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Melt inclusions at MT. Edixon (Antarctica): Chemistry, petrology and implications for the evolution of the Lanterman range

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1 **MELT INCLUSIONS AT MT. EDIXON (ANTARCTICA):**
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4 **CHEMISTRY, PETROLOGY AND IMPLICATIONS FOR THE**
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7 **EVOLUTION OF THE LANTERMAN RANGE**
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13 **Fabio Ferri**¹, Bernardo Cesare^{1*}, Omar Bartoli¹, Silvio Ferrero^{2,3}, Rosaria Palmeri⁴,
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16 Laurent Remusat⁵ and Stefano Poli⁶
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19
20 ¹ Department of Geosciences, University of Padua, Italy

21 ² Institute of Earth and Environmental Science, University of Potsdam, Germany

22 ³ Museum für Naturkunde (MfN), Leibniz-Institut für Evolutions und
23 Biodiversitätsforschung, 10115 Berlin, Germany

24
25 ⁴ Museo Nazionale dell'Antartide, Sezione Scienze della Terra, Siena, Italy

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27 ⁵ Institut de Mineralogie, Physique des Matériaux et Cosmochimie Museum National
28 d'Histoire Naturelle, Paris, France

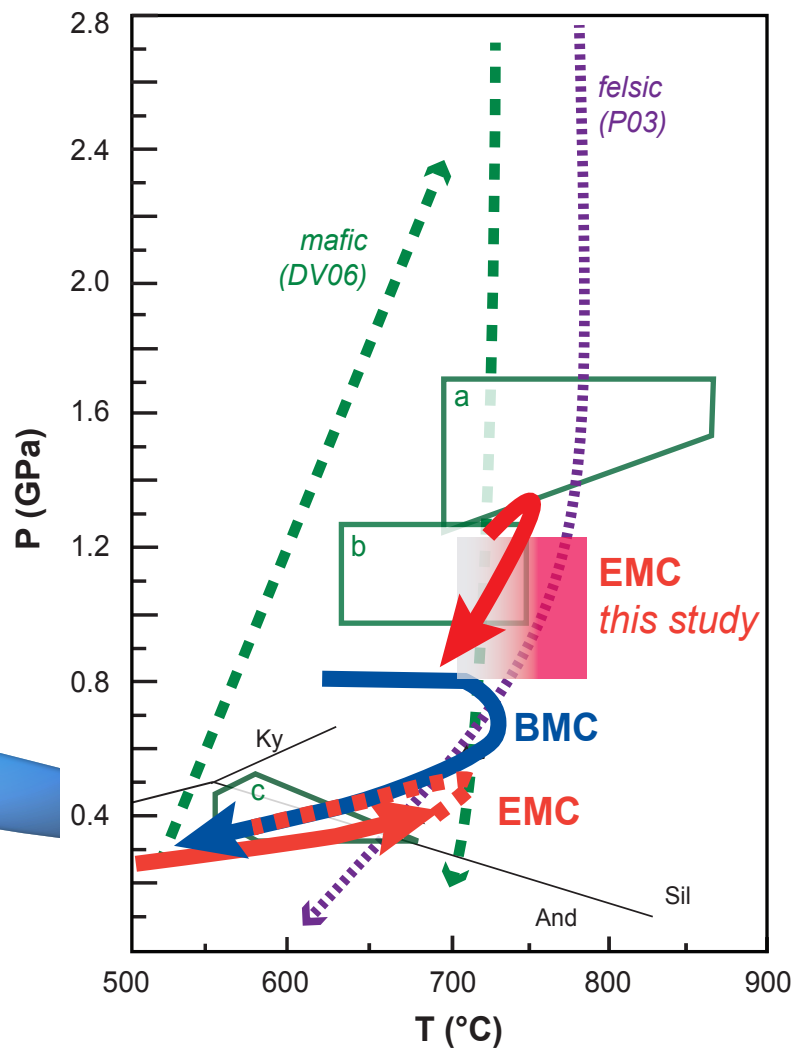
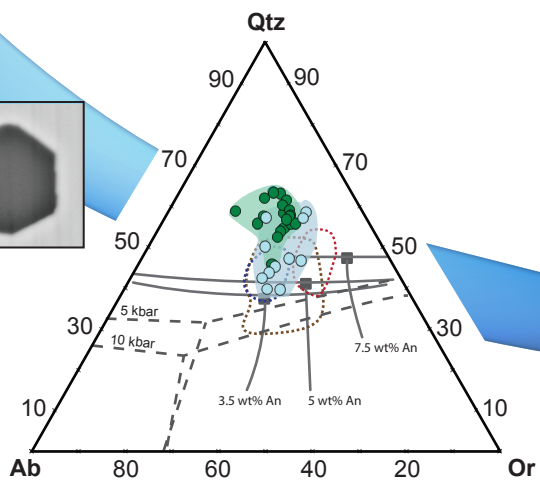
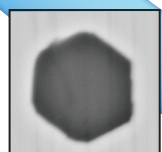
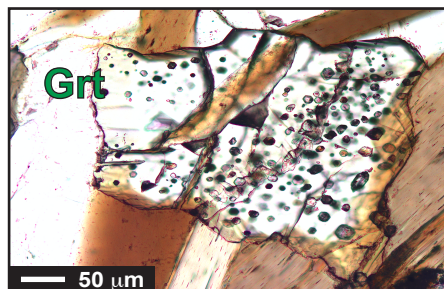
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31 ⁶ Department of Earth Sciences, University of Milan, Italy
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48 * Corresponding author. Tel: +39-049-8279148; Fax: +39-049-8272010.
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50 *E-mail address:* bernardo.cesare@unipd.it
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*Highlights (for review)

- Melt inclusions first discovered in peritectic garnets in paragneiss from the Edixon Metamorphic Complex, Lanterman Range, Antarctica
- Primary glasses from remelting of inclusions are sub-alkaline and peraluminous with 4-5 wt. % H₂O and CO₂ < 0.1 wt. %
- Anatexis at the Edixon Metamorphic Complex occurred at different crustal levels



24 ABSTRACT

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4 25 The recovery of melt inclusions in peritectic phases of metamorphic rocks brings the
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6 26 opportunity to retrieve the chemical composition of primary melts and infer melt entrapment
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9 27 conditions. Here we report the first recovery of such inclusions in a medium-grade muscovite-
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11 28 biotite-plagioclase-K-feldspar-quartz gneiss from the Edixon Metamorphic Complex (EMC)
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14 29 of the Lanterman Range in Northern Victoria Land, Antarctica.

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16
17 30 The most common inclusions are polycrystalline (PI) and fluid (FI) types coexisting in the
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20 31 same clusters and displaying frequent negative crystal shape and textures indicative of a
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22 32 primary origin. The PI contain muscovite + quartz + H₂O ± biotite ± calcite ± siderite ± albite
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25 33 while the FI are composed by CO₂ + H₂O + N₂ + CH₄ ± H₂S. The PI were remelted in a
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27 34 piston cylinder apparatus at conditions of 740-900°C at 0.8-1.0 GPa with complete remelting
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30 35 taking place at 760-780°C. The remelted glass is sub-alkaline, peraluminous (ASI ≈ 1.3-1.5)
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32 36 with CaO ≈ 1.2 - 2.1 wt% and K₂O ≈ 3.4 - 4.1 wt%, and contains ≈ 4.3-5.3 wt% of H₂O and
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34 37 400-1150 ppm of CO₂.

