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

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1 **Effect of small-scale heterogeneities on**
2 **interpretation of crustal compositions exemplified**
3 **by a layered anorthosite**

4

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10

11 **Abstract**

12

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

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20 explored on an exposed high-grade layered anorthositic body by providing a detailed field
21 map, petrological descriptions, pycnometry measurements as well as whole rock and mineral
22 analyses combined with thermodynamic phase equilibria calculations. To evaluate the results
23 of our thermodynamic calculations, densities and mineral modes obtained from the modeled
24 phase equilibria are compared to measured densities and estimated mineral modes from rock
25 samples. The proportion of mafic to ultramafic (plagioclase-poor) rocks in the mapped field
26 area amounts to 10-15% but higher proportions of these rock types in the lower crust are
27 feasible. To further study the effects of compositional variations, we have generated mixtures
28 of mafic to ultramafic and anorthositic/intermediate rocks until the average properties of these
29 mixtures are comparable to those of mafic granulites (3000-3100 kg/m³; 7.1-7.3 km/s).
30 Mixtures of anorthosite with 40-45% and of tonalite with 50-60% high-grade mafic to
31 ultramafic rocks yield average densities and seismic velocities similar to mafic granulites
32 although they still contain 50-60 vol.% plagioclase. Hence small-scale mixing of certain rock
33 types may result in the overestimation of the proportion of mafic (garnet) granulites in the
34 lithologic interpretation of crustal compositions from seismic data. Since the transition to
35 eclogite-facies in plagioclase-rich rocks is shifted to higher pressures and
36 anorthositic/intermediate eclogites yield lower densities, a lower crust with higher modal
37 amounts of plagioclase may not always provide the significant densification needed for certain
38 geodynamic settings (e.g. delamination or subsidence).

39
40 Keywords: Layered Anorthosite; Heterogeneities; Lower Crust; Densities

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42

43 1. Introduction

44

45 The composition of the lower crust plays an important role in geodynamic processes.
46 Granulite terranes and xenoliths are the two main sources providing insights in the
47 composition of lower crustal rocks. While granulite terranes suggest that felsic to intermediate
48 rocks are major constituents of the lower crust (e.g. Austrheim, 2013), the majority of
49 xenoliths as well as seismic studies indicate a predominantly mafic lower crustal composition
50 (Christensen and Mooney, 1995; Rudnick and Fountain, 1995). In addition to this
51 compositional ambiguity, granulite terranes and crustal cross-sections often show a large
52 extent of lateral and vertical heterogeneities in the form of mafic and felsic layering on varying
53 scales (Fountain and Salisbury, 1981; Harley, 1989; Baldwin et al., 2006; Miller and Snoke,
54 2009; Williams et al., 2009; Sinigoi et al., 2011). The presence of lithologic contrasts in the
55 lower crust which may be one of the causes of the observed deep crustal seismic reflectivity
56 is also corroborated by seismic modeling studies incorporating lithologies of exposed deep
57 crustal terranes and laboratory measurements. (Hale and Thompson, 1982; Hurich and
58 Smithson, 1987; Mooney and Brocher, 1987; Burke and Fountain, 1990; Reston, 1990;
59 Mooney and Meissner, 1992; Holbrook et al., 1992; Holliger and Levander, 1992;
60 Khazanehdari et al., 2000). Furthermore, Brittan and Warner (1996, 1997) have shown that
61 the size and spatial distribution of heterogeneities scaling in the range of kilometers to a few
62 hundred meters can lead to an over- or underestimation of the average velocity of the
63 heterogeneous medium. Smaller-scale heterogeneities such as small isolated lenses of
64 dense rocks, however, are generally not likely to be detected by wide angle soundings due to
65 poor lateral resolution resulting in averaged velocity over wide areas of the lower crust (e.g.

66 Kern et al., 1999; Kuusisto et al., 2006). Yet even thin-layered clusters within shear zones of
67 granulite- and/or eclogite-facies rocks can be the source of strong seismic reflectivity (e.g.
68 Fountain et al., 1994; Ji et al., 1997). In addition to poor lateral resolution, the detection of
69 heterogeneities in the lower crust may be difficult due to the non-unique lithological
70 interpretation of lower crustal rocks from seismic velocities since a variety of rock type
71 mixtures can yield the same velocities (Kuusisto et al., 2006; Kern, 2011). Additional
72 constraints can be established if the seismic data are correlated with experimentally
73 determined or calculated P- and S-wave velocities at lower crustal conditions for rock
74 samples collected from surface outcrops or from xenoliths (Christensen and Wepfer, 1989;
75 Jackson et al., 1990; Kern et al., 1999; Kuusisto et al., 2006; Kern, 2011). These findings can
76 be improved by detailed studies of the distribution and properties of heterogeneities on
77 exposed sections of the lower crust.

78 In the present paper, we provide a detailed geological map of anorthositic and mafic to
79 ultramafic rocks from the Sogndal area, Western Norway as an example for lower crustal
80 rocks exposing small scale heterogeneities. This particular area has been chosen due to the
81 accessibility of excellent outcrops and the relatively restricted lithology facilitating the
82 distinction between mafic to ultramafic (PI-poor) and anorthositic (PI-rich) rock types (we use
83 this terminology only to differentiate between PI-poor and PI-rich rocks within the layered body
84 as anorthosite is also classified as a mafic rock). By means of pycnometry, whole rock and
85 mineral compositions and geothermobarometry, we will compare measured and calculated
86 densities of these compositions. Anorthositic rocks have been reported to occur in deeper
87 levels of exposed crustal cross-sections, e.g. in the Bergen Arcs, Norway (Griffin, 1972),
88 Superior Province, Canada (Ermanovics and Davidson, 1976; Fountain and Salisbury, 1981)
89 and associated with granulite terranes (Berg, 1977; Kars et al., 1980; Glikson, 1986; Harley,

1989), but also have been found as xenoliths (Leyreloup et al., 1982). 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.

103

104 **2. Composition of the Sogndal anorthosites**

105

106 **2.1. Geological background**

107

108 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-

113 eastern part of the nappe is dominated by partially retrograded gneisses of granitic, syenitic,
114 monzonitic, dioritic and gabbroic compositions. In contrast, the central and southwestern parts
115 where the field area is located are mainly composed of high-grade anorthositic and gabbroic
116 rocks (Bryhni et al., 1983; Lundmark et al., 2007; Fig. 1a). Protolith ages of orthogneisses
117 within the Upper Jotun Nappe range between 1.66 and 1.26 Ga (Lundmark et al., 2007). The
118 rocks have been transformed to granulite-facies conditions during the Sveconorwegian
119 orogeny (1.25-0.9 Ga) followed by amphibolite-facies metamorphism $\sim 954 \pm 3$ Ma (Lundmark
120 et al., 2007). Another high-grade metamorphic event coinciding with local anatexis within
121 granitic gneiss has been dated to 934 ± 1 Ma and is interpreted to reflect post-collisional
122 elevated heat flow in the nappes (Lundmark et al., 2007). U-Pb dates of sparse zircons within
123 the massif anorthosite in the Jotun Nappe reflect events at 965 ± 4 and 913 ± 2 Ma as well as
124 a pronounced Caledonian metamorphic overprint (Lundmark and Corfu, 2008). The authors
125 interpret the older age as the protolith age of the anorthosite while the metamorphic event at
126 913 ± 2 Ma seems to be locally restricted to the outermost edge of the Western Gneiss
127 Region (Lundmark and Corfu, 2008).

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

137 the Jotun Nappe Complex (Lundmark and Corfu, 2008) and there is hence no indication for
138 these rocks having been exposed to higher pressures.

139 Due to its high aluminum content (25-30% Al_2O_3), anorthosite is interesting for
140 industrial purposes if the plagioclase contains more than 60% anorthite component enhancing
141 the solubility in acids (Wanvik, 2007). Since iron-bearing silicates would be unfavorable during
142 mining of anorthosite, studying the distribution and properties of heterogeneities in these
143 rocks may also help for resource estimates and mining strategies.

144

145 2.2. Analytical methods and modeling techniques

146

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

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

159 Rock densities of the six selected samples were calculated using the density of water
160 at 20 °C and the following equation: $\rho = 0.998 \text{ g/cm}^3 * m_d / (m_d - m_w) * 1000$ where the “dry” (m_d)

161 and “wet” (m_w) weight were measured with a balance at the Geological Museum, Oslo
162 (accuracy: one position after decimal point).

