Excursion guide

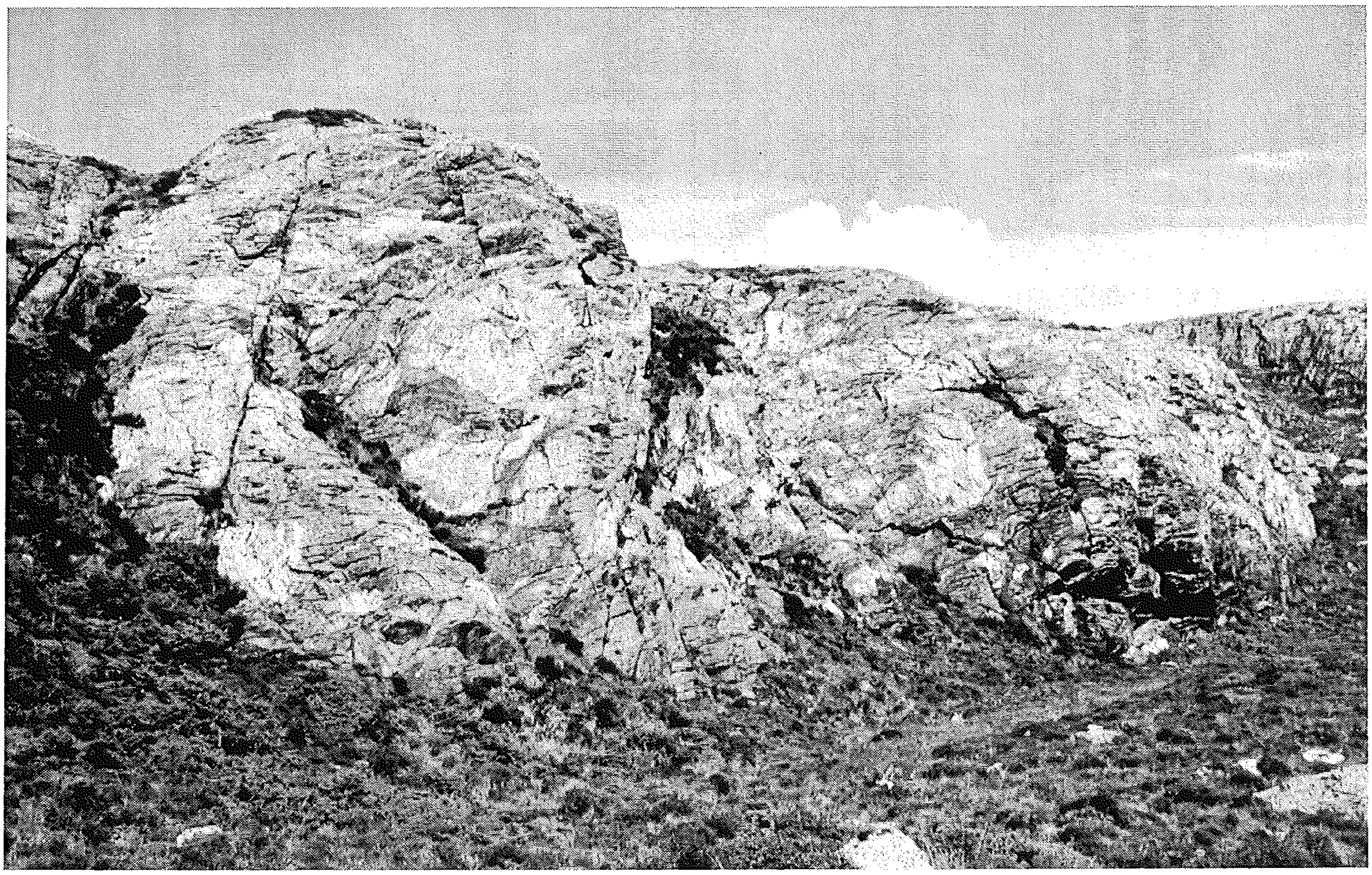
PROCESSES AND PETROPHYSICAL PROPERTIES OF CRUSTAL ROOTS

Pre-conference excursion to Holsnøy for participants at the Millennial 9th International Symposium on DEEP SEISMIC PROFILING OF THE CONTINENTS AND THEIR MARGINS.

Håkon Austrheim[1)] and David M. Fountain[2)]

[1)] Mineralogisk-Geologisk Museum, Universitetet i Oslo, Sars'gate 1, N-0562 Oslo

[2)] Dept. of Geology and Geophysics, University of Wyoming, Laramie, WY 82070-3006 USA



Eclogite breccia, a mixture of angular blocks of granulites surrounded by foliated eclogite, is an areally important rock unit on Holsnøy and in the deep crust?

1. Introduction

Below active continent-continent collision zones, the crust mantle boundary deepens to reach depths of more than 50 km under the Alps and 75 km under the Himalayas (Kissling, 1993; Roecker, 1982). The Alps and the Himalayas have developed marked root zones of ca. 20 and ca. 40 km respectively in excess of the average crustal thickness. Rocks buried into such root zones will be exposed to changing pressures, temperatures, and stress regimes and they may undergo metamorphism, partial melting and deformation. The P-T conditions prevailing in crustal root zones can stabilize eclogite-facies rocks. Eclogites have attracted metamorphic petrologists because of their ability to *record P-T-t paths* of subduction and continental collision. Such petrological work has been very successful in demonstrating the return to the surface of rocks that have experienced pressures corresponding to depths in excess of 100 km (Chopin, 1984; Schreyer, 1988).

Less attention has been paid to other important aspects of the eclogitization process. The formation of eclogite can lead to dramatic changes in *petrophysical properties*. Density changes may be greater than those met at most lithological boundaries, including the boundary between a peridotitic mantle and a gabbroic lower crust in the classical view of the Moho. These changes are likely to influence the geophysical signature (seismic velocity, reflectivity, gravimetric anomalies) of the deep crust. Geophysical properties form the basis for our present view of the structure of the deep crust and mantle and are used to constrain geodynamic models. It is therefore important to explore petrophysical changes associated with metamorphism (e.g. eclogitization) as an alternative to petrophysical changes attributed solely to compositional differences.

The changes in petrophysical properties related to metamorphism and eclogitization, in particular, must also affect the *geodynamics* of a collision or a subduction zone. The density contrast between basalt and an eclogite is at least three time greater than the density contrast resulting from cooling of the basalt drifting away from the midoceanic ridge. This latter effect may decide whether a subduction zone becomes arrested or not (Cloos, 1993). Ahrens and Schubert (1975) estimated the downward body force on a descending slab due to its eclogitization and found it to be an important component of the driving force for plate motion. Whether the crust becomes eclogitized or not must therefore have a profound influence on the geodynamics of collision zones. Mafic eclogites have densities higher than typical mantle values of 3.30 g/cm^3 and, if present in large amounts, may sink into the mantle. Such disappearance of material from the crust has been suggested for the Alps by Laubscher (1988) and for the Himalayas by LePichon et al. (1992) on the basis of mass balance calculations. The rheological changes associated with eclogitization, either temporary or permanently, may also influence the geodynamics of collision zones. Ruff and Kanamori (1983) suggested that the basalt to eclogite transition in the down- going oceanic crust is responsible for the decoupling of subduction zones below 40 km. This phase change should start at a depth of 30-35 km and could at least partially uncouple the plates by transformation plasticity. Boundy et al. (1992) suggested that crustal detachment zones may root in eclogite-facies shear zones. Backreaction of eclogites to granulites and amphibolites will affect the uplift and exhumation of deep-seated terrains as discussed by Dewey et al. (1993).

Metamorphism is a *dynamic* process that in itself can cause or influence deformation. When reacting rocks expand and contract, the volume changes will set up forces in the surrounding material (Etheridge et al., 1983; Wheeler, 1987). Such forces may result in deformation that can happen at a catastrophic rate and cause earthquakes. Rheology changes associated with phase transitions may aid the release of stresses. Several workers have connected the gabbro-eclogite transition to intermediate to deep earthquakes in subduction zones (Comte and Suarez, 1994; Hurukawa and Imoto, 1993) and collision zones (Austrheim and Boundy, 1994).

Regional metamorphism is often viewed as a slow and continuous process that takes place over millions of years following the tectonic event that brought the rocks out of their initial stability field. It is often described in terms of equilibrium following changes in P and T alone (Thompson and England, 1984) even if disequilibrium and overstepping of reactions are common features. The *role of fluids* during metamorphism and consequently in determining the petrophysical state of a crustal root zone cannot be underestimated. Ferry (1986) advocates that metamorphism occurs in pulses following fluid bursts due to devolatilization reactions. The importance of fluid is also highlighted by Fyfe et al. (1978) who suggested that crustal deformation is focused in areas of fluid overpressure. Carter et al. (1990) concluded that the rheology of the lithosphere is governed by a combination of bulk-rock flow and localized deformation in shear zones, both of which aided or controlled by rock-fluid interactions.

Judging from exposed crustal sections, granulite-facies and igneous rocks should be dominant in the lower continental crust in most shield areas (Fountain and Salisbury, 1981). Such rocks will also constitute a major part of the rocks cycled into root zones. These rocks, crystallized in the presence of low H_2O-activity fluids (CO_2 or high salinity aqueous brines), contain only small amounts of hydrous minerals. Several studies (Walther, 1994; Wayte et al., 1989; Rubie, 1990; Austrheim, 1987) have shown that granulites and igneous rocks will not react when exposed to eclogite-facies conditions unless aqueous fluids are introduced. A minimum overstepping of 5 kbars is possible at a temperature as high as 700°C and, if fluids are not available, dry rocks in the deep crust will remain metastable compared to their hydrated equivalents (Wayte et al., 1989). Therefore the metamorphic status and the geodynamics of crustal roots will be highly dependent on the availability and the composition of fluids.

