

1 **High pressure melting of eclogite and metasomatism of garnet peridotites from Monte**
2 **Duria Area (Central Alps, N Italy): a proxy for melt-rock reaction during subduction**

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14 **Highlights**

- 15 • Structural and microstructural evidence of eclogite partial melting at P-T conditions
16 corresponding to warm subduction.
- 17 • Unique case study of eclogite-derived melts interacting with garnet peridotite at high
18 pressure in the Alps (Adula-Cima Lunga unit, N Italy).
- 19 • Syn-deformation development of garnet websterites during melt-peridotite interaction in
20 subduction environments.

21

22 **Abstract**

23 In the Monte Duria area (Adula-Cima Lunga unit, Central Alps, N Italy) garnet peridotites occur in
24 direct contact with migmatized orthogneiss (Mt. Duria) and eclogites (Borgo). Both eclogites and
25 ultramafic rocks share a common high pressure (HP) peak at 2.8 GPa and 750 °C and post-peak
26 static equilibration at 0.8-1.0 GPa and 850 °C. Garnet peridotites show abundant amphibole,

27 dolomite, phlogopite and orthopyroxene after olivine, suggesting that they experienced
28 metasomatism by crust-derived agents enriched in SiO₂, K₂O, CO₂ and H₂O. Peridotites also
29 display LREE fractionation (La/Nd = 2.4) related to LREE-rich amphibole and clinopyroxene
30 grown in equilibrium with garnet, indicating that metasomatism occurred at HP conditions. At
31 Borgo, retrogressed garnet peridotites show low strain domains characterised by garnet
32 compositional layering, cut by a subsequent low-pressure (LP) chlorite foliation, in direct contact
33 with migmatized eclogites. Kfs+Pl+Qz+Cpx interstitial pocket aggregates and Cpx+Kfs thin films
34 around symplectites after omphacite parallel to the Zo+Omp+Grt foliation in the eclogites suggest
35 that they underwent partial melting at HP.

36 The contact between garnet peridotites and eclogites is marked by a tremolitite layer. The same
37 rock also occurs as layers within the peridotite lens, showing a boudinage parallel to the garnet
38 layering of peridotites, flowing in the boudin necks. This clearly indicates that the tremolitite
39 boudins formed when peridotites were in the garnet stability field. Tremolitites also show
40 Phl+Tc+Chl+Tr pseudomorphs after garnet, both crystallised in a static regime postdating the
41 boudins formation, suggesting that they derive from a garnet-bearing precursor. Tremolitites have
42 Mg# > 0.90 and Al₂O₃ = 2.75 wt.% pointing to ultramafic compositions but also show enrichments
43 in SiO₂, CaO, and LREE suggesting that they formed after the reaction between the eclogite-
44 derived melt and the garnet peridotite at HP. To test this hypothesis, we performed a
45 thermodynamic modelling at fixed P = 3 GPa and T = 750 °C to model the chemical interaction
46 between the garnet peridotite and the eclogite-derived melt. Our results show that this interaction
47 produces an Opx+Cpx+Grt assemblage plus Amp+Phl, depending on the water activity in the melt,
48 suggesting that tremolitites likely derive from a previous garnet websterite with amphibole and
49 phlogopite.

50 Both peridotites and tremolitites also show a selective enrichment in LILE recorded by amphiboles
51 in the spinel stability field, indicating that a fluid-assisted metasomatic event occurred at LP

52 conditions, leading to the formation of a chlorite foliation post-dating the garnet layering in
53 peridotites, and the retrogression of Grt-websterites in tremolitites.

54 The Monte Duria area is a unique terrane where we can observe syn-deformation eclogite-derived
55 melt interacting with garnet peridotite at HP, proxy of subduction environments.

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57

58 **Keywords**

59 Slab melting, websterite, subduction fluids, dolomite, Adula nappe.

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61

62 **1. Introduction**

63 The partial melting of metasedimentary and metagranitic rocks interacting with garnet
64 peridotites – proxy of the interaction between the subducted crust and the overlying mantle – can be
65 found in high-ultrahigh pressure (HP-UHP) terranes such as Ulten Zone in the Central Italian Alps
66 (Rampone and Morten, 2001; Tumiati et al., 2003; Scambelluri et al., 2006) and the chinese Dabie-
67 Sulu belt (Malaspina et al., 2006; 2009). Alternatively, evidence of partial melting of HP rocks
68 representative of the mafic portions of the subducted slab is very rare and its evidence is seen in the
69 production of peraluminous trondhjemite–tonalite melts (so-called adakites) forming volcanic suites
70 of andesitic to rhyodacitic composition (Drummond et al., 1996). The observation of HP partial
71 melting of eclogites has been reported only by Wang et al. (2014) in the Central Sulu orogen, while
72 to date the direct interaction between eclogite-derived melts and peridotites at HP has never been
73 observed *in situ* in orogenic peridotites.

74 Melting of mafic rocks in a subducting lithosphere at HP requires at least $T = 750\text{ °C}$ at $P =$
75 3 GPa , under H_2O saturated conditions (Schmidt and Poli, 2014), typical of low angle subduction
76 systems (Peacock and Wang, 1999) also called “warm subduction” (Syracuse et al., 2010). The
77 Monte Duria area (Adula-Cima Lunga nappe, Central Alps, Italy) is a unique terrane where we can

78 observe eclogite-derived melt interacting with garnet peridotite at P-T conditions proxy of “warm”
79 subduction paths. In this area garnet peridotites occur in direct contact with migmatized orthogneiss
80 and eclogites, sharing a common metamorphic history reaching HP peak at 2.8 GPa and 750 °C and
81 post-peak static equilibration at 0.8-1 GPa and 850 °C (Tumiati et al., 2018). In this study we will
82 show that garnet peridotites from the Monte Duria area record a multistage metasomatic event by
83 eclogite-derived adakite-like melts at HP inducing a selective enrichment in LREE, and by
84 retrograde fluids at lower pressures yielding to LILE enrichments in amphiboles. The interaction
85 between garnet peridotites and eclogite leucosomes also produces a modal metasomatism evidenced
86 by the occurrence of garnet websterite layers at the contact between eclogite and peridotite, now
87 retrogressed in the tremolite stability field.

88 The fate of crust-derived melts in the associated peridotites and their capability to transport
89 crustal components to the mantle is still matter of debate. Many occurrences show that crustal melts
90 may migrate into the mantle by porous flow, producing almost monomineralic metasomatic
91 orthopyroxene and phlogopite layers at the slab-mantle interface (e.g. Malaspina et al., 2006; 2009;
92 Endo et al., 2015). Such a mechanism, however, may limit the crust-to-mantle mass transfer unless
93 they are partitioned into residual aqueous fluids (Malaspina et al., 2009). Alternatively,
94 metasomatic melts may migrate into the mantle by focussed flow, forming a network of pyroxenite
95 veins as shown by metasomatized mantle xenoliths from arc lavas (e.g. Arai et al., 2003;
96 Kepezhinskas et al., 1995). We will show that the melt-peridotite interaction preserved in Borgo
97 outcrop of Monte Duria area occurred under a deformation regime (shearing) at HP, promoting the
98 combination of porous and focussed flow of eclogite-derived melts into the garnet peridotite. The
99 melt-enhanced deformation and flow in garnet peridotites documented in the Monte Duria area may
100 represent an exhumed example of crustal flow channels forming at the slab-mantle interface, that
101 facilitate the migration of crustal melts in the supra-subduction mantle.

102

103 **2. Geological and petrological background**

104

105 The Adula-Cima Lunga nappe (ACL) complex is the structurally highest unit of the Sub-
106 Penninic domain occurring within the Lepontine Dome (Milnes, 1974; Schmid et al., 1996). In this
107 area of the Central Alps, the ACL is comprised between the underlying Lower Penninic Simano and
108 Leventina nappes (Fig. 1) and the Middle Penninic units of Tambò, Suretta and Schams nappes
109 (Schmid et al., 1990). The base of the ACL is defined by the Misox Zone, a N-S trending
110 heterogeneous unit made by carbonatic and siliciclastic metasediments (Bündnerschiefer *Auct.*), by
111 MORB-derived amphibolites and slivers of continental basement (Stucki et al., 2003). The southern
112 limit of the ACL (Fig. 1) is represented by the Paina Fault (Fumasoli, 1974), marked in the field by
113 a thin slice of silicate-bearing marble interpreted to represent the upper contact of the ACL
114 (Fumasoli, 1974; Schmid et al., 1996). The Bellinzona-Dascio zone, occurring south of the Paina
115 Fault (Fig. 1), is either considered to be the continuation of the Misox Zone (Schmid et al., 1996),
116 or as to be part of the ACL (Berger et al., 2005). At the SE boundary of ACL the Gruf Complex is
117 separated from the ACL unit by the Forcola normal fault. The Gruf complex experienced ultrahigh
118 temperature (UHT) metamorphism and is considered to be of Permian (Galli et al., 2011) or
119 Oligocene age (Droop and Bucher, 1984; Liati and Gebauer, 2003; Nicollet et al., 2018).

120 The ACL mainly consists of ortho- and paragneiss of pre-Mesozoic age, partially retrogressed
121 eclogites, amphibolites, marbles, metasedimentary rocks of supposed Mesozoic origin (Galster et
122 al., 2012) and minor serpentinite and peridotite bodies (Evans and Trommsdorff, 1978; Fumasoli,
123 1974; Heinrich, 1986). The ACL experienced high pressure (HP) metamorphism during both the
124 Variscan (e.g. Biino et al., 1997; Herwartz et al., 2011; Liati et al., 2009) and the Alpine orogenic
125 cycle (Dale and Holland, 2003; Meyre and Frey, 1998; Nagel et al., 2002). Despite the intensive
126 Alpine metamorphism and deformation, relicts of pre-Alpine (Variscan) HP metamorphism are still
127 preserved in the central and northern sectors of the ACL (Herwartz et al., 2011; Liati et al., 2009).
128 P-T estimates of HP assemblage in eclogites from Confin, Val Large and Trescolmen, suggest
129 equilibrium conditions of $P = 1.4\text{-}2.2$ GPa and $T = 500\text{-}700$ °C (Dale and Holland, 2003; Nagel,

130 2008). Peak conditions for the Alpine eclogites increase southward from $P \approx 2.0$ GPa and $T = 600$
131 $^{\circ}\text{C}$ in the north, to $P \approx 3.0$ GPa and $T = 650\text{-}750$ $^{\circ}\text{C}$ in the south (Brouwer et al., 2005; Dale and
132 Holland, 2003; Heinrich, 1986). The highest eclogitic peak conditions were recognised in the
133 ultramafic rocks cropping out in the southern part of the nappe.