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38 38 The stable mineral assemblage is represented by muscovite + plagioclase + quartz + K-
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40 39 feldspar + garnet + melt ± rutile coexisting with excess C-O-H fluid while biotite is retrograde
41
42
43 40 (Ti 1.2 a.p.f.u.). The calculated phase equilibria and compositional isopleths in the
44
45 41 MnCNKFMASHT model system delimit the equilibrium conditions to 1.0-1.3 GPa and 680-
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47 42 730°C.

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50 43 Our study confirms that the EMC at the Lanterman Range is not a homogeneous
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52 44 metamorphic complex and that anatexis could have taken place at different crustal levels
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55 45 during the evolution of the Ross Orogen. Our results indeed suggest that, in the Lanterman
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57 46 Range, the high-pressure eclogitic rocks, the medium pressure amphibolite stage and the low-
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60 47 pressure amphibolite stage shared most of the late exhumation path.

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Key words: melt inclusions, crustal anatexis, nanogranitoids, Lanterman Range,

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6 51 Edixon Metamorphic Complex

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1. INTRODUCTION

The discovery of primary melt inclusions (MI) hosted in peritectic minerals from high grade metamorphic rocks has been described in recent papers (reviewed by Cesare et al., 2015 and Bartoli et al., 2016) and new examples are continuously adding from typical felsic crustal compositions or new mafic to ultramafic ones (Ferrero et al. 2018). The MI represent a window into the anatexis history of partially melted terranes (Acosta-Vigil et al., 2010) and may provide key microstructural and compositional information on crustal anatexis, like mechanisms and nature of the melting process, conditions of entrapment and melting, timeframes of melt production and segregation (Cesare et al., 2015). In the absence of any petrographical evidence of anatexis as pseudomorphs after melt films (Holness et al., 2011), the MI become a unique tool to infer past melting events throughout the metamorphic history of a rock. When MI coexist with fluid inclusions (FI), they are able to shed light on the fluid/melt/rock interactions during anatexis, and recent works demonstrated that melting processes are more commonly fluid-present than traditionally expected (Carvalho et al., 2018). The combination of MI and FI studies can also help understanding the exchange, transport and recycling of volatile species (especially CO₂) in crustal evolution or during subduction.

In Antarctica, the presence of felsic-nanogranite inclusions produced after cooling of MI have been documented in garnets from the Lutzow-Holm Complex (East Antarctica, Hiroi et al., 2019). Here we report the first recovery of preserved MI in metapelitic garnets from the Edixon Metamorphic Complex from the Lanterman range in Antarctica. These inclusions turn out to be an essential tool to identify anatexis in rocks lacking further evidence of partial melting and, together with the petrological study of their host rocks, provide new clues to the geodynamic evolution of this crystalline basement.

76 2. SAMPLE DESCRIPTION

77 2.1 Geological setting

78 The Ross orogenic belt is one of the Earth's great Phanerozoic mountain belts
79 resulting from convergence along the paleo-pacific margin of Antarctica in Gondwana land
80 (Cawood, 2005; Talarico et al., 2004). The Northern Victoria Land (Fig. 1A, B) represents
81 one of the major segments of the Ross orogenic belt (e.g., Federico et al., 2006, see summary
82 by Stump, 1995) and comprises the Lanterman-Mariner suture (Ricci and Tessensohn, 2003,
83 and references therein), a tectonic structure where peculiar HP-UHP metamorphic rocks have
84 been reported (Ricci et al., 1996; Palmeri et al., 2011).

85 From east to west, the Northern Victoria Land can be divided into three contrasting
86 tectonometamorphic terranes with NNW-SSE structural trends separated by major faults
87 (Bradshaw and Laird, 1983): 1) the Robertson Bay Terrane, 2) the Bowers Terrane and 3) the
88 Wilson Terrane (Fig. 1B). The Wilson Terrane represents the palaeo-Pacific margin of
89 Gondwana in the region and consists of intermediate- to high-P metamorphic rocks with local
90 eclogites close to the Lanterman-Mariner suture, (Grew et al., 1984; Di Vincenzo et al., 1997;
91 Ricci et al., 1997), and of low-P metamorphic rocks (Palmeri et al., 1991) with local remnants
92 of a polymetamorphic granulite complex (Talarico et al., 1995). The Wilson Terrane was
93 extensively intruded by the Granite Harbour Intrusive Complex, an orogenic association of
94 Cambro–Ordovician granitoids showing arc affinity (Bomparola et al., 2006; Kleinschmidt
95 and Tessensohn, 1987; Borg et al., 1990; Ghezzi et al., 1989).

96 The Lanterman-Mariner suture separates the Wilson Terrane and the Bowers Terrane
97 and is best exposed in two areas, namely the Lanterman Range to the north and the
98 Mountaineer Range to the south. From west to east, three different lithotectonic units have
99 been identified in the Lanterman Range area of the Lanterman-Mariner suture (Talarico et al.,
100 1998): i) the Edixon Metamorphic Complex, ii) the Bernstein Metamorphic Complex and iii)

101 the Gateway Hills Metamorphic Complex (Fig. 1C). The Edixon Metamorphic Complex
102 mainly consists of micaschists and gneisses containing bands and lenses of calc-silicate rocks,
103 with minor amphibolites in the southernmost outcrops of the range. This complex underwent
104 medium- to high-grade metamorphism and partial melting under low-P conditions and shows
105 close similarities to the internal sequences of the Wilson Terrane, with the presence of
106 migmatites and abundant granitoids of the Granite Harbour Intrusive Complex. The Bernstein
107 Metamorphic Complex consists of micaschists and gneisses affected by medium- to high-
108 grade metamorphism under intermediate-P conditions. In contrast to the Edixon Metamorphic
109 Complex, calc-silicate rocks, migmatites and granitoids are rare (Ghiribelli, 2000; Talarico et
110 al., 1998). The Gateway Hills Metamorphic Complex is characterized by abundant mafic and
111 ultramafic rocks, including well-preserved eclogite-facies rocks, with minor gneisses and
112 quartzites (Di Vincenzo et al., 1997; Ghiribelli et al., 2002; Palmeri et al., 2003, 2007, 2009,
113 2011). There are no granitoids in this complex (Talarico et al., 1998). Recently Di Vincenzo
114 et al. (2016) established that two eclogite types coexist along the same ridge at the Lanterman
115 Range: i) medium-grained, “colder” and undeformed eclogites, that recorded both prograde
116 and peak conditions at nearly 530 Ma (U-Pb zircon age) during ongoing regional calc-alkaline
117 magmatism, and ii) finer-grained, “hotter” and slightly deformed eclogites, that recorded only
118 metamorphic peak conditions at nearly 500 Ma and hints of the subsequent retrogression (\approx
119 480 Ma), coincident with the outboard accretion of an island arc (Bowers Terrane) onto the
120 Gondwanan margin.

