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Effect of small-scale heterogeneities on interpretation of crustal compositions exemplified by a layered anorthosite

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Abstract

The composition of the lower crust has a significant effect on geodynamic processes because it influences physical rock properties such as densities and seismic velocities. Compositional differences in lower crustal rocks are potentially large and exist on the scales of centimeters up to kilometers resulting in non-unique seismic and gravity data. While larger heterogeneities can be detected as reflections on seismic profiles, irregular small-scale compositional variations are not likely to be discovered, but will influence the averaged seismic velocities and densities of an area. The extent and effects of such small-scale heterogeneities are

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explored on an exposed high-grade layered anorthositic body by providing a detailed field map, petrological descriptions, pycnometry measurements as well as whole rock and mineral analyses combined with thermodynamic phase equilibria calculations. To evaluate the results of our thermodynamic calculations, densities and mineral modes obtained from the modeled phase equilibria are compared to measured densities and estimated mineral modes from rock samples. The proportion of mafic to ultramafic (plagioclase-poor) rocks in the mapped field area amounts to 10-15% but higher proportions of these rock types in the lower crust are feasible. To further study the effects of compositional variations, we have generated mixtures of mafic to ultramafic and anorthositic/intermediate rocks until the average properties of these mixtures are comparable to those of mafic granulites (3000-3100 kg/m³; 7.1-7.3 km/s).

Mixtures of anorthosite with 40-45% and of tonalite with 50-60% high-grade mafic to ultramafic rocks yield average densities and seismic velocities similar to mafic granulites although they still contain 50-60 vol.% plagioclase. Hence small-scale mixing of certain rock types may result in the overestimation of the proportion of mafic (garnet) granulites in the lithologic interpretation of crustal compositions from seismic data. Since the transition to eclogite-facies in plagioclase-rich rocks is shifted to higher pressures and anorthositic/intermediate eclogites yield lower densities, a lower crust with higher modal amounts of plagioclase may not always provide the significant densification needed for certain geodynamic settings (e.g. delamination or subsidence).

Keywords: Layered Anorthosite; Heterogeneities; Lower Crust; Densities
1. Introduction

The composition of the lower crust plays an important role in geodynamic processes. Granulite terranes and xenoliths are the two main sources providing insights in the composition of lower crustal rocks. While granulite terranes suggest that felsic to intermediate rocks are major constituents of the lower crust (e.g. Austrheim, 2013), the majority of xenoliths as well as seismic studies indicate a predominantly mafic lower crustal composition (Christensen and Mooney, 1995; Rudnick and Fountain, 1995). In addition to this compositional ambiguity, granulite terranes and crustal cross-sections often show a large extent of lateral and vertical heterogeneities in the form of mafic and felsic layering on varying scales (Fountain and Salisbury, 1981; Harley, 1989; Baldwin et al., 2006; Miller and Snoke, 2009; Williams et al., 2009; Sinigoi et al., 2011). The presence of lithologic contrasts in the lower crust which may be one of the causes of the observed deep crustal seismic reflectivity is also corroborated by seismic modeling studies incorporating lithologies of exposed deep crustal terranes and laboratory measurements. (Hale and Thompson, 1982; Hurich and Smithson, 1987; Mooney and Brocher, 1987; Burke and Fountain, 1990; Reston, 1990; Mooney and Meissner, 1992; Holbrook et al., 1992; Holliger and Levander, 1992; Khazanehdari et al., 2000). Furthermore, Brittan and Warner (1996, 1997) have shown that the size and spatial distribution of heterogeneities scaling in the range of kilometers to a few hundred meters can lead to an over- or underestimation of the average velocity of the heterogeneous medium. Smaller-scale heterogeneities such as small isolated lenses of dense rocks, however, are generally not likely to be detected by wide angle soundings due to poor lateral resolution resulting in averaged velocity over wide areas of the lower crust (e.g.
Kern et al., 1999; Kuusisto et al., 2006). Yet even thin-layered clusters within shear zones of granulite- and/or eclogite-facies rocks can be the source of strong seismic reflectivity (e.g. Fountain et al., 1994; Ji et al., 1997). In addition to poor lateral resolution, the detection of heterogeneities in the lower crust may be difficult due to the non-unique lithological interpretation of lower crustal rocks from seismic velocities since a variety of rock type mixtures can yield the same velocities (Kuusisto et al., 2006; Kern, 2011). Additional constraints can be established if the seismic data are correlated with experimentally determined or calculated P- and S-wave velocities at lower crustal conditions for rock samples collected from surface outcrops or from xenoliths (Christensen and Wepfer, 1989; Jackson et al., 1990; Kern et al., 1999; Kuusisto et al., 2006; Kern, 2011). These findings can be improved by detailed studies of the distribution and properties of heterogeneities on exposed sections of the lower crust.

In the present paper, we provide a detailed geological map of anorthositic and mafic to ultramafic rocks from the Sogndal area, Western Norway as an example for lower crustal rocks exposing small scale heterogeneities. This particular area has been chosen due to the accessibility of excellent outcrops and the relatively restricted lithology facilitating the distinction between mafic to ultramafic (Pl-poor) and anorthositic (Pl-rich) rock types (we use this terminology only to differentiate between Pl-poor and Pl-rich rocks within the layered body as anorthosite is also classified as a mafic rock). By means of pycnometry, whole rock and mineral compositions and geothermobarometry, we will compare measured and calculated densities of these compositions. Anorthositic rocks have been reported to occur in deeper levels of exposed crustal cross-sections, e.g. in the Bergen Arcs, Norway (Griffin, 1972), Superior Province, Canada (Ermanovics and Davidson, 1976; Fountain and Salisbury, 1981) and associated with granulite terranes (Berg, 1977; Kars et al., 1980; Glikson, 1986; Harley,
Precambrian granulitic anorthosite and mangerite of the Bergen Arcs occur on a regional scale and are frequently viewed as representative of the lower crust (e.g. Austrheim, 2013). For example, the conceptual models of Jolivet et al. (2005; see their Fig. 7) for burial and exhumation of lower crust are based on the exposed rocks in the Bergen Arcs. Anorthosite has a relatively high seismic velocity (~6.7-7.0 km/s) but low density of ~2700 kg/m³ comparable to felsic granulites (Christensen and Mooney, 1995) and therefore may have a unique effect on the physical properties of a rock type mixture. Rock properties are calculated by using phase equilibria calculations with a Gibbs free energy minimization method (Connolly, 2005) and by combining estimated mineral modes from rock samples and measured mineral compositions (Hacker and Abers, 2004). We will conclude with a study of the effects of systematically varying rock type mixtures on the average physical properties and the consequences for the lithological interpretation of lower crustal rocks and geodynamic models.

2. Composition of the Sogndal anorthosites

2.1. Geological background

The studied area is located within the Storehaugfjellet northwest of Sogndal airport in the Sognefjord region (Fig. 1a). It is part of the Caledonian Upper Jotun Nappe which represents an allochthonous slice of Proterozoic crystalline crust (e.g. Lundmark et al., 2007). Granulite-facies rocks retrograded to variable degrees and intruded by younger granitic dykes (Lundmark and Corfu, 2007) are major components of the Upper Jotun Nappe. The north-
eastern part of the nappe is dominated by partially retrograded gneisses of granitic, syenitic, monzonitic, dioritic and gabbroic compositions. In contrast, the central and southwestern parts where the field area is located are mainly composed of high-grade anorthositic and gabbroic rocks (Bryhni et al., 1983; Lundmark et al., 2007; Fig. 1a). Protolith ages of orthogneisses within the Upper Jotun Nappe range between 1.66 and 1.26 Ga (Lundmark et al., 2007). The rocks have been transformed to granulite-facies conditions during the Sveconorwegian orogeny (1.25-0.9 Ga) followed by amphibolite-facies metamorphism ~954 ± 3 Ma (Lundmark et al., 2007). Another high-grade metamorphic event coinciding with local anatexis within granitic gneiss has been dated to 934 ± 1 Ma and is interpreted to reflect post-collisional elevated heat flow in the nappes (Lundmark et al., 2007). U-Pb dates of sparse zircons within the massif anorthosite in the Jotun Nappe reflect events at 965 ± 4 and 913 ± 2 Ma as well as a pronounced Caledonian metamorphic overprint (Lundmark and Corfu, 2008). The authors interpret the older age as the protolith age of the anorthosite while the metamorphic event at 913 ± 2 Ma seems to be locally restricted to the outermost edge of the Western Gneiss Region (Lundmark and Corfu, 2008).