163 Average P - T estimates were calculated with THERMOCALC (Powell and Holland,
164 1994, 2008) using activities obtained by the program AX from selected analyses. Where a
165 clear distinction was possible, we have used core and rim compositions of adjacent mineral
166 grains.

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

171 Phase diagrams were calculated with the Gibbs free energy minimization software
172 Perple_X 6.6.6. (Connolly, 2005) using the thermodynamic dataset of Holland and Powell
173 (Holland and Powell, 1998 and updated version of 2002). For simplicity, iron was assumed to
174 be divalent and all oxides below 0.10 wt% (Table 1) were excluded resulting in the system
175 TiO_2 - Na_2O - CaO -(K_2O)- FeO - MgO -(MnO)- Al_2O_3 - SiO_2 (TiNC(K)FM(Mn)AS). For compositions
176 containing small amounts of K_2O , two separate solution models represent feldspar in order to
177 visualize the plagioclase-out reaction. Solid solution models were chosen to match the rock
178 lithologies as closely as possible (see supplement B for a detailed list of solution models).
179 Mineral modes, rock densities and seismic velocities were extracted from computed phase
180 diagrams at specified conditions. A standard error of ~3% was presumed for seismic velocities
181 (e.g. Connolly and Kerrick, 2002). For rock densities, however, we assumed a lower value of
182 ~1%.

183

184

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 $\sim 33 \text{ m}^2$ ($\sim 87\%$) of the mapped area but shows significant

209 compositional and mineralogical variation. Rocks containing high amounts of SiO_2 (>50 wt.%),
210 Al_2O_3 (25-29 wt.%) and Na_2O (~4 wt.%) are classified as anorthosites while samples with
211 lower SiO_2 (~45 wt.%) and Al_2O_3 (22-24 wt.%) but higher percentages of Fe_2O_3 (~7 wt.%)
212 and MgO (7-9 wt.%) due to higher amounts of mafic minerals are referred to as mafic
213 anorthosites (Table 1; Fig. 2). We have chosen one sample with ~95 volume percent
214 plagioclase (SH09-30; Table 2; Fig. 3a) and one sample with ~60-70 volume percent
215 plagioclase (SH09-35; Table 2; Fig. 3b) as representative end members (sample locations in
216 Fig. 1b).

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

220 In the anorthosite, garnet is present in only small amounts of 1-2 volume percent (Fig.
221 3a; Table 2) whereas the original amount of garnet in the mafic anorthosite is estimated to
222 ~15 volume percent (Fig. 3b; Table 2). Most garnet crystals, however, have been replaced by
223 symplectitic intergrowths consisting of clinopyroxene, spinel, plagioclase and orthopyroxene
224 leaving only garnet cores (Fig. 3b). Within these intergrowths, small rims of tschermakitic
225 amphibole have formed mostly around some spinel crystals. The remaining garnet cores
226 show a higher proportion of pyrope (Prp₄₀₋₅₀) compared to garnet in the anorthosite (Table 2).
227 Most garnet crystals are also partly replaced by chlorite (Fig. 3).

228 Clinopyroxene in the anorthosite has been completely replaced by chlorite and
229 epidote/clinozoisite (Fig. 3a). In the mafic anorthosite, clinopyroxene is present both as bigger
230 Al-rich (7-12 wt.% Al_2O_3) crystals (Cpx I) and as Al-poor (4-5 wt.% Al_2O_3) grains (Cpx II) in the
231 symplectites replacing garnet (Table 2; Fig. 3b).

232 The amount of spinel is also highly variable. While it is absent in the anorthosite (Fig.

233 3a) most other samples contain 1-3 volume percent spinel. Significantly higher amounts (~25
234 vol.%) and larger spinel crystals (3-5 mm) sometimes occur in contact zones between
235 anorthosite and mafic to ultramafic rocks (SH09-32; Fig. 3c). Some spinel crystals have been
236 replaced by small amounts of magnesium-rich chlorite and phlogopite indicating fluid-induced
237 alteration. Rarely, spinel is replaced by corundum. Single crystals of corundum are also
238 present in some thin sections (Fig. 3a).

239

240

241 2.5. Mafic to ultramafic rocks

242

243 The mafic to ultramafic portion of the mapped area amounts to 4.7 m² in addition to 0.3 m²
244 garnetite and 0.1 m² pyroxenite (totalling ~13% of the outcrop). The whole rock composition
245 of the homogeneous mafic sample and the garnetite yield similarly low amounts of SiO₂ (42-
246 43 wt.%) but significantly higher MgO (~11wt.%) and FeO (6-8 wt.%) compared to the
247 anorthosites (Table 1, Fig. 2). The pyroxenite, however, contains a higher percentage of SiO₂
248 (44-46 wt.%) and CaO (15-18 wt.%) but lower amounts of Al₂O₃ (~12 wt.%) than the other
249 mafic rock types. The mafic to ultramafic rocks contain 45-50 volume percent garnet and
250 clinopyroxene with less than 4 volume percent plagioclase and spinel (SH09-37; Fig. 3d).
251 Domains of garnetite (80-90 vol.% garnet; SH09-01; Fig. 3e) and pyroxenite (> 90 vol.%
252 pyroxenes; SH09-42; Fig. 3f) exist within the more homogeneous mafic rocks (Fig. 1c).

253 Garnet crystals are 2-3 mm in size (Fig. 3d) and show a zonation with Fe-enriched
254 (Alm₄₅) but Mg and Ca-depleted (Prp₃₃₋₃₆; Grs₁₆₋₁₉) rims compared to the cores (Alm₃₀₋₃₃Prp<sub>43-
255 45</sub>Grs₂₃; Fig. 4). Garnetite contains large (< 1 cm) garnet crystals (upper part of Fig. 3e).
256 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.% Al_2O_3) 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 (Ab_{~60}; 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 symplectitic 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

280 densities (e.g. Connolly and Petrini, 2002). However, both estimation methods require the
281 knowledge of a P - T range for the mineral assemblage. Hence, we have combined
282 geothermobarometric results and petrological constraints to obtain a possible P - T range for
283 most of the studied samples.

284

285 **3.1. Geothermobarometry**

286

287 The high-grade assemblage in all samples is interpreted to consist of garnet, plagioclase,
288 clinopyroxene and spinel (non symplectitic). Only the Al-rich clinopyroxene (Cpx I) is assigned
289 to be part of this assemblage because the Al-poor clinopyroxene (Cpx II) is restricted to the
290 symplectitic intergrowths. P - T conditions obtained with THERMOCALC range between 950-
291 1070 ± 200 °C at 1.0 to 1.4 ± 0.2 GPa (Table 3). The set of independent reactions used to
292 determine these P - T conditions is shown in Fig. 5.

293 Some anorthositic samples contain either isolated corundum crystals or show a partial
294 replacement of spinel by corundum. Assemblages with corundum yield lower T (800 - $850 \pm$
295 100 °C) at P of 1.1 - 1.2 ± 0.2 GPa (Table 3).

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

301

302 **3.2. P - T assessment by pseudosection calculations and mineral composition**

303

304 In addition to geothermobarometry, we have compared selected mineral compositions from
305 microprobe analyses (Table 2 and supplement A) with compositions calculated by the
306 program werami (Perple_X) within the observed mineral assemblage for three samples (Fig.
307 6; Table 3). Isopleths shown in Fig. 6 represent pyrope content in garnet as well as diopside
308 content in clinopyroxene. The resulting P - T estimate for the mafic anorthosite of $\sim 900 \pm 100$
309 $^{\circ}\text{C}$ and 1.0 ± 0.15 GPa (ellipse in Fig. 6a) is in accordance with the values obtained by
310 THERMOCALC (Table 3). For the mafic to ultramafic samples, however, the temperature
311 estimates for the high-grade assemblage of 750 ± 100 $^{\circ}\text{C}$ are significantly lower than the
312 results from THERMOCALC (Fig. 5 and 6b, c; Table 3) plotting outside of the plagioclase
313 stability field.

314

315 **3.3. Comparison of measured and computed rock properties**

316

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

328 (1995) for conditions between 30 and 40 km at average to high temperatures.

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

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

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

347

348 **3.4. Effect of retrogression on physical rock properties**

349

350 Retrogression is incomplete and inhomogeneous throughout the samples and mainly
351 localized around mafic minerals (in particular garnet). This suggests that the rocks were

352 exposed to small and varying amounts of fluids possibly moving along weak zones such as
353 cracks. Phase equilibria modeling, on the other hand, assumes complete equilibrium and
354 hence needs some modification to reproduce the partial retrogression. We added small
355 amounts of water to see if the modes of water-bearing minerals observed in the samples can
356 be approximated in the model.