121 The evolution of the Lanterman Range (Capponi et al., 1999, 2002; Crispini et al.,
122 2007; Di Vincenzo et al., 2014, 2016; Federico et al., 2010) can be summarized in a west-
123 over-east thrusting at ca. 500 Ma, followed by sinistral strike-slip shearing at 480–460 Ma
124 ago, both under amphibolite to greenschist-facies metamorphism. Large-amplitude folding of

125 the late Ross or Lachlan orogens in the late Ordovician and Cenozoic brittle tectonics
126 terminated the evolution of the suture.

127 **2.2 Petrography and bulk composition**

128 The studied rock belongs to the Edixon Metamorphic Complex of the Lanterman
129 Range, was sampled at SE of Mt. Edixon (71° 50'23.597''S, 163°39'41.35''E) and stored in
130 the PNRA (Programma Nazionale Ricerche in Antartide) rock repository at the Museo
131 Nazionale dell'Antartide (Siena, Italy) as specimen 24-12-94GO5 (from hereafter GO5). The
132 rock is a fine-grained psammitic gneiss composed of biotite, muscovite, garnet, quartz,
133 plagioclase, K-feldspar, minor ilmenite, apatite and tourmaline, and free of Al_2SiO_5
134 polymorphs (Fig. 2). Muscovite and biotite define the main foliation and show mm-sized
135 crystals with no evidence of alteration. Quartz is abundant and occurs in coarse crystals or
136 tiny aggregates of subgrains with segmented borders showing evidence of bulging. Garnet is
137 scarce, with average diameter $< 400 \mu m$ and exceptionally up to ≈ 1 mm. The sample displays
138 evidence for extensive down-temperature replacement of garnet by biotite (Figure 2B, C).
139 Overall, the mineralogical composition and texture of the studied rock would suggest a
140 medium grade metapsammite. Typical microstructural evidence of melt presence such as
141 grains with euhedral crystal faces or pseudomorphs after melt films (e.g., Holness et al., 2011)
142 are not present, and only the occurrence of melt inclusions in garnet (Fig. 2B, C) attests for the
143 former anatectic history of the rock.

144 **3. METHODS**

145 **3.1 Analytical techniques**

146 After MI recognition by optical microscopy, the mineral phases and the inclusions
147 were first characterized in detail on polished thin sections by back-scattered electron images
148 (BSEs) obtained with FESEM Sigma Zeiss at the ICMATE - CNR of Padova and with SEM
149 CamScan MX3000 at the Department of Geoscience of the University of Padova.

150 The mineral phases and the glass of the inclusions were investigated by Micro-Raman
1 spectroscopy using the HORIBA Jobin Yvon LabRAM HR 800 located at the Institute of
2 151 Earth and Environmental Science, University of Potsdam. The machine is equipped with a
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4 152 Peltier-cooled multichannel CCD detector and coupled with an Olympus BX41 petrographic
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6 153 microscope. An air-cooled NdYAG laser was used for excitation ($\lambda = 532$ nm, laser power on
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8 154 the sample of 2–3 mW) with a grating of 300 lines/mm, slit width and confocal hole set to
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10 155 100 and 200 μm , respectively. The Raman spectra of the solid and fluid phases were acquired
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12 156 with a 100 \times objective in the wavenumber range 100–4000 cm^{-1} , integrating 3 repetitions of
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14 157 60 s with spectral resolution of 10 cm^{-1} .
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21 159 The composition of mineral phases and glasses was determined using a Jeol JXA 8200
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23 160 Superprobe at the Dipartimento di Scienze della Terra, Università di Milano (Italy) using
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25 161 natural and synthetic silicates and oxides as standards. Analytical parameters for minerals
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27 162 were 15 kV accelerating voltage, 5 nA current, counting time of 30 s on peak and 10 s on
28
29 163 background. For glass analyses, the beam diameter was reduced to 1 μm and the counting
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31 164 time to 10 s on peak and 2 s on background. Sodium, K, Al, and Si were the first analyzed
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33 165 elements to limit element loss. Concentrations were corrected against leucogranitic glass
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35 166 standards with comparable H_2O contents (Morgan and London, 2005). Details concerning the
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37 167 application of correction factors and the composition of the standard glasses are given by
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39 168 Ferrero et al. (2012) and Bartoli et al. (2013a, 2013b).
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46 169 The analysis of H and C in glasses was performed using the Cameca Nano Secondary
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48 170 Ion Mass Spectrometry 50 (NanoSIMS) installed at Muséum National d'Histoire Naturelle in
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50 171 Paris (France). Polished experimental capsules with MI exposed on the garnet surface and
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52 172 standard glasses were mounted in Indium (Aubaud et al., 2007). MI were identified by
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54 173 collecting secondary ion images of Si, K and Fe. For every analysis, we first performed a pre-
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56 174 sputtering step on a $3 \times 3 \mu\text{m}^2$ surface area for 2 min with a 400 pA primary Cs^+ beam to
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175 remove gold coating, surface contamination and reach a steady-state sputtering regime. Then
176 a primary beam of 37 pA was used for data acquisition. Data were acquired by rastering a $3 \times$
177 $3 \mu\text{m}^2$ surface area and collecting only ions from the inner $1 \times 1 \mu\text{m}^2$ (beam blanking mode)
178 to reduce surface contamination. Each analysis stacked about 200 cycles, each cycle being
179 1.024 s long. The $^6\text{OH}^-$ (used as a proxy for H_2O), $^{28}\text{Si}^-$, $^{39}\text{K}^{16}\text{O}^-$ and $^{56}\text{Fe}^{16}\text{O}^-$ were recorded
180 simultaneously in multicollection mode. We checked that $^{16}\text{OH}^-/^{28}\text{Si}^-$ ratio was stable during
181 MI analyses (see Supplementary material). Secondary ions were collected by electron
182 multipliers with a dead time of 44 ns. Mass resolution was set to 10000. One inclusion was
183 large enough for replicated analyses. For NanoSIMS calibration we used a 5.5 wt% H_2O -
184 bearing leucogranitic glass from Acosta-Vigil et al. (2003), a 4.3 wt% H_2O - bearing
185 leucogranitic glass from Behrens and Jantos (2001) and an anhydrous leucogranitic glass
186 from Morgan and London (2005). The standards used for the determination of CO_2 contents
187 were four trachyandesitic standards (STR 9, 10, 11 and 13) from the Stromboli volcano
188 experimentally doped in carbon and water by Bureau et al. (2003). H and C concentrations
189 were recalculated as H_2O and CO_2 , respectively. Data corrections, using the aforementioned
190 calibration, and error calculations were performed using the R program following the
191 procedure described in Bartoli et al. (2014) and Thomen et al. (2014). During the session, the
192 vacuum in the analysis chamber remained between 2.5 and 5×10^{-10} Torr.

193 **3.2 Mineral chemistry**

194 The mean compositions of minerals are reported in Table 1 (Supplementary Material).
195 Garnet is almandine-rich (≈ 57 mol. % Alm) with slight growth zoning. From core to rim, the
196 pyrope content decreases from 12 mol. % to 8-10 mol. % while grossular and spessartine
197 components increase from 2 mol. % to 4 mol. %, and from 25 mol. % to 28-30 mol. %
198 respectively. The almandine content is constant throughout the crystal.