The anorthosite unit of the Upper Jotun Nappe consists of medium- to coarse-grained rocks showing localized deformation at shear zones (Griffin, 1971). The more pristine anorthosite shows a characteristic lilac color whereas more retrogressed areas are whiter and finer-grained due to epidotization (Griffin, 1971). The anorthosite is interlayered on all scales with rocks of mafic to ultramafic composition often in the shape of lentoid pods forming a foliation parallel to the larger-scale compositional layering (Griffin, 1971). This layered structure has been observed throughout the whole anorthosite unit (e.g. Griffin, 1971; Lundmark and Corfu, 2008). In contrast to the high-grade anorthositic and mangeritic rocks of the Bergen Arcs (e.g. Austrheim, 1987; 2013), no eclogite-facies rocks have been reported in
the Jotun Nappe Complex (Lundmark and Corfu, 2008) and there is hence no indication for these rocks having been exposed to higher pressures.

Due to its high aluminum content (25-30% Al$_2$O$_3$), anorthosite is interesting for industrial purposes if the plagioclase contains more than 60% anorthite component enhancing the solubility in acids (Wanvik, 2007). Since iron-bearing silicates would be unfavorable during mining of anorthosite, studying the distribution and properties of heterogeneities in these rocks may also help for resource estimates and mining strategies.

2.2. Analytical methods and modeling techniques

Whole rock major element analyses were performed at Actlabs (http://www.actlabs.com) on homogenized sample powders using lithium metaborate/tetraborate fusion with a Perkin Elmer SCIEX ELAN 6000 ICP/MS. For selected samples, the amount of FeO has been determined by titration (Table 1).

All mineral compositions were obtained with a Cameca SX-100 electron microprobe at the University of Oslo using an acceleration voltage of 15kV, a current of 15 nA and a beam diameter of 1 μm for most minerals except for plagioclase where the beam diameter was 10 μm. For most minerals Fe$^{3+}$ was recalculated based on stoichiometry and charge balance except for epidote where all iron is assumed to be trivalent and for amphiboles where we have used minimum and maximum ferric estimates according to (Leake et al., 1997). The average mineral compositions and standard deviations for selected samples are listed in Table 2 (representative analyses for all minerals are available as electronic supplement A).

Rock densities of the six selected samples were calculated using the density of water at 20 °C and the following equation: $\rho = 0.998 \text{ g/cm}^3 * \frac{m_d}{(m_d-m_w)}*1000$ where the “dry” ($m_d$)
and “wet” \((m_w)\) weight were measured with a balance at the Geological Museum, Oslo (accuracy: one position after decimal point).

Average \(P-T\) estimates were calculated with THERMOCALC (Powell and Holland, 1994, 2008) using activities obtained by the program AX from selected analyses. Where a clear distinction was possible, we have used core and rim compositions of adjacent mineral grains.

A set of densities and seismic velocities was generated for the estimated \(P-T\) conditions utilizing the Excel worksheet provided by Hacker and Abers (2004). The relevant mineral proportions were obtained from analyses of mineral compositions and end member percentages (Table 2).

Phase diagrams were calculated with the Gibbs free energy minimization software Perple_X 6.6.6. (Connolly, 2005) using the thermodynamic dataset of Holland and Powell (Holland and Powell, 1998 and updated version of 2002). For simplicity, iron was assumed to be divalent and all oxides below 0.10 wt\% (Table 1) were excluded resulting in the system \(\text{TiO}_2-\text{Na}_2\text{O}-\text{CaO-(K}_2\text{O)-FeO-MgO-(MnO)-Al}_2\text{O}_3-\text{SiO}_2\) (TiNC(K)FM(Mn)AS). For compositions containing small amounts of \(K_2\text{O}\), two separate solution models represent feldspar in order to visualize the plagioclase-out reaction. Solid solution models were chosen to match the rock lithologies as closely as possible (see supplement B for a detailed list of solution models). Mineral modes, rock densities and seismic velocities were extracted from computed phase diagrams at specified conditions. A standard error of \(~3\%\) was presumed for seismic velocities (e.g. Connolly and Kerrick, 2002). For rock densities, however, we assumed a lower value of \(~1\%\).
2.3. **Description of rock types**

On the regional scale, the studied area at Storehaugfjellet is mapped as anorthosite and gabbro-anorthosite (Fig. 1a). However, the anorthosite is interlayered on all scales with mafic to ultramafic rocks forming irregular shapes, rounded or lentoid pods or more continuous layers ranging from a few mm up to 20 cm in thickness with parallel alignment in a northeast-southwest direction (Fig. 1b). These heterogeneities are not restricted to the outcrop but can be observed throughout all anorthositic units within the Upper Jotun Nappe (Griffin, 1971; Lundmark and Corfu, 2008). On the kilometer scale, however, this results in a homogeneous unit (Fig. 1a). We carried out detailed mapping and sampling on a near horizontal surface within the layered anorthosite unit using a 9 by 9 meter grid (Fig. 1b) in order to obtain a quantitative estimate of the volumetric proportions of different rock types from the two-dimensional map. With a vertical section perpendicular to the mapped surface (Fig. 1c) we confirm the heterogeneous nature of the outcrop and infer that the volumetric proportions obtained from the 2D mapping can be extrapolated to a 3D grid. The resultant map was digitized in ArcGIS.

To a large extent, both major rock types, anorthosite and mafic to ultramafic rocks, exhibit mineral assemblages typical for granulite-facies metamorphism showing a lower-grade overprint to various degrees. Due to the almost monomineralic nature of the rock, large sections of the anorthosite have been preserved almost unaltered. The mafic to ultramafic rocks also comprise garnetite and pyroxenite as sub-units (Fig. 1b).

2.4. **Anorthositic rocks**

Anorthosite constitutes ~33 m² (~87%) of the mapped area but shows significant
compositional and mineralogical variation. Rocks containing high amounts of SiO2 (>50 wt.%), Al2O3 (25-29 wt.%) and Na2O (∼4 wt.%) are classified as anorthosites while samples with lower SiO2 (∼45 wt.%) and Al2O3 (22-24 wt.%) but higher percentages of Fe2O3 (∼7 wt.%) and MgO (7-9 wt.%) due to higher amounts of mafic minerals are referred to as mafic anorthosites (Table 1; Fig. 2). We have chosen one sample with ∼95 volume percent plagioclase (SH09-30; Table 2; Fig. 3a) and one sample with ∼60-70 volume percent plagioclase (SH09-35; Table 2; Fig. 3b) as representative end members (sample locations in Fig. 1b).

Both samples contain plagioclase of composition ∼An55-70. In the mafic anorthosite, plagioclase is also present in symplectites surrounding garnet. Isolated clinozoisite crystals have replaced some plagioclase and are interpreted to be secondary.

In the anorthosite, garnet is present in only small amounts of 1-2 volume percent (Fig. 3a; Table 2) whereas the original amount of garnet in the mafic anorthosite is estimated to ∼15 volume percent (Fig. 3b; Table 2). Most garnet crystals, however, have been replaced by symplectic intergrowths consisting of clinopyroxene, spinel, plagioclase and orthopyroxene leaving only garnet cores (Fig. 3b). Within these intergrowths, small rims of tschermakitic amphibole have formed mostly around some spinel crystals. The remaining garnet cores show a higher proportion of pyrope (Prp40-50) compared to garnet in the anorthosite (Table 2). Most garnet crystals are also partly replaced by chlorite (Fig. 3).

Clinopyroxene in the anorthosite has been completely replaced by chlorite and epidote/clinozoisite (Fig. 3a). In the mafic anorthosite, clinopyroxene is present both as bigger Al-rich (7-12 wt.% Al2O3) crystals (Cpx I) and as Al-poor (4-5 wt.% Al2O3) grains (Cpx II) in the symplectites replacing garnet (Table 2; Fig. 3b).

The amount of spinel is also highly variable. While it is absent in the anorthosite (Fig.
3a) most other samples contain 1-3 volume percent spinel. Significantly higher amounts (~25 vol.%) and larger spinel crystals (3-5 mm) sometimes occur in contact zones between anorthosite and mafic to ultramafic rocks (SH09-32; Fig. 3c). Some spinel crystals have been replaced by small amounts of magnesium-rich chlorite and phlogopite indicating fluid-induced alteration. Rarely, spinel is replaced by corundum. Single crystals of corundum are also present in some thin sections (Fig. 3a).

