formation. The veins in the carbonatized peridotite, displayed in Fig. 3.4 D, have a random orientation with a thickness of >1 cm and are comprised of quartz. The lenses of up to 1 m in diameter seen in Fig. 3.4 C, are by the looks an early stage of soapstone assemblage. It is from the same area as Fig. 3.2 C and Fig. 3.3 A, and the > 1 cm white veins comprised of talc ± carbonate are parallel to each other and the fractures. The low abundant veins are only present in the suggested pre-soapstone stage assemblage and not in the surrounding assemblages. This may indicate that alteration has not reached the stage where the micro rhythmic layering seen in Fig. 3.2 C and Fig. 3.3 D is formed. The same setting is seen in the block in Fig. 3.4 B, parallel fractures with white talc ± carbonate veins and few fractures perpendicular to these. This block has a darker brown weathering skin than the one in Fig. 3.4 C and has a 5 – 10 cm chromite band crosscutting the vertical fractures and veins.

The relative age relation between the structures are only visible in a few settings as crosscuttings. In Fig. 3.4 E two talc ± carbonate veins cut the numerus serpentine veins, while in Fig. 3.4 B the talc ± carbonate veins are cut by a chromite band. Hence, the suggested relative age relation consists of formation of several generations of serpentine veins, followed by formation of talc ± carbonate veins, and later formation of chromite bands.
Figure 3.4: A) Randomly orientated black serpentine veins with a dynamic thickness of 1-3cm cutting altered peridotite (talc ± magnesite). B) Chromite band perpendicularly cutting vertical talc veins. C) Lenses of up to 1m in diameter with vertical talc veins situated in the altered peridotite. D) Oxidized carbonatized peridotite with small veins of quartz. E) Numerous thin, randomly oriented black serpentine veins cut by two talc ± carbonate veins. F) Thick veins of black serpentine.
3.5 *Listwanite*

Figure 3.5: A) The dome like heap, comprised of listwanite and overlaying less altered pyroxene and gabbro. B) Ductile-deformed foliated listwanite. C) Thick vein comprised of quartz extending throughout the assemblage.
The rocks termed “listwanite” and “garben schist” in the remaining part of the field description are elaborated in chapter 5 (Results) and chapter 6 (Discussion).

Listwanite is present in a few settings in the area studied (Fig. 3.1). The largest assemblage is situated in the dome like heap displayed in Fig. 3.5 A. The heavily deformed listwanite comprise the majority of the heap, while less altered pyroxenite is situated at the top (Fig. 3.1). The heavily deformed listwanite are observed as both layered (Fig. 3.5 B) and massive. The massive parts could also be layered, if the layers are oriented in a way so that they are not visible.

The exposed layers display foliation, which have formed under ductile conditions (Fig. 3.5 B). The ductile deformation, presumably prior to the brittle deformation, is not observed in the overlaying assemblages of the less altered pyroxenites. The color of the listwanite assemblage, both green and brown, display both presence of chromite and oxidization. Quartz veins of various size and thickness are a common observed feature, with the largest observed having a thickness of 40 cm which extends throughout the assemblage (Fig. 3.5 C).

The other listwanite assemblage are grey in color (Fig. 3.6 F and Fig. 3.7 A) and even finer layered than the listwanite from the other assemblage (Fig. 3.5 B). The quartz in this section are not seen in the listwanite, but as thin veins in the garben schist comprising the layer in contact (Fig. 3.6 F). The sequence is exposed in one direction starting with listwanite, then a 7 m thick garben schist layer with internal variations (Fig. 3.6 C-E), then a 1 m thick partly washed out talc layer (Fig. 3.6 B) and finally a few meters thick layer of altered peridotite (Fig. 3.6 A). All the layers in the sequence are parallel and seem to have maintained their pre-listwanitation stratigraphy. The internal changes in the garben schist layer are gradual growth in grain size and a more greenish color towards the listwanite.
Figure 3.6: The sequence comprised of; A) A few meters thick layer of altered peridotite. B) ~1m thick, partly washed out talc layer. C)-E) The 7 meters thick garben schists layer with internal variations. F) The fine layered, grey listwanite and quartz veins intruding the garben schist in contact.
The third location where listwanite is observed is displayed in Fig. 3.7 B-D. The listwanite assemblage, located along the road (Fig. 3.1), is bluish in color and locally it exhibits parasitic folds (Fig. 3.7 D). As in the other listwanite assemblages (Fig. 3.5 and Fig. 3.6), veins of quartz are observed (Fig. 3.7 C).

Figure 3.7: A) The fine foliated, grey listwanite situated in the sequence. B) The outcrop along the road where listwanite are observed. C) Deformed blue/greenish listwanite and veins of quartz. D) Micro-scale parasitic folds in listwanite.
Chapter 4: Methods

4.1 Fieldwork and Sample Collection

The aim of the fieldwork done during two separate field trips to Leka, was to collect rock samples and establish field relationships. A total of 46 rock samples were collected, where 24 of these were collected during a fieldtrip in September 2016 and the rest in September 2017. The naming of the rock samples is based on location and year of collecting, giving the samples collected in 2016 the name LEL x-16 and the ones collected in 2017, LEL x-17.

A selection of the rock samples was further analyzed with analytical instruments in the laboratories at the University of Oslo.

4.2 Petrographic Thin Sections

To obtain thin sections of the selected rock samples, a few steps of preparation were followed. Firstly, the desired area of the rock sample to be made into a thin section had to be located. When this was located, the samples were cut with a diamond saw to obtain the correct size for further preparations. In the final steps of the making, equipment as; Thorlag grinding and polishing automat, Logitech polishing machine, diamond microsaw, Buehler phoenix 4000 polishing machine and impregnation tools were used. The product was 35 µm thick thin sections of the samples glued on to a piece of glass ready for further studying.

4.3 Optical Microscopy

Optical microscopy was used to observe the geochemical and petrogenic properties of the rock samples. The thin sections were studied in both plane polar light (PPL) and cross polar light (XPL) to identify the various minerals, textures, crystal face and relationship between the grains. PPL images of samples were taken with an Olympus TH4-200 microscope equipped with a LEICA DFC320 camera. Other observations of interest were further analyzed in the Electron Probe Microanalysis and Scanning Electron Microscope.

In the preparations of the thin sections for EMP and SEM, they were coated with carbon to make them conductive which was done with a vacuumcoater.
4.4 *Electron Microprobe (EMP)*

Electron microprobe analyses were performed with a Camera SX100 instrument equipped with five wavelength dispersive spectrometers (WDS) and an SSD energy dispersive X-ray spectroscopy. The aim for the EMP analyses was to attain chemistry of the minerals in the LEL x-17 samples. The EDS was used to do quick analyses of grains of interest to attain quantitative spectra. These spectra, together with the Back-scatter detector (BSE), were used to locate and identify minerals selected for qualitative analyses with the WDS. Minerals analyzed with the EDS and WDS and petrogenic properties in the samples are displayed with BSE-images.

For the analyses, the parameters were set to 15 kV (accelerating voltage) and 15 nA (beam current) for all the minerals except carbonates, where the beam current was set to 10 nA. For carbonates a beam focus of 10 µm was used while 0.1 µm was used for the remaining minerals. Peak count used for all the analyses was 10s. The applied calibration standards and X-ray lines used were; wollastonite (Si Kα, Ca Kα), pyrophanite (Ti Kα, Mn Kα), orthoclase (K Kα), albite (Na Kα), Fe metal (Fe Kα), fluorite (F Kα) and the synthetics; MgO (Mg Kα), Al₂O₃ (Al Kα) and Cr₂O₃ (Cr Kα). Correction of matrix was done according to the Cameca PAP procedure (Pouchon and Pichoir, 1984).

Calculations of the mineral molecule structures were done with self-made spreadsheets in Microsoft excel for each of the minerals.

4.5 *Scanning Electron Microscope (SEM)*

Energy Dispersive X-ray Spectroscopy (EDS) and BSE imaging of sample LEL 14-17 were conducted with a Hitachi SU5000 Scanning Electron Microscope.

The aim for SEM analyses of polished thin sections was to attain BSE images of aggregates and swell textures, while the EDS was used to obtain quantitative date of the mica present.

Further analyses of carbon coated fragments with a size of approximately 0,5 cm of sample LEL 14-17 was conducted to obtain 3D images of the hydrous aluminum silicate. The results from these analyses would give a good indication on which of the hydrous aluminum silicate that are present in the sample. But since none of the grains were localized such 3D images were unobtainable.
4.6 Geochemical analyses

To prepare the samples for geochemical analyses, they were milled to a desired grain size of 100-500 µm. The grain size was obtained by milling the samples at 40 Hz for 3-5 minutes depending on their hardness and to avoid contamination an agate (SiO₂) mortar was used.

To determine the loss on ignition (LOI) of each sample, a four decimal weight scale from sartorius, a warming closet and the chamber furnace CWF 1300 were used. Weighing of the samples involved four steps, where the first step was to weigh the crucible bowl, then to weigh out approximately 1.5 grams of each sample. The samples were then contained in a warming closet at 105 °C for 30 minutes, later weighed before heated to 950 °C for two hours in the chamber furnace CWF 1300. The final weight of each sample together with previous weighing were used to determine the LOI.

4.7 X-ray Fluorescence (XRF)

X-ray Fluorescence (XRF) was used to determine the whole-rock chemistry of each sample. Analyses of fusion beads were used to determine concentrations of major elements, while the concentrations of trace elements were determined by analyses of pressed pellets.

In the preparation of the samples for major element analyses the substance obtained from LOI measurements were used. For each sample, 0.6 grams of sample were weighed out and mixed with 6 grams of a flux agent consisting of 66.5 % lithium tetraborate (Li₂B₄O₇) and 33.5 % lithium metaborate (LiBO₂). Each sample was then broken down into a homogenous mixture of soluble components by fusion. The fusion was done by running the samples in the ultimate fusion glass bead maker for 20 minutes, which held a temperature of 1200 °C.

To prepare the samples for trace element analyses, 8 grams of untreated sample powder were mixed together with 2 grams of a flux agent consisting of 66.5 % lithium tetraborate (Li₂B₄O₇) and 33.5 % lithium metaborate (LiBO₂). The samples were then made into pressed pellets by applying 20T with the Specac Atlas Series Autotouch Automatic Hydraulic Press 25T.

The finished fusion beads and pressed pellets where then analyzed with a Malvern-Panalytical Axios-maX minerals Wavelength Dispersive Sequential Spectrometer (WDXRF) equipped with a 4kW Rh-tube. The Fusion bead analyses were done with the Omnian standard less setup to quantify the major elements; Al₂O₃, CaO, Cl, Co₃O₄, Cr₂O₃, CuO, Fe₂O₃, K₂O, MgO, MnO, Na₂O, NiO, P₂O₅, SiO₂, SO₃, TiO₂ and V₂O₅. While the quantification of trace elements by
analyses of pressed pellets where done by using ProTrace calibration standards. All the analyses were processed using the manufacturer’s software, PanAlytical SerQ v5.3A (2016).

4.8 X-ray Diffraction (XRD)

Clay X-ray diffraction was used to identify the clay minerals in sample LEL 14-17 and an attempt to classify the hydrous aluminum silicate.

The first step in the sample preparations were to grind down the rock sample to gravel with a size of approximately <1 cm. The gravel was then put into a beaker together with 200 mL of Na$_2$CO$_3$. After rinsing the beaker for 10 minutes in the ultrasonic it was left overnight. The next morning 400 mL of filtered water was added to the beaker, then rinsed in the ultrasonic for 10 minutes before it was left untouched for 5 hours. When the 5 hours had passed, 400 mL of the mixture were subtracted and added to a separate container. The subtracted substance was then filtered through a filter with a mask size of 0.45 µm to remove smaller sized particles. When the filtering was finished the remaining particles had a size of 0.45 – 2.00 µm. The air-dried substance from the filter with these particles were then added to a glass slide, ready for XRD analyses.

The X-ray diffraction analyses were conducted with a Bruker D8 ADVANCE diffractometer equipped with a Lynxeye 1-dimensional position-sensitive detector (PSD) and CuKα radiation ($\lambda = 0.154$ nm) operating at 40mA and 40kV. The randomly oriented air-dried clay sample were analyzed with a step size of 0.01° in the interval 2° - 65° (2θ) at a count time of 0.3 s (20).

The analyses of the sample were done at four different stages; before any treatments, after glycolate treatment and after the two heat treatments of 350 °C and 550 °C. The glycolate treatment was done by leaving the sample in a warming close (60 °C) with a glycolate component at the bottom for an hour.

The two heat treatments were both done by using a Chamber furnace. Both treatments involved a heating of the sample for approximately 1 hour.
4.9 Raman Spectroscopy

The aim for the Raman analyses were to attain Raman spectra of the hydrous aluminum silicate clay present in sample LEL 14-17. This was further used as aid in the classification of the mineral.

The Raman analyses were conducted at 300 K with illumination of the sample with light from a Spectra-Physics Millennia Pro 12sJS ND:YVO$_4$ solid-state laser operating at 200mV at 532.1 nm (Sjåstad et al., 2015). This was obtained through a neutral density optical filter (90 %) and an Olympus 50X objective giving a laser effect of 2.0 mW (Sjåstad et al., 2015). The Raman light, navigated though a 100 µm confocal pinhole and a 532 nm RazorEdge ultra steep long-pass edge filter into a Jobyn-Yvon Horiba T64000 instrument (single spectrograph), was collected in back-scattering mode (Sjåstad et al., 2015).

From the Raman analyses, low and high frequency spectra of the analyzed aluminosilicate clay were attained. The laser power of 200 mW that was used to prevent damage of the sample but high enough to produce good quality spectra. For increased precision each of the samples were ran with 4 loops of 60 seconds.