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

376

377 **3.5. Effect of mixing on rock properties**

378 To study the effect of average rock properties on heterogeneous lower crust we have
379 calculated phase diagrams for mixtures of anorthosite (whole rock composition of SH09-30,
380 Table 1) with systematically increasing amounts of mafic to ultramafic proportion (composition
381 of SH09-37, Table 1). Another batch of mixtures with a more common rock type has been
382 calculated with the whole rock composition of a tonalitic gneiss (sample D95-16 from Kern et
383 al., 1999; Table 5) in combination with the mafic composition. Densities have been obtained in
384 two different ways: 1) The oxides of the anorthositic/tonalitic compositions are mixed in the
385 given proportions with the oxides of the mafic composition (see Table 5 for whole rock
386 compositions). This approach enables the calculation of phase diagrams (Fig. 8) and the
387 extraction of densities (ρ_1 in Table 6) but assumes a homogeneous mixture. 2) Densities are
388 extracted separately from the calculated phase equilibria of the anorthosite/tonalite and mafic
389 rock at specified conditions and then are summed up in the given proportions (ρ_2 in Table 6).
390 This method assumes that the different rock types do not react with each other. The density
391 increases significantly from 5 to 40% in mixtures with anorthosite and mafic to ultramafic
392 rocks (Fig. 8a,b) and from 5 to 50% in mixtures with the tonalite and mafic composition (Fig.
393 8c,d). As before, mineral modes and densities have been extracted at two P - T conditions in
394 the stability field of the high-grade assemblage and for a P - T point representing eclogite-
395 facies conditions (650 °C at 1.5 GPa, Table 6). Temperatures of 950 °C might be too high to
396 reflect the conditions of continental lower crust but a lack of re-equilibration and therefore
397 preservation of the high temperature mineralogy of granulites at lower temperatures is likely.
398 Due to the relatively high P -wave velocity but low density of the anorthosite, mixtures with
399 relatively low mafic portions (10-25%) will yield low densities (~ 2800 - 2900 kg/m³) similar to

400 felsic granulites or mica-quartz schist (Christensen and Mooney, 1995) but too high P-wave
401 velocities (6.8-7.0 km/s) for these rock types and should therefore be recognizable as
402 mixtures. Anorthositic mixtures with 45-50% mafic to ultramafic rocks (Fig. 8c), however,
403 resemble a mafic granulite or mafic garnet granulite with densities in the range of 3000-3100
404 kg/m³ (Table 6; Fig. 9) and velocities of 7.1 to 7.3 km/s. Fig. 9 demonstrates that the
405 properties of all mixtures behave according to Birch's law (e.g. Birch, 1961a, 1961b) showing
406 the seismic velocity as an approximately linear function of the density. In addition, the Poisson
407 ratios of ~0.27 for these mixtures agree well with the range of 0.27-0.28 determined for a
408 number of mafic granulites (Kern et al., 1996, 1999). However, all values obtained from
409 thermodynamic modeling are for idealized, isotropic, homogeneous aggregates without
410 considering any textures so velocities might diverge due to the spatial orientation of minerals
411 and even small heterogeneities.

412 The difference between extracted densities from homogeneous mixtures (ρ_1 ; Table 6)
413 and averaged densities from separate rock compositions (ρ_2 ; Table 6) is mostly within 50
414 kg/m³ for mixtures with anorthosite at the two *P-T* conditions (points a and b; Table 6). For
415 eclogite-facies conditions, however, ρ_2 is significantly lower (~150-200 kg/m³) than the density
416 for the homogeneous mixture. For tonalitic mixtures and granulite-facies conditions, ρ_2 is
417 significantly higher (~100-200 kg/m³) than ρ_1 while this relation is reversed for the densities
418 representing eclogite-facies conditions.

419

420 **4. Discussion**

421 **4.1. Evaluation of thermobarometry and phase diagrams**

422

423 All samples in our field area record a high-grade assemblage of garnet, clinopyroxene,
424 plagioclase and spinel. According to Lundmark and Corfu (2008) U-Pb ages of zircons in
425 anorthosite point to a metamorphic episode at 913 ± 2 Ma affecting the layered anorthosite
426 after its intrusion (965 ± 4 Ma). Additional information about the pressure-temperature
427 evolution of the layered anorthosite is provided by coronas with garnet rims and olivine cores
428 as reported by Griffin (1971). To explain corona formation between olivine and plagioclase,
429 Griffin (1971) proposed slow cooling after the crystallization of anorthosite in combination with
430 an increase in pressure to account for the formation of garnet. Alternatively, garnet may have
431 formed by granulite-facies metamorphism which affected the rocks to varying degrees due to
432 deformation and/or fluid infiltration accounting for the preservation of coronas as well as the
433 formation of granulite-facies assemblages. This is corroborated by Sm-Nd ages of 907 ± 9 ,
434 912 ± 18 and 905 ± 37 Ma (Cohen et al., 1988) from pyroxene and garnet in coronas
435 occurring in the Lindås troctolite overlapping with the ages determined by Lundmark and
436 Corfu (2008). Therefore the authors propose that these dates either reflect isotopic re-
437 equilibration of preexisting garnet and clinopyroxene or (partial) formation of the coronas
438 during the metamorphic event (Lundmark and Corfu, 2008). Since the coronas exhibit two
439 stages of mineral reactions (Griffin, 1979) there is a possibility of the first reaction to have
440 occurred during cooling from magmatic temperatures while the second could be related to the
441 913 Ma event (Lundmark and Corfu, 2008). Based on these results, we assume that at least
442 some of the recorded high-grade assemblage is due to granulite-facies metamorphism and
443 unrelated to cooling of the anorthosite intrusion. A further discussion, however, is beyond the
444 scope of this paper since our conclusions depend only on the rock properties of the observed
445 rocks and not on the conditions and time of their formation.

446 Several *P-T* estimates obtained with THERMOCALC, especially for the garnetite

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

461

462 **4.2. Suitability of the studied rocks as a lower crustal analogue**

463 The emplacement of massif anorthosite bodies is commonly restricted to the range of 5-13
464 km but can occur at depths as great as 23-37 km (e.g. Morse, 1982) corresponding to upper-
465 middle crustal levels. In this case, however, we are studying a high-grade layered anorthosite
466 resting on top of the Caledonian orogen (Lundmark and Corfu, 2008) and representing a part
467 of the Jotun Nappe Complex. Based on the present location of the studied rocks, U-Pb
468 geochronology of the anorthosites (Lundmark and Corfu, 2008) as well as adjacent rocks
469 within the Upper Jotun Nappe (Lundmark et al., 2007) and the observed high-grade mineral
470 assemblage, we have to assume that these rocks have been exposed to high temperatures

471 and moderate pressures during one or several events since their emplacement. The result is
472 a mineral assemblage resembling granulite-facies rocks which dominate the lower crust (e.g.
473 Rudnick and Fountain, 1995). Therefore, we consider the studied rocks to be a good
474 analogue for lower crustal rocks by mineralogical aspects.

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

486

487 **4.3. Evaluation of calculation methods**

488 Densities obtained with Perple_X and the worksheet of Hacker and Abers (2004) match within
489 20-40 kg/m³ for most compositions (Table 4), approximating an aberration of ~1%. Given that
490 the standard error for densities obtained by Perple_X is ~1%, and that accessories such as
491 corundum and ilmenite have not been included in the spreadsheet, the values are in good
492 agreement with each other. Similarly, P-wave velocities obtained from the spreadsheet are 0.1
493 to 0.2 km/s lower than the values extracted from Perple_X, corresponding to a discrepancy of
494 maximal ~3% and therefore are within the error.

495 For relatively homogeneous and almost monomineralic samples such as the
496 anorthosite (SH09-30) there is a very good agreement between measured and calculated
497 densities. The significant discrepancy in density for the more heterogeneous compositions,
498 especially the mafic anorthosite (SH09-35) and contact zone (SH09-32) is a result of the
499 retrogression of high-grade minerals as well as the high variation in mineral modes due to the
500 extremely heterogeneous nature of the drill cores. Since domains of garnetite alter on very
501 small scales with the other mafic to ultramafic rocks the sample used for the whole rock
502 composition may represent a more heterogeneous mixture instead of pure garnetite which
503 could explain the discrepancy in estimated and calculated mineral modes for this rock type.