134 In the ACL garnet peridotites occur at three localities, from west to east: Cima di Gagnone
135 (Evans and Trommsdorff, 1978); Alpe Arami (Mockel, 1969); and Monte Duria (Fumasoli, 1974).
136 Chlorite harzburgite and garnet lherzolites from Cima di Gagnone represent slivers of
137 subcontinental mantle exposed at the seafloor during the formation of the Alpine Tethys and
138 subsequently involved in the Alpine subduction cycle (Evans and Trommsdorff, 1978;
139 Trommsdorff et al., 2000; Scambelluri et al., 2014). Thermobarometry of the peak garnet-facies
140 assemblage provided estimates of $750\text{-}800$ $^{\circ}\text{C}$ and $2.5\text{-}3$ GPa for the eclogitic event (Evans &
141 Trommsdorff, 1978; Nimis and Trommsdorff, 2001). Garnet peridotites from Alpe Arami have
142 been interpreted as derived either from a subducted sub-continental lithospheric mantle
143 (Trommsdorff et al., 2000) or from portions of supra-subduction mantle (Brenker and Brey, 1997).
144 Similarly, garnet and chlorite peridotites from Monte Duria Area have been interpreted as supra-
145 subduction mantle peridotites located above the subducting Alpine Tethys lithosphere during late
146 Cenozoic (Hermann et al., 2006). P-T estimates for the HP metamorphism experienced by garnet
147 peridotites of Alpe Arami and Monte Duria yielded P-T conditions of 3.2 GPa and 844 $^{\circ}\text{C}$ and 3.0
148 and 830 $^{\circ}\text{C}$, respectively (Nimis and Trommsdorff, 2001). Similar peak pressure conditions and
149 slightly lower temperature (2.8 ± 0.2 GPa and 730 ± 20 $^{\circ}\text{C}$) have been reported by Tumiati et al.
150 (2018) for the HP assemblages of garnet peridotites and associated eclogites of Monte Duria area.

151 Recently Tumiati et al. (2018) provided also evidence for a previously unknown LP-(U)HT
152 metamorphic event which post-dates the HP assemblages and pre-dates the Barrovian
153 metamorphism of the Lepontine Dome in peridotites and eclogites of Monte Duria area. Pressure-
154 temperature estimates of LP-(U)HT assemblages found in both mafic and ultramafic rocks yielded
155 $T = 850$ $^{\circ}\text{C}$ and $P = 0.8\text{-}1.2$ GPa. It is worthy of note that similar conditions are also reported for the

156 UHT stage recorded in the sapphirine-bearing granulites and charnockites of the near Gruf Complex
157 (Galli et al., 2011).

158 The peridotites and associated crustal rocks of the Monte Duria area, together with the whole
159 ACL complex, eventually experienced the post-collisional metamorphism that affected the entire
160 central Alpine nappe stack (Nagel, 2008). The Barrovian-type metamorphism increases southward
161 from upper green-schists facies in the north, to high amphibolite facies in the south (Nagel, 2008;
162 Todd and Engi, 1997), where it promoted crustal anatexis in a narrow belt close to the Insubric Line
163 (e.g. Burri et al., 2005; Rubatto et al., 2009). Partial melting was promoted by fluid infiltration and
164 occurred between 32 and 22 Ma suggesting a protracted high thermal history during Barrovian-type
165 metamorphism (Rubatto et al., 2009).

166

167 **3. Field occurrence**

168

169 The Monte Duria area is located at northwest of the Como Lake (Northern Italy) (Fig. 1a) and
170 it is part of the southern ACL. Peridotite lenses hosted by crustal rocks occur at two localities: (i)
171 Mt. Duria and (ii) Borgo (Fig. 1b). At Mt. Duria garnet peridotites are hosted by migmatitic biotite-
172 muscovite gneiss whereas at Borgo a large chlorite peridotite body is in direct contact with
173 amphibole-bearing migmatites, the latter containing several boudins of lithologically heterogeneous
174 eclogites (Tumiati et al., 2018; Fig. 2).

175 At the mesoscale, the garnet peridotites of Mt. Duria display a compositional layering
176 consisting of garnet-rich and garnet-poor layers, transposed by a chlorite foliation. Such foliation
177 becomes more penetrative close to the outermost part of the peridotite lenses (Fig. 3a, b).
178 Peridotites consist of garnet, clinopyroxene and minor amphibole porphyroclasts enclosed in a dark
179 fine-grained matrix consisting of olivine, clinopyroxene and orthopyroxene. Garnet is surrounded by
180 kelyphitic coronas and in most of the peridotite lenses is progressively replaced by chlorite

181 pseudomorphs. Garnet-pyroxenite lenses often occur within the garnet peridotite bodies of the
182 Monte Duria area (Fumasoli, 1974; Hermann et al., 2006).

183 At Borgo, an hm-sized chlorite peridotite body is in direct contact with amphibole-bearing
184 migmatites containing boudins of mafic, high- Al_2O_3 and kyanite-bearing eclogites (Fig. 2; Tumiati
185 et al., 2018). The chlorite peridotite body, along with the associated mafic rocks, is enclosed within
186 biotite-bearing migmatitic gneiss (yellow in Fig. 2). The chlorite peridotite (lilac in Fig. 2) mainly
187 consists of fine-grained olivine, orthopyroxene and rare clinopyroxene. Garnet is always replaced
188 by chlorite pseudomorphs (Fig. 3c), likely indicating that the Borgo chlorite peridotites represent
189 the retrogressed variety of the garnet peridotites of Mt. Duria. Chlorite peridotites preserve a
190 compositional layering made by alternate levels rich or poor in chlorite pseudomorphs after garnet.
191 Such layering is locally transposed by a new chlorite-bearing foliation (Fig. 3c, d).

192 The contact between the peridotite body and the associated mafic rocks at Borgo is marked by
193 the occurrence of a tremolitite layer (violet in Fig. 2) composed by more than 90 vol.% of fine-
194 grained tremolite associated with several phlogopite + chlorite + talc + tremolite pseudomorphs
195 after garnet and minor Mg-hornblende. These rocks also occur as m-scale boudins within the
196 peridotite body and show sharp contacts with the host peridotite (Fig. 4a, b, c). The garnet foliation
197 in the peridotite body wraps around the boudins and flows into the boudins necks indicating that the
198 deformation of tremolitite precursors occurred during the development of the garnet foliation in the
199 host peridotite (Fig. 4b, c). Such boudins are confined within 20 meters from the contact (Fig. 2).

200 The amphibole-bearing migmatites of Borgo display evidence for partial melting at different
201 scales. These rocks (salmon in Fig. 2) display mm- to m-thick leucosomes consisting of quartz +
202 plagioclase + K-feldspar + biotite and amphibole-rich restitic layers (Fig. 4d). Partial melting
203 occurred in a deformation regime, as displayed by the strong layered structure of amphibole-bearing
204 migmatites and the occurrence of peculiar deformation structures like lobes and cusps (Fig. 4e;
205 McLellan, 1989). These migmatites contain boudins of mafic, high- Al_2O_3 and kyanite-bearing
206 eclogites (Tumiati et al., 2018). Fine grained, dark green mafic eclogites (dark green in Fig. 2)

207 occur as m-sized boudins. These rocks show a compositional layering consisting of garnet-rich and
208 garnet-poor layers. At the rim of the boudins, such layering is crosscut by the foliation of the
209 surrounding amphibole-bearing migmatites. Larger boudins (up to 10 meters) of light green,
210 kyanite-bearing eclogites also occur (Fig. 2). As already described for the mafic eclogites, a
211 compositional layering marked by garnet-rich and garnet-poor layers represents the main fabric at
212 the mesoscale. The high- Al_2O_3 eclogite (red rim in Fig. 2) occurs as a cm-thick reddish corundum-
213 rich rim at the contact between kyanite-bearing eclogite and amphibole-bearing migmatites. These
214 rocks are garnet-free and the only HP relicts are porphyroblastic kyanite and emerald-green zoisite
215 (Tumiati et al., 2018).

216 The peridotite body and the associated mafic rocks are both separated from the surrounding
217 migmatitic biotite gneiss by a few meters thick mylonitic shear zone (Fig. 2; Fig. S-1a). The
218 migmatitic gneiss has a stromatic structure defined by alternating leucocratic bands composed by
219 quartz, plagioclase and alkali feldspar, and melanocratic domains enriched in biotite.

220

221 **4. Petrography**

222

223 Representative samples among garnet peridotites from Mt. Duria and chlorite peridotites,
224 tremolitites, mafic eclogite, kyanite-bearing eclogite, high- Al_2O_3 eclogite and amphibole-bearing
225 migmatites from Borgo (Fig. 2) have been selected for this study. All the recognised mineral
226 assemblages and the relative metamorphic conditions are summarized in Table 1.

227

228 *4.1. Peridotites*

229

230 Garnet peridotites from Mt. Duria show porphyroclastic texture with mm-sized garnet,
231 clinopyroxene (Cpx_1), orthopyroxene (Opx_1) and amphibole (Amp_1), surrounded by fine-grained
232 recrystallised matrix of olivine, orthopyroxene (Opx_2), clinopyroxene (Cpx_2), amphibole (Amp_2)

233 and Cr-spinel (Fig. 5a, b). Garnet porphyroclasts contain inclusions of clinopyroxene,
234 orthopyroxene and olivine. In sample B3A inclusions of edenitic to pargasitic amphibole and
235 dolomite have been observed in Cpx₁. Frequently, garnet is statically replaced by chlorite (Chl₁)-
236 bearing pseudomorphs surrounded by amphibole + spinel coronas (Fig. S-2a). During the
237 subsequent LP-(U)HT metamorphism (see Tumiati et al., 2018), Chl₁ decomposes into a
238 orthopyroxene + spinel assemblage (Fig. S-2b, c), while composite kelyphitic coronas consisting of
239 orthopyroxene (Opx_{Sym}) after olivine and orthopyroxene (Opx_{Sym}) + clinopyroxene (Cpx_{Sym}) +
240 spinel (Sp_{Sym}) ± amphibole (Amp_{Sym}) after garnet developed in correspondence of the previous
241 garnet-olivine grain boundaries. In kelyphites replacing olivine tiny crystals of baddeleyite and
242 srilankite occur, while sapphirine crystallises within symplectitic assemblages around garnet
243 (Tumiati et al., 2018).