2.5. Mafic to ultramafic rocks

The mafic to ultramafic portion of the mapped area amounts to 4.7 m² in addition to 0.3 m² garnetite and 0.1 m² pyroxenite (totalling ~13% of the outcrop). The whole rock composition of the homogeneous mafic sample and the garnetite yield similarly low amounts of SiO₂ (42-43 wt.%) but significantly higher MgO (~11 wt.%) and FeO (6-8 wt.%) compared to the anorthosites (Table 1, Fig. 2). The pyroxenite, however, contains a higher percentage of SiO₂ (44-46 wt.%) and CaO (15-18 wt.%) but lower amounts of Al₂O₃ (~12 wt.%) than the other mafic rock types. The mafic to ultramafic rocks contain 45-50 volume percent garnet and clinopyroxene with less than 4 volume percent plagioclase and spinel (SH09-37; Fig. 3d). Domains of garnetite (80-90 vol.% garnet; SH09-01; Fig. 3e) and pyroxenite (> 90 vol.% pyroxenes; SH09-42; Fig. 3f) exist within the more homogeneous mafic rocks (Fig. 1c). Garnet crystals are 2-3 mm in size (Fig. 3d) and show a zonation with Fe-enriched (Alm₄₅) but Mg and Ca-depleted (Prp₃₃-₃₆; Grs₁₆-₁₉) rims compared to the cores (Alm₃₀-₃₃Prp₄₃-₄₅Grs₂₃; Fig. 4). Garnetite contains large (< 1 cm) garnet crystals (upper part of Fig. 3e). Symplectites of clinopyroxene, spinel, plagioclase and orthopyroxene are common at grain
boundaries of garnet in contact with clinopyroxene and spinel (Fig. 3d and e; Fig. 4). Small amounts of amphibole have formed mostly around spinel crystals within these symplectites (Fig. 4c).

Clinopyroxene is present both as bigger (2-4 mm), brownish crystals and within the symplectitic rims surrounding garnet (Fig. 4). In the pyroxenite, clinopyroxene contains less aluminum (~6 wt.% Al₂O₃) compared to the bigger grains in the other mafic to ultramafic rocks (Table 2).

Plagioclase is present in minor amounts of ~3-5 volume %. In most samples the anorthite content is highly variable from ~An₅₄ to An₈₀ (Table 2). The plagioclase in the pyroxenite is generally more albitic (Ab₆₀; Table 2).

Spinel occurs both as individual grains and within the symplectites around garnet (Fig. 3d and e; Fig. 4). Orthopyroxene and amphibole are mostly present in the symplectic intergrowths around garnet. Chlorite forms small veins in the pyroxenite (Fig. 3f).

3. Geothermobarometry, thermodynamic modeling and determination of rock properties

Rock densities can be determined directly by measuring the weight of a sample in air and water. However, this density may be lower than the density of the high-grade metamorphic assemblage due to the lower-grade overprint and partial replacement of minerals of the original assemblage. Alternatively, the density is calculated at elevated P and T from equations of state for minerals combined with their modes and compositions (Hacker and Abers, 2004) or by computing phase diagrams from whole rock compositions and extracting
densities (e.g. Connolly and Petrini, 2002). However, both estimation methods require the
knowledge of a $P$-$T$ range for the mineral assemblage. Hence, we have combined
geothermobarometric results and petrological constraints to obtain a possible $P$-$T$ range for
most of the studied samples.

3.1. Geothermobarometry

The high-grade assemblage in all samples is interpreted to consist of garnet, plagioclase,
clinopyroxene and spinel (non symplectitic). Only the Al-rich clinopyroxene (Cpx I) is assigned
to be part of this assemblage because the Al-poor clinopyroxene (Cpx II) is restricted to the
symplectitic intergrowths. $P$-$T$ conditions obtained with THERMOCALC range between 950-
1070 ± 200 °C at 1.0 to 1.4 ± 0.2 GPa (Table 3). The set of independent reactions used to
determine these $P$-$T$ conditions is shown in Fig. 5.

Some anorthositic samples contain either isolated corundum crystals or show a partial
replacement of spinel by corundum. Assemblages with corundum yield lower $T$ (800-850 ±
100 °C) at $P$ of 1.1-1.2 ± 0.2 GPa (Table 3).

A lower grade assemblage is characterized by the replacement of garnet with
symplectitic intergrowths and the formation of orthopyroxene. Hence garnet and
clinopyroxene rim compositions have been used together with orthopyroxene for
thermobarometric calculations yielding temperatures of ~900-950 ± 100 °C at 0.8-1.1 ± 0.20
GPa (Table 3).

3.2. P-T assessment by pseudosection calculations and mineral composition
In addition to geothermobarometry, we have compared selected mineral compositions from microprobe analyses (Table 2 and supplement A) with compositions calculated by the program werami (Perple_X) within the observed mineral assemblage for three samples (Fig. 6; Table 3). Isopleths shown in Fig. 6 represent pyrope content in garnet as well as diopside content in clinopyroxene. The resulting \( P-T \) estimate for the mafic anorthosite of \( \sim 900 \pm 100 \) °C and \( 1.0 \pm 0.15 \) GPa (ellipse in Fig. 6a) is in accordance with the values obtained by THERMOCALC (Table 3). For the mafic to ultramafic samples, however, the temperature estimates for the high-grade assemblage of \( 750 \pm 100 \) °C are significantly lower than the results from THERMOCALC (Fig. 5 and 6b, c; Table 3) plotting outside of the plagioclase stability field.

### 3.3. Comparison of measured and computed rock properties

For a comparison of measured and computed densities and velocities, we have chosen two \( P-T \) points within the high-grade mineral assemblage: a high value of 950 °C and 1.0 GPa to account for the \( P-T \) results from THERMOCALC (Table 3) and a lower point of 660 °C at 0.85 GPa closer to the corundum stability field for the more anorthositic samples. Due to the dominance of plagioclase, the anorthosite (Fig. 3a; 7a) shows a good agreement (within 2% aberration) of measured and computed densities at high and lower \( P-T \) conditions (Table 4). The slightly lower density obtained by pycnometry is likely due to the replacement of clinopyroxene by secondary minerals and the marginal variation of the proportion of garnet. P-wave velocities obtained from the spreadsheet by Hacker and Abers (2004) are somewhat lower than those extracted from Perple_X but all values are well within the range for an average anorthosite (\( \rho: \sim 2760 \) kg/m\(^3\); v\( P \): 6.6-6.8 km/s) given by Christensen and Mooney.
(1995) for conditions between 30 and 40 km at average to high temperatures.

The mafic anorthosite (Fig. 7b) shows significantly higher discrepancies between measured and computed densities (Table 4). This reflects that only a fraction of the garnet is preserved in the sample (Fig. 3b) and the symplectitic intergrowths are majorly composed of less dense minerals. Due to the abundance of mafic minerals, the density is much higher than that for a normal anorthosite and rather resembles that of hornblendite (~3270 kg/m$^3$) or pyroxenite (~3280 kg/m$^3$) as given by (Christensen and Mooney, 1995). Together with the relatively high seismic velocities (7.4-7.6 km/s; Table 4), this rock mixture may likely be interpreted as pyroxenite if only geophysical data is available. The contact zone (Fig. 3c) yields even higher densities (~3300 kg/m$^3$) and slightly higher P-wave velocities (Table 4).

The mafic to ultramafic rocks (Fig. 7c and d) are characterized by exceptional high densities and velocities (~3400-3500 kg/m$^3$; 7.8-8.2 km/s; Table 4) strongly exceeding the commonly assumed values for mafic granulites (~3150 kg/m$^3$; 6.7-7.1 km/s) and rather resembling eclogite (Christensen and Mooney, 1995). Results for the pyroxenite (~3270 kg/m$^3$; 7.4-7.6 km/s) comply with the typical values of this rock type (Christensen and Mooney, 1995). The lower measured density is likely a result of the presence of secondary minerals.

In conclusion, the different rock types record significant variations in density of up to 830 kg/m$^3$ and velocity of up to 1.5 km/s on a relatively small scale.

3.4. **Effect of retrogression on physical rock properties**

Retrogression is incomplete and inhomogeneous throughout the samples and mainly localized around mafic minerals (in particular garnet). This suggests that the rocks were
exposed to small and varying amounts of fluids possibly moving along weak zones such as cracks. Phase equilibria modeling, on the other hand, assumes complete equilibrium and hence needs some modification to reproduce the partial retrogression. We added small amounts of water to see if the modes of water-bearing minerals observed in the samples can be approximated in the model.