The results from the RAMAN analyses were calibrated according to the variations between peeks from measured Tylenol and the ASTM standard E 1840 4-acetamidophenol (Tylenol) from “Raman Spectroscopy for Chemical Analysis” (McCreery, 2002).

Two peeks from the standard, respectively at 213.3 cm$^{-1}$ and 1371.5 cm$^{-1}$ were used in the calculations to attain a coefficient to succeed the calibration (Appendix).
4.10 Software

Software used in the master thesis are listed in table 4.1.

The writing of the master thesis was done in Microsoft Word with Mendeley as a reference tool. Tables used were created in Microsoft Excel and images were edited in both Adobe illustrator CS6 and Adobe Photoshop CS6.

Perple_X was used to model conditions during formation of the mineral assemblages, involving pressure, temperature and CO₂ saturation of fluids. In production of the Raman spectra, MATLAB R2018b was used. Not only to produce the Raman spectra but also as a tool in the interpretation by locating peaks and find their representative wavenumber/cm⁻¹. In the interpretation of the structural properties of the studied area QGIS 3.4 and Georient32v9 were used. QGIS 3.4 was used to attain lineaments of the studied area, including amount, direction and length. These lineaments were later put in Georient32v9 to attain a Rosediagram to visualize their properties. To interpret the XRD pattern attained from the XRD analyses, Profex and Diffrac Eva were used.

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<th>Table 4.1: Overview of Software used and their usage</th>
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<td><strong>Software</strong></td>
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Chapter 5: Results

5.1 Orientation of structural elements and sample collection

Figure 5.1: Arial photograph which display where the LEL x-17 samples and orientations were collected from in the area studied. The sequence consists of the eight samples LEL 6-17 – LEL 13-17.

The sequence consists of the eight samples LEL 6-17 – LEL 13-17. The foliation planes measured at a number of locations, plotted in the stereonet (Fig. 5.2), shows that the orientations of the listwanites follows NE-SW trending direction. These NE-SW trending low-angle shear zones are suggested by Titus et al. (2002) to have formed at a later stage/stage of deformation in the metamorphism/deformation events affecting the LOC.
Results

Fig. 5.3 display the direction and length of lineaments observed in the studied area. 196 lineaments were marked in total in GIS, and then plotted in a Rosediagram using georient (Fig. 5.4).

The rose diagram (Fig. 5.4) display that the lineaments primarily follow a NE-SW trend, and the rest except a few exceptions NW-SE. The NE-SW trending lineaments show a significant longer length of up to 196 m, while the NE-SW are no longer than up to 100 m.

Figure 5.2: Stereonet plot of the orientations attained from listwanite assemblages.
Results

Figure 5.4: Lineaments in the area studied, made with GIS.

Figure 5.3: Rosediagram made with georient, displaying length and direction of the lineaments marked in GIS.
5.2 Geochemistry

5.2.1 Major and trace element whole-rock chemistry

The major and selected trace elements attained from the LEL -16 and LEL -17 series with XRF-analyses, are listed in table 5.1. The average compositions for anorthosite and pyroxenite listed in table 5.1 are from compositional tables listed by Austhreim and Prestvik (2008) from their work associated with rodingitization of the rocks in the LOC. The samples are sorted according to protolith. The protoliths identified in the area studied are; anorthosite, gabbro, pyroxenite, and peridotite, which are distinguished by their Al-, Cr-, Ni-, Sr-, V-content and CaO/MgO ratio. Zirconium-content is less than 2 ppm in accordance an ultramafic/mafic origin. Aluminum is assumed to have been immobile during alteration of the rocks analyzed, giving the possibility to conclude the protolith of the heavy altered rocks. Based on average whole rock chemistry values attained by Austreheim and Prestvik (2008) from similar rocks in the LOC, rock samples with an Al₂O₃ < 6wt% are interpreted as pyroxenite or peridotite. These are distinguished from each other based on the Ca- and Mg-content, where the pyroxenites have a high CaO content (~15wt.%) and MgO content (~20wt.%) while the peridotites have a very low CaO content (~0.50wt.%) and very high MgO content (>35wt.%). The classification of peridotite and pyroxene is also supported by petrographically observation in section 5.3. The rock samples with Al₂O₃ contents between 15-25wt.% are interpreted to have a protolith of a gabbroic composition, while the anorthosite has an Al₂O₃ content of > 30wt.% (Table 5.1).

The trend for the protoliths relative to each other regarding the Cr, Ni, Sr and V concentrations are that the rocks with an anorthositic protolith are low in Cr, very low in Ni and V and high in Sr. The rocks with a gabbroic primary mineral assemblage have medium concentrations of Sr, Cr and V with low concentrations of Ni. The rocks with a peridotitic protolith are very low in Sr, low in V and very high in Ni and Cr. The altered pyroxenites are low-medium in Sr and Ni and very high in Cr and V. Based on the trace element concentrations relative to each other, it is in this study suggested that these concentrations may partly reflect modal variation of the primary mineralogy. Where elevated Ni-content are associated with primary olivine and pyroxene, elevated Cr-content with primary pyroxene and chromite. Relative high V-content is associated with primary pyroxene, and elevated Sr-content is associated with primary plagioclase. The minerals associated with the various trace element concentrations partly correlates with Cr and Ni distributions attained by Mercy and O’Hara (1967) from their work.
with trace element concentrations in ultramafic rocks, and trace elements concentrations attained by Austrheim and Prestvik (2008) from their analyses of rocks in the LOC.

The analyzed samples (Table 5.1) display large variations in LOI (2.91 wt.% - 32.92 wt.%), which is a result of variable extent of serpentinization and carbonation. LOI of anorthosite, gabbro and pyroxenites are controlled by the amount of carbonate, while in peridotites LOI are also controlled by the amount of serpentine and brucite (Iyer et al., 2008).
Table 5.1 Major and trace element analyses sorted according to protolith

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<th>LEL 8-16</th>
<th>LEL 11-16</th>
<th>LEL 13-16</th>
<th>LEL 5-17</th>
<th>LEL 13-17</th>
<th>LEL 14-17</th>
<th>LEL 15-16</th>
<th>LEL 16-16</th>
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<th>LEL 7-17</th>
<th>LEL 15-17</th>
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<td>Peridotite</td>
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<td></td>
<td></td>
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ᵃAnalyzed total iron.
ᵇMg/(Mg+Fe₂⁺), Fe₂⁺ as total iron.
### Table 5.1 (continued)

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<td>9.2</td>
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<td>9.6</td>
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<td>0.0</td>
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<td>38.7</td>
<td>37.4</td>
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<td>69.1</td>
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<td>49.3</td>
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<td>1259.3</td>
<td>456.3</td>
<td>1392.3</td>
<td>3066.0</td>
<td>2785.9</td>
<td>2801.0</td>
<td>2989.9</td>
<td>2796.0</td>
<td>3408.1</td>
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<td>201.6</td>
<td>210.2</td>
<td>186.8</td>
<td>141.6</td>
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<td>213.0</td>
<td>204.8</td>
<td>143.3</td>
<td>188.4</td>
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</tr>
</tbody>
</table>

<sup>a</sup> Analyzed total iron.

<sup>b</sup>Mg/(Mg+Fe²⁺), Fe²⁺ as total iron.
5.2.2 Amphibole

Composition of amphiboles in LEL x-17 samples attained from EMP-analyses are listed in table 5.2, and respectively formulae are calculated on 23 oxygens. Amphibole is observed in most of the analyzed samples except the listwanites and some of the peridotites. The composition of the amphiboles ranges between; tschermakite, tremolite, Mg-hornblende and actinolite (Fig. 6.2), with tremolite being associated with peridotites, and Mg-hornblende and actinolite with pyroxenites.

5.2.3 Carbonate

The composition of carbonates in the LEL x-17 samples from EMP analyses are listed in table 5.3. Carbonates include magnesite, dolomite and calcite. The high SiO$_2$ content in some of the carbonates are from intergrowth or grain boundaries with quartz or amphibole.

Magnesite is observed as large partly fractured euhedral grains in samples of a peridotitic protolith and the sample LEL 5-17 of a gabbroic protolith (Fig. 5.5 A and Fig. 5.13 B). The magnesites observed in sample LEL 6-17 have Mg# ranging between 0.89 – 0.87.

Calcite are observed in both the garen schists and the listwanites (section 5.3). In garen schists it forms along fractures/exsolution lamellae in amphibole and subhedral to anhedral grains in different settings throughout the samples (Fig. 5.6 – 5.9). The calcite appears to be contemporaneous with the chlorite and quartz, with the relation between them being observed as a sharp grain boundary or intergrowths of quartz and chlorite in the calcite (Fig. 5.6 – 5.9). In listwanites, calcite is observed partly or completely filling fractures/exsolution lamellae and comprising veins which typically is enclosed by either quartz or chlorite assemblages (Fig. 5.11 C and Fig. 5.12 A). It is also observed as separate subhedral to euhedral grains (Fig. 5.10 – 5.12). The average Mg# value is 0.73 but ranges from 0.59 to 0.97, with the wide range being a result of the low Fe-content and possible “contamination” form other grains. The chemical composition of calcite shows no evidence of being location nor texture dependent, but difference in brightness (Fig. 5.10 B) is recognized. Dolomite is in this study mainly associated with listwanites, where they are observed as subhedral to euhedral grains with a size of 20-50 μm (Fig. 5.10 – 5.12). The Mg# value of the dolomite ranges from 0.79 to 0.86 with an average of 0.84, composed slightly towards the Ca-endmember. The Ca content are 10-20 % higher than the Mg content and the Fe content ranges from 0.12 to 0.21 a.p.f.u.
### Table 5.2 Composition of Amphiboles in LEL x-17 samples sorted according to protolith

| Sample | LEL | LEL | LEL | LEL | LEL | LEL | LEL | LEL | LEL | LEL | LEL | LEL | LEL | LEL | LEL | LEL | LEL | LEL | LEL |
|--------|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|
| Protolith | Pyroxenite | Peri |
| Mineral | Act | Act | Act | Mg-H | Act | Act | Act | Tsc | Act | Mg-H | Act | Act | Trm |
| SiO₂ | 56.42 | 56.02 | 56.17 | 53.25 | 57.49 | 57.68 | 56.20 | 57.94 | 56.93 | 51.73 | 57.41 | 44.79 | 57.21 | 47.08 | 57.57 | 54.17 | 55.96 |
| TiO₂ | 0.03 | 0.01 | n.a. | 0.10 | 0.05 | 0.02 | 0.03 | 0.01 | 0.04 | 0.03 | 0.03 | 0.36 | 0.00 | 0.18 | 0.00 | 0.15 | 0.00 |
| Al₂O₃ | 0.25 | 0.20 | 0.20 | 6.28 | 0.24 | 0.28 | 2.59 | 0.71 | 1.73 | 3.45 | 0.56 | 14.91 | 0.35 | 11.47 | 0.46 | 4.32 | 0.06 |
| FeO (tot) | 6.70 | 6.48 | 6.40 | 5.73 | 5.47 | 5.44 | 4.73 | 4.38 | 4.80 | 5.63 | 4.49 | 6.72 | 4.26 | 6.30 | 4.46 | 5.22 | 3.72 |
| MnO | 0.11 | 0.09 | 0.13 | 0.06 | 0.15 | 0.12 | 0.11 | 0.05 | 0.08 | 0.17 | 0.10 | 0.13 | 0.09 | 0.10 | 0.09 | 0.15 |
| Na₂O | 0.07 | 0.03 | 0.03 | 0.61 | 0.03 | 0.05 | 0.21 | 0.10 | 0.14 | 0.26 | 0.04 | 2.01 | 0.05 | 1.47 | 0.06 | 0.51 | 0.02 |
| K₂O | n.a. | n.a. | 0.00 | 0.04 | n.a. | 0.00 | 0.01 | n.a. | 0.00 | 0.02 | n.a. | 0.04 | n.a. | 0.02 | n.a. | 0.01 | 0.01 |
| Cr₂O₃ | 0.24 | 0.19 | 0.22 | 1.18 | 0.22 | 0.18 | 0.30 | 0.56 | 0.36 | 0.72 | 0.26 | 0.84 | 0.24 | 0.41 | 0.24 | 0.10 | 0.06 |
| Total | 97.13 | 96.68 | 96.59 | 99.15 | 98.24 | 98.30 | 98.26 | 99.25 | 98.45 | 94.29 | 97.87 | 97.80 | 97.87 | 96.45 | 98.44 | 97.81 | 96.20 |

