

1 **Heavy minerals and garnet geochemistry of stream sediments and bedrocks from the**
2 **Almklovdalen area, Western Gneiss Region, SW Norway: implications for provenance analysis**

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23 **Abstract**

24 Detrital heavy minerals commonly document the geological setting in the source area, hence they are
25 widely used in sedimentary provenance analysis. In heavy mineral studies most commonly the 63–125
26 and 63–250 µm grain-size fractions are used. Heavy mineral data and garnet geochemistry of stream
27 sediments and bedrocks from the catchment area draining the Almklovdalen peridotite massif in SW
28 Norway reveal that a wider grain-size spectrum needs to be considered to avoid misleading
29 interpretations. The Almklovdalen peridotite massif consists mainly of dunite and harzburgite, as
30 testified by the heavy mineral suite. At the outlet of the main river the heavy mineral spectrum is very
31 monotonous due to dilution by strong influx of olivine. Heavy minerals like apatite and epidote
32 characterising the host gneisses have almost disappeared. MgO-rich almandine garnets are more
33 frequent in the coarser grain-size fractions, whereas MnO-rich almandine garnets are more frequent in
34 the finer grain-size fractions. Garnets with pyrope content exceeding 50 % are only found in the
35 500–1000 µm grain-size fraction. Therefore, the sample location and the selected grain-size fraction

36 are of paramount importance when dealing with heavy minerals and mineral geochemical data,
37 otherwise, provenance sensitive information may be missed.

38

39 **1. Introduction**

40 Heavy minerals are used to characterise, discriminate, and identify source areas (Mange and Wright,
41 2007; von Eynatten and Dunkl, 2012, and references therein). Commonly the heavy mineral
42 composition of a sediment reflects the mineralogy of the rocks exposed in the source area, however,
43 the heavy mineral assemblage may be affected by several processes during the sedimentary cycle that
44 modify occurrence and proportions of heavy mineral species (e.g., Morton and Hallsworth, 1999). For
45 instance, the stability of heavy minerals strongly depends on the specific environment and climatic
46 conditions (Pettijohn, 1941; Morton and Hallsworth, 1999; Velbel, 2007; Andò et al., 2012; Morton,
47 2012; Garzanti et al., 2013). Therefore, differences in heavy mineral ratios are not necessarily the
48 effect of different sources, but can reflect modifications which occur during the sedimentary cycle
49 (e.g., Morton and Hallsworth, 1999). When studying stream sediments it must be noted that minerals
50 can be segregated and sorted according to their grain size, density, and shape (Morton and Hallsworth,
51 1999; Garzanti et al., 2008, 2009; Resentini et al., 2013). This can happen either between different
52 heavy mineral species, but also within a single mineral group. For example, almandine-rich garnets are
53 commonly concentrated in the finer fractions in contrast to less dense garnet species (Schuiling et al.,
54 1985; Andò, 2007; Garzanti et al., 2008).

55 The heavy mineral garnet has important provenance applications, because it exists in a wide range of
56 rocks and its chemical composition depends on the composition of the source rock and on pressure and
57 temperature conditions during garnet formation (e.g., Wright, 1938; Morton, 1985; Deer et al., 1992;
58 Andò et al., 2013; Krippner et al., 2014). Although garnets from various garnet-bearing rocks often
59 show much overlap in major element geochemistry, mantle-derived garnets, for instance, can be very
60 well separated from crustal-derived garnets (e.g., Krippner et al., 2014). Therefore, garnet is used as an
61 indicator mineral in exploration, for instance, for diamond (Nowicki et al., 2003) because mantle-
62 derived garnets can be related to diamond-bearing intrusives (Grütter et al., 2004). In contrast to the
63 typical mantle-derived minerals olivine and pyroxene, garnet is more stable during alteration and

64 dispersion at the Earth's surface (e.g., Pettijohn, 1941; Velbel, 1984, 1999; Gurney, 1984; Grütter et
65 al., 2004; Morton, 2012).

66 We collected stream sediments and adjacent source rocks in the high-grade metamorphic
67 Almklovdalen area in SW Norway to study to what extent the heavy mineral suites and garnet
68 composition of the stream sediments reflect the mineralogy of the source rocks. We analysed different
69 grain-size fractions in order to test for grain-size dependency of heavy mineral assemblages and garnet
70 geochemistry which may lead to ambiguous or even wrong interpretations when a single and narrow
71 grain-size spectrum is considered only (Garzanti et al., 2009).