For the anorthosite (SH09-30) adding 0.3 wt.% H₂O at 600 °C and 0.68 GPa results in similar amounts of chlorite and zoisite as observed in the sample. However, the computed density reduction amounts to only 10 to 30 kg/m³ for the two P-T points, respectively (Table 4). Applying similar conditions to the mafic anorthosite (SH09-35) does not yield results matching the observed mineralogy. Orthopyroxene which is a minor constituent of the symplectitic intergrowths (Table 2) will only form at lower pressures (~0.6 GPa at 600 °C). Furthermore, the addition of 0.2 wt.% water results in quite high amounts of amphibole (~10 vol.%) in the pseudosection calculations which have not been observed in the sample. Reducing the amount of water to 0.1 wt.% will also reduce the modal amphibole to ~5 vol.% which is higher than the estimates (Table 2). The best fit between modeled and observed retrogressed mineral assemblages is reached at 600 °C between 0.55-0.6 GPa and ≤ 0.1 wt.% water. At these conditions, the density is lowered by ~50-100 kg/m³ but is still higher than the density obtained by pycnometry. Applying the same conditions to the contact zone (SH09-32) reduces the density by ~50-70 kg/m³ yielding a value close to the measured density (Table 4). Densities of the mafic to ultramafic rocks (SH09-37; SH09-01) are also reduced by 50-140 kg/m³ at these lower P-T conditions while the density of the pyroxenite (SH09-42) is only slightly reduced (≤ 30 kg/m³). Hence, the density reduction due to retrogression is rather insignificant (≤ 30 kg/m³) for samples with low amounts of garnet but quite high (up to ~150 kg/m³) for heterogeneous samples with abundant garnet.
3.5. Effect of mixing on rock properties

To study the effect of average rock properties on heterogeneous lower crust we have calculated phase diagrams for mixtures of anorthosite (whole rock composition of SH09-30, Table 1) with systematically increasing amounts of mafic to ultramafic proportion (composition of SH09-37, Table 1). Another batch of mixtures with a more common rock type has been calculated with the whole rock composition of a tonalitic gneiss (sample D95-16 from Kern et al., 1999; Table 5) in combination with the mafic composition. Densities have been obtained in two different ways: 1) The oxides of the anorthositic/tonalitic compositions are mixed in the given proportions with the oxides of the mafic composition (see Table 5 for whole rock compositions). This approach enables the calculation of phase diagrams (Fig. 8) and the extraction of densities ($\rho_1$ in Table 6) but assumes a homogeneous mixture. 2) Densities are extracted separately from the calculated phase equilibria of the anorthosite/tonalite and mafic rock at specified conditions and then are summed up in the given proportions ($\rho_2$ in Table 6).

This method assumes that the different rock types do not react with each other. The density increases significantly from 5 to 40% in mixtures with anorthosite and mafic to ultramafic rocks (Fig. 8a,b) and from 5 to 50% in mixtures with the tonalite and mafic composition (Fig. 8c,d). As before, mineral modes and densities have been extracted at two $P$-$T$ conditions in the stability field of the high-grade assemblage and for a $P$-$T$ point representing eclogite-facies conditions (650 °C at 1.5 GPa, Table 6). Temperatures of 950 °C might be too high to reflect the conditions of continental lower crust but a lack of re-equilibration and therefore preservation of the high temperature mineralogy of granulites at lower temperatures is likely. Due to the relatively high P-wave velocity but low density of the anorthosite, mixtures with relatively low mafic portions (10-25%) will yield low densities (~2800-2900 kg/m³) similar to
felsic granulites or mica-quartz schist (Christensen and Mooney, 1995) but too high P-wave velocities (6.8-7.0 km/s) for these rock types and should therefore be recognizable as mixtures. Anorthositic mixtures with 45-50% mafic to ultramafic rocks (Fig. 8c), however, resemble a mafic granulite or mafic garnet granulite with densities in the range of 3000-3100 kg/m$^3$ (Table 6; Fig. 9) and velocities of 7.1 to 7.3 km/s. Fig. 9 demonstrates that the properties of all mixtures behave according to Birch’s law (e.g. Birch, 1961a, 1961b) showing the seismic velocity as an approximately linear function of the density. In addition, the Poisson ratios of ~0.27 for these mixtures agree well with the range of 0.27-0.28 determined for a number of mafic granulites (Kern et al., 1996, 1999). However, all values obtained from thermodynamic modeling are for idealized, isotropic, homogeneous aggregates without considering any textures so velocities might diverge due to the spatial orientation of minerals and even small heterogeneities.

The difference between extracted densities from homogeneous mixtures ($\rho_1$; Table 6) and averaged densities from separate rock compositions ($\rho_2$; Table 6) is mostly within 50 kg/m$^3$ for mixtures with anorthosite at the two $P$-$T$ conditions (points a and b; Table 6). For eclogite-facies conditions, however, $\rho_2$ is significantly lower (~150-200 kg/m$^3$) than the density for the homogeneous mixture. For tonalitic mixtures and granulite-facies conditions, $\rho_2$ is significantly higher (~100-200 kg/m$^3$) than $\rho_1$ while this relation is reversed for the densities representing eclogite-facies conditions.

4. Discussion

4.1. Evaluation of thermobarometry and phase diagrams
All samples in our field area record a high-grade assemblage of garnet, clinopyroxene, plagioclase and spinel. According to Lundmark and Corfu (2008) U-Pb ages of zircons in anorthosite point to a metamorphic episode at 913 ± 2 Ma affecting the layered anorthosite after its intrusion (965 ± 4 Ma). Additional information about the pressure-temperature evolution of the layered anorthosite is provided by coronas with garnet rims and olivine cores as reported by Griffin (1971). To explain corona formation between olivine and plagioclase, Griffin (1971) proposed slow cooling after the crystallization of anorthosite in combination with an increase in pressure to account for the formation of garnet. Alternatively, garnet may have formed by granulite-facies metamorphism which affected the rocks to varying degrees due to deformation and/or fluid infiltration accounting for the preservation of coronas as well as the formation of granulite-facies assemblages. This is corroborated by Sm-Nd ages of 907 ± 9, 912 ± 18 and 905 ± 37 Ma (Cohen et al., 1988) from pyroxene and garnet in coronas occurring in the Lindås troctolite overlapping with the ages determined by Lundmark and Corfu (2008). Therefore the authors propose that these dates either reflect isotopic re-equilibration of preexisting garnet and clinopyroxene or (partial) formation of the coronas during the metamorphic event (Lundmark and Corfu, 2008). Since the coronas exhibit two stages of mineral reactions (Griffin, 1979) there is a possibility of the first reaction to have occurred during cooling from magmatic temperatures while the second could be related to the 913 Ma event (Lundmark and Corfu, 2008). Based on these results, we assume that at least some of the recorded high-grade assemblage is due to granulite-facies metamorphism and unrelated to cooling of the anorthosite intrusion. A further discussion, however, is beyond the scope of this paper since our conclusions depend only on the rock properties of the observed rocks and not on the conditions and time of their formation.

Several $P-T$ estimates obtained with THERMOCALC, especially for the garnetite
(SH09-01), yield high pressures (Table 3) and plot outside of the stability field of plagioclase (Fig. 9b and c). This may be due to the preservation of only garnet cores in disequilibrium with other minerals. The large range of $P$-$T$ estimates possibly reflects incomplete reactions due to partial retrogression. The formation of corundum requires lower temperatures as indicated by the phase diagrams (Fig. 7a,b), by thermobarometry (Table 3) and the fact that corundum has replaced spinel. This suggests that corundum has formed after the high-grade metamorphism. The presence of symplectitic intergrowths consisting of clinopyroxene, spinel, plagioclase and orthopyroxene surrounding garnet indicates cooling and decompression which is also in agreement with the phase diagrams (Fig. 6) and the lower thermobarometric estimates (Table 3). Similar symplectites of orthopyroxene, plagioclase and spinel forming at the expense of garnet in quartz-free mafic granulites have been described by Harley (1989) as a typical reaction texture related to isothermal decompression. The formation of amphibole predominantly around spinel within the symplectites could be related to a younger Caledonian overprint which was described for rocks of the Jotun Nappe (Lundmark et al., 2007).

4.2. Suitability of the studied rocks as a lower crustal analogue

The emplacement of massif anorthosite bodies is commonly restricted to the range of 5-13 km but can occur at depths as great as 23-37 km (e.g. Morse, 1982) corresponding to upper-middle crustal levels. In this case, however, we are studying a high-grade layered anorthosite resting on top of the Caledonian orogen (Lundmark and Corfu, 2008) and representing a part of the Jotun Nappe Complex. Based on the present location of the studied rocks, U-Pb geochronology of the anorthosites (Lundmark and Corfu, 2008) as well as adjacent rocks within the Upper Jotun Nappe (Lundmark et al., 2007) and the observed high-grade mineral assemblage, we have to assume that these rocks have been exposed to high temperatures
and moderate pressures during one or several events since their emplacement. The result is a mineral assemblage resembling granulite-facies rocks which dominate the lower crust (e.g. Rudnick and Fountain, 1995). Therefore, we consider the studied rocks to be a good analogue for lower crustal rocks by mineralogical aspects.

Furthermore, several studies have shown anorthosites to be a component in deeper levels of crustal cross-sections, at times together with mafic to ultramafic bodies (e.g. Fountain and Salisbury, 1981 and references therein). Moreover it has been shown that exposures of lower crustal rocks representing the root zone of the Norwegian Caledonides contain significant portions of anorthosite (e.g. Austrheim, 2013). We take this as an indication for the existence of high-grade anorthositic rocks at greater depths which justifies the adoption of our studied rocks as an example for lower crustal rocks but we do not claim that these rocks compose the majority of the lower crust. To make this study more representative, we have augmented the mixing models by tonalitic rocks which are a more common rock type in the lower crust but any model will be a simplification due to the extremely heterogeneous nature of the lower crust.