Structural formula based on 23(O)

| Ti | 0.003 | 0.001 | 0.000 | 0.010 | 0.006 | 0.002 | 0.003 | 0.001 | 0.004 | 0.003 | 0.003 | 0.042 | 0.000 | 0.020 | 0.000 | 0.016 | 0.000 |
| Al | 0.422 | 0.033 | 0.033 | 1.022 | 0.039 | 0.045 | 0.421 | 0.113 | 0.281 | 0.589 | 0.090 | 2.503 | 0.057 | 1.939 | 0.074 | 0.708 | 0.010 |
| Fe²⁺ (tot) | 0.788 | 0.766 | 0.756 | 0.661 | 0.632 | 0.627 | 0.545 | 0.499 | 0.551 | 0.681 | 0.518 | 0.800 | 0.492 | 0.756 | 0.512 | 0.607 | 0.438 |
| Mn | 0.013 | 0.010 | 0.015 | 0.007 | 0.018 | 0.014 | 0.012 | 0.005 | 0.010 | 0.021 | 0.012 | 0.018 | 0.011 | 0.011 | 0.011 | 0.011 | 0.017 |
| Ca | 1.933 | 1.967 | 1.966 | 1.947 | 1.994 | 1.974 | 2.001 | 1.984 | 1.970 | 1.779 | 1.983 | 1.937 | 2.043 | 1.992 | 2.012 | 1.986 | 2.017 |
| Na | 0.018 | 0.008 | 0.009 | 0.163 | 0.008 | 0.012 | 0.057 | 0.025 | 0.038 | 0.072 | 0.012 | 0.555 | 0.012 | 0.409 | 0.015 | 0.137 | 0.005 |
| K | 0.000 | 0.000 | 0.001 | 0.008 | 0.002 | 0.002 | 0.001 | 0.000 | 0.000 | 0.003 | 0.000 | 0.007 | 0.000 | 0.003 | 0.001 | 0.000 | 0.000 |
| Cr | 0.027 | 0.021 | 0.025 | 0.129 | 0.024 | 0.020 | 0.033 | 0.060 | 0.039 | 0.083 | 0.029 | 0.095 | 0.027 | 0.047 | 0.026 | 0.011 | 0.007 |
| Mg# | 0.84 | 0.85 | 0.85 | 0.85 | 0.87 | 0.87 | 0.89 | 0.90 | 0.89 | 0.87 | 0.90 | 0.80 | 0.90 | 0.82 | 0.90 | 0.87 | 0.92 |

\( \text{Mg#} = \frac{\text{Mg}^{2+}}{(\text{Mg}^{2+} + \text{Fe}^{2+})} \); Peri = peridotite, Tsc = tschermakite, Mg-H = magnesio-hornblende, Trm = tremolite, Act = actinolite

Results
Table 5.3 Composition of Carbonates in LEL x-17 samples sorted according to protolith

<table>
<thead>
<tr>
<th>Sample</th>
<th>LEL 6-17</th>
<th>LEL 9-17</th>
<th>LEL 11-17</th>
<th>LEL 12-17</th>
<th>LEL 14-17</th>
<th>LEL 17-19</th>
<th>LEL 19-21</th>
<th>LEL 14-17</th>
<th>LEL 14-17</th>
<th>LEL 14-17</th>
<th>LEL 14-17</th>
<th>LEL 14-17</th>
<th>LEL 14-17</th>
</tr>
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<tbody>
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<td>Protolith</td>
<td>Peridotite</td>
<td>Pyroxenite</td>
<td>Gabbro</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mineral</td>
<td>Mgs</td>
<td>Mgs</td>
<td>Cc</td>
<td>Cc</td>
<td>Cc</td>
<td>Cc</td>
<td>Dol</td>
<td>Dol</td>
<td>Dol</td>
<td>Dol</td>
<td>Dol</td>
<td>Cc</td>
<td>Cc</td>
</tr>
<tr>
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<td>0.08</td>
<td>0.11</td>
<td>0.03</td>
<td>1.78</td>
<td>0.00</td>
<td>0.19</td>
<td>5.88</td>
<td>0.08</td>
<td>0.03</td>
<td>0.25</td>
<td>0.02</td>
<td>0.06</td>
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<td>0.01</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
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<td>0.00</td>
<td>0.00</td>
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</tr>
<tr>
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<td>n.a.</td>
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<td>0.00</td>
<td>1.01</td>
<td>0.02</td>
<td>0.00</td>
<td>0.00</td>
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<td>0.44</td>
<td>0.38</td>
<td>0.36</td>
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<td>0.26</td>
<td>0.15</td>
<td>0.14</td>
<td>0.17</td>
<td>0.17</td>
<td>0.09</td>
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<td>MgO</td>
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<td>0.68</td>
<td>6.85</td>
<td>0.53</td>
<td>0.31</td>
<td>16.98</td>
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<td>56.21</td>
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<td>52.72</td>
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<td>0.00</td>
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<td>0.00</td>
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<td>53.24</td>
<td>53.20</td>
<td>58.96</td>
<td>57.59</td>
<td>63.56</td>
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<td>53.02</td>
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Structural formula based on 6(O)

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</tr>
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<tr>
<td>Al</td>
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<td>0.000</td>
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<tr>
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</tr>
<tr>
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<td>0.004</td>
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<tr>
<td>Mg</td>
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<td>0.005</td>
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<td>0.974</td>
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<td>0.001</td>
<td>0.000</td>
<td>0.002</td>
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<td>0.000</td>
<td>0.000</td>
<td>0.000</td>
<td>0.000</td>
</tr>
<tr>
<td>Cr</td>
<td>0.000</td>
<td>0.000</td>
<td>0.000</td>
<td>0.000</td>
<td>0.001</td>
</tr>
<tr>
<td>Mg#</td>
<td>0.89</td>
<td>0.87</td>
<td>0.59</td>
<td>0.79</td>
<td>0.97</td>
</tr>
</tbody>
</table>

Mg# = Mg²⁺/(Mg²⁺+Fe²⁺); Mgs = magnesite, Cc = calcite, Dol = dolomite
5.2.4 Chlorite

Compositions of chlorite in the LEL – 17 samples are listed in table 5.4. Based on the concentrations of Si, Al, Mg and Fe, the composition of the chlorites ranges from sheridanite to penninite, with most of them classifying as sheridanite. The chlorite in the samples contains a significant amount of Fe\(^+\), but since the Mg\(^+\) content to a high degree surpass the Fe\(^+\) content in all the samples they are considered as low Fe-chlorites.

The composition of the chlorite in samples with a gabbroic protolith compared to the ones of a pyroxenitic protolith are that they have higher Fe- and Al-contents and lower Mg- and Si-contents. Cr-content in all of the chlorites varies between 0.07 wt.% and 1.89 wt.%.

5.2.5 Other minerals

Tourmaline is observed comprising large aggregates (Fig. 5.18) in the samples LEL 5-17, 13-17 and 14-17. Chemical composition was only attained from the two latter samples (Table 5.5), however, these exhibit that the tourmalines is of an approximately similar composition.

The chemical composition of serpentines displayed in table 5.5, were attained from the samples LEL 6-17 and LEL 15-17. The two compositions of LEL 6-17 serpentines are from both the bladed serpentine\(^1\)st (Fig. 5.5 D) and the serpentine partly composing the heterogenous mixture with talc\(^2\)nd (Fig. 5.5 A). The serpentine analyzed in sample LEL 15-17 was the “darker” one composing the homogenous mixture (Fig. 5.9 D).

The plagioclase analyzed in sample LEL 13-17 (Table 5.5) classifies as andesine and is only present as a minor phase compared to the other minerals (Fig. 5.10).

Chemical analysis of the epidote in sample LEL 11-17 (Fig. 5.7 C) display that it has a composition associated with zoisite.

Mica analyzed from the samples LEL 20-17 and LEL 14-16 are rich in potassium and chromium, with the micas from LEL 20-17 being classified as Cr-muscovite while the ones from sample LEL 14-16 classifies as fuchsite.
Results
Table 5.4 Composition of Chlorite in LEL 17 samples sorted according to protolith

<table>
<thead>
<tr>
<th>Sample</th>
<th>8-17</th>
<th>9-17</th>
<th>11-17</th>
<th>12-17</th>
<th>20-17</th>
<th>13-17</th>
<th>14-17</th>
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<th>14-17</th>
<th>14-17</th>
<th>14-17</th>
<th>14-17</th>
<th>14-17</th>
</tr>
</thead>
<tbody>
<tr>
<td>Protolith</td>
<td>Pyroxenite</td>
<td>Gabbro</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mineral</td>
<td>Clh</td>
<td>Clh</td>
<td>Pnt</td>
<td>Srd</td>
<td>Srd</td>
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<td>Srd</td>
<td>Srd</td>
<td>Srd</td>
<td>Srd</td>
<td>Srd</td>
<td>Srd</td>
</tr>
<tr>
<td>SiO₂</td>
<td>29.42</td>
<td>27.68</td>
<td>35.03</td>
<td>27.81</td>
<td>26.10</td>
<td>26.36</td>
<td>27.26</td>
<td>27.08</td>
<td>27.07</td>
<td>26.98</td>
<td>27.07</td>
<td>26.50</td>
<td>27.16</td>
</tr>
<tr>
<td>TiO₂</td>
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<td>0.02</td>
<td>0.01</td>
<td>0.00</td>
<td>0.03</td>
<td>0.05</td>
<td>0.05</td>
<td>0.03</td>
<td>0.03</td>
<td>0.05</td>
<td>0.04</td>
<td>0.03</td>
<td>0.04</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>17.35</td>
<td>19.03</td>
<td>16.94</td>
<td>20.45</td>
<td>23.27</td>
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Structural formula based on 28(O)

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|Mg# = Mg²⁺/(Mg²⁺+Fe²⁺); Clh = chlinochlore, Srd = sheridanite, Pnt = penninite
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Mg# = Mg²⁺/(Mg²⁺+Fe²⁺); Gbr = gabbro, Per = peridotite, Pxy = pyroxenite
5.3 Petrography

Fig. 5.5: BSE images of the carbonated and serpentinized peridotite LEL 6-17. A) A common observation in the LEL 6-17 thin section, comprised of large fractured grains of magnesite (mgs), large unfractured grains of ferritchromite (ftc), heterogenous mixture of talc (tlc) and serpentine (srp) and replacing serpentinization. B) Replacement of ferritchromite with serpentine. C) The two ferritchromite grains from A), where Cr-spinel is situated in one of the cores. D) Bladed serpentine replacing magnesite.

Fig. 5.5 A display a setting where large fractured grains of magnesite (mgs) is situated. The fractures are often observed filled with either serpentine (srp) and/or the heterogenous mixture of talc and serpentine (tlc) which comprise a large phase in the thin section. The two grains of ferritchromite (ftc) are unfractured and Cr-spinel is observed in one of the cores (Fig. 5.5 C), implying a replacement of Cr-spinel with ferritchromite. Replacement of ferritchromite and magnesite (Fig. 5.5 B and D) with serpentine (srp), most often bladed (Fig. 5.5 D), are commonly observed.
Fig. 5.6: BSE images of the pyroxenite-garben schists LEL 8-17 (A) and LEL 9-17 (B, C and D) situated in the sequence. A) Chlorite (chl) and ilmenite (ilm) growth in parallel fractures/exsolution lamellae in the amphibole (amp). B) Growth of chlorite (chl), calcite (cc) and titanite (ttn) in parallel fractures/exsolution lamellae in amphibole as in A). C) and D) display the extensive fracturing of/exsolution lamellae in the large euhedral amphibole grains with the associated growth of chlorite, calcite, titanite and quartz.

Fig. 5.6 A display the chlorite (chl) growth in the parallel fractures/exsolution lamellae of the large euhedral amphiboles (amp) grains in sample LEL 8-17, with an elongated ilmenite (ilm) grain following the same fractures/exsolution lamellae. Fig. 5.6 B, C and D display the same fractures/exsolution lamellae associated with the mineral growth as observed in Fig. 5.6 A of LEL 8-17. However, high amounts of calcite (cc) are found in addition to the chlorite along the fractures/exsolution lamellae. The circular titanite (ttn) grain follows the fractures/exsolution lamellae in the amphibole and quartz are anhedral dark accumulation situated on the amphibole grain.
Fig. 5.7: BSE images of the pyroxenite- garnet schists LEL 10-17 (A and B) and LEL 11-17 (C and D) in the sequence. A) Display the same parallel fractures/exsolution lamellae in amphibole as observed in Fig. 5.6. B) Higher resolution BSE image of the mineral assemblage in of the veins from A), comprised of anhedral grains of quartz and fragments of the surrounding amphibole in a matrix composed of chlorite. C) Epidote (ep) and calcite (cc) growth in the fractured/exsolution lamellae amphibole, and growth of anhedral grains of quartz (qz) on the amphibole. D) Growth of chlorite (chl), calcite (cc) and quartz (qz) associated with the fractured amphibole.

Fig. 5.7 A and B display mineral growth in the continuous parallel fractured/exsolution lamellae amphibole displayed in Fig. 5.6. The two thicker veins compared to the ones displayed in Fig. 5.6, are comprised of a matrix composed of chlorite, relics of the surrounding amphibole (amp) and anhedral grains of quartz (qz). Fig. 5.7 C and D display parallel veins composed of various minerals following the fractures/exsolution lamellae in the amphibole (amp). Epidote (ep) is only observed in the one setting displayed in Fig. 5.7 C, where the epidote is observed as a fractured grain in sharp contact with anhedral grain of calcite (cc). The replacement of amphibole (amp) associated with anhedral quartz (qz) grains, chlorite (chl) and calcite growth displayed in Fig. 5.6, is a continuous replacement also observed in these two thin sections (Fig. 5.7). The coexistence of chlorite (chl) and calcite (cc) in the same vein with a sharp contact appears at first in the LEL 11-17 thin section (Fig. 5.7 D).
Results

Fig. 5.8: BSE images of the pyroxene-garnet schist sample LEL 11-17. A) Kinking of the fractures filled with either calcite (cc) and chlorite (chl) in the amphibole (amp). B) An assemblage purely comprised of calcite (cc) following the amphibole's (amp) cleavage. C) Kinking perpendicular to the fractures in the amphibole (amp), and growth of the minerals associated with the garnet schists samples; quartz (qz), chlorite (chl) and calcite (cc). D) Kinking of the parallel fractures in the amphibole.