On this excursion, we will address the processes in crustal root zones and the petrophysical properties associated with the granulite–eclogite transition based on field observations from an exposed high-pressure terrain of the Scandinavian Caledonides. The role of fluids will be a key issue.

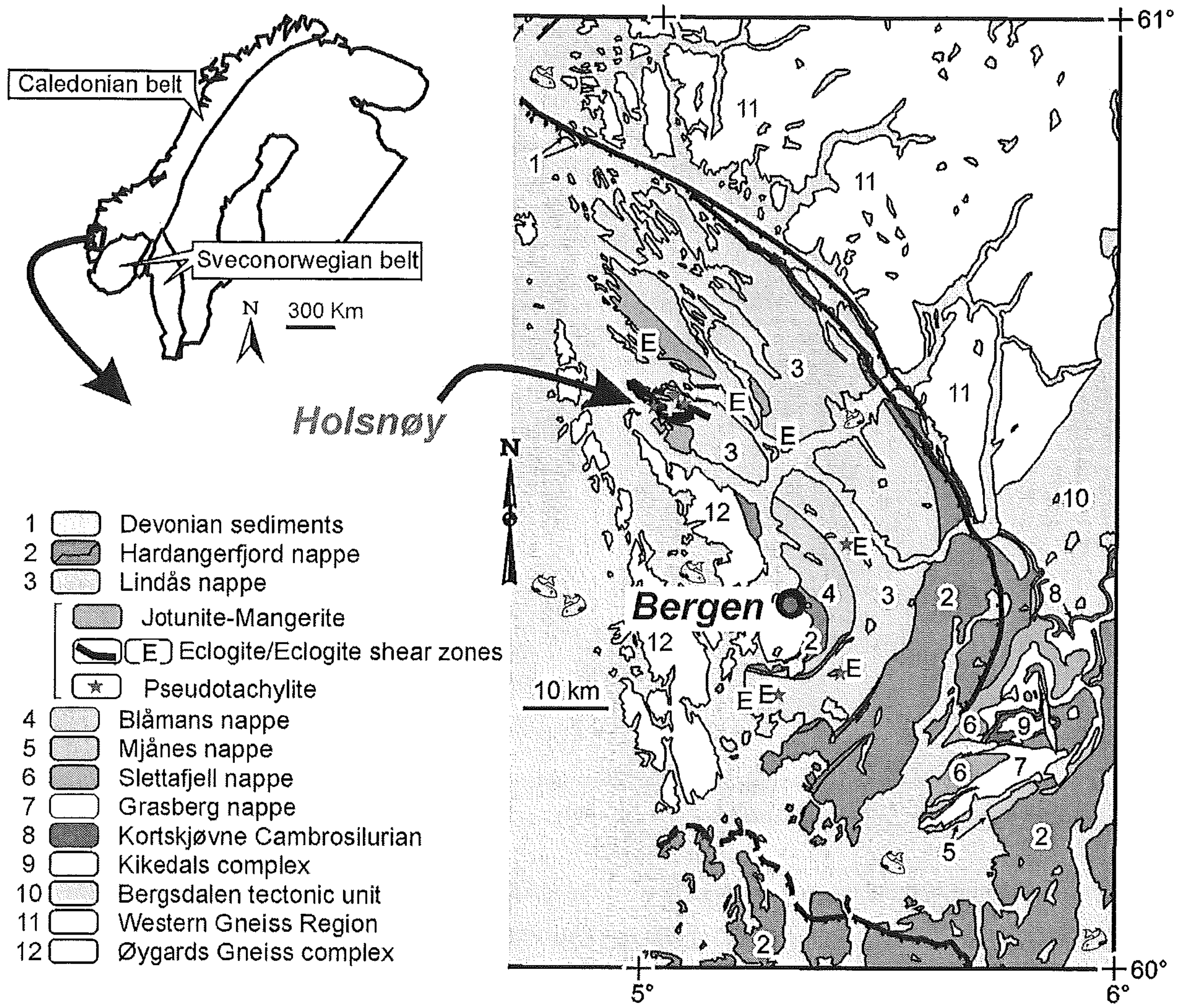


Fig. 2.1: Geologic map of the Bergen Arcs (modified from the tectonic map of the NGU 1/250000 Bergen sheet by Ragnhildsdveit (1997).

2. Metamorphism in the deep crust: The Bergen Arcs example

The Bergen Arcs (Fig. 2.1) are a series of arcuate Caledonian thrust sheets and nappes centered around Bergen, western Norway. The nappes contain both Precambrian orthogneisses and Lower Paleozoic complexes dominated by oceanic ophiolitic/island-arc lithologies. The most important unit, from both a geological and volumetric standpoint, is the Lindås Nappe, an anorthosite-mangerite-granite-charnockite complex. It originated as a layered mafic intrusion and was metamorphosed to granulite facies at 800-900°C and 1.0 GPa (a depth of approximately 30 km) in late Grenvillian time (Austrheim and Griffin, 1985). U-Pb zircon dates (Boundy et al., 1997) place peak granulite-facies metamorphism at about 945 Ma (Fig. 2.2). This metamorphism produced anhydrous mineral assemblages dominated by pyroxenes and garnet in addition to feldspar and quartz in the rocks of the complex. The granulites are strongly banded; plagioclase-rich layers alternate with mafic layers or coronas. Sm-Nd mineral isochrons (Cohen et al., 1988; Burton et al., 1995) indicate that the complex cooled to 600 to 700°C by 900 Ma (Fig.2.2). Kühn et al. (2000) obtained Rb-Sr phlogopite ages ranging between 835-882 Ma for granulite facies assemblages in lherzolites. This may indicate that the complex cooled below ca. 450°C (the assumed blocking temperature for this mineral and isotopic system) in pre-Caledonian time. However, this temperature is well below the temperature calculated for the eclogite facies event (670°C) and the authors therefor argued, in line with the field evidences, that the blocking of the isotopic is fluid controlled.

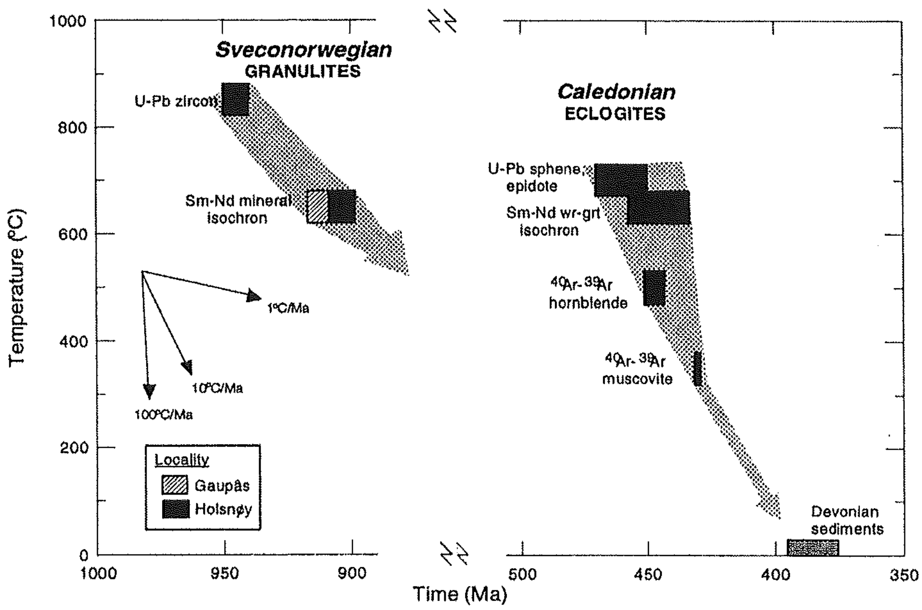


Figure 2.2: Temperature-time plot illustrating thermochronologic history of granulite facies metamorphism (left) and eclogite facies metamorphism (right) for the Lindås Nappe (from Boundy et al., 1997).

Crustal thickening and subduction during the mid-Paleozoic Caledonian orogeny brought the Lindås Nappe granulite-facies rocks into eclogite-facies conditions (Austrheim and Griffin, 1985; Austrheim, 1987). Attempts to determine the age of the eclogite facies metamorphism by various isotope systems have given ages ranging from 420 – 460 Ma (Boundy et al. 1995, 1997; Bingen et al. 1998). Boundy et al. (1997) interpreted these ages to mean that the eclogite facies metamorphism (>1.5 GPa and temperatures of about 650-700°C) occurred around 450-460 Ma and was followed by rapid cooling (Fig. 2.2). An alternative explanation is that the spread in ages reflects fluid activity over a time span of ca. 40 Ma.

However, because of retarded kinetics owing to the anhydrous nature of the granulites, conversion to eclogite-facies conditions was incomplete as is well illustrated on Holsnøy where granulites intermingle with eclogites, in shear zones and veins, at all scales (Fig. 2.3). Where eclogitization occurred, the plagioclase- and pyroxene-rich granulitic assemblage was replaced by a high pressure, hydrous eclogite assemblage of omphacite, phengitic muscovite, and zoisite/clinozoisite.