244 Chlorite peridotites display a porphyroclastic microstructure with mm-sized chloritised garnet,
245 in textural equilibrium with coarse olivine, orthopyroxene and minor clinopyroxene. Sample C2A
246 shows a finer-grained assemblage made of olivine, clinopyroxene, orthopyroxene, amphibole
247 (Amp₂) and Cr-spinel, resembling the recrystallised matrix of garnet peridotite from Mt. Duria.
248 Both samples C2A and DB113 display a strong foliation defined by syn-tectonic chlorite (Chl₃) and
249 amphibole (Amp₃), post-dating the porphyroclastic and neoblastic assemblages (Fig. S-2b).
250 Notably, sample DB113 has been collected close to the contact with the innermost tremolite
251 boudins and shows relict olivine and orthopyroxene extensively statically overgrown by a new
252 generation of porphyroblastic orthopyroxene (Opx_{Porph}), Mg-hornblende (Amp_{Porph}) and phlogopite
253 (Phl_{Porph}) (Fig. 5c). These microstructures clearly indicate a metasomatic event pre-dating the
254 chlorite foliation. Several dolomite crystals (Fig. 5d) and round-shaped brucite + calcite
255 pseudomorphs after dolomite also occur within the olivine matrix.

256

257 *4.2 Tremolite layers within the peridotite*

258

259 Tremolites display an assemblage dominated by tremolite, Mg-hornblende and minor
260 chlorite. In all the investigated samples (DB148, DB151 and DB179), tremolite forms a mosaic-like
261 texture with 120° triple junctions and straight grain boundaries. The tremolite matrix hosts older
262 mm-sized Mg-hornblende porphyroblasts (Fig. 5e). Mg-hornblende is zoned, showing dusty cores
263 with opaque minerals occurring along the cleavages, and clear inclusion-free rims. Isolated
264 phlogopite (Phl₃) + chlorite (Chl₃) + talc + tremolite hexagonal pseudomorphs after garnet have
265 been recognised in textural equilibrium with tremolite (Fig. 5f).

266

267 *4.3 Eclogite boudins and amphibole-bearing migmatites*

268

269 Mafic eclogites display a porphyroblastic microstructure with mm-sized garnet, K-feldspar,
270 quartz and zoisite embedded in a fine-grained matrix composed by Mg-hornblende (Amp₂),
271 diopside-rich clinopyroxene (Cpx₂) and albitic plagioclase (Pl₂) (Fig. 6a). Garnet is usually zoned,
272 showing dusty cores with inclusions of quartz, omphacite and minor rutile, and inclusions free rims.
273 A 30-µm inclusion of dolomite has been also observed. Garnet is occasionally replaced by
274 pseudomorphs composed by Mg-hornblende (Amp₂) and plagioclase (Pl₂). K-feldspar locally
275 displays 50-100-µm thick clinopyroxene ± Mg-hornblende coronas (Fig. 6a). Few quartz
276 porphyroblasts have irregular shape, with locally lobate-cusped grain boundaries surrounded by 50-
277 µm thick coronas composed by K-feldspar (Kfs_M) and minor clinopyroxene (Cpx_M) (Fig. 6a).
278 Zoisite commonly occurs as 100-150-µm porphyroblasts and locally surrounded by tiny, ameboid
279 allanite crystals (Fig. 6b).

280 Kyanite-bearing eclogites display a porphyroblastic microstructure with abundant garnet,
281 kyanite and quartz. Few tiny inclusions of omphacite and rutile occur within garnet and kyanite
282 porphyroblasts. The rock matrix is composed by Mg-hornblende (Amp₂), diopside-rich
283 clinopyroxene (Cpx₂) and albitic plagioclase (Pl₂). Similarly to the mafic eclogites, garnet and
284 kyanite are usually replaced by symplectitic intergrowths consisting of Mg-hornblende (Amp_{Sym})

285 and albitic plagioclase (Pl_{Sym}), whereas symplectites around kyanite are composed by anorthitic
286 plagioclase (Pl_{Sym}), spinel (Sp_{Sym}), orthopyroxene (Opx_{Sym}) and minor sapphirine (c.f. Tumiati et
287 al., 2018).

288 The high- Al_2O_3 eclogites display a strong foliation defined by syn-tectonic porphyroblastic
289 kyanite in textural equilibrium with quartz and zoisite porphyroblasts. Also these minerals are
290 embedded in a fine-grained matrix composed by diopside-rich clinopyroxene (Cpx_2) + albite-rich
291 plagioclase (Pl_2) + Mg-hornblende (Amp_2) pseudomorphs after omphacite. Kyanite is partly
292 replaced by symplectitic coronas consisting of An-rich plagioclase (Pl_{Sym}), spinel (Sp_{Sym}),
293 corundum and minor sapphirine (c.f. Tumiati et al., 2018). Similarly to K-feldspar in mafic
294 eclogites, also quartz in these samples locally displays corroded edges surrounded by fine-grained
295 intergrowths consisting of anhedral K-feldspar (Kfs_M) and clinopyroxene (Cpx_M) (Fig. 6c). The
296 $Cpx_M + Kfs_M$ intergrowths are preferentially oriented parallel to the main foliation suggesting that
297 they formed in a deformation regime when the kyanite + omphacite + quartz HP assemblage was
298 still stable. Interstitial pocket aggregates made of $Kfs_M + Cpx_M +$ plagioclase (Pl_M) + quartz (Qz_M)
299 have been also observed at quartz grain boundaries (Fig. 6d). In all the three types of eclogites, the
300 symplectitic domains are surrounded by coronas defined by porphyroblastic paragonitic hornblende
301 (Amp_3) and biotite.

302 The amphibole-bearing migmatites hosting the eclogite boudins display a layered
303 microstructure, consisting of leucocratic domains composed by quartz + plagioclase \pm Mg-
304 hornblende \pm biotite and melanocratic domains enriched in hornblende. Leucocratic domains
305 display a mosaic-like equilibrium microtexture with grain boundaries triple junctions at 120° . The
306 melanocratic domains are strongly foliated dominated by Mg-hornblende and by a fine-grained
307 matrix consisting of clinopyroxene (Cpx_2), amphibole (Amp_2) and albitic plagioclase (Pl_2) replacing
308 previous garnet and omphacite (Fig. 6e, f). These microtextures indicate that the hornblende-rich
309 domains derive from a previous eclogitic precursor.

310

311 **5. Bulk rock chemistry**

312

313 *5.1. Major elements*

314

315 The major elements composition of the investigated rocks is reported in Table 2. As shown in
316 Figure 7, the composition of Mt. Duria garnet peridotites resembles that of the reference Depleted
317 Mantle (Salters and Stracke, 2004), displaying $X_{Mg} = 0.90$, Ni = 1960-1975 ppm, and low Al_2O_3
318 (2.67-3.05 wt.%) and CaO (2.16-2.52 wt.%) concentrations. In Figure 7 the composition of
319 peridotites from other localities of the Central and Eastern Alps have been also portrayed. Our
320 samples plot close to the Alpe Arami (AA), Bellinzona-Dascio Zone (BDZ) and Ulten (UZ)
321 peridotites, whereas the ultramafic rocks from Cima di Gagnone (CdG) show lower Ni (1386-1614
322 ppm), and slightly higher Al_2O_3 (2.87-4.44 wt.%) and CaO (2.48-3.65 wt.%) concentrations.
323 Tremolitites (purple square) show $X_{Mg} = 0.91$ and Al_2O_3 concentrations comparable to mantle
324 values. Moreover, in the X_{Mg} -Ni variation diagram, tremolitites plot into the field of the ultramafic
325 compositions probably indicating that they derive from an ultramafic precursor. Despite this mantle
326 signature, they show high SiO_2 (up to ca. 55 wt.%), high CaO (12.38 wt.%) and low Ni (554 ppm)
327 concentrations (Tab. 2).

328 The bulk-rock composition of mafic, kyanite-bearing and high- Al_2O_3 eclogites indicates that
329 they all have a mafic composition ($SiO_2 = 48.96$ -51.17 wt.%; Tab. 2, Figure 7) and are compared to
330 that of N-MORB (Gale et al., 2013), layered mid ocean-ridge gabbros (Gillis et al., 2014) and
331 ophiolitic gabbros from the Bellinzona-Dascio Zone (Stucki et al., 2003). The composition of the
332 mafic eclogite overlaps that of the reference N-MORB showing $X_{Mg} = 0.61$, Ni = 139 ppm, CaO =
333 9.57 wt.% and $Al_2O_3 = 15.50$ wt.%. The other two types of eclogite display a composition
334 resembling that of reference gabbros with X_{Mg} ranging from 0.74 to 0.83 and the highest CaO
335 (11.52-11.67 wt.%) and Al_2O_3 (16.11-18.63 wt.%) concentrations.

336

337 5.2. Trace elements

338

339 The bulk rock trace elements composition of the analysed samples is reported in Table 2 and
340 portrayed in Figure 8, normalised to the Primitive Mantle (PM, McDonough and Sun, 1995). The
341 trace elements pattern of the Depleted Mantle (DM, Salters and Stracke, 2004) is also reported for
342 comparison (DM, blue bold line of Fig. 8a, b). Garnet peridotites show absolute rare earth elements
343 (REE) concentrations slightly lower than the PM, with fractionated patterns enriched in light-REE
344 (LREE) ($La_N/Nd_N = 2.4$) relative to the medium-REE (MREE) and heavy-REE (HREE) (Fig. 8a).
345 Also the trace elements composition of the studied peridotites is compared to that of peridotites
346 from Cima di Gagnone and Ulten Zone (Fig. 8a, b). The studied peridotites overlap the fractionated
347 REE patterns of Ulten peridotites, whereas Cima di Gagnone ultramafic rocks (grey area) display
348 LREE depletion with respect to MREE and HREE following the same pattern of the DM. In terms
349 of other trace elements, the composition of the investigated peridotites broadly resembles that of
350 Ulten Zone and Cima di Gagnone ultramafic rocks, showing relatively high large ion lithophile
351 elements (LILE) concentrations (i.e. $Cs = 50 \times PM$, Fig. 8b).