72

73 **2. Geological Setting**

74 The Almklovdalen area is located in the Western Gneiss Region (WGR) in SW Norway (Fig. 1). The
75 WGR comprises Precambrian basement and allochthonous cover units, metamorphosed and deformed
76 during the Caledonian orogeny induced through collisions between Baltica, Laurentia and Avalonia
77 under closure of Iapetus Ocean (Roberts and Gee, 1985; Cuthbert et al., 2000; Krabbendam et al.,
78 2000; Beyer et al., 2012). The gneisses and augen orthogneisses of the WGR are mainly of
79 granodioritic to granitic composition and are considered to represent Baltica basement (Tucker et al.,
80 1990). The gneisses are predominantly of amphibolite-facies metamorphic grade (Bryhni and
81 Andréasson, 1985; Krabbendam and Wain, 1997; Krabbendam et al., 2000), but in some parts
82 granulite-facies assemblages occur (Griffin et al., 1985; Krabbendam et al., 2000). Most of the
83 orthogneisses of the WGR are suggested to be generated during Gothian (1.7–1.5 Ga) and
84 Sveconorwegian (1.2–1.9 Ga) events (Beyer et al., 2012). During the Scandian phase (435–390 Ma)
85 of the Caledonian orogeny, rocks of the WGR suffered high-pressure (HP) to ultrahigh-pressure
86 (UHP) conditions (Griffin and Brueckner, 1980; Gebauer et al., 1985; Griffin et al., 1985; Mørk and
87 Mearns, 1986; Andersen et al., 1991; Krabbendam et al., 2000). The metamorphic grade increases
88 from SE to NW (Krogh, 1977; Cuthbert et al., 2000; Root et al., 2005; Beyer et al., 2012). The
89 temperature gradient increases from ~550 °C in the SE to >800 °C in the NW (Krogh, 1977; Griffin et
90 al., 1985; Carswell and Cuthbert, 2003). Mafic and ultramafic lenses which are enclosed within the
91 surrounding gneisses range in size from centimetres to hundreds of metres (Carswell and Cuthbert,

92 2003). A few ultramafic bodies reach sizes up to several kilometres. There are hundreds of mantle-
93 derived ultramafic bodies (Beyer et al., 2012), with some of them containing garnet-bearing
94 assemblages (Eskola, 1921; Medaris, 1984; Carswell, 1986; Medaris and Carswell, 1990; Brueckner et
95 al., 2010; Beyer et al., 2012). The ultramafic rocks show Archaean Re–Os ages which predate the
96 formation of the Proterozoic upper crusts in the region (Brueckner et al., 2002; Beyer et al., 2004).
97 One of the largest ultramafic body, with a size of $4.0 \times 3.3 \text{ km}^2$, is the Almklovdalen peridotite
98 (Medaris and Brueckner, 2003), located in the HP–UHP transition zone (Fig. 1). This ultramafic body
99 is a bowl-shaped sheet around a central gneiss area, composed of orthogneisses and paragneisses and
100 indicates three main stages of deformation. A first deformation is seen in the Proterozoic folds in the
101 garnet peridotite, a second deformation is highlighted by the Caledonian isoclinal folds in the chlorite
102 peridotite in association with recrystallisation of garnet peridotite to chlorite peridotite, and a third
103 deformation – also Caledonian in age – led to foliations and lineations in chlorite peridotite (Medaris
104 and Brueckner, 2003). The main rock types in the Almklovdalen body are anhydrous dunite and
105 harzburgite (Osland, 1997; Medaris and Brueckner, 2003; Beyer et al., 2006, 2012). They contain
106 garnet peridotite and garnet pyroxenite lenses (Medaris, 1984; Osland, 1977; Beyer et al., 2012),
107 which occur in <40 cm thick bands within the dunites and harzburgites (Cordellier et al., 1981).
108 Eclogites are also present, but only in minor amounts (Griffin and Qvale, 1985; Beyer et al., 2012).
109 The Almklovdalen body is surrounded by orthogneisses, paragneisses, anorthosites, and eclogites
110 (Beyer et al., 2012).