Because deformation and fluid, not chemical heterogeneities of the protolith, controlled the locations of eclogitization and because the eclogite mineral assemblage includes hydrous phases, Austrheim (1987) and Klaper (1990) concluded that infiltration of fluids during metamorphism was essential for the development and stabilization of the eclogite mineralogy. Outside of shear zones and veins, in zones where fluid did not penetrate, granulite-facies rocks persisted in a metastable state under eclogite-facies conditions. Jamtveit et al. (1990) used phase equilibria to deduce that the infiltrating fluid was H_2O-rich. Andersen et al. (1990) initially proposed that the eclogites interacted with a mainly N_2-rich fluid but through subsequent fluid inclusion work found evidence for a compositionally complex fluid that evolved as eclogitization proceeded (Andersen et al., 1991). At peak metamorphic conditions infiltration of a H_2O-rich fluid led to partial melting and the development of felsic extension veins consisting of plagioclase, phengite, quartz, clinozoisite, and minor calcite. These veins are found in the central parts of some shear zones (Andersen et al., 1991).

Arrest of eclogitization due to rapid exhumation and/or limited amount of fluids available to facilitate reactions allowed for preservation of a complex that provides an excellent view of the processes associated with subduction of old "dry" lower crust. These processes can best be studied on the northern part of Holsnøy, but the same relationships, including the pseudotachylytes (see below), have been observed at Gaupås some 10 km east of Bergen (Fig. 2.1)

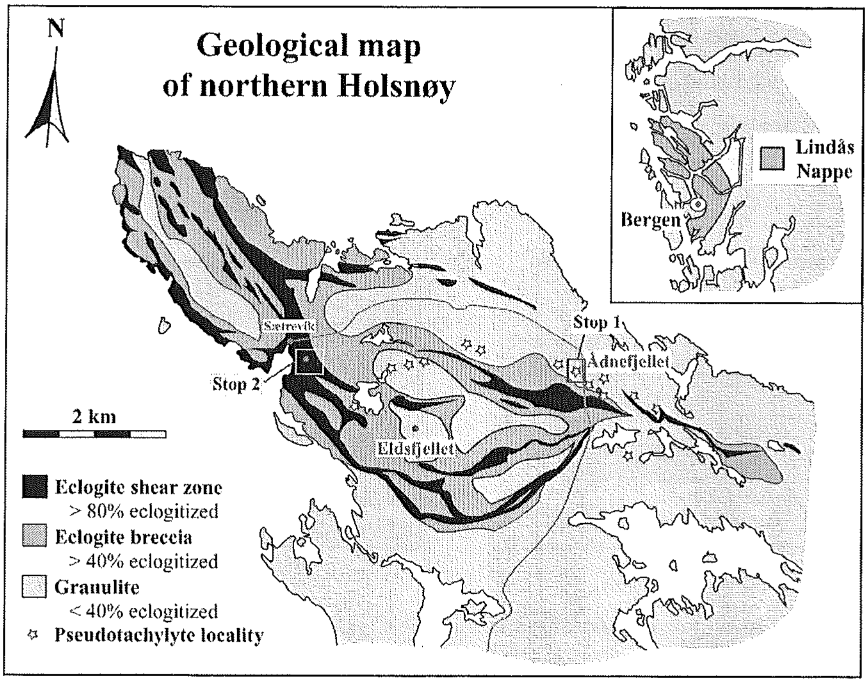


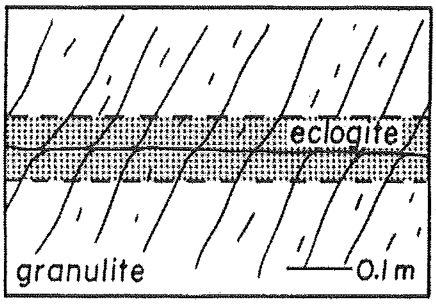
Figure 2.3: Geologic map of Holsnøy (modified after Austrheim et al., 1996; Boundy et al., 1997). Map modified by M. Lund, A. Kühn and M. Erambert.

3. Stages of eclogitization: Hundskjeften, Holsnøy

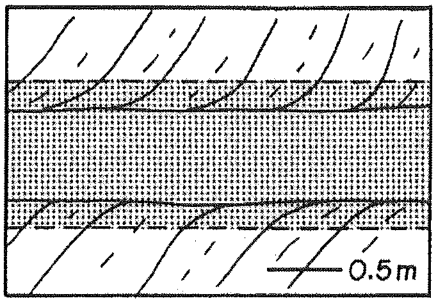
Drive to Sætrevik, northwestern Holsnøy. From the local shop at Sætrevik we will walk toward Hundskjeften and make several localities near the summerhouse at Hundskjeften. This pleasant summerhouse has become a popular base for field activities over the last decade.

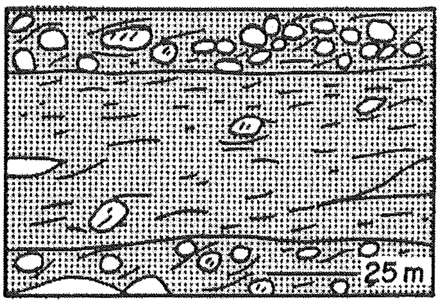
The eclogites on Holsnøy occur in three structural settings (e.g., Austrheim and Mørk, 1988; Boundy et al.; 1992; Austrheim et al., 1997), all of which are well-exposed in this area:

- cm to dm thick fractures and veins (Austrheim, 1987, 1989; Jamtveit et al., 1990; Klaper, 1991) that contain coarse-grained euhedral eclogite facies minerals including abundant hydrous phases (phengite, clinozoisite, amphibole); granulites adjacent to the veins were converted to eclogite to varying degrees as fluids migrated from the veins into the protolith [Localities A-1 and H-1].

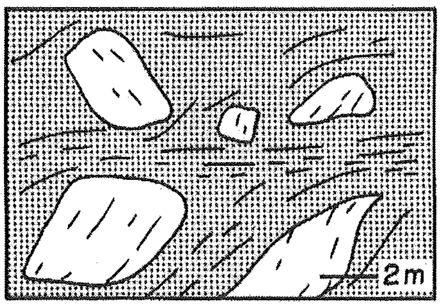


- in high-strain zones that range in thickness from centimeters to hundreds of meters (up to 100 m thick); the thicker zones can be traced laterally for several kilometers based on mapping by Boundy (Austrheim, 1989; Boundy et al., 1992; Boundy, 1995); eclogites exhibit pronounced mineral lineations and foliation [Locality H-5].





- eclogite-facies breccias in which rotated angular blocks of granulite (dm to m across) are surrounded by foliated eclogites; breccias are often bound the major high-strain zones (below) and are composed of about 40% eclogite [Localities H-2 and H-3]



3.1. Locality H-1: Jarlsberg cheese outcrop - mineralogical, geochemical, and petrophysical changes across metasomatic reaction fronts

Outcrops at this area present many excellent examples of the incipient metamorphic overprint of the Grenvillian granulite-facies complex by eclogite-facies assemblages. The cm-scale banding defined by alternating dark pyroxene/garnet and light plagioclase layers, evident in this outcrop, represent the old Precambrian structure. This structure is transected by bands of eclogites (Fig. 3.1) containing omphacite, garnet, phengite and clinozoisite. The eclogite bands are developed around central veins with phengite, quartz, and, locally, omphacite. The transformation of the granulites composed of anhydrous minerals to eclogites with abundant hydrous phases required addition of fluids to allow reactions in the anhydrous granulite facies assemblage. The central veins in the eclogite band were fluid channels from which the fluid migrated into the granulite. Note that the fronts between the dark eclogite and the light granulite are straight on outcrop scale. This is in distinct contrast to the relationships we will see at locality H-4 (see below) where the eclogite forms fingers in the granulite.

These outcrops provide an excellent opportunity to observe the geochemical, mineralogical, and petrophysical changes associated with the granulite- to eclogite-facies transition. This particular locality is but one of many that we and others drilled to investigate the processes and products of this mode of eclogitization, a mode now commonly recognized in eclogites developed in continental and oceanic rocks (e.g., Engvik et al., 2000; Pennacchioni, 1996; Zhang and Liou, 1997). We cut cores along individual bands of granulite-facies rocks that could be traced into the zone where those bands were eclogitized (Fig. 3.2). Here, we see that eclogitization has not completely obliterated the old granulite-facies structures - ghost-like granulite-facies banding can be traced into the eclogite-facies zone. This gives us confidence that we sampled the granulite-facies protolith and its eclogite-facies equivalent. The major results of our study are summarized below.

Figure 3.1: 10 cm wide, straight eclogitization fronts around eclogite-facies vein cutting granulite-facies anorthosite (Ådnefjellet, see also Location A-1).

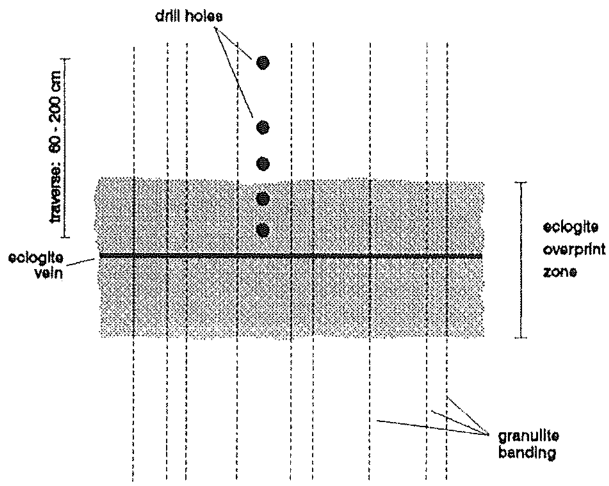


Figure 3.2: Schematic diagram of a typical sample area. A narrow vein (a few cm or less) with euhedral eclogite-facies minerals is bordered by a halo of partially to nearly fully eclogitized granulite. The granulite banding, defined by alternating plagioclase-rich and mafic layers, can be traced through the zone of eclogitized granulite. In many areas, the granulite banding is oblique to the veins, but samples were taken only in areas where the granulite banding was nearly perpendicular to the veins, as shown. For each traverse, three to eight core samples were drilled within a single compositional band beginning in uneclogitized granulite and ending in the overprint zone adjacent to the vein, for a total linear sampled distance of less than two meters.