352 Tremolites have REE concentrations up to $2.6 \times PM$ with enrichments in MREE and LREE
353 (i.e. $La_N/Er_N = 2.15$ and $Nd_N/Er_N = 2.97$) relative to HREE (Fig. 8c). These rocks do not show any
354 appreciable LILE enrichment except for relatively high Pb concentrations (up to $10 \times PM$; Fig. 8d).

355 The trace elements compositions of eclogites are compared with the composition of N-MORB
356 (dark grey line in Fig. 8e; Gale et al., 2013) and ophiolitic gabbros from Bellinzona-Dascio Zone
357 (light grey field in Fig. 8e; Stucki et al., 2003). The REE pattern of mafic eclogite broadly
358 resembles that of reference N-MORB with REE absolute concentrations up to $10 \times PM$ and a slight
359 Eu negative anomaly. Both high- Al_2O_3 and kyanite eclogites display REE concentrations almost
360 one order of magnitude lower than those of mafic eclogites (Fig. 8e), showing a slight fractionation
361 in LREE with respect to MREE and HREE ($La/Nd_N = 1.24$), falling in the patterns range of
362 Bellinzona-Dascio Zone gabbros, except for the lack of a strong positive Eu anomaly. Only the

363 high- Al_2O_3 eclogite shows a weak Eu positive anomaly, likely indicating that some cumulus
364 plagioclase was present in the igneous protolith of these rocks. The whole-rock composition of
365 eclogites also shows high LILE concentrations (i.e. Cs up to $90\times\text{PM}$), negative anomalies in Ba and
366 Nb, positive anomalies in Sr (except for the mafic eclogite) and high U/Th ratio (Fig. 8f).

367 The $\text{Cpx}_M + \text{Kfs}_M$ pocket aggregates around relict quartz and $\text{Pl}_2 + \text{Cpx}_2 + \text{Amp}_2$ symplectite
368 after omphacite in high- Al_2O_3 eclogite D9 (Fig. 6c, S-3) have strongly fractionated REE patterns
369 with enrichment in HREE relative to LREE ($\text{La}_N/\text{Er}_N = 0.11$), and a marked Eu negative anomaly
370 (Fig. 8e). It also shows high LILE concentrations ($\text{Rb} = 2645\times\text{PM}$), negative anomalies in Nb and
371 Ti, and low $\text{U/Th} = 1.5$ (Fig. 8f). The inset of Figure 8e also portrays the REE pattern of allanite
372 surrounding zoisite, which shows strong enrichment in LREE and HREE.

373

374 **6. Mineral chemistry**

375

376 *6.1 Major elements*

377

378 The major elements composition of the analysed rock forming minerals is reported in Table 3 and
379 portrayed in Figures 9 and 10.

380

381 *6.1.1. Peridotites*

382

383 In garnet peridotites, olivine has forsteritic composition with $X_{\text{Mg}} = 0.90$, comparable to that
384 of the bulk rock. Garnet is pyrope-rich ($\text{Py}_{67}\text{Alm}_{18}\text{Gr}_{15}$) with a core-to-rim increase of Al and Fe^{2+}
385 ($\text{Py}_{64}\text{Alm}_{21}\text{Gr}_{15}$) and a complementary decrease of Mg (Tab. 3), likely related to a retrograde
386 equilibration (Tab. 3). Clinopyroxenes have a diopside-rich composition (Tab. 3). In the Na-Al^(VI)
387 diagram (Fig. 9a) clinopyroxene of sample B3A displays a progressive decrease of Na and Al^(VI)
388 from Cpx_1 , to neoblastic Cpx_2 , to symplectitic Cpx_{Sym} . Cpx_1 of sample A2C2 displays a sharp

389 zonation with a core-to-rim decrease of Na and Al^(VI). The composition of the analysed rims
390 approaches that of clinopyroxene in equilibrium with post-peak minerals (Cpx₂), in agreement with
391 a retrograde equilibration of clinopyroxenes at lower pressure. Cpx₂ also shows Al-rich composition
392 with Al up to 0.07 a.p.f.u. (Tab. 3). Old coarse orthopyroxenes (Opx₁) are enstatites with X_{Mg} =
393 0.91. In the Si-Al diagram (dark grey symbols in Fig. 9b) they display an increase of Al from Opx₁
394 in equilibrium with garnet to Opx_{Sym} observed in kelyphites, and a complementary decrease in Si,
395 following the same retrogression equilibration recorded by garnet and clinopyroxene.

396 As shown in Figure 9c, Amp₁ and Amp₂ are pargasitic to edenitic in composition, while
397 symplectitic amphiboles (Amp_{Sym}) vary from tschermakite to Mg-hornblende in sample A2C2, to
398 pargasite in peridotite B3A (Tab. 3). In both Na-Al^(IV) and Al^(VI)-Al^(IV) diagrams (Fig. 9c, d)
399 amphiboles follow the pargasitic substitution, with Amp_{Sym} that shows the highest Al^(IV) and Al^(VI)
400 concentrations.

401 In chlorite peridotites olivine, orthopyroxene and Opx_{Porph} all show X_{Mg} ranging between 0.90
402 and 0.91, and clinopyroxene is diopside-rich (Di₉₀Jd₄Hd₆, Tab. 3). In the Si-Al diagram the two
403 orthopyroxene generations of chlorite peridotite overlap the composition of orthopyroxenes of
404 garnet peridotite B3A (light grey symbols in Fig. 9b). Porphyroblastic amphibole in equilibrium
405 with Opx_{Porph} and Phl_{Porph} is a hornblende (Amp_{Porph}, Tab. 3), whereas Amp₂ in the recrystallised
406 matrix and Amp₃ along the chlorite foliation vary in composition from edenite to tremolitic-
407 hornblende. In the Na-Al^(IV) and Al^(VI)-Al^(IV) diagrams (Fig. 9c, d) also amphiboles of the chlorite
408 peridotite plot along the pargasite exchange vector.

409

410 6.1.2. Tremolitite

411

412 Relict amphibole porphyroblasts in tremolitites are hornblende and their compositional
413 variation is closely related to the optical zoning (Tab. 3). The analysed dusty cores correspond to
414 Mg-hornblende while the inclusion-free rims are tremolites (Fig. 9c) with compositions resembling

415 those of retrogression tremolites, characterised by lower Al, Na and Fe content with respect to
416 porphyroblastic hornblende (Tab. 3, Fig. 9c, d).

417

418 *7.1.3. Eclogites*

419

420 In both mafic and kyanite-bearing eclogites, garnets are zoned and their core-to-rim
421 composition varies from $\text{Py}_{35}\text{Alm}_{40}\text{Gr}_{25}$ to $\text{Py}_{28}\text{Alm}_{52}\text{Gr}_{20}$ at rim (Tab. 3). The composition of
422 omphacite included in garnet corresponds to $\text{Di}_{60}\text{Jd}_{30}\text{Hd}_{10}$, whereas Cpx_2 in symplectites is
423 diopside-rich ($\text{Di}_{80}\text{Jd}_6\text{Hd}_{14}$; Tab. 3). In the Na versus $\text{Al}^{(\text{VI})}$ variation diagram (Fig. 10a)
424 clinopyroxenes show a decrease of Na and $\text{Al}^{(\text{VI})}$ from omphacites to symplectitic diopsides.

425 All amphiboles in eclogites are calcic. Amphiboles after garnet and omphacite (Amp_2) vary
426 from Mg-hornblende to actinolite, whereas coronitic Amp_3 varies in composition from pargasite to
427 Mg-hornblende to tremolite (Tab. 3). In the Na- $\text{Al}^{(\text{IV})}$ and $\text{Al}^{(\text{VI})}$ - $\text{Al}^{(\text{IV})}$ variation diagrams of Figures
428 10b and c, both Amp_2 and Amp_3 plot along the pargasite exchange vector. Amp_3 shows a variable
429 Na- $\text{Al}^{(\text{IV})}$ trend depending on its microstructural site. Coronitic amphiboles around symplectites
430 post-omphacite in fact show the highest Na concentration, comparable to some grains included in
431 garnet representing retrogressed omphacites. In mafic eclogite, porphyroblastic K-feldspar
432 corresponds to almost pure orthoclase $X_{\text{Or}} = 0.98$. Slightly lower $X_{\text{Or}} = 0.90$ characterises Kfs_M
433 occurring within the melt pockets and patchy zone aggregates together with $\text{Cpx}_M + \text{Q}_M + \text{Pl}_M$ in
434 high- Al_2O_3 eclogites (c.f. Tab. 3 and Tumiati et al., 2018).

435

436 *6.2. Trace elements*

437

438 The trace elements compositions of analysed rock forming minerals are listed in Table 4 and
439 portrayed in Figure 11 and 12.

440

441 6.2.1. *Peridotites*

442

443 Garnet from garnet peridotites shows the classic REE pattern with enrichment in HREE (up to
444 10×PM) and depletion in LREE (dark grey area in Fig. 11a). Clinopyroxenes (white area in Fig.
445 11a) and pargasites (Amp_1) of sample B3A (green circles and diamonds in Fig.11a) display REE
446 patterns characterised by relative depletions in HREE ($La_N/Er_N \approx 3$), indicating that they grew in
447 chemical equilibrium with garnet. They also show high LREE and MREE concentrations reflecting
448 the enrichment in LREE of the bulk rock (c.f. Fig. 8a). In sample A2C2 Amp_1 is zoned with REE
449 patterns (yellow circles and diamond in Fig. 11a) enriched in HREE with respect to Amp_1 of sample
450 B3A. This different HREE fractionation is in contrast with the microstructural observation that this
451 Amp_1 is in equilibrium with garnet (Fig. 5a) and likely indicates that an equilibration with post-
452 peak minerals already occurred. As shown in Tab. 4, orthopyroxene is generally depleted in all
453 incompatible elements. In terms of other trace elements, both garnet and clinopyroxene display a
454 general depletion in LILE whereas amphibole of B3A is slightly enriched in LILE with K and Pb \approx
455 10×PM (Fig. 11b). Also concerning the LILE, amphibole of sample A2C2 displays a different
456 pattern with higher LILE concentrations (up to 200×PM) with respect to amphibole of sample B3A.