111

112 **3. Samples and methods**

113 *3.1 Samples*

114 Four sand samples were collected from streams draining the Almklovdalen peridotite body (Fig. 2).
115 Loose bedload sand was sampled to cover a wide grain-size range. Stream sample AK-N20 was
116 collected ~1 km SE of Helgehornet, followed by sample AK-N19-4 further downstream (Fig. 2).
117 Stream sample AK-N19-3 was taken from the middle part of the Gusdalselva river and stream sample
118 AK-N19-1 comes from the mouth of the Gusdalselva river entering lake Gusdalsvatnet. In addition,
119 bedrock samples were collected since they represent the source for the detrital material. Bedrock

120 samples were taken in outcrop. One bedrock sample, a large pebble, was taken directly from the river
121 bed. Sample AK-N21 is an eclogite exposed south of Helgehornet and AK-EA is an eclogite pebble
122 collected at the mouth of the GUSDALSELVA river. Sample AK-N24 is a garnet peridotite and AK-N25 is
123 a gneiss exposed SE of Helgehornet. The geographic coordinates of all samples are given in Table 1.

124

125 *3.2 Methods*

126 Stream sediments were wet-sieved using a mechanical shaker to separate the different grain-size
127 fractions (63–125 μm , 125–250 μm , 250–500 μm , 500–1000 μm). We used ~ 350 grams per sample.
128 After drying, the samples were treated with acetic acid to remove the carbonate component if present.
129 The heavy mineral fractions were separated using Sodium polytungstate (SPT) with a density of 2.85
130 g/mL.

131 The heavy mineral residues were mounted on microscope slides (Mange and Maurer, 1992) using
132 Meltmount™ with refraction of 1.66 and identified under the polarising microscope. The relative
133 abundances of the heavy minerals were determined by grain counting. For that, the microscope slide
134 was moved along linear traverses and all grains between two parallel lines were counted (i.e. ribbon
135 counting; Mange and Maurer, 1992). Two-hundred and fifty translucent minerals were counted for
136 each slide. All data are given as supplementary data, i.e. data of all heavy minerals including opaque
137 minerals, micas and unknown (Supplementary Table S1), only groups of transparent minerals (see
138 Supplementary Table S2), and heavy minerals from all grain-size fractions (Supplementary Table S3).
139 Unknown mineral are probably originated by alteration of various minerals and are aggregates with no
140 well defined mineralogical composition. They are also termed as alterites (van Andel, 1950).

141 For garnet geochemical analysis, garnet selection from the stream sediments was achieved by
142 handpicking under a binocular microscope. We randomly selected garnet grains of all sizes and
143 morphological types and placed them in synthetic mounts using a bonding epoxy composed of a
144 mixture of Araldite® resin and hardener in the ratio of 5:1. Grains of different morphological types
145 were selected to consider a wide range of altered and non-altered garnets in the source area. Also
146 garnets of different colours in approximately similar amounts were selected to minimize the bias of
147 picking only one type of garnet.

148 From the bedrock samples polished thin sections were prepared. Garnet from bedrock and stream
149 samples were analysed with a JEOL JXA 8900 RL electron microprobe (EMP) equipped with five
150 wavelength dispersive spectrometers at the University of Göttingen (Department of Geochemistry,
151 Geoscience Center). Before analysis, all samples were coated with carbon to ensure conductivity.
152 Conditions included an accelerating voltage of 15 kV and a beam current of 20 nA. The counting
153 times were 15 seconds for Si, Mg, Ca, Fe, and Al, and 30 seconds for Ti, Cr, and Mn (Table 2). Matrix
154 correction was performed using ZAF corrections. We preferentially analysed garnet rims and cores.
155 From the thin sections 20 garnets from each sample were analysed and 50 garnets of each sediment
156 sample were measured. Additionally, minerals in thin sections were determined under the polarising
157 microscope. The relative abundances of the main light and heavy minerals from the bedrocks can be
158 taken from Table 3. The full database including lithology and metamorphic grade are included in
159 Supplementary Table S4.

160 From all stream sediment samples the 63–125 μm fractions were point counted and analysed by EMP.
161 For the study of grain-size dependency sample AK-N19-3 was chosen. From this sample the 125–250
162 μm and 250–500 μm grain-size fractions were point counted, too, and garnet grains from the 125–250
163 μm , 250–500 μm and 500–1000 μm size fractions were analysed by EMP.