3.1.1. Geochemistry

Major and trace element geochemical analyses of cores were carried out by the geochemistry group at Washington University and published in Rockow et al. (1997). They compared chemical compositions of granulite and its undeformed eclogitized equivalent adjacent to veins in locations where a single band of granulite could be traced and sampled as it approached the vein (such as you see at this locality). Nine separate granulite-eclogite transition zones, including this locality, located at veins in anorthositic, jotunitic, and gabbroic protoliths were analyzed. For each transition, no compositional difference between the average granulite and average eclogite composition was found at the 90% confidence level except for loss on ignition, which was consistently significantly higher in the eclogite samples. Although not significant at the 90% confidence level for any single traverse, the average eclogite concentrations of SiO_2, Na_2O, Cs, As, and Br exceed the average granulite concentrations for eight or all nine of the traverses. For most traverses, statistical analysis of the data limits any gain of SiO_2 in the eclogites to no more than a few relative percent. Other than the introduction of volatile substances, presumably an H_2O-rich fluid, eclogitization associated with vein formation was essentially isochemical.

3.1.2. Mineralogy

The metasomatic fronts reveal the progressive mineralogical changes associated with the eclogitization process. In an ongoing project we used scanning electron microscope and electron microprobe techniques to quantify the detailed mineralogical changes across the reaction fronts (Fountain, Swapp, Austrheim, and Rey, in progress) using the same cores used for the geochemical (see above) and petrophysical (see below) investigations (Fig. 3.3). The granulite-facies protoliths are generally coarse-grained (1-3 mm) with granoblastic textures. They consist of diopside (a calcic clinopyroxene), plagioclase feldspar, and garnet (Fig. 3.3). The more mafic granulites also contain an Fe-Ti oxide (ilmenite-magnetite intergrowth) and orthopyroxene in the form of exsolved bands within clinopyroxene. Granulite samples taken nearest the eclogite overprint zone show signs of incipient eclogitization; e.g., very fine grained reaction rims around plagioclase and between pyroxene and opaque phases, and feldspar clouded with needles of mica, kyanite, and zoisite (see Austrheim and Griffin (1985) and Mattey et al., (1994) for detailed descriptions of progressive mineralogical changes associated with eclogitization). The decrease in volume of the granulite-facies mineralogy and growth of the new eclogite-facies minerals approaching the eclogite band is illustrated in Fig. 3.3, an example of a traverse for an anorthositic gabbro composition protolith. Plagioclase not only diminishes in abundance with increasing reaction progress, but the remaining plagioclase becomes increasingly more sodic because calcium is taken up by clinozoisite.

The samples from the eclogite overprint zones are fine grained (<1 mm) and consist of clusters of omphacite grains partially altered to symplectite (plagioclase, jadeite-poor clinopyroxene +/- opaques) and fine-granular regions of phengitic muscovite, zoisite, quartz, and kyanite (Fig. 3.3). Proportions of these phases vary depending on the composition of the protolith and degree of reaction progress. Two generations of garnet are present. Granulite-facies relict garnet is coarse relative to the eclogite mineralogy and has corroded rims surrounded by amphibole. Second generation garnet is smaller, more euhedral, and enriched in Ca and Fe relative to the relict garnet (see Austrheim and Griffin, 1985). Eclogitization may not be complete in the eclogite-facies overprint zone as illustrated for traverse BA-44, where the eclogite still contains plagioclase, diopside, and relict garnet (Fig. 3.3). The degree of eclogitization of the samples from traverse BA-44 is illustrated in the upper portion of Figure 3.3 where we show the fraction of the sample composed of eclogite-facies minerals.

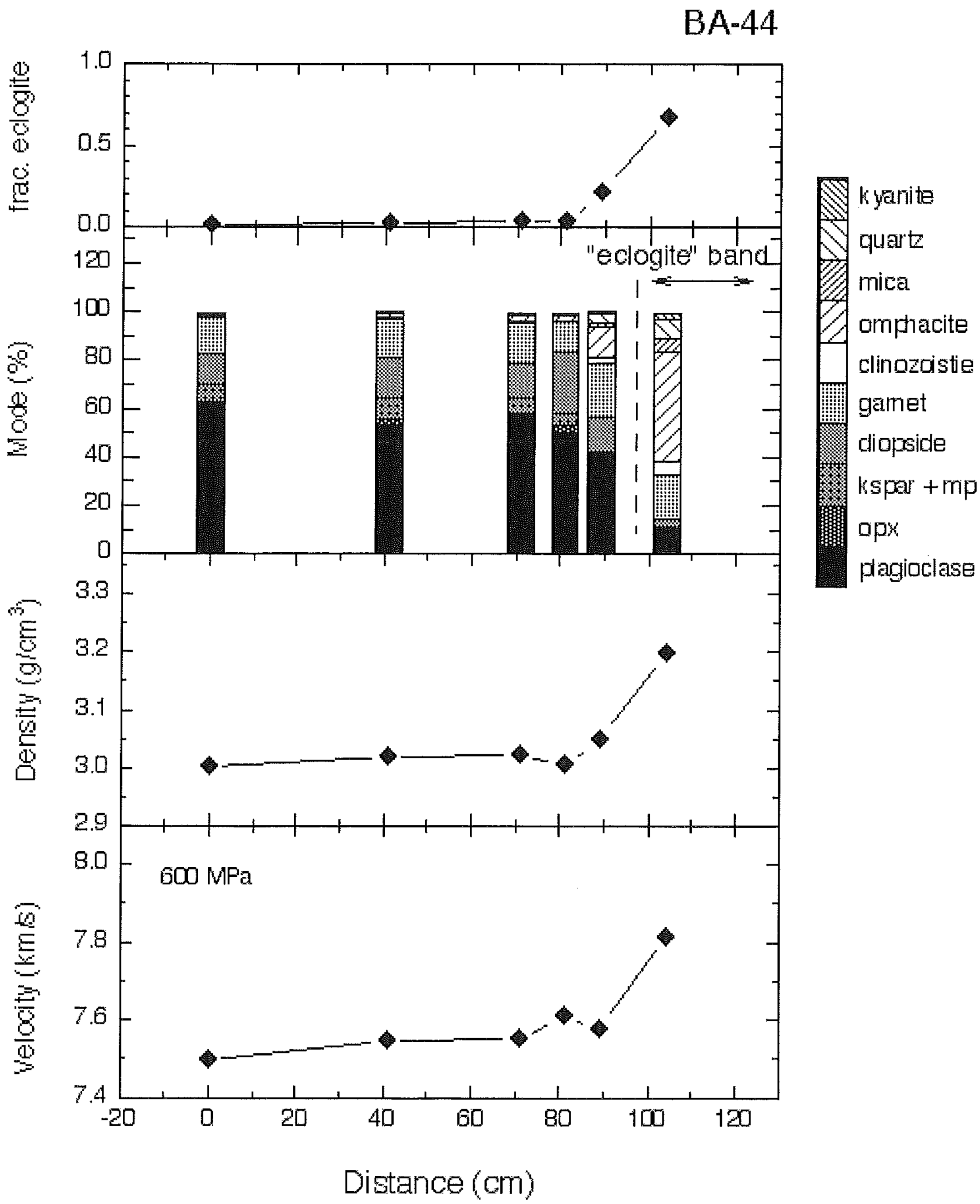


Fig. 3.3: Variation of degree of eclogitization (frac. eclogite), modal mineralogy, density, and compressional wave velocity across a granulite-facies to eclogite facies transition zone. Example if for traverse BA-44, an anorthositic gabbro protolith composition (from Fountain, Rey, Swapp, Austrheim, unpublished)

3.1.3. Physical Properties

These reaction fronts allow us to investigate the relationship between metamorphic reactions and petrophysical properties (Fountain, Rey, Grassi, and Austrheim, in progress). We measured density and compressional wave velocity (V_p) for each core and shear wave velocity (V_s) for selected cores. All velocities were determined at pressures up to 600 MPa confining pressure using the pulse transmission method (see Fountain et al. (1994) for complete description of methods).

Velocity and density vary systematically with the degree of eclogitization. For traverse BA-44 (Fig. 3.3), both V_p and density change imperceptibly approaching the eclogite zone and then increase dramatically over a relatively small distance (<30 cm). Thus, at outcrop scale, the velocity contrast associated with the granulite- to eclogite-facies transition is sharp. This conclusion, however, may not apply to seismic-scale dimensions, as is evident from localities H-1, H-2, and H-5 as well as inspection of Figure 2.3. Although some low density-low V_p phases (quartz, mica, and sodic plagioclase) increase in abundance due to the eclogite-forming reactions, most of the new minerals are high density-high-V_p. Some changes do not follow intuition. For example, the volumetric increase of omphacite relative to diopside has little effect on density (both phases have similar densities) yet has a significant effect on velocity. The progressive decrease of plagioclase causes a decrease in V_p/V_s or decrease in Poisson's ratio, assuming these rocks are isotropic, (Grassi, 1996). The variation of velocity and density correlates well with the degree of eclogitization for traverse BA-44 (Fig. 3.4). However, we have not yet completed all the mineralogical analysis, so the details of this relationship may change on further work.