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

510

511 **4.4. Assessment of heterogeneities**

512

513 Our mapped area shows mafic to ultramafic and anorthositic heterogeneities on outcrop scale
514 down to micron scale. The proportions of anorthosite and mafic rocks (including garnetite and
515 pyroxenite) have been determined to amount to ~87 and ~13%, respectively. Due to the high
516 variation of mafic minerals in the anorthositic sections, the mafic percentage may be
517 somewhat higher. In addition, the small mapped area shows an irregular distribution of the
518 mafic to ultramafic pods and bands and it is possible that other areas regionally within the

519 anorthosite unit contain more mafic minerals.

520 The level of origin within the crustal column for the anorthosites is not known but
521 seismic data and xenolith studies indicate that the lower crust becomes significantly more
522 mafic with depth (e.g. Christensen and Mooney, 1995; Kuusisto et al., 2006; Rudnick and
523 Gao, 2003). Furthermore, local positive Bouguer anomalies over a number of anorthosite
524 bodies suggest that more mafic rocks are located beneath the anorthositic rocks in the lower
525 crust or in the upper mantle (Morse, 1982 and references therein). It is therefore possible that
526 the amount of mafic to ultramafic intrusions increase in the lower parts of anorthositic bodies.

527 The observation of small scale heterogeneities in combination with the indication for a
528 mafic (plagioclase-poor) lower crust has motivated the mixing of mafic and
529 anorthositic/intermediate compositions until the average properties are comparable to mafic
530 granulites (Fig. 9). Assuming nearly dry conditions (e.g. very low amounts of hydrous
531 minerals), much lower amounts of mafic to ultramafic (40-60%) are sufficient for the mixture to
532 resemble mafic garnet granulites despite quite high amounts of plagioclase (50-60 volume
533 %).

534

535 **4.5. Consequences for geodynamic models**

536

537 Phase transitions in the lower crust, especially during formation of eclogite from granulite or
538 amphibolite have been suggested as mechanism for various geodynamic processes such as
539 basin subsidence (e.g. O'Connell and Wasserburg, 1972; Mareschal and Lee, 1983; Hamdani
540 et al., 1994; Artyushkov, 2005, 2010) or delamination (e.g. Kay and Mahlburg Kay, 1993; Gao
541 et al., 1998; Kukkonen et al., 2008). The requirement for these geodynamic settings is the
542 densification of the lowermost crust to a degree that exceeds the density of the underlying

543 mantle. In many cases, the seismic data lead to the conclusion that the lower crust is
544 predominantly composed of mafic granulites and this densification might be feasible.
545 However, we have shown that mixtures of certain rock types (anorthosites or tonalitic
546 compositions combined with mafic to ultramafic rocks) can contain far more plagioclase
547 despite the resemblance of their geophysical properties to mafic granulites. Higher amounts
548 of plagioclase component in the rock shifts the characteristic eclogite reaction (plagioclase to
549 clinopyroxene) to higher pressures (as demonstrated in Fig. 3 of Semprich and Simon, 2014).
550 Therefore the significant densification associated with eclogitization is expected at different
551 pressure.

552 In the following an example is given to illustrate some geodynamic consequences of
553 lithological heterogeneity on densification. The lowermost part of the crust is assumed to
554 consist of alternating bands of anorthosite/tonalite and mafic to ultramafic rocks similar to the
555 layered anorthosite presented in this paper. However, to ensure comparability to granulite-
556 facies lower crustal rocks higher amounts of mafic to ultramafic compositions are taken in this
557 example. In addition to a significant densification, delamination also depends on the rheology
558 and therefore the temperature of the mantle (e.g. Kay and Mählburg Kay, 1993). Delamination
559 of the lowermost eclogitized crust has been suggested for thickened crust where the crust-
560 mantle boundary shows almost no density contrast (e.g. Fennoscandian Shield; Kukkonen et
561 al., 2008). Geophysical constraints require a relatively high temperature to ensure a
562 sufficiently low viscosity of the mantle (Kukkonen et al., 2008). However, high temperature
563 reduces density. Therefore, sufficient densification and subsequent delamination is only
564 possible at higher pressure (Semprich and Simon, 2014). An increase in pressure can be
565 achieved by additional burial or by alternative pressure generating mechanisms such as
566 tectonic or reactive stress (e.g. Petrini 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: $z = P / 9.81 \text{ m/s}^2 * 3000 \text{ kg/m}^3$ 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 ($\sim 3320 \text{ kg/m}^3$). 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 $\geq 3447 \text{ kg/m}^3$. 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 $\sim 1.7\text{-}1.8 \text{ GPa}$ at $900\text{-}1000^\circ\text{C}$ to yield higher density than the mantle (Table 3; Semprich and Simon). In comparison, the anorthosite still

591 contains ~50% of plagioclase and yields a density of ~ 2970 kg/m³ at 900 °C and 2.0 GPa and
592 is therefore not dense enough to delaminate. A mix of anorthosite with 50% of exceptionally
593 dense mafic to ultramafic bands, however, reaches a cumulative density above 3220 kg/m³ at
594 pressure of ~1.8 GPa. To ensure these conditions, the rocks either have to be at a greater
595 depth of ~10-15 km or be exposed to additional tectonic pressure of ~0.3-0.5 GPa. This value
596 is compatible with tectonic overpressure obtained in geodynamic numerical simulations (e.g.
597 Petrini and Podladchikov, 2000; Schmalholz and Podladchikov, 2013; Schmalholz et al.,
598 2014).

599

600 **5. Conclusions**

601 High-grade anorthosites interlayered on all scales with mafic to ultramafic rocks forming
602 irregular shapes or more continuous bands at Sogndal may represent parts of the lower crust.
603 Geothermobarometry and a comparison of observed and modeled mineral modes suggest
604 the equilibration of the rocks to high-grade conditions of 750-1070 ± 200 °C and 0.85 to 1.4 ±
605 0.2 GPa followed by one or more episodes of lower grade overprint recorded by the formation
606 of corundum, symplectitic intergrowths as well as minor amounts of hydrous minerals.

607 The small-scale heterogeneities of anorthositic and mafic to ultramafic rocks are
608 responsible for contrasts in density of up to 830 kg/m³ and velocities of up to 1.5 km/s on the
609 cm range. Measured densities are generally lower than those extracted from the
610 thermodynamic models and calculated from mineral modes and mineral compositions which
611 is mostly due to retrogression but also a result of small-scale heterogeneities within the drill
612 cores of the sample.

613 Mixing of anorthosites and mafic to ultramafic rocks shows that a mafic portion of ~40-

614 45% is sufficient for the rock to resemble a mafic (garnet) granulite with typical densities of
615 3000-3100 kg/m³, P-wave velocities of ~7 km/s and a Poisson ratio of ~0.27. Similar results
616 can be achieved by a mixture of tonalitic gneiss with ~50-60% mafic rocks. In such a case,
617 the lithological interpretation from seismic data might overestimate the proportion of dense
618 mafic granulite in the lower crust despite the relatively high modal amount of plagioclase (~50-
619 60% volume %). As a result, the transition to eclogite is shifted to much higher pressures than
620 in mafic granulite. Geodynamic models requiring a significant densification due to phase
621 changes might hence require much higher pressures than lithostatic if similar rocks are
622 concerned.

623

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625

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634

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637 **Figure captions:**

638

639 **Figure 1:** a) Geological map of the Sognefjord area (NGU; Bryhni, I. 2003: Berggrunnskart
640 Sogndal 1417-3, M 1:50 000). The anorthosite unit is composed of banded anorthosite with

641 altering anorthositic and mafic layers of varying thickness. The circle indicates the location of
642 the field map (1b). b) Detailed field map of the high-grade anorthositic and mafic rocks in the
643 Sogndal area. Black circles and numbers indicate sample locations used in this study with
644 whole rock compositions listed in Table 1. The anorthosite unit comprises rocks with 60-95%
645 plagioclase. Mafic to ultramafic rocks contain 40-50 volume % clinopyroxene and garnet and
646 minor amounts (~2-5 volume %) of plagioclase and spinel. Garnet rich (~80-90 volume %)
647 domains within the mafic rocks are described as garnetite. Pyroxenite contains > 90 volume
648 % pyroxenes (majorly clinopyroxene). c) Vertical section perpendicular to the field map. d)
649 Photograph of a section of the field map (see box in 1b). e) Scanned sections through a
650 portion of the anorthosite (see box in 1b). Dark areas represent small mafic bands. f) Lens
651 with pyroxene core and garnet rim observed in the profile (see box 1c).