164 We measured the long and short axes of ~ 100 garnet grains from the 63–125 μm grain-size fraction of
165 each river sample, to study if the garnets are mainly separated by their short axis or mainly by their
166 long axis and if there is an appreciable input of garnets with smaller or coarser grain size, which are
167 not belonging to the specific grain size fraction. Long and short axes of garnet from bedrocks (AK-
168 N21, AK-EA, AK-N24) were also measured for comparison (Supplementary Table S5).

169 The geochemical data are presented in ternary diagrams following Mange and Morton (2007) and in
170 biplots produced using CoDaPack, an open source software for compositional data analysis (Thiό-
171 Henestrosa and Martín-Fernández, 2005; Comas-Cufí and Thiό-Henestrosa, 2011). The biplots are
172 based on principal component analysis (PCA) using centred log-ratio transformation for six major
173 element oxides, following the methodology proposed by Aitchison (1986). The biplots serve as a
174 valuable tool for estimating the potential for discrimination of a multivariate data set and its subsets
175 (e.g., von Eynatten et al., 2003). In biplots, multivariate observations are illustrated as points and

176 variables as lines. The length of the line corresponds to the variability of the respective element.
177 Length and position of the line reflects its relative influence on the respective principal component.

178

179 **4. Results**

180 *4.1. Heavy mineral analysis of all stream samples (63–125 μm)*

181 In all stream samples, olivine is the dominant heavy mineral representing between 39 and 90 % of the
182 heavy mineral suite (Fig. 3a). Pyroxene comprises between 8 and 11 % (mainly diopsitic to augitic
183 clinopyroxenes). Garnet, green calcic amphibole, epidote-group minerals (epidote, zoisite), and apatite
184 occur in different percentages. The amphiboles are dominantly blue-green with colour changing from
185 blue-green to green. Ultrastable minerals (zircon, tourmaline, and rutile) are not present or occur only
186 in traces. Other amphiboles (mainly actinolites and tremolites) and titanites occur occasionally as
187 single grains in individual samples and are grouped as ‘others’. Opaque minerals and micas are not
188 considered in the diagram to emphasize the relative concentrations of the transparent heavy minerals.

189

190 *4.2. Heavy mineral analysis of sample AK-N19-3 (63–125 μm , 125–250 μm , 250–500 μm)*

191 Olivine is the dominant heavy mineral with similar amount (59–63 %) in all grain-size fractions (Fig.
192 3b). Garnet content is higher in the coarser grain-size fractions, i.e. 28 % in the 125–250 μm fraction
193 and to 24 % in the 250–500 μm fraction compared to 9 % in the 63–125 μm fraction. In contrast,
194 pyroxene and green calcic amphibole decrease in the coarser fractions. The content of epidote-group
195 minerals remains fairly constant, whereas apatite (5 %) is only found in the 63–125 μm grain-size
196 fraction. Ultrastable minerals are almost absent in the 250–500 μm fraction. The 500–1000 μm grain-
197 size fraction is not considered here because of the very high proportion of micas and opaque minerals,
198 which amounts to almost 90 % of the entire heavy mineral spectrum.

199

200 *4.3. Grain-sizes of garnets*

201 The shortest axes of garnets from the 63–125 μm sieve fractions of the stream sediments range from
202 >40–180 μm and the longest axes range from >80–300 μm (Fig. 4; Supplementary Table S5). The

203 shortest axes of garnets from the eclogite bedrocks are between <100–700 μm and the longest axes
204 between 100 μm to >1000 μm (Fig. 4; Supplementary Table 5).

205 The garnets of sample AK-N24 (garnet peridotite) are very coarse and go beyond the camera's field of
206 view of the microscope. The shortest axis of the smallest grain found in this sample is 1000 μm and
207 the longest axis of the largest grain is about 1.4 cm (measured with a ruler).