199 The white mica is potassic ($X_{\text{Ms}} = 0.92\text{-}0.95$) with Si ranging from 3.00 to 3.04 a.p.f.u.
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2 200 (11 oxygens), Al ranging from 2.76 to 2.72 a.p.f.u. and X_{Mg} of 0.41-0.45. The TiO_2 content
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4 201 (0.03 a.p.f.u.) is typical of amphibolite-facies subsolidus muscovite (e.g., Cesare 1999,
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6
7 202 compare with Redler et al., 2011). The composition of the white mica is similar to that
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10 203 reported by Di Vincenzo et al. (2001) in sample BT5 and BT2 collected from the Edixon and
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12 204 Bernstein Metamorphic Complexes respectively and classified by the Authors as type-2
13
14 205 muscovite with selected area diffraction (SAED) pattern of 2M1 polytypes.

16 206 Biotite shows no systematic variations with microstructural occurrence, and has a
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19 207 fairly constant composition, with Ti of 0.11-0.13 a.p.f.u. and X_{Mg} of 0.44-0.45. Also for
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22 208 biotite the titanium content is typical of amphibolite-facies subsolidus conditions (Tajčmanová
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24 209 et al., 2009. Plagioclase composition has 15-23 mol. % anorthite, 76-84 mol.% albite and ca. 1
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26 210 mol.% K-feldspar. Alkali feldspar is almost pure K-feldspar with albite <1-3 mol.%.

29 211 **3.3 Remelting of inclusions**

31 212 The following nomenclature will be used: ‘remelted’ inclusions are those in which the
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34 213 crystalline assemblage was completely melted to a single glassy phase, but melt + vapor
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36 214 homogenization was not achieved; ‘rehomogenized’ are those inclusions that appear as a
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39 215 homogeneous glass without bubbles (although former accidentally-trapped minerals may be
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41 216 present). Experimental remelting of the melt inclusions was performed on hand-picked
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44 217 separates of garnets obtained from crushed rock pieces sieved at < 500 μm size. The garnet
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46 218 fragments were loaded into Au capsules filled with dry powder of silica or chemically pure
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49 219 CaCO_3 . The capsules were loaded in a salt + MgO + graphite assembly and experiments were
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51 220 performed in single stage piston cylinder apparatus at the Department of Geosciences of the
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54 221 University of Milan at condition from 740°C to 900°C at 0.8 and 1.0 GPa, with duration of
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56 222 21-68 hours (Table 2). The temperature was controlled by a K-type thermocouple with
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59 223 uncertainty of $\pm 5^\circ\text{C}$. To avoid decrepitation of inclusions, quenching was performed at
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224 experimental pressure and load was released after achieving ambient temperature. After the
225 experiments, the capsules containing garnet + SiO₂ were mounted in epoxy and polished to
226 expose rehomogenized inclusions, whereas the capsules containing garnet + CaCO₃ were first
227 opened and immersed in diluted HCl to dissolve calcite and recover the garnet grains. The
228 garnet grains were mounted in epoxy and polished (more experimental details in Bartoli et al.,
2013a and Ferrero et al., 2015).

4. RESULTS

4.1 Microstructures of melt and fluid inclusions

The inclusions preserved in the starting material and after experiments were characterized on the basis of optical microscopy BSE imaging, EDS elemental mapping and micro-Raman spectroscopy, and distinguished in fluid (FI), polycrystalline (PI) also called nanogranitoids (Cesare et al., 2015) and partially crystallized (= former melt, MI).

Inclusions have irregular distribution in some garnet hosts, but they tend to cluster at the rims of large crystals or in the core of tiny crystals (Fig. 2B, C). These microstructures are indicative of a primary entrapment (according to the classification of Roedder, 1984), and suggest that inclusions were mostly trapped during the final stage of garnet growth. The FI and PI coexist in the clusters suggesting that they belong to the same fluid inclusion assemblage and that entrapment took place during the same stage of garnet growth in a condition of fluid-melt immiscibility. Most FI and PI inclusions are < 10 μm in diameter, with isometric (negative crystal) shapes, equally abundant and have comparable size range. In PI, typical phases are albite, K-feldspar, quartz, muscovite, biotite, calcite, siderite, rutile and ilmenite. The FI contain a C-O-H fluid composed of H₂O and CO₂ (Fig. 3A). The presence of H₂O is marked by the broad band between 3100 and 3600 cm⁻¹, the CO₂ by the presence of

247 the Fermi diad (peaks at 1285 cm^{-1} and 1387 cm^{-1}), with peak distance (Δ) corresponding to
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2 248 CO_2 density of $< 0.1\text{ g cm}^{-1}$ (see Wang et al., 2011).
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4 249 After the experiments, the inclusions were carefully inspected to infer the temperature
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7 250 of entrapment from approach to remelting equilibrium (Fig. 4). At 740°C , the MI have well
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9 251 defined negative crystal shapes and contain glass (yellow arrows in Fig. 4) together with
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11 252 abundant unmelted crystals of quartz, biotite, muscovite, albite and ilmenite, indicating that
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13 253 remelting was incomplete (i.e., they are partially-remelted inclusions). At 760°C , albite and
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15 254 muscovite are completely resorbed, biotite is less frequent, and quartz is still common with
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17 255 typical round shape, suggesting partial resorption. Few inclusions are completely remelted to
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19 256 a homogeneous glass with presence of a shrinkage bubble. At 780°C experiments show the
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21 257 best approach to equilibrium with $> 40\text{-}50\%$ remelted inclusions, displaying regular walls,
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23 258 negative crystal shape and absence of offshoots. The presence of resorbed quartz is rare. At
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25 259 810°C , completely or partly remelted inclusions display offshots (white arrows in Fig. 4) and
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27 260 the presence of secondary minerals (i.e., formed during experiments or upon quenching) as
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29 261 aggregates of tiny acicular Bt. Remelted inclusions represent $> 70\%$ of the total. At 850°C
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31 262 and 900°C almost all inclusions are remelted or rehomogenized but show evidence of
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33 263 overheating (see Cesare et al., 2015; Bartoli et al., 2013c; Acosta-Vigil et al., 2016) as
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35 264 indicated by i) presence of lobate-cusate walls at the contact with the host garnet, ii) frequent
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37 265 offshoots, iii) growth of secondary garnet at the inclusion walls, iv) segmented straight walls
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39 266 after assimilation of the Grt (at 900°C). Ilmenite and rutile as trapped phases become less
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41 267 common at $T > 810^\circ\text{C}$ and almost absent at 900°C . In the Raman spectrum shown in Figure
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43 268 3C, the remelted glass has a broad band at $\approx 3400\text{ cm}^{-1}$ related to the presence of H_2O , while
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45 269 the shrinkage bubble within the glass contains a fluid mixture of $\text{CO}_2 + \text{N}_2 + \text{CH}_4$ (Fig. 3B).
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270 4.3 Major elements composition of remelted glasses