Fig. 5.8 displays the parallel fractures/exsolution lamellae in the amphibole (amp), also observed in the other garnet schists situated in the sequence (Fig. 5.6 and Fig. 5.7), are in this thin section observed kinked. Fig. 5.8 A display perpendicular kinking of the parallel fractures/exsolution lamellae, which indicates strain from a different angle. The strain, which caused the kinking, may either be from external stress, or internal stress from the mineral growth in the fractures/exsolution lamellae. The kinking displayed in Fig. 5.8 A varies from those displayed in Fig. 5.8 C and D in the way they are kinked. The displayed veins, composed of calcite (cc) and chlorite (chl), are a lot thicker in Fig. 5.8 C and D than in Fig. 5.8 A and B. Which may imply that kinking may have been caused by a larger growth of minerals in the previous fractures/exsolution lamellae. The amount of anhedral quartz grains on the amphibole correlates with the thickness of the calcite (cc) and chlorite (chl) veins, by a higher amount of quartz in areas where the veins are thicker (Fig. 5.8). As the large grains probably started as
cpx, the overall reaction leads to formation of quartz and calcite which is a part of listwanite formation.

Figure 5.9: BSE images of the pyroxenite-garben schist LEL 12-17 (A, B and C) and the carbonated peridotite LEL 15-17 (D). A, B and C) display parts of the thin section which are comprised of the same minerals as observed in the other garben schists (Fig. 5.6-5.8); titanite (ttn), quartz (qz), amphibole (amp), chlorite (chl) and calcite (cc). The calcite (cc) and chlorite (chl) in these BSE images are more well-crystallized than in the other garben schists. And calcite grains are observed to expand out of the veins where it presumably was formed. In B) the brighter parts of the amphibole (amp) are composed of the Mg-hornblende listed in table 5.2. D) fractured, partly replaced oxide with two types of serpentines (srp). The brighter serpentine is also observed crosscutting the preceding darker serpentine (srp). The amphibole (amp), tremolite (Table 5.2), is commonly observed situated in the homogenous matrix composed of the darker serpentine.
Fig. 5.9 displays the veins composed of calcite and chlorite observed situated in the fractured/exsolution lamellae amphibole is less dominant in LEL 12-17 (Fig. 5.9 A, B and C) compared to the other garben schists (Fig. 5.6-5.9), and the calcite (cc) and chlorite (chl) in LEL 12-17 are no longer only found in the fractures/exsolution lamellae in the amphibole (amp) (Fig. 5.9 C). The calcite (cc) and chlorite (chl) are in this thin section crystallized to either subhedral or euhedral grains, where the size of the calcite grains are quite large compared to the grain size of the chlorite (chl) (Fig. 5.9 B). Chlorite (chl) and quartz are in this thin section also observed to compose a heterogenous mixture (Fig. 5.9 C) in addition to how they usually appear (Fig. 5.6-5.9). Fig. 5.9 B display relics/enclosed anhedral grains of Mg-hornblende (Table 5.2) which have a brighter color compared to the surrounding dominant actinolite (Table 5.2). Anhedral grains of titanite (ttn) are as in the earlier garben schists highly abundant. Fig. 5.9 D display an area in the sample LEL 15-17 which is comprised of amphibole (amp), oxide and two generations of serpentine (srp). The oxide is observed being replace by both types of the serpentine, and the brighter bladed serpentine is observed cutting the preceding darker serpentine. The amphibole (amp) is usually observed in the

Fig. 5.10 A display one of many tourmaline (tu) aggregates situated in veins of chlorite (chl) observed in the thin section, tourmaline grains in this specific aggregate contain enclosed grains of quartz. Small rutile (ru) grains displayed, are observed sporadically spread throughout the thin section. Fig. 5.10 B display a common mineral assemblage observed in the thin section, which is comprised of varying sized grains typically composed of chlorite (chl), calcite (cc) and quartz (qz). Quartz and chlorite are observed, in addition to independent grains, as enclosed grains in the calcite. The variation in brightness observed in the calcite are presumable caused by different chemical compositions. Fig. 5.10 C display a part of the thin section where andesine is observed. Andesine is present as a minor phase compared to the other minerals; chlorite, quartz and calcite, and is only observed as small grains in the large aggregates. Fig. 5.10 D of sample LEL 14-17 display a similar mineral assemblage to the ones observed in LEL 13-17, however dolomite (dol) is more abundant. thin section as lamellae as displayed in Fig. 5.14 A.
Figure 5.10: BSE images of the gabbro-listwanites LEL 13-17 (A, B and C) and LEL 14-17 (D). A) Tourmaline (tu) commonly observed situated in veins composed of chlorite (chl). The tourmaline displayed in A), contains enclosed grains of quartz (qz). Small grains of rutile (ru) are often observed in the thin section. B and C) display typical mineral assemblages in the thin section, where chlorite (chl), calcite (calcite) and quartz comprise the major phase. In B) two types of calcite are observed, presumably with compositional variations, with enclosed grains composed of chlorite and quartz. C) display in what assemblage the andesine (plg) is observed. The assemblage is mainly composed of varies seized grains composed of calcite (cc), quartz, chlorite (chl) and andesine. D) display an assemblage observed in LEL 14-17 comprised of minerals composed of chlorite (chl), quartz (qz), calcite (cc) and dolomite (dol).
Fig. 5.11: BSE images of the gabbro-listwanite, sample LEL 14-17. A) Mineral assemblage comprised of dolomite (dol), quartz (qz), calcite (cc) and kaolinite. B) display same mineral assemblage as in A), but with chlorite (chl) in addition. The quartz (qz) is observed as separate grains, but also found to comprise the dark aggregate. C) One of multiple fractures with variable extent of calcite (cc) growth observed in the thin section. D) Relation between oxide and chalcopyrite observed as numerous small grains throughout the thin section.

Fig. 5.11 displays the LEL 14-17 listwanite that is composed of similar minerals as the LEL 13-17 gabbro-listwanite (Fig. 5.10), which include; euhedral grains of dolomite (dol), quartz (qz), calcite (cc) and chlorite (chl) (Fig 5.11 A and B). The common observation of calcite growth in large fractures (Fig. 5.11) is not observed in the LEL 13-17 listwanite (Fig. 5.10). And the kaolinite displayed in Fig. 5.11 A and B are only observed in this thin section. Kaolinite is observed sporadically spread throughout the thin section and commonly in sharp contact with minerals of various compositions. Fig. 5.11 D display the alteration of chalcopyrite to iron oxide.
Fig. 5.12: BSE images of the gabbro-listwanite LEL 14-17 (A) and pyroxenite derived listwanite LEL 20-17 (B, C & D). A) display where kaolinite is situated in a different part of the LEL 14-17 thin section. The part of the thin section is comprised of calcite (cc), quartz (qz), chlorite (chl) and kaolinite. B and C) display common mineral assemblages in the LEL 20-17 thin section. The mineral assemblages are comprised of dolomite (dol), quartz (qz), calcite (cc), chlorite (chl) and ±Cr-mica. Cr-mica comprise a minor phase in the thin section. D) Display sulfide (core) altered to oxide. This is a typical oxidation of sulfide that occur in the supergene zone.

Fig. 12 A displays a mineral assemblage comprised of sporadically spread euhedral kaolinite grains, elongated subhedral chlorite (chl) grains, calcite (cc) growth in the fracture and anhedral to subhedral grains which is in some settings are replaced by quartz (qz). In addition to the subhedral grains of quartz observed in sharp contact with surrounding minerals, the quartz comprises the darker aggregate following the fracture. Fig. 5.11 B display a mineral assemblage comprised of the same minerals (except kaolinite) as observed in LEL 14-17, with subhedral to euhedral grains composed of dolomite (dol), quartz (qz), chlorite (chl) and calcite (cc). Fig. 5.12 C display a mineral assemblage where Cr-muscovite is observed in the thin section. The Cr-muscovite is present as subhedral to euhedral grains and mainly observed in contact with quartz (qz). Fig. 5.12 D display the partial replacement of chalcopyrite with oxide.
Fig. 5.13: PPL-images of the altered peridotite LEL 7-17 (A) and listwanitized gabbro LEL 5-17 (B). A) display the formation of secondary amphibole (amp) in the talc-rich altered peridotite. The secondary formed amphibole is highly abundant and acts as long elongated euhedral grains. The amphibole grains are observed cutting previous textures, commonly composed of talc (tlc) which comprise the major phase in the thin section. B) Large magnesite (mgs) grain situated in an assemblage composed of chlorite (chl). Magnesite is the only carbonate observed in the thin section.

Fig. 5.13 A display the talc (tlc) and amphibole (amp) which comprise the majority of the thin section. The amphibole, observed cutting the previous formed talc, is highly abundant and acts as long elongated euhedral grains. With the amphibole cutting preceding textures it is identified as formation of a secondary amphibole. Fig. 5.13 B display a large magnesite (mgs) grain situated in an assemblage composed of chlorite (chl). The displayed magnesite grain is one of many which compose the carbonates observed in the thin section.
Fig. 5.14: PPL-images of the altered peridotites LEL 15-17 (A and B) and LEL 6-17 (C and D). A) display the amphibole (amp) lamellae commonly observed in a homogenous matrix composed of serpentine. B) Magnesite (mgs) grain observed enclosing serpentine which also compose the surrounding matrix. C and D) display the formation of magnesite and serpentine. In D), the replacement may exhibit a pseudomorphic replacement texture of the olivine grain replaced.

Fig. 5.14 A display the commonly observed amphibole (amp) in serpentine (srp). In Fig. 5.14 B an anhedral grain of magnesite enclosing serpentine which also compose the surrounding assemblage is displayed. Fig. 5.14 C and D display the formation of serpentine and magnesite, which presumably is a simultaneously formation where olivine is replaced. In Fig. 5.14 D the replacement may exhibit a pseudomorphic replacement texture, where the original grain boundary of the replaced olivine is kept.
The map display that there is high concentration of Al in the kaolinite, medium to high in the aggregate to the top left and medium in the chlorite (chl) (Fig. 5.16 A and B). The only minerals containing Fe are chlorite (chl) and dolomite (dol), with a medium concentration in chlorite and low in dolomite (Fig. 5.16 C). The Fe and Mg maps display a zonation of Fe and Mg in the dolomite, where the parts with higher concentrations of either Fe or Mg are poorer in the other element (Fig. 5.16 C and E). The higher concentration of Fe and Mg tend to be located close to the core of the dolomite grains. Ca is highly concentrated in the calcite (cc) and medium in the dolomite. Mg in the mapping are medium to highly concentrated in the chlorite and medium in the dolomite. Area with the highest Si content is quartz (qz), medium concentration in the aggregate to the top left and low concentrations in the chlorite and kaolinite.
Figure 5.16: Color map of the BSE image displayed in Fig. 5.15, displaying the distribution of major elements in the minerals observed. A and B) aluminum distribution, C) iron distribution, D) calcium distribution, E) magnesium distribution and F) silicon distribution. The color tables to the right in the images display the concentration of the element associated with each color.
Fig. 5.17 displays the observed swell-textures comprised of kaolinite (kao) and chlorite (chl) in sample LEL 14-17.
Fig. 5.18 display the commonly observed large aggregates of tourmaline (tu) situated in chlorite (chl) in sample LEL 14-17. The tourmaline aggregates are comprised of numerous euhedral grains of tourmaline, and the chlorite are of a fibrous texture.
Fig. 5.19 display a setting where paragonite (Pa) is formed. It is observed situated in a thick vein mainly composed of the fibrous textured paragonite. In PPL the veins of paragonite are dark with a mica like glare. In the top right of Fig. 5.19 two grains of kaolinite (ASi) is displayed, the right most grain has two thin parallel veins of presumably chlorite (Chl) and can remind of the swell texture displayed in Fig. 5.17.
5.4 Phase diagram

Figure 5.20: T-XCO₂ diagram of the gabbro-listwanite LEL 14-17 with CO₂ – H₂O saturated fluids, 1kbar pressure, and the elements Na₂O – MgO – Al₂O₃ – SiO₂ – CaO – FeO. The colored stability area (orange) for the mineral assemblage observed in LEL 14-17, is stable between 90 °C and 220 °C at XCO₂ = 0. The top temperature for the stability area decreases significantly together with an increase in saturation of CO₂ in fluids which also leads to a shrinking in size of the stability area.
Fig. 5.20 displays the T-XCO₂ diagram attained from the gabbro-listwanite LEL 14-17.

**CO₂ – H₂O saturated fluid metasomatism**

T-XCO₂ diagram for sample LEL 14-17 where made with CO₂ – H₂O saturated fluids, 1kbar pressure, and the elements Na₂O – MgO – Al₂O₃ – SiO₂ – CaO – FeO. The molar bulk composition used; Na₂O (1.4 wt.%) – MgO (5.56 wt.%) – Al₂O₃ (18.13 wt.%) – SiO₂ (33.37 wt.%) – CaO (11.37 wt.%) – FeO (4.41 wt.) (Table 5.1). Since Perplex mineral data base do not include tourmaline, the estimated amount of tourmaline (10 %) was subtracted from the molar bulk composition. The FeO value used was attained from recalculation of the wt.% Fe₂O₃ in the sample (Table 5.1).

The pressure used was 1kbar, but a diagram with identical settings were also made with 3kbar. These showed very small variations, just a small increase of temperature for the stability areas, resolving in use of only the P = 1kbar diagram.

The low K₂O (0.10 wt.%) and MnO (0.09 wt.%) concentrations where not included in the phase diagram, since they complicated the mineral assemblage with non-present minerals. This together with ankerite and siderite being excluded were the only form of manipulation.