208

209 *4.4. Geochemistry of garnets*

210 Biplots with the major element oxides SiO_2 , Al_2O_3 , CaO , MgO , FeO , and MnO as variables are used to
211 differentiate garnets from the different bedrock types and from the sediments (Fig. 5). These diagrams
212 are optimal in the sense that most of the total variability is illustrated in two dimensions (i.e. 89–99%
213 in this case). SiO_2 and Al_2O_3 show the lowest variation in all of the three biplots (Figs. 5a, 5b, 5c). The
214 small distance between SiO_2 and Al_2O_3 indicates relatively constant $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratios typical for all
215 garnet varieties with aluminum in the crystallographic Y-site (i.e. endmembers almandine, spessartine,
216 pyrope, grossular).

217 The major element data of the garnets from the bedrocks show highest variability for MnO and MgO
218 (spread in opposite directions and strongly controlling PC1) and a moderate variability for CaO and
219 FeO along with fairly constant CaO/FeO (Fig. 5a). The three bedrock samples are clearly separated
220 from each other in the biplot, with garnets from the peridotite being distinct due to relative high MgO
221 content. Garnets of the eclogites accumulate on the left side of the biplot indicating relatively lower
222 MgO content, with garnets of sample AK-N21 appearing to have slightly higher relative MnO content
223 (i.e. higher MnO/CaO ratios), when compared to eclogite sample AK-EA (Fig. 5a).

224 The detrital garnets show a comparably high variability for MnO , MgO and CaO (spread in different
225 directions) and show much overlap between both different samples and different grain-size fractions
226 (Fig. 5b, 5c). The CaO/FeO value is no longer constant, particularly for the coarser grain-size fraction
227 (Fig. 5c). CaO and MnO have the highest impact on PC2, which in turn has much higher impact on the
228 total variability of the detrital garnets (22 and 26%) compared to the bedrocks (7%; Fig. 5). This
229 implies higher relevance of CaO/MnO ratios for the variability of the detrital garnets. The detrital
230 garnets of the different grain-size fractions of sample AK-N19-3 show only little contrast. However,

231 few garnets of the coarser grain-size fractions (250–1000 μm) show overlap with the garnet peridotite
232 (Fig. 5c).

233 Garnet composition from stream sediments and bedrocks is illustrated in the classical ternary
234 classification diagram using almandine + spessartine, pyrope, and grossular as poles, and the
235 discrimination fields A, B, Bi, Ci, Cii, and D (Mange and Morton, 2007; Fig. 6). This diagram has
236 widely been applied in a number of garnet provenance studies (e.g., Whitham et al., 2004; Morton et
237 al., 2005; Mange and Morton, 2007; Meinhold et al., 2010; Morton et al., 2011; Krippner et al., 2015).
238 All of the garnets of samples AK-EA and almost all of the garnets of sample AK-N21 plot in field Ci,
239 the field for garnets derived from high-grade mafic rocks, such as eclogites. Garnets of sample AK-
240 N24 plot in field Cii, the field for garnets derived from ultramafic rocks (Fig. 6a).

241 For comparison, all garnets from the stream sediments (63–125 μm grain-size fraction) and from the
242 bedrocks are plotted together in one ternary diagram (Fig. 6b). The garnets from the stream sediments
243 overlap fields B, Ci, A, and Cii, with almost 90 % of garnets plotting in field Ci. They show a high
244 degree of overlap with garnets measured in the eclogites. However, many detrital garnets are not
245 comparable to the garnets derived from the eclogites as they show a wider distribution and are
246 possibly derived from other source rocks than the measured eclogites. Interestingly, no detrital garnets
247 are comparable with those garnets measured in the garnet peridotite (Fig. 6b).

248 The garnets from the different grain-size fractions from sample AK-N19-3 overlap fields B, Bi, A, Ci,
249 Cii, with almost 90 % plotting in field Ci. Many garnets show a high degree of overlap with garnets
250 measured in the eclogites, but many garnets tend to higher MgO composition and also towards higher
251 CaO or FeO+MnO composition. Two single garnets of the 500–1000 μm grain-size fractions show
252 overlap with the garnets measured in the peridotite (Fig. 6c). Overall, the detrital garnets show a
253 distinct higher variability than the garnets from the bedrocks, hence point to even other source rocks
254 than the measured ones.

255
256 With increasing grain size the MgO/MnO value increases on average (Fig. 7). About 6% of the garnets
257 of the coarser grain-size fractions (250–1000 μm) lie above line 2 comparing to 1% of the finer grain-
258 size fraction (125–250 μm). None of the garnets of the 63–125 μm grain-size fraction lie above line 2.