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2 271 In order to determine the composition of the melt produced during anatexis of the
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5 272 metapsammite, and to infer the temperature of entrapment of inclusions, we conducted EMP
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7 273 glass analyses on completely or partially remelted glasses to explore their potential interaction
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10 274 with the host garnet as function of experimental temperature (Table 3). We excluded analyses
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12 275 with clear signal of contamination either by minerals preserved in the inclusions or by host
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14 276 garnet e.g. $\text{SiO}_2 < 60 \text{ wt\%}$ and $> 80 \text{ wt\%}$, or $\text{Al}_2\text{O}_3 > 20 \text{ wt\%}$ or $(\text{FeO}_t + \text{MgO} + \text{TiO}_2) > 7$
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16 277 wt\% . The difference to 100% of the totals may reflect not only the fluid content but also
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19 278 interaction of the microprobe beam with the shrinkage bubble if present below the polished
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22 279 surface of the MI (see discussion below on $\text{H}_2\text{O}-\text{CO}_2$ content), so results of the analyses were
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24 280 recalculated on anhydrous basis (100 wt%). Figure 5 compares the chemical composition of
25
26 281 the glasses recovered in all the experiments. The average MnO vs. FeO_t content is
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29 282 respectively 0.5 wt% and 1.6-1.7 wt% up to 780°C, and systematically increases from 810°C
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31 283 to 900°C. The $\text{FeO}_t + \text{MgO} + \text{TiO}_2$ is 1-3 wt% up to 780°C with SiO_2 increasing from 66 wt%
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34 284 at 740°C to 72-76 wt% at 760°C and 75-79 wt% at 780°C. At $T > 810^\circ\text{C}$, the $\text{FeO}_t + \text{MgO} +$
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36 285 TiO_2 is inversely proportional to SiO_2 (lower SiO_2 , higher $\text{FeO}_t + \text{MgO} + \text{TiO}_2$). The CaO vs.
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39 286 SiO_2 has similar behavior for most temperatures with the SiO_2 increasing from 760°C to
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41 287 780°C and decreasing from 780°C to 810°C, 850°C and 900°C (Fig. 5C). No temperature
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44 288 dependence is visible on the Na_2O vs. SiO_2 plot, except at 900°C where the SiO_2 is
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46 289 systematically lower for comparable Na_2O contents. The Al_2O_3 is inversely proportional to
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49 290 SiO_2 at 760-810°C but values get dispersed at $T > 850^\circ\text{C}$ resulting in an ASI number > 1 , with
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51 291 the exception of few inclusions remelted at 850-900°C. The systematic variations reported in
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53 292 Fig. 5A-C and E may result from contamination of melt by mixtures of Grt-Qz and Feld-Qz
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56 293 (see directions of exchange vectors reported in Fig. 5B and E) suggesting that host interaction
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294 is not negligible at $T > 810^{\circ}\text{C}$. As a consequence, we consider the glass compositions obtained
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2 295 at $760\text{-}780^{\circ}\text{C}$ as most representative of the anatectic melt trapped within garnets.
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4 296 In Fig. 6 the composition of glasses remelted at $760\text{-}780^{\circ}\text{C}$ is plotted on the
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6
7 297 pseudoternary normative diagrams Qz-Ab-Or and An-Ab-Or. Compositions at 760°C include
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10 298 remelted glasses with and without daughter quartz. The MI cluster above the cotectic lines at
11
12 299 $0.5\text{-}1.0\text{ GPa}$, which is consistent with the displacement of cotectic lines towards Qz apex as
13
14 300 effect of increasing Al and Ca content in the granite model system (see Bartoli et al., 2016).
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16 301 The composition of MI at 760°C differ by lower SiO_2 which is related to the widespread
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18
19 302 presence of relict unmelted quartz in the inclusions. However, few glasses at 760°C overlap
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22 303 with glasses at 780°C suggesting that equilibrium melt composition are presumably achieved
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24 304 within the interval $760\text{-}780^{\circ}\text{C}$ and represent the closest approximation to the original
25
26 305 anatectic melt. Displacement of the MI composition from the minimum melt in the Qz-Ab-Or
27
28
29 306 model system has been explained by Wilke et al. (2017) as due to the presence of extra
30
31 307 components in natural systems, mostly Ca, Fe and Ti. Wilke et al. (2017) report a
32
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34 308 displacement of the eutectic point to Or-enriched composition at similar Qz amount when
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36 309 melts have $\approx 3.5\text{-}7\%$ normative An. Such An content corresponds to that of remelted glasses
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39 310 clustering at $2\text{-}7\%$ with few analyses up to $11\text{-}16\%$. The MI show variation in the Qz/feldspar
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41 311 ratio as indicated by their distribution along the directories $\text{Ab}/\text{Or} = 0.38$ and $\text{Ab}/\text{Or} = 0.5$
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43
44 312 collinear with the Qz apex. Such trends differ from those of Bartoli et al. (2016) who reported
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46 313 compositional alignments of remelted granitic glasses parallel to the Qz-feldspar join and
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48
49 314 interpreted them as due to slow diffusion of Si and Al compared to alkalis in granitic melts
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51 315 (Acosta-Vigil et al., 2006, 2017). In the An-Ab-Or projection, most glass compositions plot in
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53 316 the granite field close to the boundary with Qz-monzonite, with few Qz-monzonite
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56 317 compositions. Figure 6 also reports the composition of MI recovered in garnet of
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58 318 metasediments from Ojén (Bartoli et al., 2013b, 2013c) and in plagioclase of Grt-Bt-Sil
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319 enclaves from El Hoyazo (Acosta-Vigil et al., 2010; Baroli et al., 2016) equilibrated at similar
320 metamorphic conditions (700°C and 0.5 GPa for Ojén, 700-750°C and 0.5-0.7 GPa for El
321 Hoyazo, see below). In terms of Ab/Or ratio, the composition of glasses remelted at 760°C
322 and 780°C are similar but those at 780°C have higher normative Qz and An. The glasses plot
323 in the calcic field of the modified alkali-lime index (MALI = Na₂O + K₂O – CaO) vs. SiO₂
324 (wt%) after Frost et al. (2001) and in the rhyolite field of the TAS classification (Fig. 7A, B).

4.4 Volatile content of remelted glasses

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326 The H₂O and CO₂ concentration in glasses was determined by nanoSIMS only on MI
327 remelted at 760°C and 780°C (Table 4 – Supplementary Material) and performed on selected
328 homogeneous portion of the glass exposed on the surface of polished garnets.

329 The H₂O content of individual analyses is reported in Fig. 8 and is between 2.2-8.0
330 wt% at 760°C and 3.0-6.5 wt% at 780°, while the corresponding CO₂ is 250-850 ppm at
331 760°C and 500-2200 ppm at 780°C. In Fig. 8B and C the MI analyses are grouped according
332 to host Grt and display similar variability except for garnets ANT9-1 and ANT8-1 displaying
333 higher scatter for H₂O and CO₂, respectively. The averages at 760°C and 780°C (white circles
334 in Fig. 8B, C) are similar for H₂O with 4.3 ± 1.7 wt% at 760°C, and 5.3 ± 1.0 wt% at 780°C,
335 and slightly differ for CO₂ with 407 ± 184 ppm at 760°C, and 1147 ± 519 ppm at 780°C. In
336 Figure 3C the measured H₂O in remelted glasses is compared with maximum and minimum
337 water contents expected in model primary granitic melts (Holtz et al., 2001) and will be
338 discussed below. The Fig. 3D compares the H₂O contents with the values predicted for H₂O-
339 undersaturated haplogranitic melts (Holtz et al., 2001) while the solidus is from
340 thermodynamic modelling (see below).