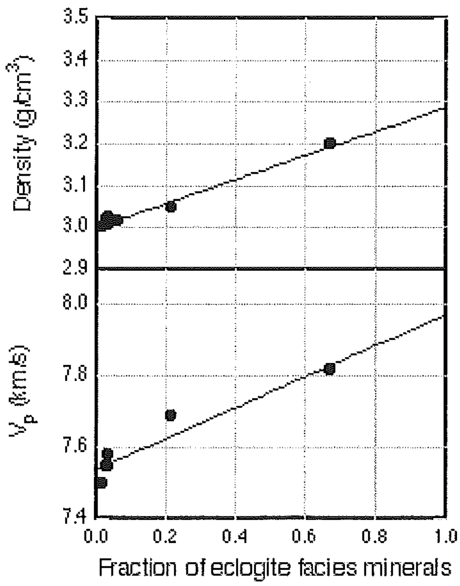


Figure 3.4: Variation of Vp and density as a function of degree of eclogitization for the BA-44 traverse depicted in Fig. 3.3 (Fountain, Rey, Swapp, and Austrheim, unpublished data).

3.2. Locality H-2: Eclogite breccia, part 1 - mixture of granulites and eclogites

The term eclogite breccia, as defined by Austrheim and Mørk (1998) and well displayed in the cliff in front of us, refers to zones of foliated eclogite that enclose and wrap around blocks of granulite-facies rocks, which occur as angular, lensoidal blocks typically less than 5 m across. Such mixtures of granulites and eclogites outcrop over an area of 1000 m^2 and are a characteristic and dominant unit on northern Holsnøy. Breccias, containing about 40% eclogite, are commonly developed adjacent to major eclogite-facies shear zones (see Locality H-5) such as this locality where eclogite breccias bound the Hundskjeften shear zone. The presence of such lithologies in the deep crust is surprising and, unfortunately, gives us one more degree of freedom to interpret seismic velocities measured in present day deep crust and upper mantle. Austrheim and Mørk (1988) assumed that the velocity would be a linear function of the volumetric ratio of the two rock types. But it is possible that an abrupt change may occur at a certain percentage of eclogitization, that depends on the seismic wavelength and the spatial distribution of eclogites and granulites at depth. If a marked increase in velocity occurs at a low degree of eclogitization, the result will be a crust with a high velocity and a relatively low density.

3.3. Locality H-3: Eclogite breccia, part 2 - rheology of deep crust

The rheology of the crust and upper mantle are generally regarded as functions of mineralogy and geothermal gradient. Most published lower crust/upper mantle depth–viscosity profiles are based on models where the rheology of the lower crust is controlled by feldspar and the upper mantle by olivine. The rheological properties of eclogite-facies rocks and the change in properties associated with the granulite–eclogite-facies transition are generally not considered. This outcrop and locality H-2 demonstrate that the eclogites have low viscosities compared to their granulite-facies protolith.

Here you will observe rotated blocks of granulite floating in the eclogite matrix. The granulite-facies banding is completely disrupted in the eclogitized matrix. The remnants of the mafic bands are recognized as mafic boudins within the eclogite. The eclogite gives the impression of having flowed as a viscous fluid.

3.4. Locality H-4: Eclogite fingers outcrop - deep crust hydration processes

Introduction of fluids into dry granulites is a prerequisite for eclogitization in the Lindås Nappe. The mechanism by which dry impermeable deep crust is hydrated remains controversial. On a large scale, fluid introduction requires the presence of highly permeable channelways, such as faults or fractures. At Localities A-1 and H-1 we observed that eclogitization fronts are parallel to a central veins and extend only 5 cm into the wall rock. At this locality we observe that the front separating dry and wet rock is morphologically unstable with the fingers extending several m into the granulites. Microtextures in the eclogite fingers demonstrate that relict granulite-facies garnet is fractured, suggesting that microfracturing allowed hydration (Fig. 3.5). Field relationships, microtextural data, and a simple network model indicate that the volume change associated with the volatilization reaction led to fracturing and that transport of fluid into the initially dry rock was accelerated by perturbation of the local stress fields caused by the associated volatilization reactions (Jamtveit et al., in review). Further, the morphology of the reaction fronts was found to depend on the anisotropy in the external stress field.

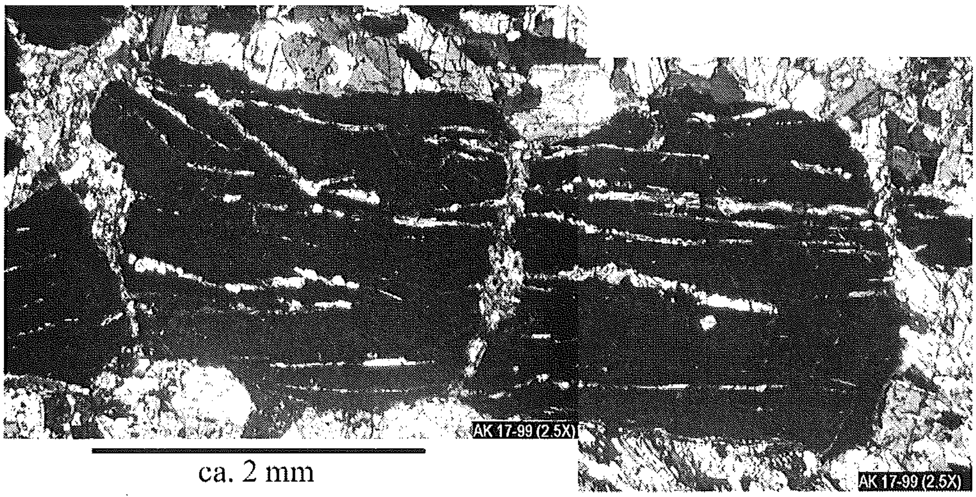


Figure 3.5: Microphotograph (crossed nicols) of fractured garnet from an eclogite facies "finger". The fractures are sub-parallel oriented and filled with eclogite facies minerals, omphacite, kyanite, phengite and amphibole.

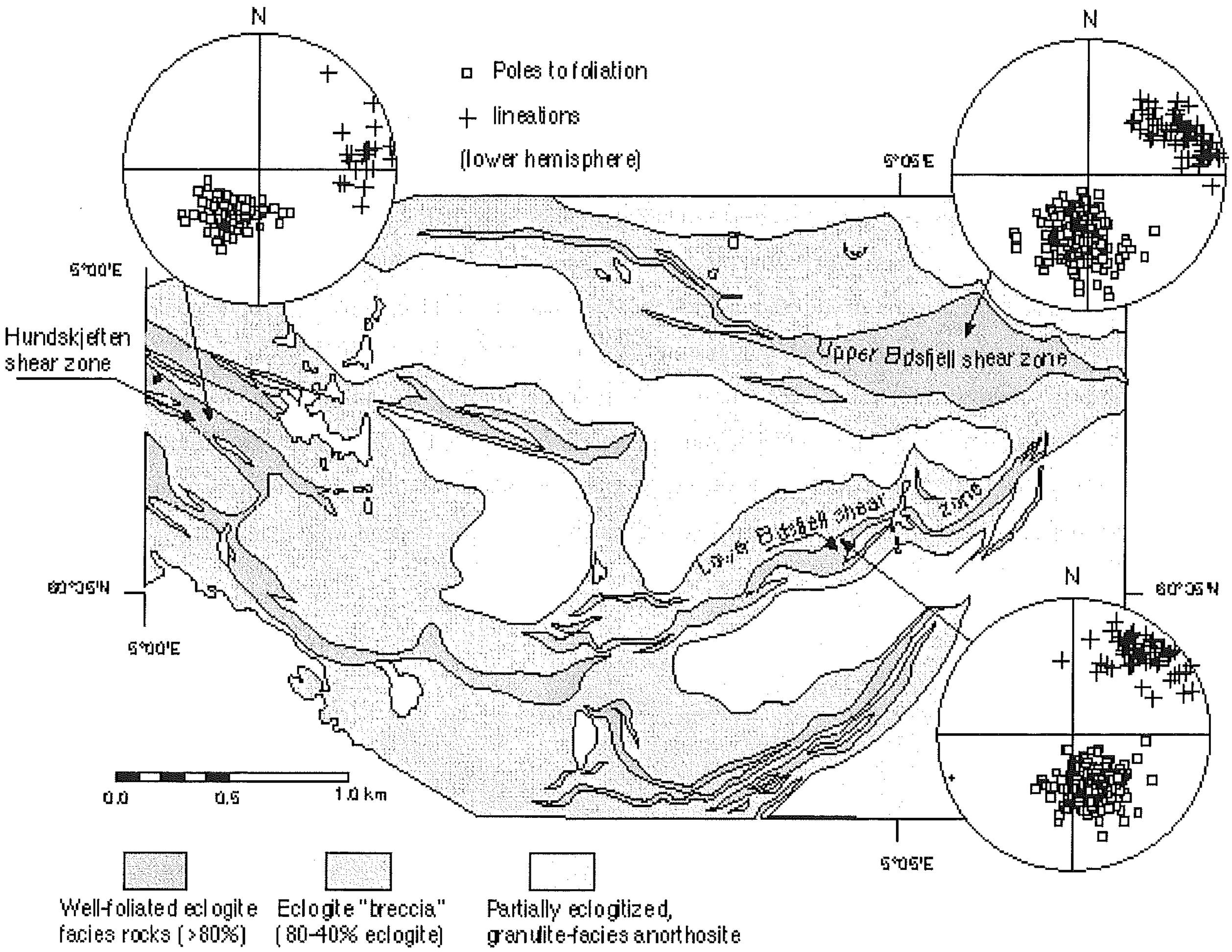


Figure 3.6. Geologic map of a portion of Holsnøy (from Boundy et al., 1992; Fountain et al., 1994; Rey et al., 1998)) showing eclogite-facies shear zones and structural data (squares = poles to foliation; + = lineations). Outcrop widths are not indicative of true thickness because topography is not shown.