259 In contrast, 18% of the garnets of the finer grain-size fractions (63–250 μm) and only 4% of the
260 coarser grain-size fractions (250–1000 μm) lie below line 1. This is because the content of MnO is
261 overall higher in the finer grain-size fractions (63–250 μm) than in the coarser grain-size fractions. In
262 contrast, the MgO content of the coarser grain-size fractions (250–1000 μm) is higher on average
263 compared to the finer grain-size fractions (Fig. 7). MnO contents exceeding 2 wt.% are only evident in
264 the garnets of the finer grain-size fractions (63–250 μm) (Supplementary Table S4).

265

266 5. Discussion

267 Heavy mineral analysis of stream samples from the Almklovdalen area in SW Norway reveal an
268 increase of olivine from upstream (AK-N19-4) to downstream (AK-N19-1) from ~40 to 90 % (Fig. 3a)
269 and a decrease of green calcic amphibole from ~13 to ~1 %. The content of ultrastable minerals
270 decreases, too. The content of garnet slightly increases in the middle part of the river and decreases at
271 the downstream end of the GUSDALSSELVA river where only 1 % garnet is found (sample AK-N19-1).
272 Apatite grains decrease strongly from upstream to downstream. In sample AK-N19-4, approximately
273 11 % of the heavy mineral assemblage consists of apatite minerals, in sample AK-N19-1, at the
274 downstream end of the GUSDALSSELVA river, apatite grains were not found. Epidote-group minerals
275 commonly occur in the upper section of the river (~10 %) but at the downstream end of the
276 GUSDALSSELVA river they are strongly depleted (~1 %). The pyroxene content is nearly constant in all
277 samples (around 10%). Most of the pyroxenes are clinopyroxenes, probably of diopsitic to augitic
278 composition and likely derived from the lherzolites (Beyer et al., 2006). However, there are also some
279 orthopyroxenes identified which are probably of enstatitic composition.

280 Chlorite peridotite (dunite, harzburgite) covers a large area in the downstream part of the GUSDALSSELVA
281 river, including a huge quarry (Fig. 2), thus explaining the downstream increase in olivine
282 concentration. In the upper part of the river, garnet peridotites, eclogites and gneisses are more
283 frequent than in the downstream part of the river. Therefore, in samples AK-N19-3, AK-N19-4, and
284 AK-N20 the influx from those rocks is higher than in sample AK-N19-1. Further downstream most of
285 the heavy minerals are diluted by the high input of olivine. Apatite is found in the gneisses (Table 3).
286 There is a major input of apatite recorded in sample AK-N19-4, which is not surprising since this part

287 of the river is draining the gneisses upstream from the sample location. The strong decrease of apatite
288 minerals is likely due to the dilution process described before but may also result from partial
289 dissolution, because apatite becomes unstable under acidic conditions (Morton, 2012). The vegetation
290 of the study area and its surroundings consist mainly of coniferous forest and indicates a rather acidic
291 environment in which apatite is prone to dilution.

292 The main lithologies in the study area are dunite and harzburgite with minor garnet peridotite,
293 eclogite, and gneiss. The major garnet-bearing rocks are garnet peridotites and eclogites. In the
294 sampled gneiss no garnet grains were found, but garnet-bearing gneisses are known from the
295 surrounding area; hence, they also provide a source for the detrital garnets in the stream sediments.
296 The detrital garnets of the 63–125 μm grain-size fraction show a high degree of overlap with the
297 garnets analysed in the eclogites (Fig. 6b). However, many of the detrital garnets cannot be directly
298 linked to the studied eclogites. A possible explanation is that the several eclogite bodies exposed in the
299 area cover a range of different garnet compositions, which is also evident from the two measured
300 eclogites as seen in the biplots, because they can be clearly separated from each other (Fig. 5a). This
301 separation is only possible when FeO and MnO, respectively almandine and spessartine, are not
302 combined. That is why a discrimination of both eclogites cannot be determined in the ternary
303 diagrams. Garnets with a high MnO and a low MgO content can be probably linked to the garnet-
304 bearing gneisses, which suffered lower grade amphibolite-facies metamorphism (Bryhni and
305 Andréasson, 1985; Krabbendam and Wain, 1997; Krabbendam et al., 2000). There is no overlap of
306 detrital garnets with the garnets from the garnet peridotite. This suggests that the eclogites and the
307 gneisses are the main source of the detrital garnets. Due to the fact that the grain size of the garnets in
308 the garnet peridotite is coarser (the short axis of the smallest grain is $>1000 \mu\text{m}$) than the garnets in the
309 studied 63–125 μm grain-size fraction, it is not very likely that these garnets occur in this grain-size
310 fraction. Of course, coarse-grained source rock garnets, crushed through hydraulic or other mechanical
311 processes, can occur in the finer grain-size fractions, but less likely than garnets with an original finer
312 size distribution in the source rocks. In contrast, garnets of original finer size distribution in the source
313 rocks cannot be expected in the coarse grain-size fractions. Therefore, it can be assumed that lower
314 grade metamorphic garnets, possibly derived from the garnet-bearing gneisses, are generally finer,