4.3. Evaluation of calculation methods

Densities obtained with Perple_X and the worksheet of Hacker and Abers (2004) match within 20-40 kg/m³ for most compositions (Table 4), approximating an aberration of ~1%. Given that the standard error for densities obtained by Perple_X is ~1%, and that accessories such as corundum and ilmenite have not been included in the spreadsheet, the values are in good agreement with each other. Similarly, P-wave velocities obtained from the spreadsheet are 0.1 to 0.2 km/s lower than the values extracted from Perple_X, corresponding to a discrepancy of maximal ~3% and therefore are within the error.
For relatively homogeneous and almost monomineralic samples such as the anorthosite (SH09-30) there is a very good agreement between measured and calculated densities. The significant discrepancy in density for the more heterogeneous compositions, especially the mafic anorthosite (SH09-35) and contact zone (SH09-32) is a result of the retrogression of high-grade minerals as well as the high variation in mineral modes due to the extremely heterogeneous nature of the drill cores. Since domains of garnetite alter on very small scales with the other mafic to ultramafic rocks the sample used for the whole rock composition may represent a more heterogeneous mixture instead of pure garnetite which could explain the discrepancy in estimated and calculated mineral modes for this rock type.

We conclude that the applied calculation methods are suitable to model phase equilibria from whole rock compositions and that discrepancies of measured and modeled rock properties are mainly due to retrogression and very small-scale heterogeneities within the drill cores of the samples. Since we are interested in the high-grade assemblage as representative lower crust, we have used the least retrogressed samples (anorthosite and mafic to ultramafic rock) for the computation of mixing models.

### 4.4. Assessment of heterogeneities

Our mapped area shows mafic to ultramafic and anorthositic heterogeneities on outcrop scale down to micron scale. The proportions of anorthosite and mafic rocks (including garnetite and pyroxenite) have been determined to amount to ~87 and ~13%, respectively. Due to the high variation of mafic minerals in the anorthositic sections, the mafic percentage may be somewhat higher. In addition, the small mapped area shows an irregular distribution of the mafic to ultramafic pods and bands and it is possible that other areas regionally within the
The level of origin within the crustal column for the anorthosites is not known but seismic data and xenolith studies indicate that the lower crust becomes significantly more mafic with depth (e.g. Christensen and Mooney, 1995; Kuusisto et al., 2006; Rudnick and Gao, 2003). Furthermore, local positive Bouguer anomalies over a number of anorthosite bodies suggest that more mafic rocks are located beneath the anorthositic rocks in the lower crust or in the upper mantle (Morse, 1982 and references therein). It is therefore possible that the amount of mafic to ultramafic intrusions increase in the lower parts of anorthositic bodies.

The observation of small scale heterogeneities in combination with the indication for a mafic (plagioclase-poor) lower crust has motivated the mixing of mafic and anorthositic/intermediate compositions until the average properties are comparable to mafic granulites (Fig. 9). Assuming nearly dry conditions (e.g. very low amounts of hydrous minerals), much lower amounts of mafic to ultramafic (40-60%) are sufficient for the mixture to resemble mafic garnet granulites despite quite high amounts of plagioclase (50-60 volume %).

4.5. Consequences for geodynamic models

Phase transitions in the lower crust, especially during formation of eclogite from granulite or amphibolite have been suggested as mechanism for various geodynamic processes such as basin subsidence (e.g. O’Connell and Wasserburg, 1972; Mareschal and Lee, 1983; Hamdani et al., 1994; Artyushkov, 2005, 2010) or delamination (e.g. Kay and Mahlburg Kay, 1993; Gao et al., 1998; Kukkonen et al., 2008). The requirement for these geodynamic settings is the densification of the lowermost crust to a degree that exceeds the density of the underlying
mantle. In many cases, the seismic data lead to the conclusion that the lower crust is predominantly composed of mafic granulites and this densification might be feasible. However, we have shown that mixtures of certain rock types (anorthosites or tonalitic compositions combined with mafic to ultramafic rocks) can contain far more plagioclase despite the resemblance of their geophysical properties to mafic granulites. Higher amounts of plagioclase component in the rock shifts the characteristic eclogite reaction (plagioclase to clinopyroxene) to higher pressures (as demonstrated in Fig. 3 of Semprich and Simon, 2014). Therefore the significant densification associated with eclogitization is expected at different pressure.

In the following an example is given to illustrate some geodynamic consequences of lithological heterogeneity on densification. The lowermost part of the crust is assumed to consist of altering bands of anorthosite/tonalite and mafic to ultramafic rocks similar to the layered anorthosite presented in this paper. However, to ensure comparability to granulite-facies lower crustal rocks higher amounts of mafic to ultramafic compositions are taken in this example. In addition to a significant densification, delamination also depends on the rheology and therefore the temperature of the mantle (e.g. Kay and Mahlburg Kay, 1993). Delamination of the lowermost eclogitized crust has been suggested for thickened crust where the crust-mantle boundary shows almost no density contrast (e.g. Fennoscandian Shield; Kukkonen et al., 2008). Geophysical constraints require a relatively high temperature to ensure a sufficiently low viscosity of the mantle (Kukkonen et al., 2008). However, high temperature reduces density. Therefore, sufficient densification and subsequent delamination is only possible at higher pressure (Semprich and Simon, 2014). An increase in pressure can be achieved by additional burial or by alternative pressure generating mechanisms such as tectonic or reactive stress (e.g. Petrič and Podladchikov, 2000; Schmalholz and
Podladchikov, 2013; Schmalholz et al., 2014). Only the effect of heterogeneity on densification is illustrated here and rheology aspects are not considered further. The following pressures and temperatures have been chosen in order to compare them to previously published data for mafic granulite (Semprich and Simon, 2014), but as delamination is not only dependent on the density, it may not occur even if the density of the lowermost crust exceeds that of the mantle. A lithostatic pressure of 1.4 GPa corresponds to a depth of ~50 km using the approximation: \( z = \frac{P}{9.81 \text{ m/s}^2} \ast 3000 \text{ kg/m}^3 \) and may represent conditions of thickened crust suitable for delamination. Two cases, for respectively low and high temperature, are considered.

At temperatures of 550 at 1.4 GPa, a mafic garnet granulite representative of the lower crust (Hölttä et al., 2000) yields a density of 3380 kg/m\(^3\) assuming complete equilibration (Table 2; Semprich and Simon, 2014), exceeding the mantle density (~3320 kg/m\(^3\)). A homogeneous and equilibrated mixture of anorthosite/tonalite with 50% mafic to ultramafic rocks at the same conditions also exceeds the mantle density with values of of ≥ 3447 kg/m\(^3\). This would be sufficiently high density for delamination. However, some authors have argued, that rocks may not undergo mineral transformations at temperatures below 750 °C and pressures up to 3 GPa without a catalyst such as a fluid or deformation (e.g. Austrheim, 1987; Hacker, 1996; Austrheim, 1998; Krabbendam et al., 2000; John and Schenk, 2003; Peterman et al., 2009; Austrheim, 2013). If this was the case the anorthositic or tonalitic bands would preserve significant amounts of low density phases. The cumulative density of the layered body may then be too low for delamination.

At higher temperature eclogitization is shifted to higher pressure. A mafic garnet granulite, for example, requires pressures of ~ 1.7-1.8 GPa at 900-1000°C to yield higher density than the mantle (Table 3; Semprich and Simon). In comparison, the anorthosite still
contains ~50% of plagioclase and yields a density of ~ 2970 kg/m\(^3\) at 900 °C and 2.0 GPa and is therefore not dense enough to delaminate. A mix of anorthosite with 50% of exceptionally dense mafic to ultramafic bands, however, reaches a cumulative density above 3220 kg/m\(^3\) at pressure of ~1.8 GPa. To ensure these conditions, the rocks either have to be at a greater depth of ~10-15 km or be exposed to additional tectonic pressure of ~0.3-0.5 GPa. This value is compatible with tectonic overpressure obtained in geodynamic numerical simulations (e.g. Petrini and Podladchikov, 2000; Schmalholz and Podladchikov, 2013; Schmalholz et al., 2014).

5. Conclusions

High-grade anorthosites interlayered on all scales with mafic to ultramafic rocks forming irregular shapes or more continuous bands at Sogndal may represent parts of the lower crust. Geothermobarometry and a comparison of observed and modeled mineral modes suggest the equilibration of the rocks to high-grade conditions of 750-1070 ± 200 °C and 0.85 to 1.4 ± 0.2 GPa followed by one or more episodes of lower grade overprint recorded by the formation of corundum, symplectitic intergrowths as well as minor amounts of hydrous minerals.