652

653 **Figure 2:** Plots of whole rock wt% of SiO₂ versus MgO, Al₂O₃, CaO, Na₂O, Fe₂O₃ and K₂O,
654 respectively for anorthosite, mafic anorthosite, contact zone, mafic to ultramafic rock,
655 pyroxenite and garnetite (whole rock compositions listed in Table 1). Anorthosite is
656 characterized by high amounts of SiO₂ (>50 wt.%), Al₂O₃ (25-29 wt.%) and Na₂O (~4 wt.%).
657 Mafic anorthosite shows lower amounts of SiO₂ (~45 wt.%) and Al₂O₃ (22-24 wt.%) but higher
658 percentages of Fe₂O₃ (~7 wt.%) and MgO (7-9 wt.%). The mafic to ultramafic compositions
659 yield similarly low amounts of SiO₂ (42-43 wt.%) but significantly higher MgO (~11wt.%) and
660 FeO (6-8 wt.%) compared to anorthosite. The pyroxenite, however, contains a higher
661 percentage of SiO₂ (44-46 wt.%) and CaO (15-18 wt.%) but lower amounts of Al₂O₃ (~12
662 wt.%) than the other mafic rock types.

663

664 **Figure 3:** a) Scanned thin section of anorthosite (sample SH09-30). Aggregates of chlorite
665 and clinozoisite/epidote are interpreted to represent former clinopyroxene. Garnet is partly
666 replaced by chlorite. b) Thin section of mafic anorthosite (sample SH09-35). Garnet is partly
667 replaced by chlorite and symplectitic intergrowths of clinopyroxene, spinel, plagioclase,
668 orthopyroxene and amphibole. c) Thin section scan of a contact zone between anorthosite
669 and mafic rocks (sample SH09-32) with unusually high amounts of spinel. d) Scan of mafic to
670 ultramafic composition (sample SH09-37). e) Thin section of a garnet-rich section within the
671 mafic bands (SH09-01). f) Section of an area dominated by pyroxene (SH09-42). Whole rock
672 compositions are listed in Table 1 and plotted in Fig. 2 while average mineral compositions
673 are shown in Table 2.

674
675 **Figure 4:** a) Symplectitic intergrowths of clinopyroxene, spinel, plagioclase and
676 orthopyroxene replacing garnet in the garnetite (sample SH09-01). b) Symplectites between
677 two garnet crystals in the mafic to ultramafic composition (SH09-37). c) BSE image of the
678 symplectites replacing garnet in the mafic to ultramafic rock. Garnet shows a zonation with
679 higher almandine (Alm_{45}) component at the rim compared to the core ($\text{Alm}_{30-33}\text{Prp}_{43-45}\text{Grs}_{23}$).

680
681 **Figure 5:** Independent set of reactions used for thermobarometry estimates with
682 THERMOCALC for the granulite-facies assemblage ($\text{Grt}+\text{Cpx}+\text{Pl}+\text{Spl}$; see also Table 3)
683 exemplary shown for one P - T point of the mafic to ultramafic rock (sample SH09-37; used
684 mineral data of assemblage [1] in Table 3). The black square represents the resulting P - T
685 estimate with error bars while the red square depicts the P - T estimates obtained from a
686 comparison of measured and calculated mineral compositions with Perple_X (compare to Fig.
687 6). Also shown are the quartz-coesite reaction and the aluminosilicate triple point for reference.

688

689

690 **Figure 6:** a) Pseudosection diagram for the mafic anorthosite (SH09-35). Red lines represent
691 isopleths of pyrope content in garnet within the relevant granulite-facies stability field
692 (Grt+Cpx+Pl+Sp) obtained from Perple_X ($\text{MgO}/[\text{MgO}+\text{FeO}+\text{CaO}]$). Thick red lines indicate
693 values obtained from EMP measurements (Table 2; supplement A) where core measurements
694 are indicated by (c). Green lines represent diopside content in clinopyroxene
695 ($\text{MgO}/[\text{MgO}+\text{FeO}+2\text{Na}_2\text{O}]$). The resulting P - T range for the assemblage is represented by the
696 ellipse. Black squares represent P - T estimates obtained with THERMOCALC (see Table 3
697 and Fig. 5). b) Pseudosection diagram for the mafic to ultramafic rock (SH09-37) c)
698 Pseudosections for the garnetite (SH09-01).

699

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

706

707 **Figure 8:** Phase and density diagrams for anorthositic (a,b) and tonalitic (c,d) mixtures with
708 increasing amount of mafic component. Mixtures with higher mafic proportion yield a
709 significant density increase of up to 200-300 kg/m^3 . White circles represent specified P - T
710 conditions used for the extraction of rock properties (Table 6).

711

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.

736

737 **Table 4:** Calculated and estimated mineral modes, densities and seismic velocities for
738 anorthositic and mafic to ultramafic samples. The relevant P - T conditions used for the
739 calculation are also shown in Fig. 7.

740

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

745

746 **Table 6:** Calculated mineral modes and densities for mixtures of anorthosite/tonalite with
747 increasing proportion of mafic to ultramafic rocks. Some corresponding density diagrams are
748 shown in Fig. 8 including the relevant P - T points.

Table 1.

	SH09-30 Anorthosite	SH09-31 Anorthosite	SH09-36 Anorthosite	SH09-38 Anorthosite	SH09-40 Anorthosite	SH09-43B Anorthosite	SH09-33 Mafic anorthosite	SH09-35 Mafic anorthosite	SH09-43A Mafic anorthosite
SiO ₂	50.99	51.95	50.22	51.57	52.64	51.54	45.64	44.86	44.15
TiO ₂	0.10	0.13	0.12	0.15	0.12	0.08	0.28	0.27	0.24
Al ₂ O ₃	29.10	28.73	25.77	25.67	26.82	27.95	22.06	21.48	24.63
Fe ₂ O ₃	1.17 ^T	1.08 ^T	4.90 ^T	3.05 ^T	2.73 ^T	1.54 ^T	7.58 ^T	7.47 ^T	7.15 ^T
FeO	-	-	-	-	-	-	-	-	-
MgO	0.72	0.82	4.52	2.43	2.32	1.43	8.77	8.69	7.23
MnO	0.01	0.01	0.07	0.04	0.04	0.02	0.09	0.09	0.09
CaO	12.22	12.07	11.46	11.32	11.2	12.43	13.95	13.84	12.68
Na ₂ O	4.09	4.40	3.43	4.00	4.30	3.90	2.03	1.98	2.14
K ₂ O	0.17	0.18	0.12	0.15	0.18	0.12	0.07	0.06	0.36
P ₂ O ₅	0.02	0.02	0.04	0.02	0.02	0.01	0.01	0.01	< 0.01
LOI	1.05	0.45	0.23	0.35	0.45	0.69	0.30	0.23	1.15
Total	99.64	99.84	100.96	98.75	100.82	99.71	100.78	98.98	99.82

	SH09-32 Contact zone	SH09-37 Mafic to ultramafic	SH09-42 Pyroxenite	SH09-44 Pyroxenite	SH09-01 Garnetite
SiO ₂	38.77	42.44	48.58	46.26	43.27
TiO ₂	0.41	0.68	0.59	0.55	0.42
Al ₂ O ₃	26.94	19.09	11.42	12.59	19.12
Fe ₂ O ₃	9.05 ^T	4.71	1.15	9.29 ^T	4.42
FeO	-	7.86	5.09	-	6.00
MgO	8.36	10.97	11.82	12.48	11.21
MnO	0.11	0.19	0.06	0.10	0.14
CaO	13.88	12.98	18.84	15.56	14.37
Na ₂ O	1.52	0.85	1.60	1.41	1.12
K ₂ O	0.03	0.01	0.01	0.11	< 0.01
P ₂ O ₅	< 0.01	< 0.01	< 0.01	0.01	< 0.01
LOI	0.18	-0.84 ^a	-0.11 ^a	1.16	-0.63 ^a
Total	99.25	99.78	99.16	99.96	100.07

^T Total iron

Table 2.