315 because almandine garnets with higher MnO content are more frequent in the finer grain-size fractions
316 (Fig. 7). In the ternary diagram, only two detrital garnets of the very coarse 500–1000 μm grain-size
317 fraction overlap with the garnets from the peridotites (Fig. 6c). In the biplots (Fig. 5c), this number is
318 slightly higher and includes grains from the 250–500 μm , too. Therefore, a general coarse grain size of
319 garnets from the garnet peridotites exposed in this area can be assumed, as observed in the bedrock
320 sample AK-N24, too. The fact that only grains from the coarse grain-size fractions show overlap with
321 the garnet peridotite (Figs. 5c, 6c) is likely due to the inheritance of grain size from source rock to
322 sediment. Alternatively, the peridotites contribute only little detritus to the sediment and the number of
323 50 garnets measured from the 63–125 μm grain-size fraction is insufficient to detect them. As an
324 effect of hydraulic sorting during settling, heavy minerals concentrate in specific grain-size classes
325 (Rubey, 1933; Rittenhouse, 1943; Garzanti et al., 2008; Resentini et al., 2013). The specific
326 endmembers of garnet show different densities, with the highest density of $\sim 4.3 \text{ g/cm}^3$ for almandine
327 garnets, 4.2 g/cm^3 for spessartine garnets and 3.6 g/cm^3 for pyrope garnets (Deer et al., 1992).
328 Commonly, garnets with a higher density are found in the fine tail of the grain-size distribution in
329 contrast to less dense garnet species (Schuiling et al., 1985; Andò, 2007; Garzanti et al., 2008). Due to
330 the fact that the less dense pyrope-rich almandine garnet is more frequent in the coarse grain-size
331 fraction, hydraulic sorting according to their density can be excluded and it is more likely that this is
332 an effect of grain-size distribution in the source rocks.

333 The short axes of the garnets lie within the expected interval of 63–125 μm , except of a few grains.
334 The long axes show a wider distribution in their sizes (>80 –260 μm) (Fig. 4; Table 2). Most of the
335 long axes (38–86 %) are longer in size than the sieved grain-size fraction (Fig. 4; Table 2). This means
336 that a wet-sieved sample is only separated by the short axes of the minerals. Therefore, it is possible
337 that grains with a short axis of 120 μm and a long axis of 480 μm , for instance, occur in the 63–125
338 μm grain-size fraction although their average grain size may be around 300 μm . The highest
339 concentration of garnet grains occurs in the 125–500 μm grain-size fraction (Fig. 3b). This also points,
340 in contrast to e.g. apatite, to a generally coarser garnet grain size in the source rocks.

341

342 **6. Conclusions**

343 The heavy mineral assemblages reflect the geological situation in the area of Almklovdalen. The
344 dominant heavy mineral is olivine. From upstream to downstream the content of olivine strongly
345 increases, whereas the content of all other heavy minerals found in the samples strongly decreases.
346 This is because in the upstream part of the sampled river the diversity of potential source rocks is
347 higher than in the downstream part of the river where the proportion of dunite and chlorite peridotite is
348 much higher. Garnets with high MgO content are more frequent in the coarse grain-size fractions,
349 which likely result from the inheritance of grain size from source rock to sediment. However, only
350 very few garnets of the 250–1000 μm grain-size fractions show full overlap with the garnets from the
351 garnet peridotite sample. In contrast, garnets with high MnO content are more frequent in the fine
352 grain-size fractions. Therefore, analysing a wide grain-size window is of paramount importance, as
353 also discussed in other studies (e.g., Garzanti et al., 2009; Krippner et al., 2015), because we can miss
354 information contained in other grain-size fractions.