The mineral assemblage in sample LEL 14-17 observed in EMP, SEM and optical microscopy consists of pa – chl – kao – q – cc – dol. The colored stability area is the only one showing a similar mineral assemblage, except the substitution of clinochlore with daphnite. The substitution is explained with Perplex not taking in account that dolomite can contain notable amounts of FeO and assume that the present FeO are contained in the chlorite. The colored stability area (orange, Fig. 5.20) for the mineral assemblage observed in LEL 14-17 is stable between 90 °C and 220 °C at XCO₂ = 0, with the top temperature decreasing signifyingly together with an increase in saturation of CO₂ in fluids. This decrease leads to a smaller stability area for the correct mineral assemblage with higher XCO₂ value.
5.5 RAMAN spectrometry results

Figure 5.21: BSE image and optical image with transmitted light taken with Raman spectrometer. The laser beam was set to 1 μm during the Raman spectrometer analyses.

Figure 5.21 displays the grain of clay analyzed to attain Raman spectra of the hydrous aluminum silicate present in sample LEL 14-17. The grain analyzed, together with chlorite, comprise the swell texture displayed in Fig. 5.17 and Fig. 5.21. There were several attempts in attaining Raman spectra of various hydrous aluminum silicate grains.

Figure 5.22: 1st attained Raman spectra of the lattice regions of the analyzed hydrous aluminum silicate, 100-1200cm⁻¹.
But difficulties with localization and interference of other minerals in attained spectra, resulted in that good quality spectra were only attained from the grain displayed in Fig. 5.21.

Fig. 5.22 and Fig. 5.23 display the attained Raman spectra of the lattice regions of the hydrous aluminum silicate grain analyzed.

The interval 100 – 1200 cm$^{-1}$ show a common spectra for hydrous aluminum silicate, where the bands represents bend, stretch, translation, deformation and symmetry of the units in hydrous aluminum silicate (Frost, 1995; Frost and Shurvell, 1997). Results attained by Frost (1995) and Frost and Shurvell (1997) from their Raman analyses on hydrous aluminum silicate, show variations in placement, splitting and intensity of the bands for the four hydrous aluminum silicates. However, these properties are highly sample dependent (Frost, 1995).

The sample dependence together with use of different instruments and calibrations leads to variations from sample to sample, making classification based on results from this interval very difficult. Leaving it to represent a Raman spectrum for hydrous aluminum silicate in the interval 100 – 1200 cm$^{-1}$ with unobtainable characterizations of the four possibilities.
Fig. 5.24 display a Raman spectrum of the OH stretching region attained from the analyzed hydrous aluminum silicate grain. Maximums/peaks in this interval represents present crystallographically OH groups, where the amount of these together with their frequency and intensity are used to distinguish the four hydrous aluminum silicates apart.

Due to insecurity whether the three peaks at 3662, 3666 and 3670 cm$^{-1}$ are to be recognized as a maximum or not, a minimum of three maximums at 3619, 3652 and 3685 cm$^{-1}$ are taken in account at first. Based on a minimum of three peaks halloysite can be excluded as a possibility based on its commonly two peak OH stretching region Raman spectra (Frost, 1995; Frost and Shurvell, 1997).

Of the three remaining possibilities nacrite and dickite are characterized by three maximums and kaolinite by four (Frost, 1995). Based the insecurity around the peak at 3666 cm$^{-1}$, the frequency of the maximums is taken in account at first in distinguishing the three. The frequency of the three maximums in dickite are usually located at around ~3620, ~3655 and ~3710-3730 cm$^{-1}$, ~3600, ~3635 and ~3695 cm$^{-1}$ for nacrite, ~3620, ~3652, ~3668 and ~3695 cm$^{-1}$ for kaolinite (Bish and Johnston, 1993; Frost, 1995; Johnston et al., 1998, 2008; Johnston, 2006). The frequencies of the maximums exclude both nacrite and dickite as possibilities based on mis fits of their hydroxy bands compared to the results. Nacrite has a mis fit on the two low-frequency hydroxy bands and dickite on the high-frequency.
With kaolinite being the only viable option based on frequencies of the maximums, leads to the assumption that the band at 3666 cm$^{-1}$ are a peak based on its frequency and when taken in account result in four maximums in the OH stretching region.

Since the intensity of the OH bands are highly dependent on the crystal orientation and scattering geometry (Johnston et al., 1998), interpretation based on the intensity is rather difficult. However, numerous different studies (Bish and Johnston, 1993; Frost, 1995; Johnston et al., 1998, 2008; Johnston, 2006) suggests that the high-frequency hydroxy band of kaolinite are the one of highest intensity and longest wavelength in its OH stretching region, while dickite and nacrite has its OH band with equal properties at a low frequency.

It is important to take in account that these results were from analyses of a single grain, leaving up the possibility that the other hydrous aluminum silicate grains can be either halloysite, dickite or nacrite.

When comparing the results with earlier studies done on the subject, a few parameters can raise incurrence about the viability of the interpretation. Use of different instruments, calibrations, grain orientation, temperature and pressure have a significant impact on the results.
5.6 XRD results

Fig. 5.25 and Fig. 5.26 display the X-ray diffraction pattern of clays attained from XRD analyses of sample LEL 14-17. Fig. 5.25 display the XRD pattern between the 2-theta angle values 4.4 – 13.25 and Fig. 5.26 the values between 17.6 – 29.0. Peaks in both the intervals are marked with their representative Angstrom (Å) value.

In the determination of which clay minerals that are represented in the XRD pattern the characteristics of the peaks are utilized. The amount and position of peaks are associated with the Miller index of the representative clay mineral which have a signature response to treatments (Poppe et al., 2001; Jeans, 2002).

By comparing the characteristics of the peaks from the attained results with peaks attained by Poppe et al. (2001) and Jeans (2002), the three clay minerals chlorite, hydrous aluminum silicate and illite-smectite were identified.

Chlorite are represented by the peaks at 001/14.10 Å, 002/7.06 Å, 003/4.71 Å and 004/3.53 Å (Fig. 5.25 and Fig. 5.26). The peaks response to treatments were similar to those described by Poppe et al. (2001) and Jeans (2002), were change of the peaks only followed the 550 °C heat treatment. The change is observed as an increase in intensity for the peak at 14.10 Å and a decrease in the other three (Fig. 5.25 and Fig. 5.26).

Hydrous aluminum silicate was identified by the two peaks at 001/7.13 Å and 002/3.57 Å. As with the chlorite, the peaks only responded to the 550 °C heat treatment. However, instead of slight change in intensity the hydrous aluminum silicate became amorphous (Poppe et al., 2001; Jeans, 2002).

The peaks at 001/9.58 Å, 002/4.80 Å and 003/3.20 Å were interpreted as illite-smectite (Fig. 5.27 and Fig. 5.28). Not only do the position of the peaks fit the description from earlier studies (Poppe et al., 2001; Jeans, 2002), but also the high intensity of the 001 and 003 peaks together with no response to either of the treatments. Yet, a minor shift to lower Angstrom values after the 550 °C heat treatment is observed (Fig. 5.25 and Fig. 5.26).

Since the remaining peaks at 16.5 Å, 14.5 Å, 3.34 Å and 3.18 Å can’t be correlated to any of the crystallography planes in the identified clays, interpretation of these are a lot harder. However, the peaks at 3.34 Å, 14.5 Å and 3.18 Å were addressed to illite-smectite since the XRD pattern of illite-smectite attained by Poppe et al. (2001), shows peaks of approximately same characteristics (Fig. 5.25 and Fig. 5.26). The peak at 16.5 Å which show an expansion
after glycolate treatment is addressed to the XRD pattern of halloysite, this is far from certain but none of the other identified clays seem to show the same characteristic. The peak could also be addressed to minor smectite in the sample, but the peak does not show the shifting and high intensity of the 001 peak which is usual for the smectite.
Figure 5.25: X-ray diffraction pattern, between the 2-theta angle values 4.4 – 13.25, of clays attained from XRD analyses of sample LEL 14-17. Peaks in the interval are marked with their representative Angstrom (Å) value.
Figure 5.26: X-ray diffraction pattern, between the 2-theta angle values 17.6 – 29.0, of clays attained from XRD analyses of sample LEL 14-17. Peaks in the interval are marked with their representative Angstrom (Å) value.
Chapter 6: Discussion

6.1 Introduction

The mineralogical and geochemical composition of the analyzed rocks from Leka demonstrates that the extent of carbonation and serpentinization is high. Earlier works (Prestvik, 1980; Furnes et al., 1988; Iyer et al., 2008; Michels et al., 2018), indicate that the studied area is situated at the transition of the layered crustal sequence and metagabbros in the LOC. The layered sequence consists of gabbro, anorthosite, peridotite and pyroxenite. All these rock types can, based on whole rock chemistry, be identified as protolith to the samples studied here (Table 5.1). By assuming that Al was immobile, rocks with more than 10 wt.% Al$_2$O$_3$ is most likely derived from gabbro. Very high content of Al$_2$O$_3$ in excess of 30 wt.% points to an anorthositic protolith. Rocks with very high Ni and high Cr content is most likely derived from peridotite/dunite. The remaining samples has been interpreted to be derived from pyroxenite. From Fig. 6.1 the possible primary mineralogy of these rocks can be inferred.

A comparison of whole rock composition with less altered rocks from the same stratigraphic level suggest that the main compositional changes are increase in LOI. However, we also notice that some of the samples must have undergone metasomatism especially with respect to CaO. Alteration products contributing to an increased LOI are mainly serpentine and carbonate (Austrheim and Prestvik, 2008; Iyer et al., 2008; Bjerga, 2014). H$_2$O in serpentine is approximately 10 wt.% LOI (Austrheim and Prestvik, 2008; Iyer et al., 2008), meaning that a totally serpentinized rock will have a maximum LOI content of ca. 10 wt.% A higher LOI content will require the presence of carbonate, since brucite has not been detected. The whole-rock chemistry (Table 5.1) of the rocks shows a large range in LOI (2.19 – 32.92 wt.%) which are a result of the variable extent of alteration with a substantial amount of carbonate in the samples with the higher LOI contents. The rocks with a higher LOI are localized in the heavily deformed parts of the studied area, which are seen to be in sharp contact with surrounding less altered rocks (Chapter 3).

The sequence, consisting of the samples LEL 6-17 to LEL 13-17 was at first believed to have maintained their pre-listwanitation stratigraphy and displaying a stage-wise listwanitation (Fig. 3.6). This was later contradicted by their whole-rock chemistry implying that the rocks are alteration products of different protoliths (Table 5.1).
6.2 Primary composition of the analyzed samples

The variation in chemistry displayed in table 5.1, is caused partly by alteration but also by a modal variation in the primary minerals; olivine, pyroxenes, plagioclase and chromite.

Based on ionic radius the trace elements Ni, Sr, Cr and V and the major elements Ca, Mg and Al can be assigned to the primary minerals as shown in table 6.1. Ni-content tend to be mainly controlled by the amount of olivine and pyroxene in primary mineral assemblage, with primary olivine being the main Ni bearing phase (Table 6.1). Elevation in Cr-content are associated with both primary pyroxene and chromite, where the highest concentrations (Table 5.1) are associated with primary chromite (Table 6.1). In addition to elevated Cr-content, primary pyroxene is also associated with relative high V-content. Elevated Sr-content is associated with primary plagioclase (Table 6.1).

For a more thorough elaboration of the recognized trend, please look to section 5.2.1.

Based on the recognized trend (section 5.2.1) regarding whole-rock chemistry (Table 5.1) and petrographically observation of present mineral assemblage (section 5.3), the studied rocks are located in the classification diagram of ultramafic and mafic rocks as shown in Fig. 6.1.

<table>
<thead>
<tr>
<th>Element</th>
<th>Attributor</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nikkel</td>
<td>Ol&gt;Opx&gt;Cpx</td>
</tr>
<tr>
<td>Strontium</td>
<td>Ant</td>
</tr>
<tr>
<td>Cromium</td>
<td>Chr&gt;Px</td>
</tr>
<tr>
<td>Vanadium</td>
<td>Px</td>
</tr>
<tr>
<td>Calcium</td>
<td>Ant&gt;Cpx</td>
</tr>
<tr>
<td>Magnesium</td>
<td>Ol=Opx</td>
</tr>
<tr>
<td>Aluminum</td>
<td>Ant&gt;Cpx</td>
</tr>
</tbody>
</table>

Ant: anorthite, Chr: chromite
Opx: orthopyroxene, Ol: olivine,
Px: pyroxene, Cpx: clinopyroxene
The suggested composition of primary mineral assemblage of the studied rocks is displayed in Fig. 6.1. The protoliths (Table 5.1), further separated by individual chemical compositions (elaborated in section 5.2.1 and section 6.2), are suggested to have had primary compositions related to those displayed in Fig. 6.1 by the colored areas. The sample with an anorthositic protolith is suggested to have been almost purely composed of anorthite. The analyzed gabbros of a gabbronoritic composition with varying properties of primary plagioclase and primary clinopyroxene. The elevated Ni- and Cr-content in the listwanitized gabbro LEL 5-17 (Table 5.1) may suggest that it had higher proportions of primary Opx compared to the other gabbros analyzed. The pyroxenitic protoliths are suggested to have had a composition resembling websterite (Fig. 6.1), with small amounts of primary plagioclase and/or primary olivine. Composition of the peridotitic protoliths is suggested to have ranged between dunite (sample LEL 16-16) and olivine ortho/clinopyroxene (sample LEL 7-17) (Table 5.1).
6.3 Alteration

As previously mentioned, the intensive alteration of the analyzed samples has led to a complete obliteration of primary mineral assemblages. Primary minerals which includes olivine, plagioclase, clinopyroxene, chromite and orthopyroxene, are all altered following different reactions.

The alterations found to have occurred to the analyzed samples are; amphibolitization, chloritization, serpentinization and carbonation. These are all elaborated and discussed in the following sections.