The small-scale heterogeneities of anorthositic and mafic to ultramafic rocks are responsible for contrasts in density of up to 830 kg/m\(^3\) and velocities of up to 1.5 km/s on the cm range. Measured densities are generally lower than those extracted from the thermodynamic models and calculated from mineral modes and mineral compositions which is mostly due to retrogression but also a result of small-scale heterogeneities within the drill cores of the sample.

Mixing of anorthosites and mafic to ultramafic rocks shows that a mafic portion of ~40-
45% is sufficient for the rock to resemble a mafic (garnet) granulite with typical densities of 3000-3100 kg/m³, P-wave velocities of ~7 km/s and a Poisson ratio of ~0.27. Similar results can be achieved by a mixture of tonalitic gneiss with ~50-60% mafic rocks. In such a case, the lithological interpretation from seismic data might overestimate the proportion of dense mafic granulite in the lower crust despite the relatively high modal amount of plagioclase (~50-60% volume %). As a result, the transition to eclogite is shifted to much higher pressures than in mafic granulite. Geodynamic models requiring a significant densification due to phase changes might hence require much higher pressures than lithostatic if similar rocks are concerned.

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Figure captions:

Figure 1: a) Geological map of the Sognefjord area (NGU; Bryhni, I. 2003: Berggrunnskart Sogndal 1417-3, M 1:50 000). The anorthosite unit is composed of banded anorthosite with
altering anorthositic and mafic layers of varying thickness. The circle indicates the location of
the field map (1b). b) Detailed field map of the high-grade anorthositic and mafic rocks in the
Sogndal area. Black circles and numbers indicate sample locations used in this study with
whole rock compositions listed in Table 1. The anorthosite unit comprises rocks with 60-95%
plagioclase. Mafic to ultramafic rocks contain 40-50 volume % clinopyroxene and garnet and
minor amounts (~2-5 volume %) of plagioclase and spinel. Garnet rich (~80-90 volume %)
domains within the mafic rocks are described as garnetite. Pyroxenite contains > 90 volume
% pyroxenes (majorly clinopyroxene). c) Vertical section perpendicular to the field map. d)
Photograph of a section of the field map (see box in 1b). e) Scanned sections through a
portion of the anorthosite (see box in 1b). Dark areas represent small mafic bands. f) Lens
with pyroxene core and garnet rim observed in the profile (see box 1c).

Figure 2: Plots of whole rock wt% of SiO₂ versus MgO, Al₂O₃, CaO, Na₂O, Fe₂O₃ and K₂O,
respectively for anorthosite, mafic anorthosite, contact zone, mafic to ultramafic rock,
pyroxenite and garnetite (whole rock compositions listed in Table 1). Anorthosite is
characterized by high amounts of SiO₂ (>50 wt.%), Al₂O₃ (25-29 wt.%) and Na₂O (~4 wt.%).
Mafic anorthosite shows lower amounts of SiO₂ (~45 wt.%) and Al₂O₃ (22-24 wt.%) but higher
percentages of Fe₂O₃ (~7 wt.%) and MgO (7-9 wt.%). The mafic to ultramafic compositions
yield similarly low amounts of SiO₂ (42-43 wt.%) but significantly higher MgO (~11wt.%) and
FeO (6-8 wt.%) compared to anorthosite. The pyroxenite, however, contains a higher
percentage of SiO₂ (44-46 wt.%) and CaO (15-18 wt.%) but lower amounts of Al₂O₃ (~12
wt.%) than the other mafic rock types.
**Figure 3:** a) Scanned thin section of anorthosite (sample SH09-30). Aggregates of chlorite and clinozoisite/epidote are interpreted to represent former clinopyroxene. Garnet is partly replaced by chlorite. b) Thin section of mafic anorthosite (sample SH09-35). Garnet is partly replaced by chlorite and symplectitic intergrowths of clinopyroxene, spinel, plagioclase, orthopyroxene and amphibole. c) Thin section scan of a contact zone between anorthosite and mafic rocks (sample SH09-32) with unusually high amounts of spinel. d) Scan of mafic to ultramafic composition (sample SH09-37). e) Thin section of a garnet-rich section within the mafic bands (SH09-01). f) Section of an area dominated by pyroxene (SH09-42). Whole rock compositions are listed in Table 1 and plotted in Fig. 2 while average mineral compositions are shown in Table 2.

**Figure 4:** a) Symplectitic intergrowths of clinopyroxene, spinel, plagioclase and orthopyroxene replacing garnet in the garnetite (sample SH09-01). b) Symplectites between two garnet crystals in the mafic to ultramafic composition (SH09-37). c) BSE image of the symplectites replacing garnet in the mafic to ultramafic rock. Garnet shows a zonation with higher almandine (Alm45) component at the rim compared to the core (Alm$_{30-33}$Pr$_{43-45}$Grs$_{23}$).

**Figure 5:** Independent set of reactions used for themobarometry estimates with THERMOCALC for the granulite-facies assemblage (Grt+Cpx+Pl+Spl; see also Table 3) exemplary shown for one $P$-$T$ point of the mafic to ultramafic rock (sample SH09-37; used mineral data of assemblage [1] in Table 3). The black square represents the resulting $P$-$T$ estimate with error bars while the red square depicts the $P$-$T$ estimates obtained from a comparison of measured and calculated mineral compositions with Perple_X (compare to Fig. 6). Also shown are the quartz-coesite reaction and the alumosilicate triple point for reference.
Figure 6: a) Pseudosection diagram for the mafic anorthosite (SH09-35). Red lines represent isopleths of pyrope content in garnet within the relevant granulite-facies stability field (Grt+Cpx+Pl+Spl) obtained from Perple_X (MgO/[MgO+FeO+CaO]). Thick red lines indicate values obtained from EMP measurements (Table 2; supplement A) where core measurements are indicated by (c). Green lines represent diopside content in clinopyroxene (MgO/[MgO+FeO+2Na2O]). The resulting P-T range for the assemblage is represented by the ellipse. Black squares represent P-T estimates obtained with THERMOCALC (see Table 3 and Fig. 5). b) Pseudosection diagram for the mafic to ultramafic rock (SH09-37) c) Pseudosections for the garnetite (SH09-01).

Figure 7: a) Major phase fields and density diagram for the anorthosite (SH09-30; Fig. 3a). White circles with label a and b represent the conditions at which rock properties have been extracted (Table 4). b) Density diagram for the mafic anorthosite (SH09-35; Fig. 3b). c) Densities for the mafic to ultramafic composition (SH09-37; Fig. 3d). e) Density diagram for the garnetite (SH09-01; Fig. 3e). This figure shows exceptional density differences between anorthositic and mafic to ultramafic compositions.

Figure 8: Phase and density diagrams for anorthositic (a,b) and tonalitic (c,d) mixtures with increasing amount of mafic component. Mixtures with higher mafic proportion yield a significant density increase of up to 200-300 kg/m³. White circles represent specified P-T conditions used for the extraction of rock properties (Table 6).
Figure 9: Density versus seismic P-wave velocity plot for selected mixture of anorthositic/intermediate and mafic to ultramafic rocks. TG represents mixtures with tonalitic gneiss (sample D95-16 from Kern et al., 1999), while An stands for mixtures with anorthosite. Numbers represent the amount of mafic rocks in the mixture (An45 has 45% of the whole-rock composition of sample SH09-37). Results for two P-T conditions are shown: 950 °C and 1.0 GPa (indicated by “h” and squares) and 660 °C and 0.85 GPa (indicated by “l” and diamonds). The field of mafic granulites and mafic garnet granulites represent data for 30-40 km at average to high temperatures including the standard deviation as given by Christensen and Mooney (1995). Anorthositic mixtures with 45-50% mafic to ultramafic rocks and tonalitic mixtures with 55-60% mafic rocks show rock properties similar to mafic granulites or mafic garnet granulites with densities in the range of 3000-3100kg/m³ and velocities between 7.1 to 7.3 km/s.

Table 1: Whole rock compositions for the selected samples (locations in Fig. 2; see Fig. 3 for whole-rock plots).

Table 2: Average EMP measurements including standard deviations (representative analyses listed as electronic supplement), recalculated structural formulae, end member proportions and mineral estimates (visual estimates from thin section) for anorthositic and mafic to ultramafic samples.

Table 3: Representative P-T estimates for granulite-facies assemblages and lower grade overprint. Numbers in brackets following the minerals represent analyses listed in supplement A.
Table 4: Calculated and estimated mineral modes, densities and seismic velocities for anorthositic and mafic to ultramafic samples. The relevant $P$-$T$ conditions used for the calculation are also shown in Fig. 7.