457 In chlorite peridotites (DB113), porphyroblastic amphibole (Amp_{Porph}) in equilibrium with
458 Opx_{Porph} and Phl_{Porph} (Fig. 5c) has been analysed and its REE pattern is enriched in LREE with
459 respect to HREE (Fig. 11c; Tab. 4). Its LILE concentrations reaches only up to 10×PM with a
460 strong Ba negative anomaly. Opx_{Porph} growing at the expenses of olivine shows LREE contents just
461 above the detection limit, with $La_N/Ce_N = 1.9$ (Fig. 11c, Tab. 4). Chl_3 from the retrogressed chlorite
462 foliation represents the major host of fluid mobile elements displaying high Cs, Rb and K (Cs up to
463 300×PM), a negative anomaly in Ba and positive U/Th ratios (Fig. 11d).

464 In tremolitites, the REE patterns of Mg-hornblende porphyroblasts and tremolite neoblasts
465 closely match that of the bulk-rock (c.f. Fig. 8c) showing higher LREE (maximum La = 5×PM)
466 with respect to the HREE (Fig. 11e). Concerning the LILE, both amphibole generations show

467 concentrations up to 10×PM with a strong Ba negative anomaly and a relatively high U/Th ratio
468 (Fig. 11f).

469

470 6.2.2. Eclogites

471

472 Garnets from all eclogite samples have HREE-enriched and LREE-depleted patterns with REE
473 normalised concentrations up to 20×PM. Garnets from mafic eclogites display variable HREE
474 concentrations with Lu varying from 4 to 14×PM but no important differences are observed
475 between analysed cores and rims (Fig. 12a). By contrast, garnets of kyanite-bearing eclogites show
476 a relative enrichment in MREE and a sharp core-to-rim zonation with progressive depletion in
477 HREE (cores: Lu = 8×PM; rims: Lu = 2-4×PM; Fig. 12b). In terms of other trace elements garnet is
478 depleted in LILE except for Cs and Pb (Tab. 4). Zoisite from mafic eclogite D6 is almost
479 homogeneous in composition and a slight difference in Yb concentration can be observed between
480 core and rim (Fig. 12c). Zoisite has REE absolute concentrations up to 11000×PM with a slight
481 fractionation between LREE and HREE (e.g. $La_N/Yb_N = 5.3 \times PM$). Allanite growing around zoisite
482 (Fig. 6b) shows REE concentrations up to 40000×PM, with strong enrichment in LREE (Fig. 12d).
483 Amp₂ of the lower pressure paragenesis was too small to be analysed. We therefore portray in
484 Figure 12e and 12f only the patterns of coronitic Amp₃. All amphiboles display a relative
485 enrichment in MREE with respect to HREE, resembling that of garnet from kyanite-eclogite (Fig.
486 12e). Amp₃ represents the major host of fluid mobile elements with spikes in Rb and K and high
487 U/Th ratio and is likely the responsible for the LILE enrichment of the bulk rock (Fig. 12f).

488

489 7. Discussion

490

491 Numerous field-based, petrological and geochemical studies have described the chemical
492 interaction between mantle peridotites and felsic continental crust via aqueous fluids, melts and

493 supercritical liquids released at HP and UHP conditions, during prograde or retrograde
494 metamorphism along a subduction pathway (e.g. Rampone and Morten, 2001; van Roermund et al.,
495 2002; Scambelluri et al., 2006; Janák et al., 2006; Malaspina et al., 2006, 2017; Tumiati et al., 2007;
496 Scambelluri et al., 2008; Malaspina et al., 2010; Gudelius et al., 2019). Concerning the Alpine belt,
497 the best examples of suprasubduction garnet peridotites metasomatised by crust-derived agents at
498 HP are represented by occurrences in the Eastern and Central Alps at Pohorje (Janák et al., 2006),
499 Ulten Zone (Rampone and Morten, 2001; Scambelluri et al., 2006; Tumiati et al., 2007; Gudelius et
500 al., 2019), Alpe Arami (Nimis and Trommsdorff, 2001; Paquin and Altherr, 2001) and Monte Duria
501 (Fumasoli, 1974; Evans and Trommsdorff, 1978; Hermann et al., 2006). They all show evidence for
502 metasomatism by the presence of metasomatic phases like amphibole, phlogopite, dolomite and
503 REE-minerals (Tumiati et al., 2005; Lavina et al., 2006; Sapienza et al., 2009; Malaspina and
504 Tumiati, 2012) and by the whole rock enrichment of some incompatible elements such as LREE
505 (Rampone and Morten, 2001; Scambelluri et al., 2006; Tumiati et al., 2007; Gudelius et al., 2019).
506 Similarly to Ulten zone amphibole+garnet peridotites, Monte Duria garnet peridotites and the
507 retrogressed chlorite peridotites show abundant porphyroblasts of pargasitic amphibole, dolomite,
508 phlogopite and porphyroblastic orthopyroxene forming at the expenses of a previous olivine (Fig.
509 5a-d), indicating that they experienced metasomatism and interaction with crust-derived agents,
510 enriched in SiO₂, Al₂O₃, K₂O, CO₂ and H₂O. Similar features in metasomatised suprasubduction
511 peridotites have been described by Malaspina et al. (2006) and Endo et al. (2015) who demonstrated
512 that replacive orthopyroxene derives by the reaction of the peridotite with a slab-derived silicate
513 melt at HP/UHP. In addition, experimental results on olivine solubility in COH fluids (Tiraboschi et
514 al., 2018) and mass balance modelling of solid-solution equilibrium between slab-derived aqueous
515 fluids and garnet peridotite (Campione et al., 2017) indicate that also H₂O-rich fluids released from
516 a subducted eclogite are able to crystallise metasomatic orthopyroxene in the mantle. Whether the
517 metasomatic agent responsible for the modal metasomatism of Monte Duria peridotites is a C-

518 bearing silicate melt or an aqueous COH fluid (Hermann et al., 2006) must be searched in the
519 associated crustal rocks and in their reconstructed P-T path.

520 Further evidence of metasomatism recorded by Monte Duria garnet peridotites is the
521 peculiar fractionation of LREE ($La_N/Nd_N = 2.4$; Fig. 8a) related to the LREE enrichment in Cpx_1
522 and Amp_1 , both crystallised in the garnet stability field, as shown by the relative depletion in HREE
523 (Fig. 11a). The “spoon-shaped” LREE pattern of Monte Duria peridotites strongly resembles that of
524 the Ulten Zone peridotites (pink area in Fig. 8a), which has been interpreted as acquired by the
525 interaction with a hydrous melt (Rampone and Morten, 2001; Scambelluri et al., 2006).
526 Interestingly, Monte Duria peridotites also show a selective enrichment in some LILE (Fig. 8b),
527 which were mainly incorporated by those amphiboles equilibrated in the spinel stability field.
528 Indeed, as shown in Figure 11a and b, the most LILE enriched Amp_1 (yellow symbols) shows
529 HREE enriched patterns indicating non-equilibrium with garnet. This observation suggests that the
530 LREE and LILE enrichment occurred at different pressure conditions (garnet and spinel stability
531 fields) and likely by different metasomatic agents, silicate melt and aqueous fluids, respectively.

532

533 *7.1 The HP partial melting of Borgo eclogites*

534

535 In the Monte Duria area, garnet peridotites occur embedded in low-grade migmatitic gneiss
536 (Mt. Duria) or in direct contact with variably migmatised HP mafic rocks (Borgo). Bulk rock and
537 mineral phase assemblages indicate that these mafic migmatites derive from an eclogite precursor,
538 with a basaltic or gabbroic protolith (Tab. 2 and Fig. 7 and 8e). Peridotites and HP mafic
539 migmatites of Borgo share a common metamorphic evolution, reaching peak conditions at 2.5-3
540 GPa and 750-800 °C and post-peak static equilibration at 0.8-1 GPa and 850 °C (Tumiati et al.,
541 2018). Field occurrence of mafic rocks in Borgo indicates that partial melting started in a
542 deformation regime that is neither related to the shear zone at the contact with migmatitic gneiss
543 (Fig. 2 and Fig. S-1a), nor to the static re-equilibration at HT conditions, recorded by sapphirine-

544 baddeleyite-srilankite and sapphirine-corundum assemblages in coronas around garnet (peridotites)
545 and kyanite (eclogites), respectively (Tumiati et al., 2018). Moreover, the strong layered structure
546 of melanosome and leucosome containing zoisite, omphacite and garnet of amphibole-bearing
547 migmatites (Fig. 4d), along with the occurrence of peculiar deformation structures like lobes and
548 cusps, characteristics of syn-melting deformation (Fig. 4e; McLellan, 1989), strongly suggest that
549 Borgo eclogitic rocks underwent partial melting during prograde-to-peak HP metamorphic
550 conditions. Field structures are supported by microstructural evidence in eclogite boudins enclosed
551 in amphibole-bearing migmatites. Both mafic (D6) and high- Al_2O_3 (D9) eclogites show thin films
552 of clinopyroxene + K-feldspar around both quartz relicts and symplectite aggregates post-
553 omphacite (Fig. 6a,c), together with interstitial pocket aggregates of K-feldspar + plagioclase +
554 quartz + clinopyroxene (Fig. 6d). The reported mineral assemblages closely resemble the ones
555 reported in multiphase melt inclusions trapped by zircons from eclogites of Eastern Papua new
556 Guinea, which underwent partial melting during the beginning of hexumation (DesOrmeau, 2018).
557 Also the microstructural features constraining the direct observation of HP partial melting of
558 eclogites are in agreement with the ones reported in the Central Sulu Orogen by Wang et al.
559 (2014), who recognised melt droplets of leucosome composed of quartz-plagioclase-K-feldspar
560 formed along grain boundaries in the eclogites of the General's Hill. Interestingly, our $\text{Cpx}_M + \text{Kfs}_M$
561 rims (Fig. 6c) and $\text{Kfs}_M + \text{Pl}_M + \text{Qz}_M + \text{Cpx}_M$ pockets (Fig. 6d) are preferentially oriented parallel to
562 the HP foliation of the high- Al_2O_3 eclogites, suggesting that the crystallisation of these
563 microstructures was sin-kinematic to the HP deformation regime that led to the formation of the Ky
564 + Omp + Qz + Zo \pm Kfs assemblage. Moreover, similar evidence of viscous shearing syn-kinematic
565 melt-rock interaction during the infiltration of a felsic melt in mafic rocks has been reported in the
566 Seiland Igneous Province, northern Norway (Degli Alessandrini et al., 2017). In this work a
567 polyphase mixture of clinopyroxene + orthopyroxene + plagioclase + quartz + ilmenite \pm K-
568 feldspar wrap porphyroclasts of orthopyroxene, clinopyroxene and plagioclase along the mylonitic

569 foliation and pools of former melt are preserved as K-feldspar surrounding rounded plagioclase and
570 quartz grains.