6.3.1 Replacement of primary pyroxene

Amphibolitization, a H$_2$O-consuming metamorphism, is a process commonly observed in ophiolites (Mével, 1987; Hopkinson and Roberts, 1995; Berger et al., 2005; Austrheim and Prestvik, 2008; Iyer et al., 2008; Plissart et al., 2009). Earlier work (Berger et al., 2005; Iyer et al., 2008; Plissart et al., 2009) describes the amphibolitization of ophiolites as a retrograde H$_2$O-rich metamorphism occurring as an ocean-floor processes in slow-spreading ridge environments under amphibolite facies conditions (Mevel, 1988; Cannat et al., 2007; Dilek et al., 2007). The amphibolitization, which primarily led to replacement of clinopyroxene (R1) (Austrheim and Prestvik, 2008; Iyer et al., 2008), may also have led to replacement of orthopyroxene (R2) (Iyer et al., 2008) and a more sodic composition of plagioclase (R3) (Berger et al., 2005; Plissart et al., 2009) following e.g.,

\[
\begin{align*}
CaMgSi_2O_6 + 6HCl & \rightarrow Ca_2Mg_3Si_8O_{22}(OH)_2 + 3CaCl_{2(aq)} + 2SiO_{2(aq)} + 2H_2O \\
CPx & \text{Tremolite} \\
4CaMgSi_2O_6 + 5Mg_2Si_2O_6 + 2H_2O & \rightarrow Ca_2Mg_5Si_8O_{22}(OH)_2 + 2Mg_2SiO_4 \\
CPx & \text{Opx} \quad \text{Tremolite} \quad \text{Olivine} \\
CPx + Plg(An_{56-80}) + Fe - Ti oxide & \rightarrow Mg - Hbl + Plg(An_{35-63}) + Ttn \\
\end{align*}
\]  

(R1) 

(R2) 

(R3)
The analyzed amphiboles plot as (Fig. 6.2); Tschermakite, tremolite (Iyer et al., 2008), Mg-hornblende and actinolite (Berger et al., 2005; Iyer et al., 2008; Plissart et al., 2009). The amphibole is also observed to participate in further alterations (e.g. in carbonation of the pyroxenites, Fig. 5.6).

In addition to the early stage amphibolitization (Iyer et al., 2008), primary clinopyroxene is suggested to also have been serpentinized in peridotites following e.g. (R4), made by Austrheim and Prestvik (2008);

\[
100g \ Cpx_{pr \ m} + 12.5g \ H_2O + 12.5g \ MgO + 1g \ FeO \rightarrow \\
Mg_3Si_2O_5(OH)_4 + 18.5g \ SiO_2 + 24g \ CaO + 1.5g \ Al_2O_3 + 0.2g \ Na_2O
\]  \hspace{1cm} (R4)

This is suggested based on the observation of amphibole lamellae in serpentine which comprise the major phase in sample LEL 15-17 (Fig. 5.14 A). The amphibole lamellae, presumably formed at the cleavage of Cpx, are left unaltered during serpentinization while unamphibolitized Cpx are serpentinized. The variations in Al_2O_3-content in serpentines from sample LEL 6-17 and LEL 15-17 (Table 5.5), may indicate that formation of the ones elevated in Al_2O_3 involved replacement of primary clinopyroxene following e.g. (R4) made by Austrheim and Prestvik (2008).
Discussion

Chloritization of primary Cpx is also suggested to have occurred, however, since formation of chlorite also is associated with replacement of plagioclase in this study, chloritization of primary Cpx is elaborated in section 6.4.2.

6.4 **Carbonation ±serpenetization**

Petrographically observations (section 5.3) and geochemical analyses (section 5.2) display that all of the four protoliths are carbonatized, while serpenetization is only observed to have occurred in the peridotites.

6.4.1 **Peridotite**

The samples LEL 6-17, LEL 7-17 and LEL 15-17, which are all heavily altered, display large petrographic (section 5.3) and geochemical variations (Table 5.1).

Mineral assemblage observed in sample LEL 6-17 consists of Mgs + Srp + Tlc + Ferritchromite + Cr-Spl, where magnesite appears as large fractured euhrdral grains which is occasionally being replaced by bladed serpentine (Fig. 5.5 D). Serpentine is also found in the heterogenous mixture with talc which comprise a large phase in the sample (Fig. 5.5). Complete and partly replacement at the rim of unfractured Cr-spinel with ferritchromite are a commonly observed feature (Fig. 5.5).

Sample LEL 7-17 are composed of primarily talc and small amounts of serpentine, carbonate and secondary amphibole. The serpentine is observed in small amounts of the close to homogenous fibrous talc, carbonate as small anhedral grains and the amphibole as elongated anhedral grains (Fig. 5.13 A). The growth of the amphibole indicate that this is a secondary amphibole assemblage, as it crosscuts the talc’s texture (Fig. 5.13 A).

Mineral assemblage in sample LEL 15-17 consists of Tre + Srp + Mgs + Fe-Oxide. The serpentine is observed in BSE images as either dark grey aggregates or as brighter grey needles (Fig. 5.9 D). The two types of serpentine are assumed to represent two generations of serpentinization, where the darkest serpentine presumably formed first. The age relationship between the two is interpreted based on the observation of the brighter serpentine cutting the darker which also contain amphibole lamellae (Fig. 5.9 D and Fig. 5.14 A). The amphibole, present as amphibole lamellae and heavily deformed grains (Fig. 5.9 D and Fig. 5.14 A), classifies as tremolite (Fig. 6.2). Carbonate in the sample are highly abundant (~10 %) and classifies as magnesite and a Mg-rich dolomite with inclusions (Table 5.3).
Apart from the suggested formation of serpentine from primary Cpx (R4), serpentine are mainly formed from breakdown of olivine (Fig. 6.3).

Breakdown of primary olivine occurs as illustrated in (Fig. 6.3) following either (R5), (R6), (R7) or (R8) (Halls and Zhao, 1995; Iyer et al., 2008; Power et al., 2013; Bjerga, 2014). (R5) is a reaction where olivine is completely hydrated to form serpentine and brucite (Fig. 6.3), while (R8) is a pure carbonation of olivine where magnesite and quartz are formed. Formation of magnesite and quartz directly from olivine is a replacement commonly associated with listwanite formation (Halls and Zhao, 1995);

\[
2Mg_2SiO_4 + 3H_2O \rightarrow Mg_3Si_2O_5(OH)_4 + Mg(OH)_2
\]
\[
\text{Olivine} \quad \text{Serpentine} \quad \text{Brucite}
\]

\[
Mg_2SiO_4 + 3CO_2 \rightarrow 2MgCO_3 + SiO_2
\]
\[
\text{Olivine} \quad \text{Magnesite} \quad \text{Quartz}
\]

(R6) and (R7) (Fig. 6.3) are reactions where breakdown of olivine occurs following interactions with binary fluids (H\(_2\)O+CO\(_2\)) which results in the formation of either Srp + Mgs (R6) or Tlc + Mgs (R7) (Power et al., 2013; Bjerga, 2014);

\[
2Mg_2SiO_4 + CO_2 + 2H_2O \rightarrow Mg_3Si_2O_5(OH)_4 + MgCO_3
\]
\[
\text{Olivine} \quad \text{Serpentine} \quad \text{Magnesite}
\]

(R6)
Minerals formed following (R5), (R6) and (R7) can further be carbonatized and dehydrated following the reactions (Fig. 6.3) (Power et al., 2013; Bjerga, 2014):

\[
2Mg_2SiO_4 + CO_2 + 2H_2O \rightarrow 2Mg_3Si_4O_{10}(OH)_2 + MgCO_3
\]

Olivine Talc Magnesite \hspace{1cm} (R7)

\[
Mg(OH)_2 + CO_2 \rightarrow MgCO_3 + H_2O
\]

brucite magnesite \hspace{1cm} (R9)

\[
2Mg_3Si_2O_5(OH)_4 + 3CO_2 \rightarrow Mg_3Si_4O_{10}(OH)_2 + 3MgCO_3 + 3H_2O
\]

Serpentine Talc Magnesite \hspace{1cm} (R10)

\[
Mg_3Si_4O_{10}(OH)_2 + MgCO_3 + 3CO_2 \rightarrow 4MgCO_3 + 4SiO_2 + H_2O
\]

(R11)

Which of the numerous reactions (Fig. 6.3) to have occurred to form the present mineral assemblages are hard to interpret. However, a few observations (1-5) may indicate which reactions and in what sequence they occurred.

**Observations (1-5):**

1) The absence of brucite in present mineral assemblages imply that either (R5) has not occurred, or that brucite formed following (R5) are completely replaced following (R9).

2) The relation between magnesite and serpentine in sample LEL 6-17 (Fig. 5.14 A and B), exhibit either a simultaneously formation of serpentine and magnesite following replacement of olivine (R6) or serpentinization of previously formed magnesite.

3) Serpentine observed enclosed by magnesite in sample LEL 15-17 (Fig 5.14 B) could be a result of two possible scenarios; carbonation of brucite (R9) formed from hydration of olivine (R5), or a simultaneously formation of serpentine and carbonate from olivine (R6). The serpentine enclosed by magnesite (Fig. 5.14 B), is the previously described dark serpentine (Fig. 5.9 D) suggested to involve replacement of primary clinopyroxene (R4). If this suggestion is true, the age relation between the magnesite and serpentine would be complicated, as they are formed following alteration of two different primary minerals (Cpx and Ol).

4) Observation of brighter grey bladed serpentine, fitting the descriptions of antigorite according to Iyer et al. (2008), crosscutting previous textures and replacing magnesite indicate a later stage of serpentinization (Fig. 5.9 D, Fig. 5.5 A and D). However, which reaction this serpentinization followed are hard to interpret.

5) Quartz are not observed in either of the present mineral assemblages, indicating that complete carbonation of the analyzed peridotites following (R8) and (R11) has not occurred.
The explanation behind the absence of brucite are of great importance regarding the sequence of alteration: If brucite was formed \((R5)\) it would imply a complete hydration prior to carbonation (Power et al., 2013; Bjerga, 2014). In addition, if formation of brucite occurred or not, influences the interpretation of the age relation between the magnesite and serpentine in observation; 3). As magnesite may be a result of carbonation of brucite \((R9)\) and not necessarily indicate serpentinization post carbonation.

With quartz and brucite being absent and presence of a significant amount of serpentine and carbonate in the analyzed samples, it is strongly suggested that reacting fluid in this study have been of a binary composition \((\text{H}_2\text{O}+\text{CO}_2)\). Textures discussed in observations 2) and 3) may indicate that there has been a shift in \(f(\text{CO}_2)\) during replacement, leading to carbonation instead of serpentinization or the other way around. The observation 4) of bladed antigorite (Fig. 5.5) cross-cutting previous textures indicate an elevation in temperature and shift in \(f(\text{CO}_2)\) (Wicks and Whittaker, 1977; Iyer et al., 2008). The observed formation of a secondary amphibole in LEL 7-17 (Fig. 5.13 A) may indicate either a temperature or pressure increase according to phase diagrams made by Iyer et al. (2008) and Bjerga (2014).

The observed replacement of Cr-spinel with ferritchromite (Fig. 5.5 C), indicates that magnetite most presumably was formed following \((R13)\), made by Iyer et al. (2008). However, it may also have formed following \((R12)\), made by Iyer et al. (2008). The magnetite formed may later have been broken down during oxidization and/or Fe leaching, processes described in other studies to have occurred in the LOC (Iyer et al., 2008; Bjerga, 2014) and in other ophiolites (Plissart et al., 2009);

\[
\begin{align*}
\text{Fayalite} & \rightarrow \text{Magnetite} \\
\text{Cr - spinel} + \text{Fe}^{3+} + \text{H}_2\text{O} & \rightarrow \text{Ferritchromite} + \text{Magnetite} + \text{Al}^{3+} + \text{H}_2
\end{align*}
\]

\((R12)\)

\((R13)\)

6.4.2 Gabbro-Listwanite

Mineral assemblages in the analyzed gabbro-listwanites (sample LEL 5-17, 13-17 and 14-17) consists of carbonate, chlorite, quartz, tourmaline, pyrite/oxide, mica and ±hydrous aluminum silicate (section 5.3). The whole-rock chemistry (Table 5.1) and the suggested primary mineral assemblage (Fig. 6.1) of the analyzed gabbros, imply that they have carried a significant amount of primary plagioclase and substantially amount of primary clinopyroxene.
Amphibolitized (R3) and/or primary clinopyroxene in gabbros are interpreted to have been completely replaced during a later stage carbonation following reactions e.g. (Kishida and Kerrich, 1987; Plissart et al., 2009);

\[
2 \text{Czo} + 3 \text{Act} + 10 \text{CO}_2 + 8 \text{H}_2\text{O} \rightarrow 3 \text{Chl} + 10 \text{Cal} + 21 \text{Qtz} \quad (R14)
\]

\[
\text{Primary Cpx + fluids} \rightarrow \text{Chlorite + Carbonate + Quartz} \quad (R15)
\]

If (R15) occurred, it would be a direct carbonation of primary minerals associated with listwanitation according to Halls and Zhao (1995).

However, the variations in present mineral assemblage (section 5.3) and in composition (Table 5.1), are mainly associated with the replacement of plagioclase.

An early replacement of primary plagioclase may have occurred, which would presumably have led to formation of a less anorthitic plagioclase e.g. as described by Plissart et al. (2009). However, if one assumes immobility of Al, the $\text{Al}_2\text{O}_3$ content (18.14 wt.\% - 2098 wt.\%) of the analyzed gabbros (Table 5.1) indicate an approximately similar amount of primary plagioclase.