Sample	SH09-30 (anorthosite)							SH09-35 (mafic anorthosite)																	
Mineral	Pl	sd	Grt	Chl	sd	Ep	sd	Pl	sd	Grt	sd	Cpx (I)	sd	Cpx (II)	sd	Opx	sd	Spl	sd	Amp	sd	Chl	sd	Bt	sd
Analyses	3		1	3		6		9		3		3		4		3		6		8		2		2	
SiO ₂	52.27	0.28	39.45	27.14	0.36	38.63	0.44	52.4	1.62	40.34	0.65	47.52	0.6	50.98	0.65	51.98	0.59	0.03	0.01	42.37	1.29	29.58	0.21	36.57	1.53
TiO ₂	0	0	0.19	0.04	0.03	0.18	0.06	0	0	0.05	0.04	1.12	0.26	0.45	0.13	0.03	0.01	0.01	0.01	0.54	0.27	0.38	0.52	0.72	0.02
Al ₂ O ₃	29.87	0.16	22.24	20.72	0.35	26.62	1.11	29.29	1.04	22.69	0.47	9.69	2.79	4.92	0.34	4.29	0.57	62.08	1.05	16.31	1.21	21.74	0.57	18.95	0.57
Cr ₂ O ₃	0	0	0	0.02	0.01	0.01	0.02	0.01	0.01	0	0	0.02	0.02	0	0	0.02	0.01	0.04	0.02	0.01	0.01	0.01	0	0.02	0.02
Fe ₂ O ₃	-	-	-	-	-	8.37 ^a	0.62	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
FeO	0.08	0.06	16.79	17.56	0.37	-	-	0.23	0.17	15.22	3.87	5.35	0.57	5.82	0.17	16.89	0.65	23.19	2.24	8.90	0.58	11.25813	0.50	5.74	0.13
MnO	0.02	0.03	0.44	0.21	0.05	0.04	0.01	0.02	0.02	0.42	0.38	0.1	0.02	0.14	0.03	0.24	0.09	0.13	0.04	0.08	0.02	0.13	0.02	0.02	0
MgO	0	0	9.22	20.24	0.54	0.27	0.43	0	0	13.17	1.6	11.9	0.87	13.97	0.3	25.47	0.38	13.24	1.28	13.70	0.78	21.19	0.85	21.43	0.50
CaO	12.95	0.2	11.46	0.12	0.02	23.95	0.23	12.66	1.22	8.39	1.94	22.66	0.63	22.7	0.27	0.27	0.05	0.10	0.07	12.13	0.08	1.18	0.35	0.38	0.41
Na ₂ O	4.15	0.14	0.02	0.03	0.02	0.02	0.01	4.4	0.67	0.01	0	0.91	0.31	0.67	0.08	0	0	0	0.01	1.91	0.19	0.25	0.27	0.19	0.08
K ₂ O	0.08	0.01	0	0.01	0	0.01	0.01	0.1	0.07	0	0	0.01	0.01	0	0	0.01	0	0	0.01	0.61	0.44	0.37	0.30	8.24	1.94
Total	99.42		99.81	86.36		98.1		99.11		100.29		99.28		99.65		99.2		98.82		96.54		86.08		92.25	
volume %	95		1	3		1		68		2		14		4		1		2		1		8		-	

Sample Mineral Analyses	SH09-32 (contact zone)											SH09-37 (mafic to ultramafic rock)											
	Pl 10	sd	Grt 2	sd	Cpx 6	sd	Spl 7	sd	Amp 10	sd	Chl 1	Pl 6	sd	Grt 13	sd	Cpx 14	sd	Opx 2	sd	Spl 6	sd	Amp 7	sd
SiO ₂	51.33	2.35	39.67	0.33	46.16	1.30	0.02	0.02	40.72	0.92	22.68	48.73	2.26	39.33	0.35	47.18	1.47	52.62	0.51	0.02	0.02	40.77	0.36
TiO ₂	0.01	0.01	0.09	0	1.64	0.33	0.02	0.01	1.48	1.40	0.04	0.01	0.01	0.09	0.04	1.52	0.41	0.08	0.00	0.05	0.02	2.50	0.80
Al ₂ O ₃	30.37	1.58	22.23	0.06	10.82	1.90	62.24	0.66	17.00	1.33	30.24	31.34	1.38	22.02	0.14	8.59	1.63	3.24	0.46	59.94	0.67	15.22	0.99
Cr ₂ O ₃	0.01	0.02	0	0	0.03	0.03	0.03	0.02	0.03	0.03	0	0	0.01	0.03	0.02	0.04	0.03	0.03	0.03	0.21	0.06	0.05	0.02
Fe ₂ O ₃	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
FeO	0.09	0.04	17.54	0.26	5.72	0.85	21.34	1.05	9.03	1.56	11.70	0.27	0.12	20.09	1.84	6.94	0.49	17.11	0.99	26.08	1.32	10.14	0.68
MnO	0.01	0.01	0.82	0.01	0.09	0.04	0.11	0.02	0.10	0.04	0.04	0.01	0.01	0.53	0.11	0.07	0.02	0.12	0.04	0.05	0.02	0.04	0.03
MgO	0.01	0.01	9.16	0.19	11.78	1.06	15.04	0.80	13.18	1.24	20.39	0	0	10.25	0.78	11.84	0.89	26.01	1.08	11.74	0.82	13.12	0.42
CaO	13.69	1.93	10.56	0.11	21.75	1.11	0.02	0.03	11.98	0.31	0.02	15.55	1.67	7.79	1.02	22.31	0.45	0.39	0.08	0.19	0.03	12.08	0.23
Na ₂ O	3.86	1.09	0.04	0	0.97	0.16	0.02	0.01	2.49	0.20	0.03	2.99	0.99	0.02	0.01	0.90	0.09	0.04	0.05	0.01	0.02	2.85	0.08
K ₂ O	0.03	0.02	0	0	0.03	0.06	0	0	0.58	0.39	0.02	0.02	0.02	0.01	0.00	0.01	0.01	0.03	0.04	0.00	0.00	0.10	0.06
Total	99.40		100.11		99.00		98.83		96.58		85.15	98.93	0	100.15		98.71		99.69		98.30		96.88	
volume %	35		10		28		25		2		-	2		48		48		0.3		1		0.7	

Sample Mineral Analyses	SH09-01 (garnetite)														SH09-42 (pyroxenite)									
	Pl	sd	Grt	sd	Cpx (I)	sd	Cpx (II)	sd	Opx	sd	Spl	sd	Amp	sd	Pl	sd	Cpx	sd	Opx	sd	Amp	sd	Chl	sd
	22	sd	25	sd	25	sd	17	sd	18	sd	16	sd	24	sd	6	sd	4	sd	4	sd	4	sd	2	sd
SiO ₂	51.62	2.33	39.95	0.44	47.40	1.16	50.72	0.69	52.37	0.28	0.11	0.25	41.50	0.79	57.92	0.70	50.45	0.41	53.82	0.74	47.56	6.08	29.66	3.36
TiO ₂	0.01	0.01	0.07	0.04	1.01	0.23	0.49	0.12	0.05	0.02	0.03	0.02	1.47	1.07	0.01	0.01	0.64	0.02	0.07	0.01	1.37	1.67	0.00	0.00
Al ₂ O ₃	30.09	1.45	22.26	0.27	9.86	1.73	5.15	0.70	3.55	0.35	60.58	1.81	15.53	0.95	25.74	0.53	5.91	0.77	2.61	1.16	9.43	4.83	19.58	1.56
Cr ₂ O ₃	0.01	0.01	0	0.02	0.04	0.03	0.01	0.02	0.01	0.01	0.18	0.13	0.02	0.02	0.01	0.02	0.31	0.02	0.18	0.03	0.32	0.20	0.01	0.02
Fe ₂ O ₃	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
FeO	0.25	0.13	17.72	2.66	6.28	0.69	6.56	0.59	17.91	0.98	24.59	3.77	9.90	0.61	0.25	0.17	5.35	0.15	15.82	0.72	7.49	1.12	31.97	3.71
MnO	0	0.01	0.44	0.19	0.08	0.03	0.11	0.04	0.17	0.04	0.07	0.03	0.06	0.03	0	0.01	0.05	0.03	0.10	0.01	0.03	0.01	0.03	0.04
MgO	0.01	0.01	10.81	1.15	11.59	0.87	13.41	0.42	24.88	0.58	12.39	2.44	13.25	0.56	0.06	0.15	13.56	0.30	26.58	0.49	16.81	2.62	5.07	1.50
CaO	13.39	1.83	8.61	1.46	22.35	0.39	22.42	0.27	0.39	0.08	0.16	0.16	12.08	0.59	8.48	0.44	22.38	0.28	0.52	0.14	12.04	0.49	1.25	0.02
Na ₂ O	4.06	1.04	0.01	0.01	0.96	0.19	0.83	0.09	0.02	0.02	0.02	0.02	2.68	0.21	6.97	0.26	1.04	0.09	0.01	0.01	1.95	1.06	0.22	0.23
K ₂ O	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0	0.01	0.01	0.00	0.01	0.26	0.17	0.01	0.00	0	0	0.00	0.00	0.05	0.02	0.14	0.17
Total	99.47		99.92		99.58		99.71		99.35		98.14		96.76		99.46		99.69		99.71		97.04		87.94	
volume %	5		82		7		2		1		2		-		3		93		1		1		2	

^a All analyzed FeO recalculated to Fe₂O₃ (wt% FeO*1.111)

Table 3.