4.5 Thermodynamic modelling

341
342 Constraints on the P-T conditions of melting and inclusions entrapment were obtained
343 by phase equilibria modelling (Fig. 9). Calculations were done in the model chemical system

344 MnNCaKFMASHT using the Perple_X software (Connolly, 2009) with the thermodynamic
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2 345 database of Holland and Powell (2003 revision). We used the solution models of White et al.
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4 346 (2007) for melt, Holland and Powell (1998) for garnet, Tajčmanová et al. (2009) for biotite,
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7 347 Coggon and Holland (2002) for white mica, Newton et al. (1980) for plagioclase, and
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10 348 Thompson and Hovis (1979) for K-feldspar. The bulk rock composition of the rock was
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12 349 obtained from XRF analysis and is reported as inset in Fig. 9. The bulk composition has no -
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14 350 residual character and is considered representative of the original protolith with no necessity
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17 351 of melt reintegration (see Bartoli, 2017) because i) melt inclusions are the only petrographic
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19 352 evidence of anatexis (absence of melt pseudomorphs), ii) the composition of white mica, and
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22 353 that of biotite as well, is comparable to that of primary muscovite from amphibolite-facies
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24 354 subsolidus metapelites in the area (Di Vincenzo et al., 2001). The amount of H₂O used in the
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26 355 calculation was assumed to be the L.O.I. (loss of ignition) from the XRF analysis (=1.5 wt%).
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29 356 The presence of CO₂ + H₂O ± carbonates (calcite and siderite) in PI and FI point to a C-O-H
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32 357 fluid coexisting with the rock and melt during anatexis (Connolly and Cesare, 1993). Slight
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34 358 variations in H₂O activity (tested from 0.8 to 1.0 but not reported here) resulted in limited
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36 359 displacement of the solidus at P > 0.5 GPa, where the reaction slope is high and positive (see
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38
39 360 later and Fig. 9).

41 361 The petrographic study shows that the stable mineral assemblage is represented by Ms
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43 362 + Pl + Kf + Grt + Qz +Liq ± Rt coexisting with excess C-O-H fluid as indicated by the
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46 363 presence of FI and PI within the same cluster. In the pseudosection of Fig. 9, such assemblage
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49 364 is stable at P > 0.5 GPa within a triangular field limited by the fluid-present solidus curve and
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51 365 by the Ms-out/Als-in curve to temperatures of 690°C < T < 820°C. The Fig. 10
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53 366 (Supplementary Material) reports the calculated compositional isopleths for garnet,
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56 367 plagioclase and biotite. In larger garnets from cores to rims, the X_{Pyp} slightly increases and
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58 368 X_{Sps} slightly decreases suggesting a ≈ 50°C temperature increase during the garnet growth,
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369 while smaller garnets are homogeneous with composition similar to that of large garnet rims.
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2 370 The X_{Alm} in garnet has a slightly reversed tendency (higher temperature in the core and lower
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4 371 at the rim). The combination of garnet isopleths with the X_{An} in plagioclase, delimit a P-T
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7 372 space corresponding to approximately 680°C to 730°C at 1.0-1.3 GPa (Fig. 9). The low Ti
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10 373 content in biotite (0.12 a.p.f.u.) is not compatible with such temperatures and supports the
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12 374 petrographic observation of post peak recrystallization of biotite at $T < 600^\circ\text{C}$.

14 375 The calculated isomodes for the phases biotite, garnet and melt reported in Fig. 9B
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16 376 and muscovite (not reported for clarity) indicate that biotite and muscovite are partially
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19 377 consumed while limited amount of garnet is formed during progressive metamorphism at P-T
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22 378 evolving from condition 1 to conditions 2 in Fig. 9. Modelled muscovite decreases from 12 to
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24 379 8 wt%, biotite from 22 to 20 wt% and garnet increases from 1.5% to 2-4 wt%. The melt
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26 380 produced is limited to 2-4 wt%. The calculated isomodes are in agreement with the modal
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29 381 phase proportions constrained by mass balance calculation and corresponding to 7.9 wt% of
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31 382 muscovite, 28.1 wt% of biotite, 3.4 wt% of garnet and ≈ 2 wt% of melt.

35 383 5. DISCUSSION

38 384 5.1 Conditions of MI entrapment

40 385 The present study is an example of successful remelting of inclusions that adds to
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43 386 previous case studies of metasedimentary rocks undergoing anatexis (e.g., Cesare et al, 2015;
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45 387 Bartoli et al., 2016; Tacchetto et al., 2019; Carvalho et al., 2018). Previous works report MI
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48 388 entrapped under fluid present (Bartoli et al., 2013b) or fluid-absent melting of felsic (e.g.,
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50 389 Cesare et al., 2015) or ultramafic rocks (Ferrero et al., 2018), and in a range of temperatures
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52 390 from upper-amphibolite to UHT conditions. The present case study yields new data on the
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55 391 onset of low-T anatexis of metasedimentary crust under possibly fluid-present conditions,
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57 392 which are likely to be more common than previously thought (see Carvalho et al., 2018 and
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59
60 393 discussion therein).

394 The P-T-fluid conditions of melt entrapment within the host mineral can be
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2 395 reconstructed combining the microstructures produced during remelting experiments at
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4 396 increasing temperature with the proportion of remelted inclusions, the compositional variation
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7 397 of major elements in glass and the thermodynamic modelling. In remelting experiments,
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9 398 microstructural changes include i) type of minerals coexisting with glass (daughter minerals
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11 and/or trapped phases), ii) presence of offshoots (suggesting decrepitation), iii) presence of
12 399 irregular boundaries between MI and Grt (indicative of host interaction), iv) crystallization of
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14 400 new phases (indicative overheating), v) appearance of zoning/growth structures (indicative of
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16 401 intense host interaction). Since glasses remelted at 780° are mostly free of daughter minerals,
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18 402 have no visible cracks/offshoots, and display limited or no chemical interaction with the host
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20 403 garnet (e.g., low MnO + FeO_t in Fig. 5), they can be confidently interpreted as representative
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22 404 of entrapment conditions. However, at 760°C, where most inclusions have quartz which has
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24 405 not been re-homogenized and hence lower SiO₂, few remelted inclusions resemble in texture
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26 406 and composition those typically recovered at 780°C. One may question about the effect of
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28 407 kinetics on remelting conditions and the counter play of temperature and run duration on the
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30 408 achievement of equilibrium for remelting. Preliminary experiments at 780°C performed at
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32 409 different durations (hours to days, unpublished data) seem to confirm that run duration has an
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34 410 effect on remelting and that the chemical signature of the glasses recovered at lower
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36 411 experimental duration partly overlaps (though not completely) with the composition of
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38 412 homogeneous glasses recovered at 760°C. Even though the kinetics dilemma is out of the
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40 413 scope of this work, it is reasonable to consider the interval 760-780°C at 0.8-1.0 GPa as
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42 414 representative of the entrapment conditions. Temperatures above 810°C are not representative
43
44 415 due to evident and intense glass-host interaction. Thermodynamic calculations indicate that
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46 416 the wet solidus of the rock locates at ≈690-710°C between 0.5 GPa and 1.5 GPa, and that the
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48 417 stability field of the assemblage Ms + Pl + Kf + Grt + Qz + Liq ± Rt is between 700-780°C.
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419 The temperature intervals indicated by compositional isopleths of Grt and Pl and the mineral
420 isomodes (Fig. 9) are more consistent with $\approx 700\text{-}730^\circ\text{C}$ and $\approx 730\text{-}760^\circ\text{C}$ respectively. The
421 discrepancy between calculated and experimental conditions is maximum $60\text{-}80^\circ\text{C}$ and can be
422 ascribed to diverse sources of uncertainty in phase equilibria calculations, such as the quality
423 of thermodynamic data and, in turn, the quality of mineral and melt solution models (White et
424 al., 2011). For example, the current melt model requires improvements to reproduce properly
425 natural processes (see discussion in Bartoli et al., 2013b), as already pointed out in previous
426 works (e.g., Bartoli et al., 2013a and 2016; Grant, 2009 and White et al., 2011).