Table 6.2. Mineralogy of gabbroic samples

<table>
<thead>
<tr>
<th>Aluminum-bearing minerals</th>
<th>Sodium-bearing minerals</th>
</tr>
</thead>
<tbody>
<tr>
<td>Carbonate</td>
<td>Magnesite</td>
</tr>
<tr>
<td>Calcite</td>
<td>Dolomite</td>
</tr>
<tr>
<td>LEL 5-17</td>
<td>-</td>
</tr>
<tr>
<td>LEL 13-17</td>
<td>+++</td>
</tr>
<tr>
<td>LEL 14-17</td>
<td>++</td>
</tr>
</tbody>
</table>

Aluminum-bearing minerals associated with the replacement of plagioclase in this study are chlorite, kaolinite (Fig. 5.17), tourmaline (Fig. 5.18), paragonite (Fig. 5.19) and andesine (Fig. 5.10 C), with the last three mentioned also being the associated sodium-bearing minerals (Table 6.2). Formation of carbonate and chlorite are also associated with replacement of primary pyroxene (Berger et al., 2005; Plissart et al., 2009). However, an interesting observation is the abundance of the various aluminum-bearing and sodium-bearing minerals (Table 6.2). During replacement, the assumed immobile aluminum, is interpreted to either form; chlorite, kaolinite, tourmaline, paragonite and/or plagioclase (andesine). Which minerals formed, are suggested to correlate with $\text{Mg}^{2+}$, $\text{Ca}^{2+}$, $\text{Fe}^{3+}$, $\text{Na}^{+}$ concentrations and LOI in the samples (Table 5.1). The suggested correlation between Mg, Na, Fe, LOI and present mineral assemblage observed (section 5.3), is displayed in Fig. 6.4.

To better express the suggested correlation, Ca was excluded in the illustration. The exclusion of Ca does affect the mineral-lines, as Ca could form calcite and together with Mg form dolomite, and hence LOI would not necessarily lead to a less abundant chlorite. However, the
aim with the figure is to display how element concentrations and LOI affect formation of aluminum-bearing minerals, and Ca is only a minor component in tourmaline (Table 5.5).

The compositions of chlorite (Table 5.4) and tourmaline (Table 5.5) are similar in the samples LEL 13-17 and LEL 14-17, however composition of these to minerals were not attained from sample LEL 5-17. Hence, regarding the viability of the illustration, the composition of these two minerals in sample LEL 5-17 was assumed to be of similar compositions to those analyzed in sample LEL 13-17 and LEL 14-17.

The previously discussed varying replacement of plagioclase in the gabbro-listwanite, is suggested to have followed reactions e.g.;

\[
\text{LEL 14} - 17: \text{Plg} + \text{CO}_2 + B + M g^{2+} + F e^{x+} \\
\rightarrow \text{carbonate + kaolinite + tourmaline + paragonite + chlorite (R16)}
\]
Discussion

\( LEL\ 13 - 17: Plg + CO_2 + B + Mg^{2+} + Fe^{x+} \rightarrow \text{carbonate} + \text{tourmaline} + \text{chlorite} + \text{andesite} \) (R17)

The Mg\(^{2+}\) and Fe\(^{x+}\) needed to form tourmaline, chlorite, magnesite and dolomite, may either have been excess elements from replacement of e.g. pyroxenes, or have come from the reacting fluids.

6.4.3 Pyroxenite

The suggested primary mineral assemblages in the pyroxenites (Fig. 6.1) are close to websteritic composition. This suggestion is supported by the observation of the highly abundant Ca-rich amphibole (Fig. 5.6-5.9), leading to the assumption that the mineral assemblage is a product of amphibolitization of primarily clinopyroxene and lesser orthopyroxene (Berger \textit{et al.}, 2005; Austrheim and Prestvik, 2008; Plissart \textit{et al.}, 2009).

Fig. 5.6 – 5.9 display the mineral assemblages in the samples LEL 8-17 to LEL 12-17 consisting of Amp + Qtz + Chl/Zoi + Cc + Ilm/Ttn, where the extent of carbonation affects the proportions of the minerals. The least altered sample LEL 8-17 consist mainly of amphibole with minor amounts of chlorite, calcite, quartz and ilmenite (Fig. 5.6), while the most altered sample LEL 12-17 (Fig. 5.9) show replacement of ilmenite with titanite, decreasing amounts of amphibole and increasing amounts of chlorite/zoisite, calcite and quartz. The formation of calcite along the amphiboles cleavage/exsolution lamellae may indicate where the CO\(_2\)-enriched fluids entered the amphibole during the replacement of amphibole with chlorite/zoisite, calcite and quartz (R14), made by Kishida and Kerrich (1987). As the extent of carbonation increases, the veins of chlorite/zoisite and calcite expands. Amphibole and quartz are observed in some of the veins, assumed to be a result of complete replacement of that part of the amphibole. The observed kinking in sample LEL 11-17 (Fig. 5.8) are suggested to be a result of external pressure.

The replacement observed in the less altered pyroxenites are suggested to display the replacement of amphibole described by Hopkinson and Roberts (1995), Berger \textit{et al.} (2005) and Plissart \textit{et al.} (2009) following the reaction (R14), made by Kishida and Kerrich (1987). The reaction, described to occur in gabbros, are also observed in the pyroxenites in this study.

The complete carbonation of pyroxenites are represented by the mineral assemblage in sample LEL 20-17 (Fig. 5.12), which fulfills the requirements to be classified as listwanite according to Halls and Zhao (1995). The mineral assemblage consists of Chl + Dol + Qtz + Cc + Phg +
Chalcopyrite, which are interpreted to be a result of complete replacement of amphibole following the reaction (R14). The formation of Cr-muscovite in sample LEL 20-17, indicate a K+ metasomatism, an metasomatism associated with interaction of seawater during ocean-floor metamorphism (Mével, 1987; Bach \textit{et al.}, 2001; Rüpke \textit{et al.}, 2002; Honnorez, 2003; Berger \textit{et al.}, 2005; Plissart \textit{et al.}, 2009). The high LOI (24.59 wt.%) and the low Al2O3-content (Table 5.1), indicate that the majority of Mg2+ is situated in dolomite and that chlorite comprise a minor phase.

6.4.4 \textit{Anorthosite}

Sample LEL 3-16, which is the only sample classified as anorthosite, was only analyzed with XRF. The whole-rock chemistry attained (Table 5.1) indicate that the sample was very rich in primary anorthite (Fig. 6.1), hence its high Al-content. However, the low Ca-content (11.51 wt.%) indicate that Ca2+ has been leached out of the sample during alteration. The suggested Ca-migration are elaborated in the next section (section 6.5).

6.5 \textit{Fluids sources and elements migration}

The fluids reacting during the ocean-floor alteration of the LOC are described in other studies (Iyer \textit{et al.}, 2008), and in studies of similar scenarios (Hopkinson and Roberts, 1995; Hopkinson \textit{et al.}, 2004; Berger \textit{et al.}, 2005; Boschi \textit{et al.}, 2006; Plissart \textit{et al.}, 2009), to presumably have gained access through either; oceanic fractures, detachment faults, sub-seafloor hydrothermal vent systems or direct interaction with bottom seawater. What scenario and which fluid source that led to alteration of the samples analyzed in this study, are hard to interpret. However, 87Sr/86Sr isotope analyses conducted by Gussone \textit{et al.} (In review) indicate that seawater in the rodiningites may compose 5 \% of the fluid source. Formation of Cr-muscovite (Fig. 5.12) in LEL 20-17, fuchsite in LEL 14-16 (Table 5.5) and illite-smectite in LEL 14-17 (section 5.6) supports that parts of the fluid source was composed of seawater, as these two formations are commonly associated with low-T seafloor K+ metasomatism (Melson and Thompson, 1973; Hart \textit{et al.}, 1974; Kishida and Kerrich, 1987; Bach \textit{et al.}, 2001; Plissart \textit{et al.}, 2009; Böhlke \textit{et al.}, 2010). Formation of smectite, as in sample LEL 14-17 (section 5.6), is in recent studies (Marschall, 2017) together with seafloor serpentinization associated with boron migration. Possible fluid sources during the boron migration are according to Marschall (2017), either seawater and/or hydrothermal vent fluids. The findings in (Marschall, 2017) may be used to suggest that boron migration in this study occurred in an ocean-floor setting with seawater and/or hydrothermal
vent fluids as fluid source. However, predominance of calcite over zoisite indicates that reacting fluids where enriched in CO$_2$, opposite to low f(CO$_2$) associated with bottom seawater (Kitajima et al., 2001; Plissart et al., 2009). According to Bjerga (2014), the main fluid source for serpentinization and carbonation of the upper parts of the Leka ophiolite was devolatilization of the lower parts of the complex. The main fluids source described by Bjerga (2014), is a possible fluid source for the alteration found in this study. However, in which setting the devolatilization occurred is difficult to determine.

Table 6.3
Molar proportions of cations, calculated from wt.% displayed in table 5.1.
Further used to calculate *ratio

<table>
<thead>
<tr>
<th>Sample</th>
<th>Protolith</th>
<th>Anorthosite</th>
<th>Gabbro</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>LEL 3-16</td>
<td>LEL 5-17/</td>
<td>LEL 13-17/</td>
</tr>
<tr>
<td>Cations</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Si</td>
<td>0.7468</td>
<td>0.6021</td>
<td>0.7826</td>
</tr>
<tr>
<td>Ti</td>
<td>0.0005</td>
<td>0.0013</td>
<td>0.0014</td>
</tr>
<tr>
<td>Al</td>
<td>0.6501</td>
<td>0.4115</td>
<td>0.3874</td>
</tr>
<tr>
<td>Fe</td>
<td>0.0226</td>
<td>0.1101</td>
<td>0.0870</td>
</tr>
<tr>
<td>Mn</td>
<td>0.0004</td>
<td>0.0011</td>
<td>0.0017</td>
</tr>
<tr>
<td>Mg</td>
<td>0.0910</td>
<td>0.3939</td>
<td>0.3139</td>
</tr>
<tr>
<td>Ca</td>
<td>0.2053</td>
<td>0.0244</td>
<td>0.0936</td>
</tr>
<tr>
<td>Na</td>
<td>0.0952</td>
<td>0.0441</td>
<td>0.0615</td>
</tr>
<tr>
<td>K</td>
<td>0.0022</td>
<td>0.0005</td>
<td>0.0005</td>
</tr>
<tr>
<td>P</td>
<td>0.0000</td>
<td>0.0000</td>
<td>0.0000</td>
</tr>
<tr>
<td><strong>SUM</strong></td>
<td>1.8141</td>
<td>1.5889</td>
<td>1.7294</td>
</tr>
<tr>
<td><strong>ratio</strong></td>
<td>0.7813</td>
<td>0.2268</td>
<td>0.6430</td>
</tr>
</tbody>
</table>

*ratio: (Na+K+2*Ca)/Al ratio
*SUM: SUM molar proportion of cations
Example: Calculation of *ratio with an anorthosite purely composed of anorthite

If one assumes immobility of aluminum during alteration, table 6.3 display the (Na+K+2*Ca/Al) ratio in the gabbro-listwanites and the altered anorthosite. A scenario, where all the aluminum is attributed by primary plagioclase (CaAl$_2$Si$_2$O$_8$-NaAlSi$_3$O$_8$) and replaced during isochemical metamorphism, the *ratio would be 1. The *ratio represents a minimum value, as Ca$^{2+}$ and Na$^+$ also could have been attributed by primary Cpx. However, *ratio calculated from cation molar proportions in the gabbro-listwanites and anorthosite (Table 6.3) strongly imply that migration of Na$^+$, K$^+$ and Ca$^{2+}$ have occurred to a variable extent.

The *ratio for each of the samples, imply that minor amounts of Na$^+$, K$^+$ and Ca$^{2+}$ has been removed from the anorthosite (LEL 3-16) and to a larger extent from the gabbro-listwanite LEL 13-17. LEL 5-17 imply an even stronger removal of Na$^+$, K$^+$ and Ca$^{2+}$, while LEL 14-17 indicate a substantial addition of Na$^+$, K$^+$ and Ca$^{2+}$. However, as previously mentioned Ca$^{2+}$ and Na$^+$
could have been attributed by primary Cpx, which means that the elements not necessarily have been added. As the composition of primary plagioclase is mainly associated with anorthite, the large extraction and addition of elements displayed in Table 6.3 are mainly associated with Ca\textsuperscript{2+} migration.

Ca-metasomatism of the LOC have been described in earlier studies (Gussone et al., In review; Austrheim and Prestvik, 2008; Iyer et al., 2008) as a Ca-migration associated with serpentinization and rodingitization, where rodingites acts as sinks for calcium released during serpentinization (Gussone et al., In review; Iyer et al., 2008). An Lu-Hf age referenced by Gussone et al. (In review) dates the rodingites to an age identical to the formation age of the LOC (497±2 Ma), suggesting that this alteration took place at ocean floor. If the Ca-migration in this study is to be associated with rodingitization and serpentinization described by Iyer et al. (2008) and Gussone et al. (In review), is difficult to determine. Especially since when the Ca-migration occurred, either prior to/ or during carbonation, is not clear.