571 The partial melting of MORB systems at HP has been experimentally studied by several works
572 (e.g. Lambert and Wyllie, 1972; Schmidt and Poli 1998; Yaxley and Green, 1998; Rapp et al.,
573 1999; Schmidt et al., 2004; Kessel et al., 2005; Klimm et al., (2008); Liu et al., 2009), giving
574 indication on the P-T conditions at which eclogites undergo fluid absent melting in
575 subduction zones. As pointed out by Schmidt and Poli (2014), fluid-saturated melting of K-bearing
576 basaltic rocks can be achieved at 650 °C and 1.5 GPa, reaching 750 °C at 3 GPa and ending at 950
577 °C and 5.5 GPa by the addition of an aqueous fluid from the external system (Lambert and Wyllie,
578 1972; Schmidt and Poli, 1998; Schmidt et al., 2004). At higher pressures, fluid absent melting of K-
579 bearing MORB is mainly controlled by the dehydration of phengite, which starts at 850 °C and 2-
580 2.5 GPa and continues up to 900-950 °C at 3 GPa (Hermann and Green, 2001; Liu et al., 2009). In
581 Figure 13 we portrayed the peak conditions recorded by our samples (green is Mt. Duria garnet
582 peridotite, red is Borgo eclogite) compared with the P-T conditions of wet and dry basalt solidi,
583 along with the phengite dehydration melting for a MORB composition. The stable paragenesis at
584 the peak conditions (750 °C and 3 GPa; Tumiati et al., 2018) is represented by garnet + omphacite +
585 kyanite + zoisite + K-feldspar + quartz/coesite (Tab. 1 and Fig. 6; Fig. 6d of Tumiati et al., 2018).
586 The presence of K-feldspar relict porphyroblasts instead of phengite suggests that the eclogites were
587 almost dry at peak conditions. Indeed, the K-feldspar composition of Borgo eclogites ($X_{Or} = 0.98$)
588 indicates its stability with a garnet + omphacite + kyanite + quartz at average H₂O content of 0.05
589 wt.% at P = 3 GPa and T = 750 °C (Tumiati et al., 2018). These observations, therefore, preclude
590 the possibility that our eclogites underwent phengite dehydration melting, because the temperature
591 equilibration conditions recorded by the peak mineral assemblage are too low.

592 As shown in Figure 13, the P-T peak conditions of Borgo eclogites are slightly beyond the wet
593 basalt solidus, thus suggesting that the partial melting occurred at fluid present conditions. It must
594 be considered that Monte Duria area garnet peridotites record a first hydration stage in a static

595 regime given by chlorite pseudomorphic crystallisation on garnet (Chl_1) predating the development
596 of a penetrative LP-LT chlorite foliation (Chl_3) cutting the garnet layering (Fig. 3c, d, S-2a, d).
597 Moreover, some garnets of Mt. Duria peridotites show Chl_1 pseudomorphs partly overgrown by
598 orthopyroxene + spinel symplectites, in turn surrounded by spinel-amphibole kelyphitic corona
599 (Fig. S-2a, b, c), suggesting that a hydration event occurred before the LP-(U)HT metamorphic
600 event at 850 °C and 0.8-1 GPa (Tumiati et al., 2018). Because at the peak conditions of both
601 peridotites and eclogites chlorite is stable in an ultramafic system (Fumagalli and Poli, 2005; grey
602 curve of Fig. 13), the hydration event likely occurred at HP led to the chloritisation of garnets in the
603 peridotites and flushed the associated eclogites, therefore triggering partial melting.

604 Melting of mafic crust at high pressures produces the so-called adakites, which are
605 characterised by a peculiar geochemical imprint, with high SiO_2 , Al_2O_3 , La/Yb and Sr/Y, coupled
606 with low Y and Yb concentrations (Drummond et al. 1996; Schmidt and Poli, 2014). The
607 comparison between the major element compositions of the melt pockets in eclogite of Borgo (D9
608 in Tab. 2; Fig. 6c,d and S-3) and the composition of archean adakite and low-Al trondhjemite-
609 tonalite-dacite (TTD) of Drummond et al. (1996) show some similarities between the two sample
610 groups. In terms of major element composition, a significant difference is the CaO content, which is
611 higher in our melt pockets, likely due to the occurrence of several clinopyroxene grains, and the
612 $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratio, which is $\gg 1$ for archean adakite and ≈ 0.5 for our leucosome. This difference is
613 due to the fact that omphacite is still stable (in the presence of K-feldspar) in the residual eclogite
614 (Tumiati et al., 2018) that retains Na_2O from the melt, with $\text{Na}^{\text{cpx/melt}}$ partition coefficient close to
615 unity at 3 GPa (Schmidt et al. 2004), supporting that the partial melting of Borgo mafic rocks
616 occurred at high pressure. In terms of trace elements, instead, Borgo leucosome shows very
617 different La/Yb = 0.1, Sr/Y = 8.16, Zr/Sm (8.2) vs La/Sm (0.24) ratios with respect to adakites *l.s.*
618 of Drummond et al. (1996). These differences are mainly due to the strong partitioning of LREE
619 and Sr into the residual allanite and zoisite, respectively, in the resitic portions of eclogites as

620 shown in Figure 8e, f. This partitioning results in a relative enrichment of HREE of our leucosome
621 despite the occurrence of garnet (together with zoisite and allanite) in the residue.

622

623 *7.2 Melt/rock interaction and formation of garnet websterites*

624

625 The contact between the chlorite peridotite body of Borgo and the HP mafic migmatites is
626 marked by the occurrence of a tremolitite layer (Fig. 2, S-1b). Such rocks also occur as variably
627 stretched layers within the peridotite body (Fig. 4a-c) showing sharp contacts with the host
628 peridotite and a marked boudinage parallel to the garnet layering of the peridotite. The now
629 retrogressed garnet-bearing foliation of the peridotite, in fact, wraps the boudins and is deflected
630 into the necks (Fig. 4b, c), indicating that the tremolitite boudinage occurred during a deformation
631 event when Borgo peridotites were in the garnet stability field. These layers display relict Mg-
632 hornblende extensively overgrown by tremolite and contain $\text{Phl}_3 + \text{Chl}_3 + \text{Tc} + \text{Tr}$ pseudomorphs
633 after garnet (Fig. 5e, f). As shown in Fig. 5e, the crystallisation of tremolite and chlorite occurred in
634 a static regime together with the pseudomorphic replacement of garnets. This indicates that the Phl_3
635 + $\text{Chl}_3 + \text{Tc} + \text{Tr}$ mineral assemblage of the tremolitite layers postdates the boudinage deformation
636 event, suggesting that they derive from a precursor that during the formation of the peridotite
637 garnet-foliation was more rigid than the host peridotite. Two hypotheses may explain these
638 structural and microstructural observations. The first one is that the tremolitites of Borgo derive
639 from a Mg-hornblende-bearing precursor as indicated by the presence of relict Mg-hornblende
640 porphyroblasts (Fig. 5e). The alternative hypothesis is that the tremolitites derive from a garnet
641 websterite with a relatively high clinopyroxene modal content. In both cases, these hybrid layers,
642 characterised by high Mg# (>90, Fig. 7a) and high CaO (12.38 wt%, Fig. 7b) formed in the garnet
643 stability field, and subsequently retrogressed in tremolitites during the LP-LT re-equilibration at
644 fluid-present conditions.

645 The peculiar hybrid composition of the tremolitite layers strongly suggests that they formed
646 after the reaction between the eclogite leucosome and the garnet peridotite at HP (Fig. 7). To test
647 this hypothesis, we modelled the mineral assemblage that would be thermodynamically stable when
648 the pristine garnet peridotite (composition of A2C2, Tab. 2) chemically mixes with a felsic adakite-
649 like melt (HP melt pockets in D9 eclogite, Tab. 2). The result is shown in Figure 14a, where the
650 weight fraction (X) of different degrees of chemical mixing between the garnet peridotite (X = 0)
651 and the eclogite leucosome (X = 1) is plotted versus the activity of water ($\log a_{\text{H}_2\text{O}}$) at fixed P-T
652 conditions corresponding to the peak recorded by both peridotite and eclogite. We also assume that
653 most of infiltrating melt is consumed in the reactions. This assumption is supported by field
654 observations that show no evidence of preserved eclogite-derived melt within the peridotite bodies.

655 At X = 0 (no melt infiltration), and $a_{\text{H}_2\text{O}} = 0.02$ the garnet peridotite is represented by the
656 olivine + orthopyroxene + clinopyroxene + garnet + phlogopite mineral assemblage. At higher
657 water activities ($a_{\text{H}_2\text{O}} > 0.4$) pargasitic amphibole is stable, while at the maximum water activity
658 ($a_{\text{H}_2\text{O}} > 0.9$) chlorite is stable together with pargasite and phlogopite. This scenario is supported by
659 the petrographic and microstructural evidence that Chl_1 forms at HP and at water saturated
660 conditions after the prograde-to-peak Amp_1 formation (Tab. 1, Fig. 5a, b, Fig. S-2a). Additional
661 clino- and orthopyroxene are expected to form when the peridotite composition is mixed with
662 increasing weight fraction of leucosome, reaching the maximum modal proportion at X = 0.45 (31
663 Cpx vol% and 40 Opx vol%). In addition, for X = 0.45 and $a_{\text{H}_2\text{O}} > 0.4$ chlorite and olivine are not
664 stable. The presence of high modal amounts of porphyroblastic orthopyroxene, amphibole and
665 phlogopite growing at the expenses of relict olivine and orthopyroxene in Borgo peridotite (Fig. 5c)
666 close to the contact with inner tremolitite layers (location of DB113 in Fig. 2) indicates mixing of
667 more than 35 wt% of leucosome with the garnet peridotite (Fig. 14a). Therefore, from a structural
668 and thermodynamic point of view, the tremolitite layers likely derive from a previous garnet
669 websterite or garnet amphibole-phlogopite-bearing websterite, depending on the water activity of
670 the melt infiltrating the garnet peridotite. It is worth noting that the trace element composition of

671 relict hornblende of the tremolitite DB148 and the metasomatic porphyroblastic amphibole of
672 DB113 peridotite are identical, both in terms of REE (Fig. 11c, e) and LILE (Fig. d, f).