Sample	Water-free assemblages	T (°C)	sd	P (GPa)	sd	Independent set of reactions
SH09-35	Grt(22)-Pl(21)-Cpx-I(42)-Spl(32) [1]	970	347	1.3	0.35	$3 \text{ Di} + 3 \text{ CaTs} = \text{Prp} + 2 \text{ Grs}$ $2 \text{ Di} + \text{An} + \text{Spl} = \text{Prp} + \text{Grs}$ $2 \text{ Hed} + \text{An} + \text{Hc} = \text{Grs} + \text{Alm}$ $\text{Grs} + 3 \text{ An} + 3 \text{ Hc} = \text{Alm} + 6 \text{ CaTs}$
	Grt(24)-Pl(36)-Cpx-I(43)-Spl(36) [2]	1065	292	1.08	0.25	$2 \text{ Di} + \text{An} + \text{Spl} = \text{Prp} + \text{Grs}$ $\text{Di} + 2 \text{ An} + 2 \text{ Spl} = \text{Prp} + 3 \text{ CaTs}$ $2 \text{ Hed} + \text{An} + \text{Hc} = \text{Grs} + \text{Alm}$ $\text{Hed} + 2 \text{ An} + 2 \text{ Hc} = \text{Alm} + 3 \text{ CaTs}$
	Grt-Pl-Cpx-Spl [3]	900	100	1	0.15	Prp in Grt and Di in Cpx compositions (see Fig. 9a)
	Grt(24)-Pl(36)-Cpx-I(43)-Spl(36)-Crn [4]	805	112	1.13	0.22	$2 \text{ Di} + \text{An} + \text{Spl} = \text{Prp} + \text{Grs}$ $\text{Prp} + \text{Grs} + 4 \text{ Crn} = 3 \text{ An} + 3 \text{ Spl}$ $\text{Di} + 2 \text{ An} + 2 \text{ Spl} = \text{Prp} + 3 \text{ CaTs}$ $2 \text{ Hed} + \text{An} + \text{Hc} = \text{Grs} + \text{Alm}$ $3 \text{ Hed} + 2 \text{ Crn} = \text{Grs} + \text{Alm}$
	Grt(23)-Pl(21)-Cpx-II(18)-Opx(28)-Spl(30) [5]	904	194	1.1	0.18	$\text{En} + \text{MgTs} = \text{Prp}$ $3 \text{ Di} + 3 \text{ MgTs} = 2 \text{ Prp} + \text{Grs}$ $3 \text{ Di} + 3 \text{ CaTs} = \text{Prp} + 2 \text{ Grs}$ $2 \text{ Di} + \text{An} + \text{Sp} = \text{Prp} + \text{Grs}$ $2 \text{ Hed} + \text{An} + \text{Hc} = \text{Grs} + \text{Alm}$ $\text{Grs} + 3 \text{ An} + 3 \text{ Hed} = \text{Alm} + 6 \text{ CaTs}$ $\text{Grs} + \text{Fs} = 2 \text{ Hed} + \text{CaTs}$
SH09-32	Grt(41)-Pl(34)-Cpx(39)-Spl(29) [1]	890	118	1.17	0.13	$3 \text{ Di} + 3 \text{ CaTs} = \text{Prp} + 2 \text{ Grs}$ $2 \text{ Di} + \text{An} + \text{Spl} = \text{Prp} + \text{Grs}$ $2 \text{ Hed} + \text{An} + \text{Hc} = \text{Grs} + \text{Alm}$ $\text{Grs} + 3 \text{ An} + 3 \text{ Hc} = \text{Alm} + 6 \text{ CaTs}$
	Grt(41)-Pl(34)-Cpx(39)-Spl(29)-Crn [2]	852	68	1.17	0.13	

						$\begin{aligned} \text{Gr} + 2 \text{Crn} &= 3 \text{CaTs} \\ 3 \text{Di} + 3 \text{CaTs} &= \text{Prp} + 2 \text{Grs} \\ 2 \text{Di} + \text{An} + \text{Spl} &= \text{Prp} + \text{Grs} \\ 2 \text{Hed} + \text{An} + \text{Hc} &= \text{Grs} + \text{Alm} \\ 3 \text{Hed} + 2 \text{Crn} &= \text{Grs} + \text{Alm} \end{aligned}$
	Grt ^c (66)-Cpx ^c (67)-Pl(55)-Spl(63) [1]	1041	223	1.18	0.21	$\begin{aligned} 3 \text{Di} + 3 \text{CaTs} &= \text{Prp} + 2 \text{Grs} \\ 2 \text{Di} + \text{An} + \text{Spl} &= \text{Prp} + \text{Grs} \\ 2 \text{Hed} + \text{An} + \text{Hc} &= \text{Grs} + \text{Alm} \\ \text{Grs} + 3 \text{An} + 3 \text{Hc} &= \text{Alm} + 6 \text{CaTs} \end{aligned}$
	Grt ^c (80)-Cpx ^c (81)-Pl(85)-Spl(84) [2]	1066	353	1.25	0.33	$\begin{aligned} 3 \text{Di} + 3 \text{CaTs} &= \text{Prp} + 2 \text{Grs} \\ 2 \text{Di} + \text{An} + \text{Spl} &= \text{Prp} + \text{Grs} \\ 2 \text{Hed} + \text{An} + \text{Hc} &= \text{Grs} + \text{Alm} \\ \text{Grs} + 3 \text{An} + 3 \text{Hc} &= \text{Alm} + 6 \text{CaTs} \end{aligned}$
SH09-37	Grt-Pl-Cpx-Spl [3]	750	100	0.85	0.15	Prp in Grt and Di in Cpx compositions (see Fig. 9b)
	Grt ^r (87)-Cpx ^r (92)-Opx(91)-Pl(95)-Spl(84) [4]	947	223	0.83	0.23	$\begin{aligned} \text{En} + \text{MgTs} &= \text{Prp} \\ 3 \text{Di} + 3 \text{MgTs} &= 2 \text{Prp} + \text{Grs} \\ 2 \text{Di} + \text{An} + \text{Spl} &= \text{Prp} + \text{Grs} \\ \text{Grs} + 3 \text{An} + 3 \text{Spl} &= \text{Prp} + 6 \text{CaTs} \\ \text{Grs} + 3 \text{Fs} &= \text{Alm} + 3 \text{Hed} \\ 2 \text{Hed} + \text{An} + \text{Hc} &= \text{Grs} + \text{Alm} \\ \text{Gr} + 3 \text{An} + 3 \text{Hc} &= \text{Alm} + 6 \text{CaTs} \end{aligned}$
	Grt ^c (57)-Cpx-I(56)-Pl(58)-Spl(51) [1]	1181	351	1.49	0.34	$\begin{aligned} 3 \text{Di} + 3 \text{CaTs} &= \text{Prp} + 2 \text{Grs} \\ 2 \text{Di} + \text{An} + \text{Spl} &= \text{Prp} + \text{Grs} \\ 2 \text{Hed} + \text{An} + \text{Hc} &= \text{Grs} + \text{Alm} \\ \text{Grs} + 3 \text{An} + 3 \text{Hc} &= \text{Alm} + 6 \text{CaTs} \end{aligned}$
SH09-01	Grt ^c (65)-Cpx-I(67)-Pl(69)-Spl(75) [2]	955	159	1.22	0.17	$\begin{aligned} 3 \text{Di} + 3 \text{CaTs} &= \text{Prp} + 2 \text{Grs} \\ 2 \text{Di} + \text{An} + \text{Spl} &= \text{Prp} + \text{Grs} \\ 3 \text{Hed} + 3 \text{CaTs} &= 2 \text{Grs} + \text{Alm} \\ 2 \text{Hed} + \text{An} + \text{Hc} &= \text{Grs} + \text{Alm} \end{aligned}$