The source of CO\textsubscript{2} for the observed carbonation are not clear, however as previously mentioned, heating of lower stratigraphically parts may be the main fluid source suggesting that CO\textsubscript{2} consumed to form carbonation are from prograde metamorphism of carbon-bearing minerals (Johnson et al., 2008; Bjerga, 2014). The prior carbon-bearing minerals could have been enriched in CO\textsubscript{2} during formation of the LOC at the spreading ridge in a supra-subduction and back-arc setting (Furnes et al., 1988). Volatile CO\textsubscript{2} are a common feature addressed to subduction zones, where marine sediments and/or hydrothermally altered oceanic crust of the down-dropping plate are heated which leads to an upward volatile-migration containing CO\textsubscript{2} (Furnes et al., 1988; Kerrick and Connolly, 1998). The volatile CO\textsubscript{2} in an arc-magmatic affinity are then intruding the overlaying plate where the LOC are interpreted to have been sited during formation (Furnes et al., 1988). A possible scenario is that carbonates formed at high-T, destabilizes during the decreasing temperatures following regression leading to an enriching of interacting fluids with CO\textsubscript{2}, or heating of lower parts of the lithology witch breaks up the carbonate resulting in upward-migrating fluids enriched in CO\textsubscript{2} (Johnson et al., 2008).
6.6 Listwanite

The term listwanite was first introduced by Rose (1837, 1842), and later elaborated by Halls and Zhao (1995) to involve rocks of certain characteristics. The characteristics described by Halls and Zhao (1995) needs to be fulfilled before one can classify a rock as listwanite. The term listwanite is often a misused term, as it may often be used on rocks that do not fulfill the characteristics described by Halls and Zhao (1995). The most common misinterpretation is of rocks which are carbonatized and serpentinized, but do not reach the stage where it may be classified as listwanite. An example in this study is the soapstone (Fig. 3.2 C) which is carbonatized to form talc and magnesite, but quartz is not present and consequently the soapstone is not a listwanite. The characteristics described by Halls and Zhao (1995), involves mainly index mineral e.g. carbonates, pyrite, Cr-mica and quartz and color of the assemblage, which is to be green (Rose, 1837, 1842) because of fuchsite. The rocks in this study classified as listwanite fulfill these characteristics to variable extent. The whole-rock chemistry (Table 5.1) imply that K+ metasomatism has led to formation of Cr-muscovite, fuchsite and the presumably smectite-illite, which is a common metasomatism associated with listwanites (Halls and Zhao, 1995). However, Cr-mica is only present in the pyroxenite-listwanites and not the gabbro-listwanites. This is suggested to correlate with the amount of primary pyroxene, which is assumed together with chromite to have been the Cr-attributors in the rocks analyzed (section 6.2). A higher abundance of primary pyroxene in the pyroxenite-listwanites than in the gabbro-listwanites (section 6.2) is suggested to have led to an alteration more sufficient in Cr, which led to formation of Cr-mica.

In which listwanite Cr-mica are formed do also correlate with the color of the assemblages, where the pyroxenite-listwanite assemblages exhibit a green color while the gabbro-listwanite assemblages typically exhibit a more greyish color (Chapter 3). But with this said, the classification of listwanites are often related to rocks of a peridotitic composition associated with serpentinization (Halls and Zhao, 1995), and not of gabbro- and pyroxenite-listwanites. Hence, the color associated with gabbro- and pyroxenite-listwanite assemblages is a less documented feature.

Based on the findings regarding listwanites in this study, it is strongly suggested that they are of the rare type derived from gabbro and pyroxenite, where a high abundance of primary plagioclase in the gabbro-listwanites has led to formation of tourmaline and occasionally kaolinite. The T-XCO₂ diagram (Fig. 5.20) attained from sample LEL 14-17 display a stability
area for the observed mineral assemblage between 90 °C and 220 °C at XCO₂ = 0, with the maximum temperature for the stability area decreasing significantly together with an increase in saturation of CO₂ in fluids. Based on the correlation between formation of tourmaline, ±kaolinite and the extent of carbonation previously discussed (section 6.4.2), it is suggested that the carbonation of the analyzed samples presumably occurred concurrent and thus in the same tectonic setting, where temperature and f (CO₂) were the controlling factors during carbonation.

6.7 Structures

The location where listwanites are situated in the area studied (Fig. 5.1), is associated with the large NE-SW trending faults or shear zones described by Furnes et al. (1988) and Titus et al. (2002), which are suggested by Titus et al. (2002) to have formed during a late-stage brittle deformation of the LOC post-obduction. The context between the listwanite formation and NE-SW trending faults are hard to interpret, however if not concurrent, one of them has to precede the other. The lineaments (Fig. 5.3) display that the general orientation of linear surface-structures (Gabrielsen et al., 2002) are NE-SW oriented, in addition to the listwanites, large veins of carbonated peridotite are observed following these NW-SW trending faults (Fig. 5.1). The listwanitation (Fig. 3.1), associated with ocean-floor processes in this study, are suggested to have preceded the brittle deformation, and possibly even played a vital role in the deformation of the LOC. The suggestion is based on the ductile deformation to have affected the foliated listwanite oriented perpendicular to the NW-SW trending faults (Fig. 3.5 B and Fig. 3.7), which indicate that the studied listwanites were formed simultaneously or prior to the ductile deformation. The formation of the faults and fractures associated with post-obduction deformation (Titus et al., 2002), are in this study suggested to maybe have occurred during the ocean-floor phase and/or during the emplacement of the LOC. The volume increase associated with serpentinization and carbonation (Iyer et al., 2008), and fluids paths derived from e.g. detachment faults and oceanic fractures (Hopkinson and Roberts, 1995; Hopkinson et al., 2004; Boschi et al., 2006; Plissart et al., 2009), indicate a possible extensive deformation occurring prior to obduction. Serpentinization are commonly observed in present-day detachment faults in oceanic settings (Boschi et al., 2006; Karson and Lawrence, 2007; Plissart et al., 2009), which may indicate that alteration of the rocks in the LOC followed faults formed in oceanic settings. And that the observed alteration following the NE-SW trending faults are caused by
the volume increase following serpentinization and carbonation to have occurred as ocean-floor metamorphism.

Two theories to why the alteration is concentrated around the NE-SW trending faults are suggested in this study:

1) The dome-like heap which exhibit the ductile-deformed foliated listwanite (Fig. 3.5), are suggested to display a deformation associated with contraction. Contraction of the LOC are described by Furnes et al. (1988), to have occurred during the Caledonian orogeny. During the contraction, formation of NE-SW trending faults along the heavily altered zones which are weaker, or along tectonically weakened zones may have occurred.

2) The formation of the faults in the LOC could have occurred during the complex tectonics prior to emplacement. Contraction, extension, sinistral-dextral movements and mantle interference (e.g. mantle plumes) are observed in modern ocean environments (Vegas et al., 1990; Alavi, 1991) to form fault- and fracture-systems in un-obducted oceanic crusts sited in the Iberia plate.

More work is required to decide if either of the models are valid explanations for the association of the altered rocks in this study and the NE-SW trending faults and shear zones described by Furnes et al. (1988) and Titus et al. (2002).
Chapter 7: Conclusions

Concluding remarks based on the findings in the study.

- Alteration are found to have occurred to samples of all the four protoliths; peridotite, gabbro, pyroxenite and anorthosite. The alteration, which includes amphibolitization, serpentinization, carbonation, chloritization and listwanitisation, has affected the rocks to a various extent and led to formation of varying present mineral assemblages.

- Formation of listwanite is only found to have occurred to the rocks with protoliths of gabbroic and pyroxenitic compositions, where primary plagioclase in the gabbro-listwanites has led to an aluminum-rich present mineral assemblage. The high aluminum content and boron is what has resulted in formation of tourmaline and occasionally kaolinite.

- Present mineral assemblages observed in the listwanite are suggested to have been formed simultaneously, and with the attained T-XCO$_2$ of the sample LEL 14-17, carbonation and complete listwanitation has occurred under temperatures possible as low as and maybe even lower that 90 °C.

- The association between listwanites and the NE-SW trending faults and shear zones, are in this study suggested to be a result of preceding listwanitation, where later deformation has led to formation of the associated NE-SW trending faults and shear zones along these weakened zones.

- The findings in this study imply that the listwanites in the LOC are partly controlled by primary mineral assemblage, especially regarding primary plagioclase in gabbro-listwanites, and element migration of mainly Ca$^{2+}$, Na$^+$ and K$^+$. 

- In this study it is suggested that listwanite has formed in an ocean-floor setting as it indicates reaction with seawater, addition of boron associated with sea-floor serpentinization and formation of smectite, and Ca$^{2+}$, Na$^+$ migration which may correlate with sea-floor serpentinization and rodingitization.
Conclusions
Chapter 8: Future work

The complex alteration-history of the LOC complicates the interpretation of the findings in this study. Even though, this work has improved the understanding of listwanite in the LOC and conditions during listwanitation, there is this a lot more that needs to be clarified through future work.

Interesting topics for future work could include $^{87}\text{Sr}/^{86}\text{Sr}$, $\delta^{44/40}\text{Ca}$, oxygen and Lu-Hf isotope measurements, which would determine fluid source, calcium source and age of the listwanites. This could further be used in determining in which setting listwanite was formed, and age relation between the associated listwanites and NE-SW trending faults found in this study.

Analyzes of a less altered gabbro from the same area studied, would improve the understanding regarding evolution of gabbro-listwanites, involving aluminum phases and if the suggested Ca migration occurred prior to/ or during listwanitation.

Test if gold is enriched in the Leka listwanite by applying analytical methods with a lower detection limit than XRF.
Future work
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Appendix

Table 5.6
Composition of Titanite, Talc, Oxide and Cr-spinel sorted according to mineral

<table>
<thead>
<tr>
<th>Sample</th>
<th>Titanite</th>
<th>Talc</th>
<th>Oxide</th>
<th>Cr-spinel</th>
</tr>
</thead>
<tbody>
<tr>
<td>9-17</td>
<td>LEL</td>
<td>LEL</td>
<td>LEL</td>
<td>LEL</td>
</tr>
<tr>
<td>6-17</td>
<td>Per</td>
<td>Pxy</td>
<td>Per</td>
<td>Per</td>
</tr>
</tbody>
</table>

| SiO₂    | 30.76    | 58.26 | 0.02  | 1.85    | 0.09    | 0.02 |
| TiO₂    | 38.05    | 0.03  | 0.48  | 50.42   | 0.04    | 0.09 |
| Al₂O₃   | 1.41     | 0.64  | 0.49  | 0.98    | 0.32    | 22.88 |
| FeO (tot)| 0.42  | 4.02  | 62.28 | 40.39   | 90.99   | 33.43 |
| MnO     | 0.02     | 0.01  | 0.37  | 3.33    | 0.00    | 0.33 |
| MgO     | 0.54     | 31.20 | 0.22  | 0.67    | 0.04    | 4.94 |
| CaO     | 27.75    | 0.05  | 0.02  | 0.55    | 0.09    | 0.02 |
| Na₂O    | 0.01     | 0.00  | n.a.  | 0.01    | n.a.    | n.a. |
| K₂O     | 0.00     | 0.00  | 0.01  | 0.00    | 0.00    | n.a. |
| Cr₂O₃   | 0.32     | 0.18  | 31.41 | 0.09    | 1.71    | 37.28 |
| Total   | 99.28    | 94.39 | 95.29 | 98.29   | 93.25   | 98.98 |

Structural formula based on:

<table>
<thead>
<tr>
<th>5(O)</th>
<th>11(O)</th>
<th>3(O)</th>
<th>4(O)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Si</td>
<td>1.011</td>
<td>3.823</td>
<td>0.001</td>
</tr>
<tr>
<td>Ti</td>
<td>0.940</td>
<td>0.001</td>
<td>0.012</td>
</tr>
<tr>
<td>Al</td>
<td>0.055</td>
<td>0.050</td>
<td>0.019</td>
</tr>
<tr>
<td>Fe²⁺ (tot)</td>
<td>0.012</td>
<td>0.221</td>
<td>1.705</td>
</tr>
<tr>
<td>Mn</td>
<td>0.001</td>
<td>0.001</td>
<td>0.010</td>
</tr>
<tr>
<td>Mg</td>
<td>0.026</td>
<td>3.052</td>
<td>0.011</td>
</tr>
<tr>
<td>Ca</td>
<td>0.977</td>
<td>0.003</td>
<td>0.001</td>
</tr>
<tr>
<td>Na</td>
<td>0.001</td>
<td>0.000</td>
<td>0.000</td>
</tr>
<tr>
<td>K</td>
<td>0.000</td>
<td>0.000</td>
<td>0.000</td>
</tr>
<tr>
<td>Cr</td>
<td>0.008</td>
<td>0.009</td>
<td>0.813</td>
</tr>
<tr>
<td>Mg#</td>
<td>0.68</td>
<td>0.93</td>
<td>0.01</td>
</tr>
</tbody>
</table>

Mg# = Mg²⁺/(Mg²⁺+Fe²⁺); Per = peridotite, Pxy = pyroxenite
Calculation done to attain the coefficient used for calibrating the Raman spectra.

The coefficient -2.5678 do technically only apply for the interval between the two peaks, but since the standard to not contain peaks in the intervals <200 cm\(^{-1}\) and >3326 cm\(^{-1}\), coefficient for these intervals are unattainable. Leading to calibration of the whole attained Raman spectra with the coefficient -2.5678.

\[
\begin{align*}
  a + 215.7 \times b &= 213.3 \\
  a + 1373 \times b &= 1371.5 \\
  1: a &= (213.3 - 215.7 \times b) \\
  2: (213.3 - 215.7 \times b) + 1373.0 \times b &= 1371.5 \\
  -215.7 \times b + 1373.0 \times b &= 1371.5 - 213.3 \\
  1157.3 \times b &= 1158.2 \\
  b &= \frac{1158.2}{1157.3} \\
  &\approx 1.000778 \\
  1: a &= 213.3 - (215.7 \times 1.000778) \\
  a &= -2.5678
\end{align*}
\]