673 A metasomatic origin for clinopyroxenites/websterites with a similar major element
674 compositions has been also proposed for group D-Ronda pyroxenites of the Betic Cordillera,
675 Southern Spain (Garrido and Bodinier, 1999), for the supra-subduction San Jorge and Santa Isabel
676 pyroxenites of Solomon Islands, West Pacific (Berly et al., 2006) and for peridotite-hosted garnet
677 clinopyroxenites from the Granulitgebirge in the Bohemian Massif, central Europe (Borghini et al.,
678 2018). These pyroxenites are characterised by high Mg#, CaO > 10 wt% and Al₂O₃ ≈ 2 wt%
679 similarly to our tremolitite (Tab. 2, Fig. 5).

680 It is worth noting that Borgo peridotite shows several dolomite crystals in equilibrium with
681 amphibole and chlorite (Fig. 5d), indicating that the melt interacting with the peridotite carried
682 additional CaO and CO₂. The water-assisted retrogression of the hybrid garnet ±amphibole-bearing
683 websterite has been modelled in Figure 14b, where the T-X section of the peridotite-leucosome
684 mixing in the representative range of 0 < X < 0.5 has been calculated at H₂O saturated condition at P =
685 0.7 GPa, interpreted as the beginning of retrogression after the HT event (Fig. 13). Even
686 considering the uncertainties of some solid solution models (e.g. amphibole, see Tumiati et al.,
687 2013), this pseudosection indicates that tremolitic amphibole (80 to 90 mol.% of tremolite) forms
688 when metasomatic websterites equilibrate at LP-LT conditions.

689 As summarised in Table 1 and supported by structural, microstructural, chemical and
690 thermodynamic evidences, a HP hydration event formed static chlorite on garnets of the Monte
691 Duria peridotites (Fig. 3c, S-2) and flushed the associated eclogites. At 3 GPa and 750 °C water
692 saturated eclogites underwent partial melting (Fig. 4d, e), producing Ca-rich leucosomes (Fig. 6a, c,
693 d), which reacted with the associated garnet peridotite forming garnet-amphibole-phlogopite
694 websterites (Fig. 4a-c and 5c) and dolomite in the chlorite peridotite (Fig. 5d). The melt-rock
695 interaction occurred at high deformation regimes, yielding to the formation of garnet layering in the
696 peridotite and the consequent boudinage of the websterite, particularly in the layers within the

697 garnet peridotite (Fig. 2 and 4b). The garnet peridotite and associated (partially melted) eclogites
698 subsequently underwent decompression in the spinel stability field (Fig. 5b) forming symplectites
699 after garnet and omphacite (Fig. 6e, f; Tumiati et al., 2018). Water-absent LP-HT re-equilibration at
700 1 GPa and 850 °C crystallised sapphirine-baddeleyite-srilankite coronas around garnet in the
701 peridotites and sapphirine-corundum coronas around kyanite in eclogites (Tumiati et al., 2018).
702 Finally, a LP-LT water-present deformation event formed a chlorite foliation cutting the garnet
703 layering in the peridotites (Fig. 3c, d), yielded to the hydration of the amphibole-phlogopite bearing
704 websterites with the crystallisation of tremolite and $\text{Phl}_3 + \text{Chl} + \text{Tc} + \text{Tr}$ pseudomorphs after garnet
705 (Fig. 5e, f) and the formation of amphibole (Amp_3) coronas around symplectites after omphacite in
706 the eclogites.

707

708 **8. Deformation-induced melt/peridotite reaction at the slab-mantle interface**

709

710 The fate of slab-derived melts in the overlying mantle and their capability to transport
711 crustal components to the mantle wedge is still poorly known. As shown by Spandler and Pirard
712 (2013), crustal melts may migrate into the mantle by porous flow, focussed flow and diapiric flow,
713 depending on their composition, and therefore reactivity, with ultramafic rocks and also by porosity
714 of the mantle. Porous flow produces a strong metasomatism, forming almost monomineralic
715 orthopyroxene and phlogopite layers at the slab-mantle interface, limiting the mass transport of
716 crustal elements except those filtered into residual aqueous fluids (Malaspina et al., 2006; 2009).
717 Focussed flow is represented by the direct transfer of slab melt from the top of the slab to the
718 mantle wedge, forming a network of pyroxenite veins, limiting the melt-peridotite reaction only to
719 minor interaction products (pyroxenes and amphiboles) at the wall rock (e.g. Arai et al., 2003;
720 Kepezhinskas et al., 1995). Finally, diapiric flow may occur after the detachment of sediments +
721 serpentinite rich mélangé due to buoyancy of diapirs which undergo partial melting in the mantle

722 wedge (e.g., Marschall and Schumacher, 2012; Tumiati et al., 2013). It must be specified that all
723 these models consider melting of sediment-like systems of the subducting slab.

724 The direct evidence of eclogite melting during subduction has been reported by Wang et al.
725 (2014) and the interaction of the mantle with mafic melts at high pressure was experimentally
726 studied by Wang et al. (2016) at 1 GPa and 1200 °C and Perchuk et al. (2018) at 2.9 GPa and 750-
727 900 °C. In these studies the reaction with mafic melts produces harzburgite and then
728 orthopyroxenite layers. Interestingly, experiments performed at P-T conditions closed to the peak
729 recorded by eclogites and garnet peridotites of Monte Duria (3 GPa and 750 °C) show reaction
730 zones at the metabasite-peridotite interface with formation of orthopyroxene ± magnesite ± garnet
731 and layers of newly-formed omphacite ± garnet ± phlogopite + orthopyroxene + magnesite by
732 focused flow (Perchuk et al., 2018). Similar mineral associations occur at the contact between
733 Borgo peridotite and inner tremolitite layers (Fig. 5c, d) and in the modelled mineral assemblage of
734 tremolitites before the later hydration (Fig. 14).

735 The interaction between mafic melts and peridotite in a deformation regime (i.e. during the
736 formation of extensional shear zones) has been widely studied in natural case studies, and
737 experimentally and theoretically modelled, in lithospheric and sub-oceanic mantle (e.g. Garrido and
738 Bodinier, 1999; Liang et al., 2010; Gysi et al., 2011; Baltzell et al. 2015). Regardless of the
739 composition of the melts, experiments and numerical modelling evidenced the role of shear rates in
740 the “compaction-decompaction” porosity bands in the peridotite, favouring melt migration and
741 solubility gradients at increasing deformation (Liang et al., 2010; Baltzell et al. 2015). Following a
742 similar mechanism, the development of alternating pyroxenite (l.s.) bands in the peridotite of Borgo
743 (Fig. 2 and 4a-c) may have been favoured by melt-peridotite interaction during shearing. This
744 interpretation is reported in Figure 15, which portrays a conceptual model of the interaction of
745 eclogite-derived melt and overlying peridotite during the shearing induced by slab subduction
746 and/or mantle corner flow. Evidence of syn-deformation melt-rock interaction is given by the
747 occurrence of boudinage of our metasomatic layers, parallel to and wrapped by the garnet foliation

748 of the host peridotite, indicating a shearing induced weakening of the garnet peridotite. Such
749 mechanism has been reported by a recent work of Tommasi et al. (2017), who demonstrated that the
750 occurrence of hydrous melts during shearing strongly changes the deformation processes in mantle
751 peridotites. Following these authors (Tommasi et al., 2017), accomodation of deformation by
752 peridotite results from stress-controlled dissolution and precipitation and advective transport of
753 chemical components by the melts. Moreover, the presence of hydrous melts also favours fast grain
754 boundary migration producing strong rheological weakening of the garnet peridotite. The focussed
755 flow of slab melts into the mantle, instead of porous flow, may therefore occur when the overlying
756 mantle peridotite undergoes shearing and weakening. In conclusion, the application of well studied
757 syn-kinematic melt-rock reaction models in sub-continental and oceanic mantle environment to
758 subduction environments may represent a further step to unravel the mechanism of initiation of
759 mass transport from subducted eclogites to the supra-subduction mantle wedge occurring at the
760 slab-mantle interface.

761

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959

960

961 **Figure captions**

962

963 Fig. 1

964 (a) Tectonic scheme of the Adula-Cima Lunga Nappes modified after Burri et al. (2005). (b)

965 Detailed geological scheme of the Monte Duria area with the locations of Mt. Duria and Borgo
966 outcrops modified from Tumiati et al. (2018). At: Antigorio; Mg: Maggia; Sm: Simano; LL:
967 Leventina-Lucomagno; Tb: Tambò; Su: Suretta; SSB: Southern Steep Belt.

968

969 Fig. 2

970 (a) Borgo outcrop. PDT: retrogressed garnet peridotite; AG: amphibole-bearing migmatites; ME:
971 mafic eclogites; E: kyanite eclogites; HAE: high- Al_2O_3 rim between kyanite eclogites and host

972 amphibole-bearing migmatites. (b) Detailed geological map of the Borgo outcrop (from Tumiati et
973 al., 2018) with samples locations.

974

975 Fig. 3

976 Structures of ultramafic rocks of Mt. Duria and Borgo. (a) Garnet peridotite lens on the southern
977 ridge of Monte Duria. (b) Detail of garnet layering transposed by a secondary chlorite foliation in
978 garnet peridotite from Monte Duria. (c) Chlorite-bearing pseudomorphs replacing garnets in
979 chlorite peridotite of Borgo. (d) Retrogressed garnet layering transposed by a younger chlorite
980 foliation in chlorite peridotite of Borgo.

981

982 Fig. 4

983 Lithologies and structures of ultramafic and mafic rocks of Borgo. (a) Tremolitite boudin within the
984 chlorite peridotite, with samples location. (b) Detail of tremolitite boudin with garnet layering
985 wrapping around the boudins and flowing into the boudins neck. (c) Garnet layering wrapping the
986 tremolitite boudin and $\text{Phl}_3 + \text{Chl} + \text{Tr} + \text{Tc}$ pseudomorphs replacing garnet in tremolitites. (d)
987 leucosomes and melanosomes in amphibole-bearing migmatites; e) lobes and cusps structures in
988 amphibole-bearing migmatites.