Table 5: Whole rock compositions of tonalitic gneiss (composition from Kern et al., 1999), a representative mafic granulite (composition from Hölttä et al., 2000) and homogeneous mixtures of anorthosite (An)/tonalitic gneiss (TG) with mafic to ultramafic rocks (see Fig. 8 and 9). Numbers indicate the proportion of mafic component in the mixture.

Table 6: Calculated mineral modes and densities for mixtures of anorthosite/tonalite with increasing proportion of mafic to ultramafic rocks. Some corresponding density diagrams are shown in Fig. 8 including the relevant $P$-$T$ points.
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<th>SH09-36 Anorthosite</th>
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<td>0.18</td>
<td>0.12</td>
<td>0.15</td>
<td>0.18</td>
<td>0.12</td>
<td>0.07</td>
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<td>0.02</td>
<td>0.04</td>
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<td>0.01</td>
<td>0.01</td>
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<td>99.84</td>
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<td>100.82</td>
<td>99.71</td>
<td>100.78</td>
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<table>
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<th>SH09-32 Contact zone</th>
<th>SH09-37 Mafic to ultramafic</th>
<th>SH09-42 Pyroxenite</th>
<th>SH09-44 Pyroxenite</th>
<th>SH09-01 Garnetite</th>
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<td>9.05&lt;sup&gt;T&lt;/sup&gt;</td>
<td>4.71</td>
<td>1.15</td>
<td>9.29&lt;sup&gt;T&lt;/sup&gt;</td>
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<td>5.09</td>
<td>-</td>
<td>6.00</td>
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<td>11.82</td>
<td>12.48</td>
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<td>&lt; 0.01</td>
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<tr>
<td>P₂O₅</td>
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<td>&lt; 0.01</td>
<td>&lt; 0.01</td>
<td>0.01</td>
<td>&lt; 0.01</td>
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<tr>
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<td>-0.11&lt;sup&gt;a&lt;/sup&gt;</td>
<td>1.16</td>
<td>-0.63&lt;sup&gt;a&lt;/sup&gt;</td>
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<tr>
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<td>99.78</td>
<td>99.16</td>
<td>99.96</td>
<td>100.07</td>
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<sup>T</sup> Total iron
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<tr>
<th>Mineral</th>
<th>SH09-30 (anorthosite)</th>
<th>SH09-35 (mafic anorthosite)</th>
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<td>Grt</td>
</tr>
<tr>
<td>Analyses</td>
<td>3</td>
<td>sd</td>
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<tr>
<td>TiO2</td>
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<td>0</td>
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<tr>
<td>Al2O3</td>
<td>29.87</td>
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<tr>
<td>Cr2O3</td>
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<td>0</td>
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<tr>
<td>Fe2O3</td>
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<tr>
<td>MnO</td>
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<td>0.03</td>
</tr>
<tr>
<td>MgO</td>
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<td>0.2</td>
</tr>
<tr>
<td>CaO</td>
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<td>2.35</td>
</tr>
<tr>
<td>Na2O</td>
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<td>0.14</td>
</tr>
<tr>
<td>K2O</td>
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<td>0.05</td>
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<tr>
<td>Total</td>
<td>99.42</td>
<td>99.81</td>
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* All analyzed FeO recalculated to Fe2O3 (w/r Fe/O = 1.11)
<table>
<thead>
<tr>
<th>Sample</th>
<th>Water-free assemblages</th>
<th>T (°C)</th>
<th>sd</th>
<th>P (GPa)</th>
<th>sd</th>
<th>Independent set of reactions</th>
</tr>
</thead>
</table>
| Grt(22)-Pl(21)-Cpx-I(42)-Spl(32) [1] | 970 | 347 | 1.3 | 0.35 | 3 Di +3 CaTs = Prp + 2 Grs  
2 Di + An + Spl = Prp + Grs  
2 Hed + An + Hc = Grs + Alm  
Grs + 3 An + 3 Hc = Alm + 6 CaTs |
| Grt(24)-Pl(36)-Cpx-I(43)-Spl(36) [2] | 1065 | 292 | 1.08 | 0.25 | 2 Di + An + Spl = Prp + Grs  
Di + 2An + 2 Spl = Prp + 3 CaTs  
2 Hed + An + Hc = Grs + Alm  
Hed + 2 An + 2 Hc = Alm + 3 CaTs |
| Grt-Pl-Cpx-Spl [3] | 900 | 100 | 1 | 0.15 | Prp in Grt and Di in Cpx compositions (see Fig. 9a) |
| SH09-35 | | | | | |
| Grt(24)-Pl(36)-Cpx-I(43)-Spl(36)-Crn [4] | 805 | 112 | 1.13 | 0.22 | 2 Di + An + Spl = Prp + Grs  
Prp + Grs + 4 Crn = 3 An + 3 Spl  
Di + 2 An + 2 Spl = Prp + 3 CaTs  
2 Hed + An + Hc = Grs + Alm  
3 Hed + 2 Crn = Grs + Alm |
| Grt(23)-Pl(21)-Cpx-II(18)-Opx(28)-Spl(30) [5] | 904 | 194 | 1.1 | 0.18 | En + MgTs = Prp  
3 Di + 3 MgTs = 2 Prp + Grs  
3 Di + 3 CaTs = Prp + 2 Grs  
2 Di + An + Sp = Prp + Grs  
2 Hed + An + Hc = Grs + Alm  
Grs + 3 An + 3 Hed = Alm + 6 CaTs  
Grs + Fs = 2 Hed + CaTs |
| SH09-32 | | | | | |
| Grt(41)-Pl(34)-Cpx(39)-Spl(29) [1] | 890 | 118 | 1.17 | 0.13 | 3 Di +3 CaTs = Prp + 2 Grs  
2 Di + An + Spl = Prp + Grs  
2 Hed + An + Hc = Grs + Alm  
Grs + 3 An + 3 Hc = Alm + 6 CaTs |
| Grt(41)-Pl(34)-Cpx(39)-Spl(29)-Crn [2] | 852 | 68 | 1.17 | 0.13 | 3 Di +3 CaTs = Prp + 2 Grs  
2 Di + An + Spl = Prp + Grs  
2 Hed + An + Hc = Grs + Alm  
Grs + 3 An + 3 Hc = Alm + 6 CaTs |
<table>
<thead>
<tr>
<th>Sample</th>
<th>Inclusions</th>
<th>Mineral Composition</th>
<th>X-ray</th>
<th>Vickers</th>
<th>Zeta</th>
<th>Alpha</th>
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</thead>
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<td>SH09-37</td>
<td>Grt-Pl-Cpx-Spl</td>
<td>1</td>
<td>750</td>
<td>100</td>
<td>0.85</td>
<td>0.15</td>
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<tr>
<td>SH09-01</td>
<td>Grt-Cpx-I-Pl-Spl</td>
<td>2</td>
<td>955</td>
<td>159</td>
<td>1.22</td>
<td>0.17</td>
</tr>
</tbody>
</table>

**Chemical Equations**

- **Grt (66)-Cpx (67)-Pl(55)-Spl(63)**
  - Gr + 2 Crn = 3 CaTs
  - 3 Di + 3 CaTs = Prp + 2 Grs
  - 2 Di + An + Spl = Prp + Grs
  - 2 Hed + An + Hc = Grs + Alm
  - 3 Hed + 2 Crn = Grs + Alm

- **Grt (80)-Cpx (81)-Pl(85)-Spl(84)**
  - 3 Di + 3 CaTs = Prp + 2 Grs
  - 2 Di + An + Spl = Prp + Grs
  - 2 Hed + An + Hc = Grs + Alm
  - Grs + 3 An + 3 Hc = Alm + 6 CaTs

- **Grt (87)-Cpx (92)-Opx (91)-Pl (95)-Spl (84)**
  - En + MgTs = Prp
  - 3 Di + 3 MgTs = 2 Prp + Grs
  - 2 Di + An + Spl = Prp + Grs
  - Grs + 3 An + 3 Spl = Prp + 6 CaTs
  - Grs + 3 Fs = Alm + 3 Hed
  - 2 Hed + An + Hc = Grs + Alm
  - Gr + 3 An + 3 Hc = Alm + 6 CaTs

- **Grt (57)-Cpx-I (56)-Pl (58)-Spl (51)**
  - 3 Di + 3 CaTs = Prp + 2 Grs
  - 2 Di + An + Spl = Prp + Grs
  - 2 Hed + An + Hc = Grs + Alm
  - Grs + 3 An + 3 Hc = Alm + 6 CaTs

**Notes**

- Prp in Grt and Di in Cpx compositions (Fig. 9b)

---

**SH09-37**

- Grt (66)-Cpx (67)-Pl(55)-Spl(63)
  - 1041 | 223 | 1.18 | 0.21

**SH09-01**

- Grt (87)-Cpx-I (67)-Pl(69)-Spl(75)
  - 955 | 159 | 1.22 | 0.17

---

**SH09-37**

- Grt (80)-Cpx (81)-Pl(85)-Spl(84)
  - 1066 | 353 | 1.25 | 0.33

**SH09-01**

- Grt (57)-Cpx-I (56)-Pl (58)-Spl (51)
  - 1181 | 351 | 1.49 | 0.34
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<tr>
<th>Composition</th>
<th>wt%</th>
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<th>SiO2</th>
<th>TiO2</th>
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<td>127</td>
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<tr>
<td>Grt(81)-Cpx-II(79)-Opx(82)-Pl(70)-Spl(75)</td>
<td>895</td>
<td>119</td>
<td>0.85</td>
<td>0.13</td>
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</tbody>
</table>

3 Di + 3 CaTs = Prp + 2 Grs
2 Di + An + Spl = Prp + Grs
2 Hed + An + Hc = Grs + Alm
Grs + 3 An + 3 Hc = Alm + 6 CaTs

Prp in Grt and Di in Cpx compositions (see Fig. 9c)

6 Fs + 3 An + 3 Hc = Grs + 5 Alm

C analyses of mineral cores
r analyses of rims
Table 4.