427 **5.2 Fluid regime and volatile content of the primary melt**

428 There are few lines of evidence suggesting that partial melting of the investigated
429 rocks occurred in the presence of a fluid phase, including i) the coexistence of primary FI and
430 PI within the same cluster, and ii) the major element composition of primary melt.

431 The abundant FI coexisting with PI in the same cluster indicate the presence of a C-O-
432 H fluid dominated by H_2O and CO_2 . Graphite was not observed within the FI but carbonates
433 (calcite and siderite) are common within garnets. Recently Carvalho et al. (2018) discussed
434 the recovery of C-O-H bearing FI in metasedimentary rocks from Ivrea at amphibolite to
435 granulite conditions and argued that “a (graphitic) rock undergoing melting will always show
436 the presence of a free C-O-H fluid, in variable amounts and with variable $a_{\text{H}_2\text{O}}$ depending on
437 pressure, temperature and the bulk composition of the system”. According to Carvalho et al.
438 (2018), the fluid regime during anatexis might be characterized by CO_2 -dominated fluid in
439 processes dominated by dehydration of OH-bearing phases. In the Edixon Complex, where
440 the fluid was likely H_2O dominated, conditions of entrapment are compatible with the
441 beginning of partial melting and potentially with both muscovite and biotite fluid-present
442 conditions. The calculated isomodes suggest that muscovite and biotite are consumed at the
443 solidus to produce limited amounts of melt and peritectic garnet, always in the absence of

1 444 aluminum silicate. Since the muscovite fluid-present melting (without biotite) would produce
2 445 peritectic aluminum silicate instead of garnet (Weinberg and Hasalova, 2015), the rock
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4 446 assemblage is more consistent with a muscovite + biotite fluid-present melting reaction
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7 447 (Weinberg and Hasalova, 2015). Increasing $a_{\text{H}_2\text{O}}$ should reduce garnet stability (Conrad et al.,
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9 448 1988; Gardien et al., 2000), but the experiments of Alonso-Perez et al. (2009) on Si-rich melts
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11 449 indicated the opposite, i.e. that stability of peritectic garnet increases with increasing H_2O in
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14 450 melt.

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17 451 The H_2O content of glasses at 760°C is more scattered ($\approx 2\text{-}8$ wt%) than at 780°C ($\approx 4\text{-}$
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19 452 7 wt%) but in both cases it is close to the values predicted for H_2O saturation at 0.8 GPa for
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21 453 haplogranitic melts (≈ 7.6 wt%; Holtz et al., 2001, Fig. 8D). The lower H_2O contents of some
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24 454 glasses may be explained as effect of heterogeneous distribution of the fluid (see discussion in
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26 455 Bartoli et al., 2014) and/or by the presence of a diluting carbonic species in the COH fluid,
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28
29 456 namely CO_2 . The H_2O content of glasses determined as difference to 100 from EMPA
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31 457 analyses may yield similar contents, though Acosta-Vigil et al. (2016) report deviations up to
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34 458 30%. In Table 3, the average difference to 100 total decreases with increasing experimental
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36 459 temperatures from ≈ 14 wt% at 740°C to $\approx 12\text{-}11$ wt% at $760\text{-}780^\circ\text{C}$ and ≈ 10 wt% at 810-
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39 460 900°C . All values overestimate the measured H_2O in glasses. Possible explanations may
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41 461 account for the presence of light and incompatible elements as B, which is abundant in first
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43 462 anatectic products or micropores and would be consistent with the presence of tourmaline in
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45
46 463 the rock. Concerning other volatiles, the Cl and F content of the glasses was below the
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49 464 detection limits (Table 3). This inconsistency requires further investigation.

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51 465 The maximum CO_2 content of the remelted glasses from NanoSIMS (with all C
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53 466 calculated as CO_2) is $200\text{-}500$ ppm at 760°C and $500\text{-}2000$ ppm at 780°C (Fig. 8A). Such
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56 467 difference may be explained considering that i) the CO_2 solubility in felsic melts slightly
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58 468 increases with increasing T at least at 0.5 GPa (Tamic et al., 2001), ii) the CO_2 solubility and
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469 speciation are affected by the glass composition, with total CO₂ increasing with increasing
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2 470 SiO₂ and decreasing with increasing Na₂O, and molecular CO₂ prevailing on the (CO₃)²⁻
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4 471 species in SiO₂ rich- and Na₂O poor- compositions (Brooker et al., 1999). At 780°C, glasses
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6 472 have higher SiO₂-CaO and lower Na₂O with respect to glasses produced at 760°C, so the CO₂
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8 473 speciation is expected to be dominated by molecular CO₂. Even though most of the analyzed
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10 474 MI appeared homogeneous and bubble-free on polished surface, the presence of shrinkage
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12 475 bubbles deep within the inclusion cannot be excluded. Since the fluid phase preserved in the
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14 476 shrinkage bubble is mostly CO₂-rich (Fig. 3B) in agreement with Wallace et al. (2015), this
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16 477 may account for an additional CO₂ reservoir. If (slightly) higher resorption of the shrinkage
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18 478 bubbles takes place at increasing temperature, this may explain the higher CO₂ content of
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20 479 glasses remelted at 780°C vs. 760°C.
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26 480 Granitic melt produced by water-present melting of metasedimentary rocks are
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28 481 expected to have higher ASI values than those produced by muscovite or biotite dehydration
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30 482 melting (Acosta-Vigil et al., 2003), to be enriched in Na₂O and CaO (Weinberg and Hasalova,
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32 483 2015), and to contain higher proportion of MgO and FeO at increasing water content or water
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34 484 activity (e.g. Conrad et al., 1988; Holtz and Johannes, 1991) at constant P–T (Patiño Douce
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36 485 and Beard, 1996). This is partly consistent with the glass composition at 780°C, with ASI
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38 486 value of 1.57, MALI (= Na₂O + K₂O - CaO) value of $\approx 4.0 \pm 1.0$ wt%, CaO content of ≈ 1.27
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40 487 ± 0.7 wt.% and FeO_{tot} + MgO > 2.1 wt%, supporting the hypothesis of water-saturated partial
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42 488 melting.
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48 489 **5.3 Anatexis and evolution of the Edixon Metamorphic Complex**