Grt ^c (103)-Cpx-I(101)-Pl(71)-Spl(105) [3]	948	127	1.19	0.13	$3 \text{ Di} + 3 \text{ CaTs} = \text{Prp} + 2 \text{ Grs}$ $2 \text{ Di} + \text{An} + \text{Spl} = \text{Prp} + \text{Grs}$ $2 \text{ Hed} + \text{An} + \text{Hc} = \text{Grs} + \text{Alm}$ $\text{Grs} + 3 \text{ An} + 3 \text{ Hc} = \text{Alm} + 6 \text{ CaTs}$
Grt-Pl-Cpx-Spl [4]	750	100	0.85	0.15	Prp in Grt and Di in Cpx compositions (see Fig. 9c)
Grt(81)-Cpx-II(79)-Opx(82)-Pl(70)-Spl(75) [5]	895	119	0.85	0.13	$2 \text{ Di} + \text{An} + \text{Spl} = \text{Prp} + \text{Grs}$ $\text{Grs} + 3 \text{ An} + 3 \text{ Spl} = \text{Prp} + 6 \text{ CaTs}$ $6 \text{ En} + 3 \text{ An} + 3 \text{ Spl} = 5 \text{ Prp} + \text{Grs}$ $\text{Prp} + 3 \text{ An} + 3 \text{ Spl} = \text{Grs} + 6 \text{ MgTs}$ $2 \text{ Hed} + \text{An} + \text{Hc} = \text{Grs} + \text{Alm}$ $\text{Grs} + 3 \text{ An} + 3 \text{ Hc} = \text{Alm} + 6 \text{ CaTs}$ $6 \text{ Fs} + 3 \text{ An} + 3 \text{ Hc} = \text{Grs} + 5 \text{ Alm}$

^c analyses of mineral cores

^r analyses of rims

Table 4.

Sample	SH09-30 (anorthosite)			SH09-35 (mafic anorthosite)			SH09-32 (contact zone)		
	Rock ^b	950 °C; 1.0 GPa (a)	660 °C; 0.85 GPa (b)	Rock ^b	950 °C; 1.0 GPa (a)	660 °C; 0.85 GPa (b)	Rock ^b	950 °C; 1.0 GPa (a)	660 °C; 0.85 GPa (b)
Grt (vol%)	1	2	3	16 ^e	22.5	24	11	14	17
Cpx (vol%)	4 ^c	3	3.5	14	45	42	29	46	47
Pl (vol%)	95	94 ^d	93 ^d	68	28.5	29	35	25	21
Spl (vol%)	-	<1	acc	2	4	4	25	14	13
Crn (vol%)	acc	-	acc	acc	-	acc	acc	-	2
ρ (kg/m ³)	2700	2740 ± 30	2760 ± 30	2870	3250 ± 30	3270 ± 30	3210	3302 ± 30	3365 ± 30
ρ (kg/m ³); H ^a	-	2740 ± 30	2760 ± 30	-	3220 ± 30	3250 ± 30	-	3260 ± 30	3320 ± 30
v_P (km/s)	-	6.7 ± 0.2	6.8 ± 0.2	-	7.5 ± 0.2	7.6 ± 0.2	-	7.6 ± 0.2	7.8 ± 0.2
v_P (km/s); H ^a	-	6.6 ± 0.2	6.6 ± 0.2	-	7.4 ± 0.2	7.5 ± 0.2	-	7.5 ± 0.2	7.7 ± 0.2

Sample	SH09-37 (mafic to ultramafic)			SH09-01 (garnetite)			SH09-42 (pyroxenite)		
	Rock ^b	950 °C; 1.0 GPa (a)	660 °C; 0.85 GPa (b)	Rock ^b	950 °C; 1.0 GPa (a)	660 °C; 0.85 GPa (b)	Rock ^b	950 °C; 1.0 GPa (a)	660 °C; 0.85 GPa (b)
Grt (vol%)	48	44	47	84 ^e	34	34.5	-	1	13
Cpx (vol%)	48	52	47	9	60	55.5	93	81	85
Pl (vol%)	2	1	2	5	2	5	3	7	-
Spl (vol%)	1	3	3	2	4	5	-	-	-
Crn (vol%)	-	-	-	-	-	-	-	-	-
ρ (kg/m ³)	-	3540 ± 35	3570 ± 40	3480	3480 ± 35	3490 ± 35	3150	3270 ± 30	3380 ± 30
ρ (kg/m ³); H ^a	-	3510 ± 35	3550 ± 40	-	3450 ± 35	3470 ± 35	-	3230 ± 30	3360 ± 30
v_P (km/s)	-	8.0 ± 0.2	8.2 ± 0.3	-	8.0 ± 0.2	8.1 ± 0.2	-	7.6 ± 0.2	7.9 ± 0.2
v_P (km/s); H ^a	-	7.9 ± 0.2	8.1 ± 0.2	-	7.8 ± 0.2	8.0 ± 0.2	-	7.4 ± 0.2	7.7 ± 0.2

^a Densities and velocities obtained by the spreadsheet from Hacker and Abers (2004)

^b Mineral estimates from 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.

	Tonalitic gneiss	An40	An45	An50	TG50	TG55	TG60	Granulite
SiO ₂	60.7	47.6	47.1	46.7	51.6	50.7	49.7	47.9
TiO ₂	0.9	0.3	0.4	0.4	0.8	0.8	0.8	1.0
Al ₂ O ₃	16.9	25.1	24.6	24.1	18.0	18.1	18.2	14.3
FeO [†]	5.6	5.5	6.1	6.6	8.9	9.2	9.5	12.2 ^a
MgO	2.2	4.8	5.3	5.9	6.6	7.0	7.5	6.6
MnO	0.1	0.1	0.1	0.1	0.2	0.2	0.2	0.3
CaO	4.5	12.5	12.6	12.6	8.7	9.2	9.6	11.2
Na ₂ O	3.9	2.8	2.6	2.5	2.4	2.2	2.1	2.9
K ₂ O	3.1	0.1	0.1	0.1	1.6	1.4	1.2	0.9

[†] Total iron calculated as Fe₂O₃* 0.899 + FeO

^a Measured as Fe₂O₃

Table 6.

Anorthosite with	40% mafic (An40)			45% mafic (An45)			50% mafic (An50)		
	950 °C; 1.0 GPa (a)	660 °C; 0.85 GPa (b)	650 °C; 1.5 GPa (c)	950 °C; 1.0 GPa (a)	660 °C; 0.85 GPa (b)	650 °C; 1.5 GPa (c)	950 °C; 1.0 GPa (a)	660 °C; 0.85 GPa (b)	650 °C; 1.5 GPa (c)
Grt (vol.%)	17	19	13	19	22	16	21	24	19
Cpx (vol.%)	19	19	63	21	21	62	24	24	60
Pl (vol.%)	62	60	-	58	55	-	53	49	-
Spl (vol.%)	1	-	-	1	-	-	2	-	-
Ky (vol.%)	-	-	18	-	-	18	-	-	17
ρ_1 (kg/m ³)	3010 ± 30	3060 ± 30	3440 ± 30	3050 ± 30	3100 ± 30	3460 ± 35	3090 ± 30	3140 ± 30	3470 ± 35
ρ_2 (kg/m ³)	3060 ± 30	3080 ± 30	3250 ± 30	3100 ± 30	3125 ± 30	3290 ± 30	3140 ± 30	3165 ± 30	3320 ± 30

Tonalitic gneiss with	50% mafic (TG50)			55% mafic (TG55)			60% mafic (TG60)		
	950 °C; 1.0 GPa (a)	660 °C; 0.85 GPa (b)	650 °C; 1.5 GPa (c)	950 °C; 1.0 GPa (a)	660 °C; 0.85 GPa (b)	650 °C; 1.5 GPa (c)	950 °C; 1.0 GPa (a)	660 °C; 0.85 GPa (b)	650 °C; 1.5 GPa (c)
Grt (vol%)	5	21	37	6	23	38	6	25	40
Cpx (vol%)	13	17	45	14	18	45	15	20	45
Fsp (vol%)	61	52	16	59	49	14	57	47	12
Opx (vol%)	20	8	-	20	8	-	21	8	-
Ky vol%	-	-	2	-	-	3	-	-	3
ρ_1 (kg/m ³)	2990 ± 30	3105 ± 30	3430 ± 30	3010 ± 30	3135 ± 30	3450 ± 30	3030 ± 30	3160 ± 30	3470 ± 30
ρ_2 (kg/m ³)	3170 ± 30	3200 ± 30	3340 ± 30	3200 ± 30	3240 ± 30	3370 ± 30	3240 ± 30	3270 ± 30	3395 ± 30

a **Fig. 2** **b**

Fig. 3

Fig. 4

Fig. 7

SH09-37

Fig. 8

Fig. 10 Anorthositic:

Mixtures with anorthosite:

Mixtures with tonalite:

Fig. 11

Fig. 12