<table>
<thead>
<tr>
<th>Sample</th>
<th>SH09-30 (anorthosite)</th>
<th>SH09-35 (mafic anorthosite)</th>
<th>SH09-32 (contact zone)</th>
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<td></td>
<td>Rock b</td>
<td>950 °C; 1.0 GPa (a)</td>
<td>950 °C; 1.0 GPa (a)</td>
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<tr>
<td></td>
<td></td>
<td>660 °C; 0.85 GPa (b)</td>
<td>660 °C; 0.85 GPa (b)</td>
</tr>
<tr>
<td>Grt (vol%)</td>
<td>1</td>
<td>2</td>
<td>16</td>
</tr>
<tr>
<td>Cpx (vol%)</td>
<td>4 c</td>
<td>3</td>
<td>14</td>
</tr>
<tr>
<td>Pl (vol%)</td>
<td>95</td>
<td>94 d</td>
<td>93 d</td>
</tr>
<tr>
<td>Spl (vol%)</td>
<td>-</td>
<td>&lt;1 acc</td>
<td>2</td>
</tr>
<tr>
<td>Crn (vol%)</td>
<td>acc</td>
<td>acc</td>
<td>acc</td>
</tr>
<tr>
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<td>2760 ± 30</td>
</tr>
<tr>
<td>ρ (kg/m³); H*</td>
<td>-</td>
<td>2740 ± 30</td>
<td>2760 ± 30</td>
</tr>
<tr>
<td>v_P (km/s); H*</td>
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<td>6.8 ± 0.2</td>
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<tr>
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<td>-</td>
<td>6.6 ± 0.2</td>
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<th>SH09-42 (pyroxenite)</th>
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<td>950 °C; 1.0 GPa (a)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>660 °C; 0.85 GPa (b)</td>
<td>660 °C; 0.85 GPa (b)</td>
</tr>
<tr>
<td>Grt (vol%)</td>
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<td>84</td>
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<td>9</td>
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<tr>
<td>Spl (vol%)</td>
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<td>2</td>
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<tr>
<td>Crn (vol%)</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
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<td>-</td>
<td>7.9 ± 0.2</td>
<td>7.8 ± 0.2</td>
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a Densities and velocities obtained by the spreadsheet from Hacker and Abers (2004)
b Mineral estimates form thin section; density measured with pycnometer on a section of rock drill cores
c Estimate of Cpx based on the amount of Ep+Chl
d Sum of Pl and Kfs which have been listed separately
e Extrapolation from symplectitic intergrowths
Table 5.

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<th>TG55</th>
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<td>0.8</td>
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<td>0.1</td>
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<td>11.2</td>
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<td>2.6</td>
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<td>2.4</td>
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<td>2.9</td>
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<td>0.1</td>
<td>1.6</td>
<td>1.4</td>
<td>1.2</td>
<td>0.9</td>
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³ Total iron calculated as Fe₂O₃* 0.899 + FeO
* Measured as Fe₂O₃
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<th>Anorthosite with</th>
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<th>45% mafic (An45)</th>
<th>50% mafic (An50)</th>
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</thead>
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<tr>
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<tr>
<td>Grt (vol.%)</td>
<td>1.0 GPa (a) 0.85 GPa (b) 1.5 GPa (c) 1.0 GPa (a) 0.85 GPa (b) 1.5 GPa (c) 1.0 GPa (a) 0.85 GPa (b) 1.5 GPa (c) 1.0 GPa (a) 0.85 GPa (b) 1.5 GPa (c)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cpx (vol.%)</td>
<td>19 19 13 19 22 16 21 24 24 24 60</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pl (vol.%)</td>
<td>62 60 - 58 55 - 53 49 -</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Spl (vol.%)</td>
<td>1 - 1 - - 2 - - -</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ky (vol.%)</td>
<td>- - 18 - - - - -</td>
<td></td>
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</tr>
<tr>
<td>$p_1$ (kg/m$^3$)</td>
<td>3010 ± 30 3060 ± 30 3440 ± 30 3050 ± 30 3100 ± 30 3460 ± 35 3090 ± 30 3140 ± 30 3470 ± 35</td>
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</tr>
<tr>
<td>$p_2$ (kg/m$^3$)</td>
<td>3060 ± 30 3080 ± 30 3250 ± 30 3100 ± 30 3125 ± 30 3290 ± 30 3140 ± 30 3165 ± 30 3320 ± 30</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Tonalitic gneiss with</th>
<th>50% mafic (TG50)</th>
<th>55% mafic (TG55)</th>
<th>60% mafic (TG60)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>950 °C; 660 °C; 650 °C; 950 °C; 660 °C; 650 °C; 950 °C; 660 °C; 650 °C; 950 °C; 660 °C; 650 °C;</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Grt (vol.%)</td>
<td>1.0 GPa (a) 0.85 GPa (b) 1.5 GPa (c) 1.0 GPa (a) 0.85 GPa (b) 1.5 GPa (c) 1.0 GPa (a) 0.85 GPa (b) 1.5 GPa (c) 1.0 GPa (a) 0.85 GPa (b) 1.5 GPa (c)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cpx (vol.%)</td>
<td>13 17 45 14 18 45 15 20 45</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fsp (vol.%)</td>
<td>61 52 16 59 49 14 57 47 12</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Opx (vol.%)</td>
<td>20 8 - 20 8 - 21 8 -</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ky vol. (%)</td>
<td>- - 3 - - - - -</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$p_1$ (kg/m$^3$)</td>
<td>2990 ± 30 3105 ± 30 3430 ± 30 3010 ± 30 3135 ± 30 3450 ± 35 3030 ± 30 3160 ± 30 3470 ± 35</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$p_2$ (kg/m$^3$)</td>
<td>3170 ± 30 3200 ± 30 3340 ± 30 3200 ± 30 3240 ± 30 3370 ± 30 3240 ± 30 3270 ± 30 3395 ± 30</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Fig. 4

(a) Chl, Grt, Crn, Chl + Ep (Cpx), PI

(b) PI, Spl, Cpx-Spl-PI-Opx-Amp, Chl, Grt

Scale: 1 cm
a) **Fig. 9** SH09-35 (mafic anorthosite)

![Diagram](image)


b) **SH09-37** (mafic to ultramafic)

![Diagram](image)


c) **SH09-01** (gametite)

![Diagram](image)

Fig. 10  Anorthosites:

SH09-30 (anorthosite)

1: Grt Cpx Pl Opx Spl Kfs Ilm

SH09-35 (mafic anorthosite)

1: Cpx Pl Opx Spl Ilm; 2: Grt Cpx Pl Opx Ol Spl Ilm; 3: Grt Cpx Pl Spl Crn Ilm; 4: Grt Cpx Pl Spl Rt; 5: Grt Cpx Pl Rt

Mafic to ultramafic rocks:

SH09-37 (mafic to ultramafic)

1: Cpx Pl Opx Ol Spl Ilm; 2: Grt Cpx Pl Opx Ol Spl Ilm; 3: Grt Cpx Pl Spl Crn Ilm; 4: Cpx Pl Opx Ol Spl Ilm; 5: Grt Cpx Pl Opx Ol Spl Ilm

SH09-01 (garnetite)

1: Cpx Pl Opx Ol Spl Ilm; 2: Grt Cpx Pl Opx Ol Spl Ilm; 3: Grt Cpx Pl Spl Crn Ilm; 4: Cpx Pl Opx Ol Spl Ilm; 5: Grt Cpx Pl Opx Ol Spl Ilm
Mixtures with anorthosite:

Fig. 11

Mixtures with tonalite:

1: Grt Cpx Pl Opx Spl Kfs Ilm; 2: Grt Cpx Pl OI Spl Kfs Ilm