3.5. Locality H-5: Hundskjeften shear zone – structure and petrophysical characteristics of eclogite-facies high strain zones

The Hundskjeften shear zone is one of several major shear zones transecting the granulites on Holsnøy (Fig. 3.6). Within the shear zones the eclogitization process is nearly complete and only remnants of granulite-facies garnets can be found. Shear zones are surrounded by eclogite breccias (see Localities H-2 and H-3) that consist of granulite- and eclogite-facies rocks in about equal proportion. Breccias are composed of zones of foliated eclogite that enclose and wrap around blocks of granulite-facies rocks, which occur as angular, lensoidal blocks typically less than 5 m across. These breccia zones are flanked above and below by granulite-facies anorthosites and gabbroic anorthosites.

The eclogite-facies shear zones (the dark horizontal band evident in this outcrop) exhibit a pronounced mm-scale layering defined by alternating omphacite/garnet- and kyanite/zoisite-rich layers and a strong shape-fabric foliation defined by aligned omphacite, kyanite, zoisite, and phengite. Shear zone foliations generally have a north to northeast dip of about 10° to 30° (Fig. 3.6). Lineations in the shear zones are defined by rod-shaped aggregates of omphacite and garnet, elongate relict corona structures, and mineral lineations. Rey et al. (1998) identified a variety of macroscopic and microscopic kinematic indicators in the shear zones: sigmoid-shaped mineral clusters, S and C planes, asymmetric pressure shadows around garnets, tiling microstructures, asymmetric crystallization tails. Rey et al. (1998) interpreted these as evidence of top to the east northeast sense-of-shear for the Hundskjeften shear zone. Although these features indicate that the shear zones are normal faults in their present orientation, shear zone orientation during metamorphism and deformation was likely different.

Universal-stage measurements of omphacite crystallographic preferred orientation (CPO) show that omphacite *b*-axis maxima are approximately normal to the foliation and *c*-axis girdles lie within the foliation plane (Boundy et al., 1992). Weak *c*-axis maxima within the girdles are approximately parallel to the lineation. An example from the Eldsfjell shear zone is shown in Fig. 3.7.

The geometry of the shear zones and anticipated large contrast in physical properties across them drew us to these shear zones in order to construct velocity profiles through deep crustal structures. We focused on the nearby shear zones on Eldsfjell, taking advantage of topography to build a traverse through two thick shear zones. Compressional wave velocity measurements to 600 MPa for shear zone samples are reported in Fountain et al. (1994). We found that V_p in shear zone samples ranges from 8.3 to 8.5 km/s, values generally higher than reported from reaction fronts presumably because shear zone eclogites tend to be more completely reacted. Velocities within shear zones are not uniform (Fig. 3.8) due to the effects of intra-shear zone compositional variations, degree of reaction (see above), and anisotropy (see below). The velocity and density contrast between strained eclogites is large as illustrated by one of our hypothetical profiles in which we show a one-dimensional profile through the two Eldsfjell shear zones bounded by granulite-facies rocks (Fig. 3.9). More realistically, because the shear zones are bounded by eclogite breccias, the velocity, density, and impedance contrasts across the shear zone boundaries may be less (Fig. 3.10). However, the granulites within the breccias are matrix-supported blocks within a high-velocity matrix. How do P- and S- waves propagate through a unit like this?

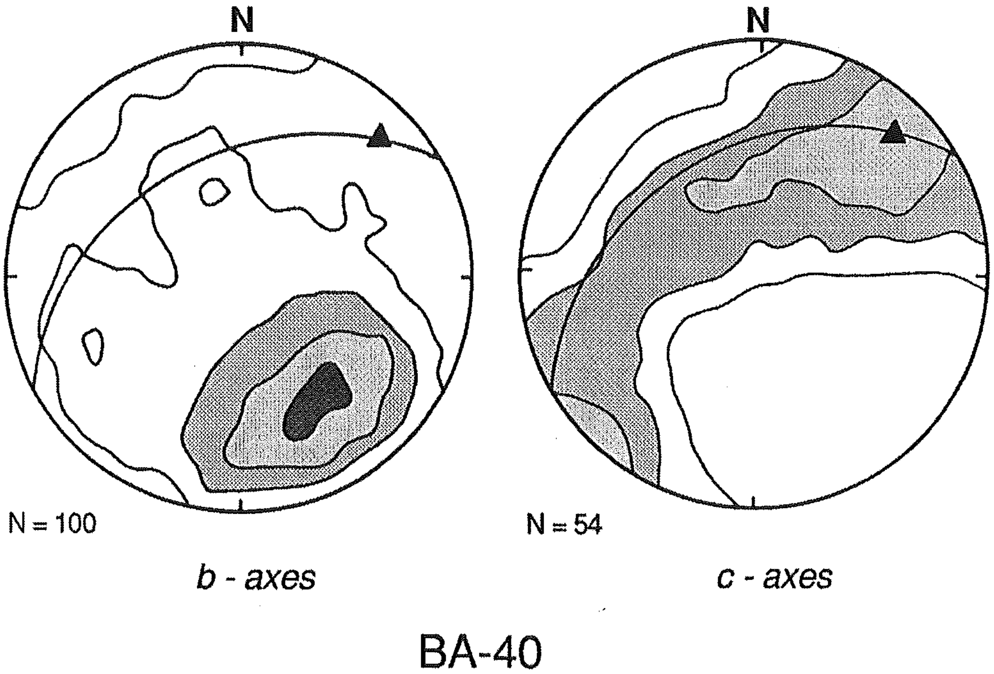


Figure 3.7: CPO for omphacite from eclogite-facies shear zones. Lower hemisphere projection (from Boundy et al., 1992).

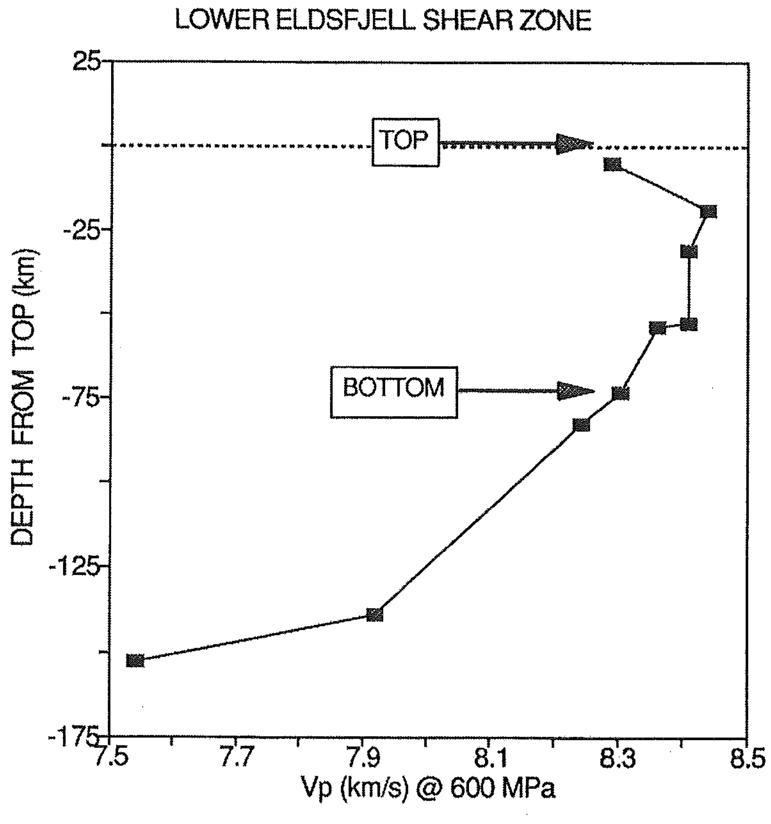


Figure 3.8: P-wave velocity profile for an eclogites facies shear zone.

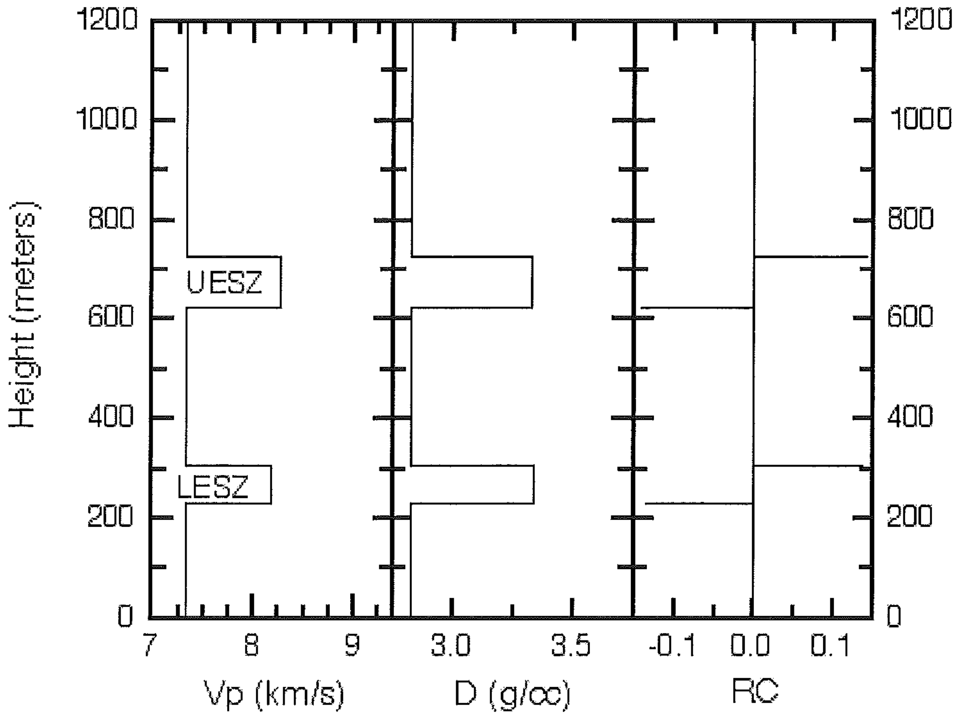


Figure 3.9: Variation of Vp, density, and reflection coefficients (RC) through the two Eldsfjell shear zones for the case where the shear zones is bounded by granulite-facies rocks. From Fountain et al. (1994).