989

990 Fig. 5

991 Photomicrographs and back-scattered electron (BSE) images of peridotites and tremolitites. (a)
992 Cross polarised light image of Mt. Duria garnet peridotite B3A showing garnet porphyroclasts and
993 coarse orthopyroxene (Opx_1) and clinopyroxene (Cpx_1) and amphibole (Amp_1) in a matrix of finer
994 grained matrix; (b) Cross polarised light image of neoblastic orthopyroxene (Opx_2), clinopyroxene
995 (Cpx_2) and spinel (Sp_2) in sample B3A. (c) Plane polarised light image of porphyroblastic
996 orthopyroxene ($\text{Opx}_{\text{Porph}}$), phlogopite ($\text{Phl}_{\text{Porph}}$) and amphibole ($\text{Amp}_{\text{Porph}}$) statically growing at the
997 expenses of relict olivine and orthopyroxene in chlorite peridotite DB113 from Borgo. (d) BSE

998 image of dolomite in textural equilibrium with olivine and chlorite (Chl_1) in sample DB113. (d)
999 BSE image of relict Mg-hornblende overgrown by tremolite and minor chlorite (Chl_3) in tremolitite
1000 DB148 from Borgo. (f) Plane polarised light image of phlogopite (Phl_3) + Chl_3 + Tr + Tlc after
1001 garnet in tremolitite DB179 from Borgo.

1002

1003 Fig. 6

1004 BSE images of eclogites from Borgo. (a) Quartz porphyroblasts surrounded by Cpx_M + Kfs_M
1005 coronas and K-feldspar porphyroblasts surrounded by symplectitic clinopyroxene (Cpx_2) in mafic
1006 eclogite D6. (b) Zoisite porphyroblast surrounded by tiny, bright allanite crystals in mafic eclogite
1007 D6. (c) Melt pocket of recrystallised Cpx_M + Kfs_M around relict quartz and $\text{Pl}_2 + \text{Cpx}_2 + \text{Amp}_2$
1008 symplectite after omphacite in high- Al_2O_3 eclogite D9. (d) Melt pocket of recrystallised Kfs_M + Pl_M
1009 + Qz_M + Cpx_M around relict quartz in high- Al_2O_3 eclogite D9. (e) Pl_2 + Amp_2 symplectite replacing
1010 garnet in Na-rich Pl_2 + Amp_2 + Cpx_2 symplectite after omphacite in amphibole-bearing migmatite
1011 D5. (f) Pl_2 + Cpx_2 + Amp_2 symplectite after omphacite in amphibole-bearing migmatite D5.

1012

1013 Fig. 7

1014 Mg# versus Ni (ppm) plot (a) and CaO versus Al_2O_3 wt.% concentrations (b) of peridotites from
1015 Mt. Duria and tremolitite and eclogites from Borgo (data from Table 2) compared with Depleted
1016 Mantle (Salters and Stracke, 2004), average N-MORB (Gale et al., 2013), average composition of
1017 Mean Upper Crust, Mean Shallow Gabbro and Mean Lower Gabbro (Gillis et al., 2014), peridotites
1018 from Ulten Zone (UZ; Tumiati et al., 2003), Cima di Gagnone (CdG; Scambelluri et al., 2014),
1019 Alpe Arami (AA; Ernst, 1978) and peridotite and Mg-gabbro from the Bellinzona-Dascio Zone
1020 (BDZ; Stucki et al., 2003). The dashed black line separates mafic compositions on the left from the
1021 ultramafic compositions on the right (modified after Malaspina et al., 2006).

1022

1023 Fig. 8

1024 Primitive Mantle normalised REE and other trace element patterns of the investigated samples (data
1025 from Table 2). Elements are presented in order of increasing compatibility (left to right) during
1026 melting in the upper mantle (Gale et al., 2013). Normalising values are from McDonough and Sun
1027 (1995). (a) and (b) trace elements pattern of A2C2 and B3A garnet peridotites from Monte Duria,
1028 compared with the reference Depleted Mantle (blue solid line DM; Salters and Stracke, 2004),
1029 garnet peridotites from Cima di Gagnone (grey area CdG; Scambelluri et al., 2014), garnet
1030 peridotites from Ulten Zone (pink area UZ; Scambelluri et al., 2006 and Tumiati et al., 2007). (c)
1031 and (d) trace elements pattern of tremolite DB151. (e) and (f) trace elements pattern of D6, D9,
1032 B5A eclogites and interstitial $Cpx_M + Kfs_M$ pocket aggregate in high- Al_2O_3 eclogite D9 compared
1033 with average N-MORB pattern (solid grey line; Gale et al., 2013), Mg-gabbro from Bellinzona-
1034 Dascio Zone (light grey area BDZ; Stucki et al., 2003) and the trace element pattern of allanite from
1035 eclogite D6.

1036

1037 Fig. 9

1038 (a) Compositional variation of clinopyroxenes (octahedral Al versus Na) in garnet peridotites from
1039 Mt. Duria and chlorite peridotites from Borgo. (b) Al versus Si content of porphyroclastic,
1040 neoblastic and symplectitic orthopyroxenes of garnet peridotites from Mt. Duria and
1041 porphyroblastic metasomatic orthopyroxene in chlorite peridotite from Borgo. (c) and (d)
1042 Compositional variations of amphiboles from garnet and chlorite peridotites. Tetrahedral aluminium
1043 (Al^{IV}) is plotted with respect to total Na and octahedral aluminium (Al^{VI}); Ts = tschermakite, Ed =
1044 edenite, Pg = pargasite, Tr = tremolite.

1045

1046 Fig. 10

1047 Octahedral Al versus Na compositional variation of clinopyroxenes (a) and tetrahedral aluminium
1048 versus total Na and octahedral aluminium compositional variation of amphiboles in eclogites
1049 boudins from Borgo. Abbreviations same as in Figure 9.

1050

1051 Fig. 11

1052 Primitive mantle normalised REE and other trace elements patterns of (a) and (b) porphyroclastic
1053 garnets, clinopyroxenes and amphiboles of garnet peridotites from Mt. Duria; (c) and (d)
1054 porphyroblastic metasomatic amphibole of chlorite peridotites from Borgo; (e) and (f) relict
1055 hornblendes and tremolites from tremolite boudins at the contact between chlorite peridotite and
1056 amphibole-bearing migmatites at Borgo.

1057

1058 Fig. 12

1059 Primitive mantle normalised REE and other trace elements patterns of garnets (a) and (b), zoisite
1060 (c), allanite (d) and amphiboles (e) and (f) from all eclogite types at Borgo. The different colour
1061 labels of allanite in (d) indicate different point analyses of the same samples because of the
1062 scattered normalised concentrations of MREE and HREE.

1063

1064 Fig. 13

1065 Peak P-T conditions of mafic (red square) and ultramafic (green square) rocks in the Monte Duria
1066 area (from Tumiati et al., 2018). In the P-T space are reported: i) basalt wet solidus (Schmidt and
1067 Poli, 1998); ii) zoisite and amphibole-out curves (Schmidt and Poli, 1998); iii) phengite dehydration
1068 melting curves in basalt ((1) Hermann and Green, 2001; (2) Liu et al., 2009); iv) basalt dry solidus
1069 (Lambert and Wyllie, 1972); v) chlorite-out curve in the peridotite system (Fumagalli and Poli,
1070 2005); vi) top slab geotherms of warm subduction zones (orange field; Syracuse et al., 2010) and of
1071 cold subduction zones (grey field, Arcay et al., 2007). Note that the peak conditions of the coupled
1072 eclogites and garnet peridotites overlap the “warm” subduction P-T path suggesting that Borgo
1073 eclogites can be considered as a proxy for partial flush melting of a mafic crust in a thermal regime
1074 characterised by a low angle subduction zone.

1075

1076 Fig. 14

1077 a) $\log a_{\text{H}_2\text{O}}-X$ diagram calculated at fixed $P = 3 \text{ GPa}$ and $T = 750^\circ\text{C}$ for garnet websterite forming
1078 after the melt-peridotite reaction. $X = 0$ corresponds to garnet peridotite (A2C2, Tab. 2), while $X = 1$
1079 corresponds to the bulk of leucosome pockets composition (D9, Fig. S-3, Tab. 2). Dashed lines and
1080 corresponding black numbers in the lower part of the section are the isomodes of clinopyroxene
1081 (vol.%) increasing from peridotite to leucosome composition, whereas orthopyroxene increases
1082 from 19 to 40 vol.% in the range $0 < X < 0.45$ and decreases from 37 to 2 vol.% in the range $0.5 < X < 1$.
1083 b) T-X diagram calculated at H_2O saturated condition of the websterites forming after the reaction
1084 between peridotite and eclogite leucosome in the range of interest ($0 < X < 0.45$) at 0.7 GPa,
1085 representing the beginning of retrogression. Dashed grey lines and corresponding black numbers are
1086 the isopleths of the tremolite molar content in the amphibole solid solution (mol.%).

1087

1088 Fig.15

1089 Conceptual model for garnet websterite formation at the slab-mantle interface in a deformation
1090 regime during subduction (not to scale). See text for explanation. Isotherms in the inset are relative
1091 to warm subduction environments (Peacock and Wang, 1999).

1092

1093 Tab. 1

1094 Mineral assemblages stable at peak, decompression, low pressure and high-temperature and the
1095 retrogression stages, reconstructed from microstructures of Monte Duria peridotites and eclogites.

1096

1097 Tab. 2

1098 Major (oxide wt.%) and trace (ppm) element compositions of whole-rocks of peridotites, eclogites,
1099 metasomatic layer and migmatite leucosome from Monte Duria area.

1100

1101 Tab. 3

1102 Representative WDS microprobe analyses (oxide wt.%) and reconstructed formulae of garnet (Grt),
1103 clinopyroxene (Cpx, Omp), orthopyroxene (Opx), olivine (Ol), amphiboles (Amp, Hbl, Tr),
1104 plagioclase (Pl), K-feldspar (Kfs) and biotite (Bt) in selected samples of peridotites and eclogites
1105 from Monte Duria area.

1106

1107 Tab. 4

1108 Trace element composition (ppm) of representative minerals of selected samples of peridotites and
1109 eclogites from Monte Duria area.