355

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360

361 **Appendix A. Supplementary data**

362 Supplementary data associated with this article can be found, in the online version, at xxx.

363

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533

534 **Tables**

535

536 **Table 1.** Geographic coordinates (WGS 84) of all samples (river sediments and bedrocks) from the
537 Almklovdalen area.

538

539 **Table 2.** Operating conditions of the electron microprobe for garnet analyses. *Count time* on the peak (in s),
540 *Bckg time* counting time on background position (in s), *DL* detection limit (in ppm).

541

542 **Table 3.** Mineralogy of the bedrock samples.

543

544 **Figure captions**

545

546 **Fig.1.** Map of the outer Nordfjord and Stradlandet area of the Western Gneiss Region showing the
547 distribution of ultramafic bodies. Dashed lines mark the limits of the HP and UHP zones and the
548 extent of the mixed HP/UHP zone (adapted from Carswell and Cuthbert, 2003). The inset shows the
549 location of the study area in the SW part of the Caledonides in Norway.

550

551 **Fig. 2.** Geological map of the Almklovdalen area (adapted from Medaris and Brueckner, 2003)
552 showing sample locations. The eclogite occurrences have sizes up to several meters and are enclosed
553 within both the gneisses and peridotites.

554

555 **Fig. 3. a)** Transparent heavy mineral suite of the 63–125 μm grain-size fraction of the analysed
556 samples. **b)** Heavy mineral distribution of all grain-size fractions of samples AK-N19-3.
557 Epidote-group: epidote, zoisite; ultrastable minerals: zircon, rutile, tourmaline; others: titanite, other
558 amphibole (excluding green calcic amphibole).

559

560 **Fig. 4.** Grain-size distribution of the short and the long axes of garnets from stream sediments of the
561 63–125 μm grain-size fraction (upper part) and of the eclogite samples AK-N21 and AK-EA (lower
562 part).

563

564 **Fig. 5.** Compositional biplot of garnets based on major elements Al_2O_3 , SiO_2 , CaO , FeO , MnO , and
565 MgO . Axes are first and second principal components (PC). **a)** Compositional biplot of garnet grains
566 from the bedrocks. Almost 73% of the variability is explained by the first PC and 13% by the second
567 PC. **b)** Compositional biplot of detrital garnet grains from the bedrocks and from all stream sediments
568 of the 63–125 μm grain-size fraction. Almost 51% of the variability is explained by the first PC and
569 20% by the second PC. **c)** Compositional biplot from sample AK-N19-3 of the 63–125 μm , 125–250
570 μm , 250–500 μm , and 500–1000 μm grain-size fractions. Almost 49% of the variability is explained
571 by the first PC and 20% by the second PC.

572

573 **Fig. 6.** Composition of garnets in the ternary classification diagram of Mange and Morton (2007) with
574 almandine+spessartine, grossular and pyrope as poles. **a)** Composition of garnets from the bedrock
575 samples. **b)** Composition of detrital garnets of the 63–125 μm grain-size fraction and from the bedrock
576 samples. **c)** Composition of detrital garnets of the 125–250 μm , 250–500 μm , and 500–1000 μm
577 grain-size fraction and from the bedrock samples. A – mainly from high-grade granulite-facies
578 metasedimentary rocks or charnockites and intermediate felsic igneous rocks, B – amphibolite-facies
579 metasedimentary rocks, Bi – intermediate to felsic igneous rocks, Ci – mainly from high-grade mafic
580 rocks, Cii – ultramafics with high Mg (pyroxenites and peridotites), D – metasomatic rocks, very low-
581 grade metamafic rocks and ultrahigh-temperature metamorphosed calc-silicate granulites.

582

583 **Fig. 7.** Binary diagram showing CaO/MnO vs. MgO/MnO garnet values of all grain-size fractions of
584 sample AK-N19-3.