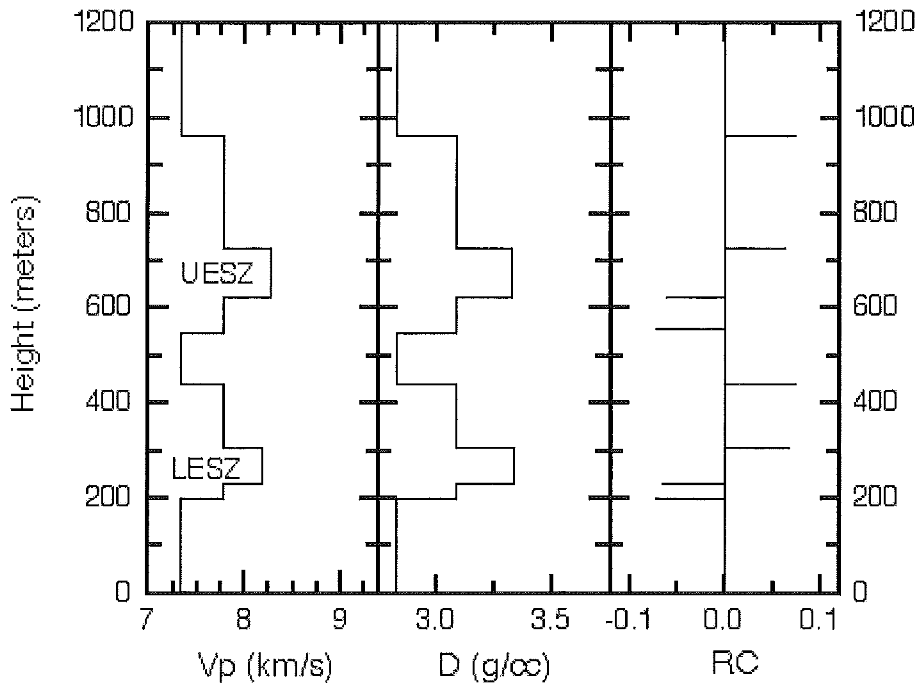


Figure 3.10: Variation of Vp, density, and reflection coefficients (RC) through the two Eldsfjell shear zones for the case where the shear zones is bounded by eclogite breccias. From Fountain et al. (1994).

Deformed eclogites from high strain zones exhibit P- and S-wave anisotropy, anisotropy that can, in some cases, are comparable to that reported from laboratory measurements on ultramafic rocks. P-wave anisotropy for Eldsfjell shear zone samples ranges from 1 to 7% (based on 3 minicores cut parallel to the main fabric elements) and, when present, generally exhibits a transversely isotropic pattern. Minimum V_p is generally normal to foliation and, for samples for which omphacite CPO is known, parallel to *b*-axis maxima. The fast propagation direction is parallel to foliation and, for samples for which omphacite CPO is known, lies within the *c*-axis girdles. Mauler et al. (2000) report similar P-wave anisotropy ranges and patterns are reported for non-retrogressed eclogites collected from the Monviso ophiolite complex, western Alps.

Our assumption, and the usual assumption in the literature, is that omphacite CPO is most likely responsible for anisotropy in eclogites. Single-crystal measurements of omphacite elastic constants (Bhagat et al., 1992) show that omphacite is indeed strongly anisotropic. However, the V_p pattern for omphacite differs significantly from orthogonal pattern of olivine because omphacite is monoclinic. Consequently, the velocity minimum and maximum do not bear a simple relationship to the crystallographic axes. Because of this monoclinic symmetry, proper modeling of anisotropy of an omphacite aggregate with a strong CPO is not easily accomplished from universal stage measurements as determination of the sign of the crystallographic axes is not possible. Thus, we (Grassi, Abalos, and Fountain, in progress) have been unsuccessful in our attempts to model anisotropy patterns for strained eclogites from Holsnøy and Cabo Ortegal from standard petrofabric data. Mauler et al. (2000) used electron backscatter diffraction methods and was able to construct complete CPO patterns for the Monviso eclogites but failed to find a clear quantitative relationship between omphacite CPO and P-wave anisotropy. The Monviso eclogites are similar to the Holsnøy eclogites in that other strongly anisotropic phases (e.g., clinozoisite, mica, kyanite – the elastic constants of some of these phases have not been measured or the CPO is not usually/easily determined), compositional layering, and shape fabric may also strongly influence P-wave anisotropy. It also important to stress that our studies and those of other researchers have only measured velocities parallel to the main macroscopic fabric elements of strained eclogites. Because of the importance of low-symmetry minerals in these rocks, velocity maxima and minima may not correspond to these macroscopic elements. This is partly born out by Mauler et al. (2000) and some calculations for Holsnøy and Cabo Ortegal samples.

Strained eclogites also exhibit varying degrees of shear wave splitting (or shear wave birefringence). Grassi (1996) reported splitting measurements for four deformed eclogites from the Eldsfjell shear zones for which omphacite CPO results were available (Boundy et al., 1992) and one undeformed eclogite. Splitting for the undeformed eclogite was very low and variable for the strained eclogites (Fig. 3.11). Maximum splitting was observed for propagation directions within the foliation plane and perpendicular to lineation. A few samples showed splitting comparable to that observed in laboratory measurements of ultramafic rocks.

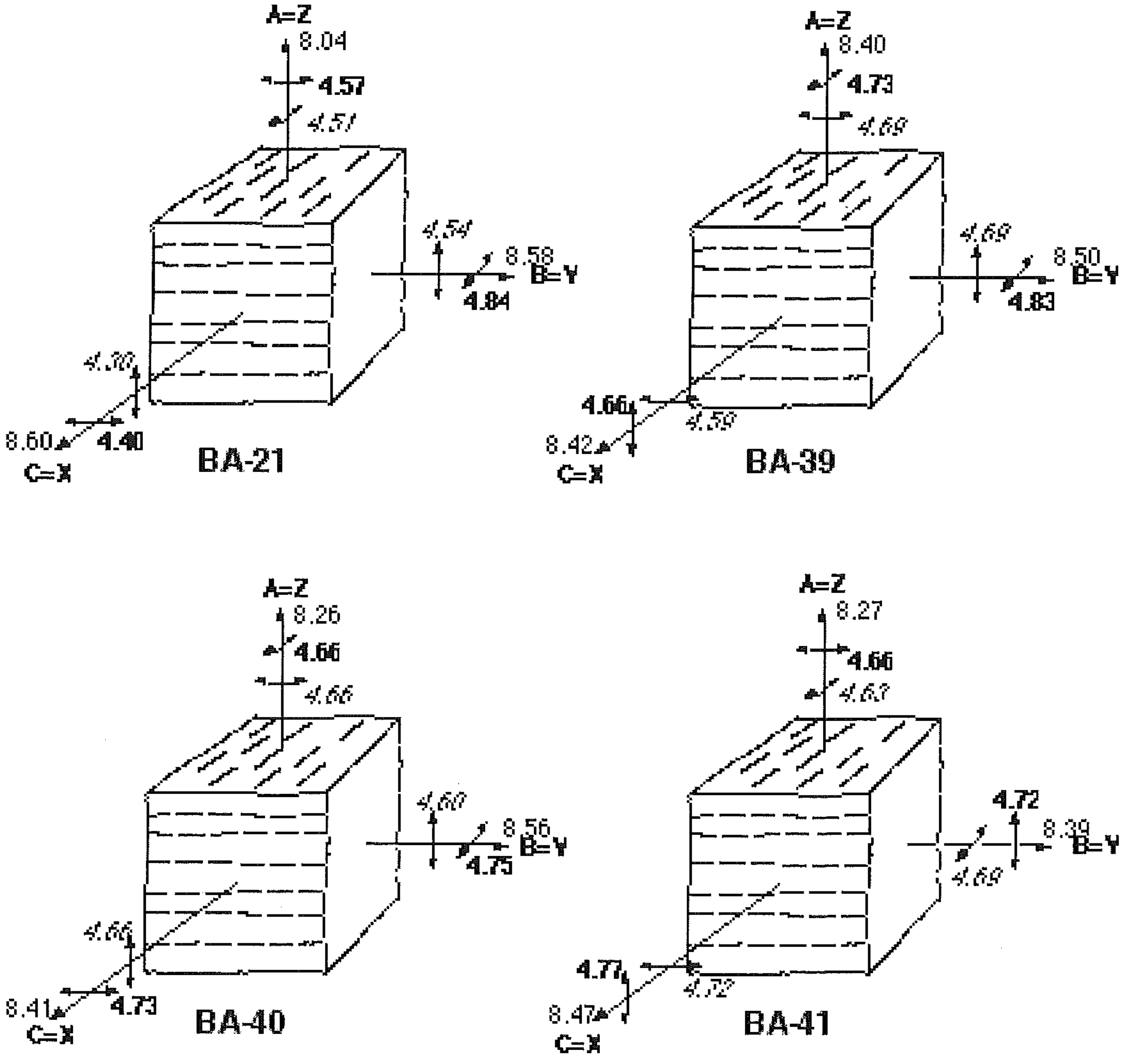


Figure 3.11: P-wave velocities, S-wave velocities, and S-wave polarization directions for samples from the eclogite shear zones. Form Grassi (1996).

4. Evidence for fluid-mediated eclogitization and deep crustal earthquakes: Ådnefjell west, Holsnøy

Drive to the area between Ådnefjell and Eldsfjell. We will take a few hundred meters walk, partly through bushy terrain, into an area displaying the incipient stages of eclogitization of the granulite-facies gabbroic anorthosite. We will visit three localities in this area.

4.1. Locality A-1: Eclogite-facies veins and reaction fronts outcrops – the beginnings of eclogitization

The low degree of eclogitization here permits observation of pre-Caledonian structures and mineralogy of the granulite-facies complex. A cm-scale banding defined by alternating dark pyroxene/garnet and light plagioclase layers represent the old Precambrian structure. This structure is transected by bands of eclogites (Fig. 3.1) containing omphacite, garnet, phengite and clinozoisite. The eclogite bands are developed around central veins with phengite, quartz, and, locally, omphacite. The transformation of the granulites composed of anhydrous minerals to eclogites with abundant hydrous phases requires addition of fluids. The central veins in the eclogite band represent the fluid channels from which the fluid migrated into the granulite. Note that the fronts between the dark eclogite and the light granulite are straight on outcrop scale. This contrasts with the relationships at Hundskjeften (Locality H-4) where the eclogite forms fingers in the granulite.

4.2. Locality A-2: Eclogite-facies pseudotachylytes, part 1 – evidence for deep earthquakes

Here, the dry granulite is shattered by cm thick dark flinty veins. At outcrop scale these veins have all the characteristics of pseudotachylytes: spatial relationships to faults, internal flow banding, and intrusive relationships with the wallrock. Note how the small blocks of granulites are rotated and how the pseudotachylyte veins can be followed into the adjacent eclogite where their color changes from black to green. Note also that displacement occurs along the thinner veins while the thicker veins represent intrusive frictional melts. Detailed photographs of these features can be found in Boundy and Austrheim (1998).

4.3. Locality A-3: Eclogite-facies pseudotachylytes, part 2 – evidence for deep earthquakes

The pseudotachylyte veins on this outcrop are associated with ultramylonites and cataclasites. The outcrop display areas where the granulite is totally crushed. A sketch of this outcrop is shown in Fig. 4.1. Pseudotachylytes are typically interpreted as frictional melts formed during seismic faulting. Austrheim and Boundy (1994) and Austrheim et al. (1996) demonstrated that these pseudotachylyte veins contain mineral assemblages and textures indicative of formation at eclogite-facies conditions. In addition to the Ådnefjell area, we observed the same spatial relationship between pseudotachylytes and fluid-mediated eclogitization of granulites on more than 50 outcrops. Single pseudotachylyte veins can be followed for at least 50 m and the measured displacement is up to 50 cm. The pseudotachylytes are formed in dry granulites enclosed by eclogite-facies shear zones. On a larger scale, the pseudotachylyte veins are found over at least 100 km^2 and preferentially along major eclogite-facies shear zones.

The close association between the eclogites and the pseudotachylytes led us to speculate that pseudotachylytes formed as the competent granulite blocks were loaded due to movement on the ductile eclogite-facies shear zones/and or due to volume changes caused by the eclogitization. We observed that the shattering of the rocks enhanced reactions by allowing fluid to reach new volume of the granulites as discussed by Austrheim et al. (1996, 1997). Perhaps seismic faulting is an integral part of the eclogitization process.

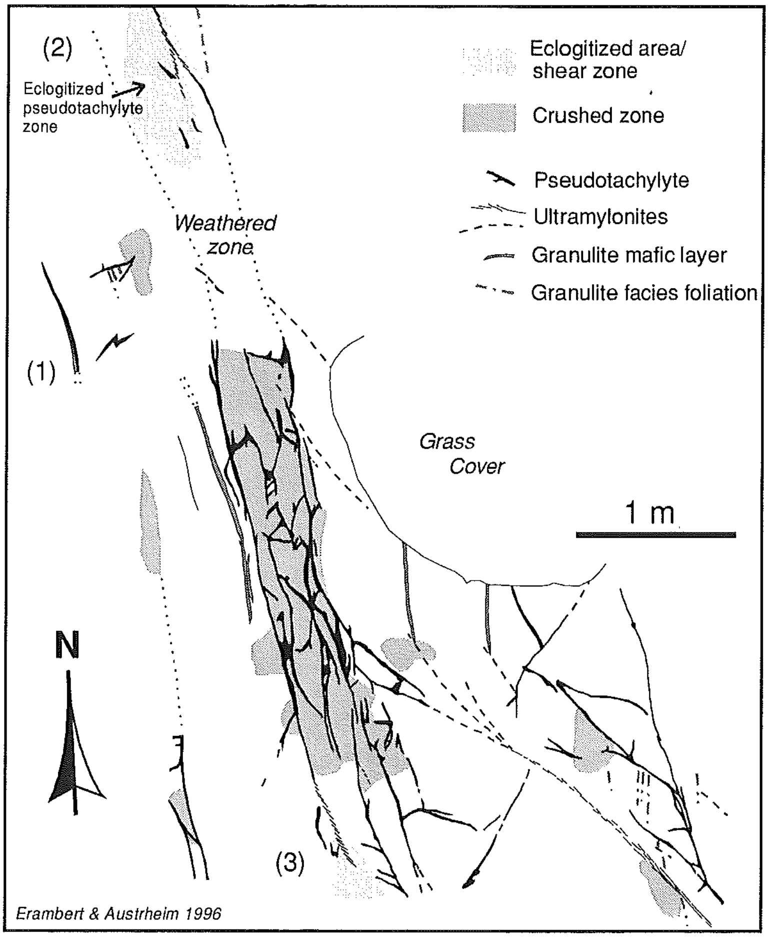


Figure 4.1: Field sketch showing relationship between pseudotachylytes, cataclasites, and eclogite-facies shear zone. A pseudotachylyte zone, about 0.5-1m thick, (central part) is bounded by pseudotachylyte veins and ultramylonites. It is oriented parallel to the granulite-facies layering, a common feature in the area. Veins often form at the contact of the mafic layers (1). Pseudotachylytes may pass into cataclasites or ultramylonites. The crushed zone is penetrated by a network of small pseudotachylyte veins, cataclasites and ultramylonites. Preferential eclogitization of pseudotachylyte zone is observed in (2), promoted by fluid infiltration along the fractures. (3) Tip of eclogite-facies shear zone.

Concluding remarks

The field evidence presented here demonstrates the importance of fluids in promoting reactions and hence controlling petrophysical properties of crustal root zones. This fluid-mediated eclogitization can result in a partially eclogitized crust, with a mixture of rock types and metamorphic facies. Similar relationships have been observed in several exposed deep crustal sections as summarized by Austrheim (1998). The field observations indicate that an anhydrous crust may increase its thickness with at least 20 km (corresponding to the pressure difference between the granulite facies and the eclogite facies) and still preserve its initial mineralogy. Such thickening is known to take place under the Alps where the crust-mantle boundary is found at a depth of 55 km.

The occurrence or absence of earthquakes in a given region is generally taken to be indicative of the material properties and particularly of the rheology in the focal region. If our interpretation of earthquakes as due to eclogitization processes is correct, some earthquakes may also record ongoing metamorphic processes. Nearly all intermediate-deep earthquakes occur in thin zones in areas where a slab of oceanic lithosphere has been subducted in the last 10 to 20 Ma.. Hori et al. (1985) deduced from measured seismic velocities and thickness of the subcrustal seismic zone below SW Japan that this zone is made up of untransformed basaltic oceanic crust and that eclogitization does not take place in the slab until a depth of 50-60 km. This is in agreement with the observation from Holsnøy that eclogitization is postponed relative to the crossing of the reaction boundaries and that earthquake is released at the front of eclogitization in a metastable crust.

Fluid-mediated eclogitzation is also likely to actively influence the geodynamics of collision and subduction zones. The eclogitization process leads to densification of the crust and at the same time produces rocks which are rheologically weak. This may be an ideal situation for fractionation of crust and sinking of dense material. Ductility enhancement by transformation plasticity or other processes should favor delamination of deep crust. The depths at which this fractionation will start obviously depend on access of fluid relative to the convergence rate of the plates.

The observations from Holsnøy and other collision zones suggest that modeling their metamorphic and petrophysical evolution in terms of T and P alone is an oversimplification. The evolution of a collision zone is strongly dependent on the fluid budget and fluid transport. This is particularly true, as the material recycled during collision is likely to be dry. The Holsnøy example demonstrate that geodynamic models of collision zones must consider fluids in addition to P and T and suggests that different collision zones may develop in various ways depending on fluid availability. Unfortunately, we are still in the infancy of understanding fluid regimes at depth in collision and subduction zones.

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