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50 490 The tectonometamorphic history of the Edixon Metamorphic Complex and its role in
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52 491 the evolution of the Wilson Terrane has been primarily constrained by investigations along
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54 492 the western Lanterman Range but also southward within a wider belt including Salamander
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56 493 Range and Niagara Icefalls-Mountaineer Range regions which are exposed, as the Lanterman
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494 Range, close to the Wilson-Bowers tectonic boundary (e.g., Ghiribelli, 2000; Talarico et al.,
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2 495 1998; 2004). In general, the EMC contains no record of an early HP metamorphic stage, but
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4 496 preserves a broad metamorphic zonation from low/medium grade to high grade up to anatexis.
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7 497 However, previous studies showed the presence of rocks with contrasting P-T paths in
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9 498 different localities of the EMC (Castelli et al., 2003; Ghiribelli, 2000; Talarico et al., 1998,
10
11 499 2004). Ghiribelli (2000) recognized, at the western Lanterman Range and at the Salamander
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13 500 Range, that the rocks of the EMC show close similarities in lithological assemblages and
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15 501 metamorphic field gradients with other metasedimentary sequences characterized by an
16
17 502 andalusite-sillimanite prograde counterclockwise path with isobaric cooling, and abundant
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19 503 Grt-Crd migmatites outcropping at the Deep Freeze Range (Palmeri et al. 1991, 1994;
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21 504 Palmeri, 1997). The latter low-pressure belt (> 200 km wide) located in an internal position of
22
23 505 the Wilson Terrane is characterized by the presence of the Granite Harbor Intrusive complex
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25 506 (Borg et al., 1990; Ghezzo et al., 1989). The extensive and prolonged magmatic underplating
26
27 507 of the Granite Harbor Intrusives bodies in the Wilson Terrane (from 520-530 to 500 Ma,
28
29 508 Giacomini et al., 2007; Rocchi et al., 2004) possibly triggered, in different times, the anatexis
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31 509 event and, at the same time, promoted the juxtaposition of migmatites with contrasting P-T
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33 510 paths likely formed during the same Ross orogenic cycle (Palmeri, 1997).
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41 511 In other localities, as at the Mountaineer Range region, a different P-T path has been
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43 512 proposed (Castelli et al., 2003; Talarico et al., 1998), involving an early decompression path
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45 513 following a moderate thickening, magmatic intrusions between 500 to 480 Ma (the Granite
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47 514 Harbour intrusives), and a final cooling-unloading path has been inferred (Castelli et al.,
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49 515 2003).
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53 516 In this paper, the MI preserved in peritectic garnet and phase equilibria modelling
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55 517 demonstrate that the portion of the EMC close to Mt. Edixon experienced anatexis above
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57 518 700°C and up to 760-780°C at ≈ 1.0 GPa, corresponding to the P-T conditions of entrapment.
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2 519 In the NW part of the Lanterman Range (Carnes Crag) Grt-Crd migmatites yield T-P
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4 520 conditions at $\approx 700^\circ\text{C}$ and $\approx 0.5\text{ GPa}$ (Ghiribelli, 2000). Different conditions, however, were
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6 521 found in the southern portion of the EMC (Mountaineer Range), where the highest
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8 522 metamorphic grade shows temperature at $\approx 700\text{-}750^\circ\text{C}$ and pressure from ≈ 0.5 up to 0.6 GPa
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10 523 (Castelli et al., 2003). The spread in low to medium pressure rock types and contrasting P-T
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12 524 paths at the boundary with the Wilson-Bowers terranes is not uncommon (Ricci et al., 1997;
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14 525 Ricci and Tessensohn, 2003), and supports the interpretation that anatexis could have taken
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16 526 place at different crustal levels during the evolution of the tectono-metamorphic stages of the
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18 527 Ross Orogen.

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21 528 The latter is in keeping with the multiple and diachronous generations of white micas
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23 529 evidenced by Di Vincenzo et al., (2001) in the different metamorphic complexes of the
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25 530 Lanterman Range. The different white micas have been interpreted as testifying different P-T
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27 531 deformation stages developed during the evolution of the three metamorphic complexes (Di
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29 532 Vincenzo et al., 2001).

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32 533 The white micas in the investigated sample GO5 have no evidence of retrogression,
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34 534 they are texturally and compositionally similar to type-2 white mica of Di Vincenzo et al.,
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36 535 (2001), in fact they grow along the main amphibolite-facies foliation, and show a light
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38 536 enrichment in celadonic component ($\text{Si} = 3.1\text{ a.p.f.u.}$ and $\text{Ti} = 0.022\text{ a.p.f.u.}$). Moreover, data
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40 537 collected in this paper indicate that white micas are compatible with fluid-present melting as
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42 538 reconstructed by combination of thermodynamic modelling and MI remelting experiments.
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44 539 Since the composition of white micas in the Wilson Terrane evolves with time and the type-2
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46 540 mica dates to ca. $480\text{-}490\text{ Ma}$, we infer that the latter age could be referred to melting in the
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48 541 Lanterman Range.

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51 542 Even though it is not possible to reconstruct a proper P-T-relative time path based on
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53 543 the single sample GO5, the equilibrium conditions reconstructed from the remelting
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1 544 experiments and thermodynamic modelling allow extending the evolution of the EMC up to
2 545 ca. ≈ 1.0 GPa at 760-780°C. The latter data suggest two possibility: i) the EMC at the
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4 546 Lanterman Range is not an homogeneous metamorphic complex characterized by a low-
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7 547 pressure metamorphic imprint as postulated previously (Ghiribelli, 2000) but, similar to the
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9 548 HP-UHP belt (GHMC), it represents a mélange of different crustal slices; ii) the outcrop,
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11 549 from which the study sample comes from could be the extension towards the southern
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13 550 Lanterman Range of the intermediate-pressure belt of the BMC. Whatever the solution of the
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15 551 latter problem is, combining (Fig.10) geochronological data collected at the Lanterman Range
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17 552 (Di Vincenzo et al., 2016 and references therein), previous P-T paths (Di Vincenzo et al.,
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19 553 2001; Ghiribelli 2000; Talarico et al., 2004) and the results for sample GO5, the docking
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21 554 between the BMC and/or the different tectonic slices of the EMC might have been before 480
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23 555 Ma. The latter age is widespread in the area and it was interpreted as the Early Ordovician
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25 556 final “closure” of micas at the Lanterman Range (Di Vincenzo et al., 2001) and mainly
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27 557 referred to the exhumation process of the different complexes, whereas in the inner Wilson
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29 558 Terrane that age correspond to the youngest undeformed calc-alkaline granitoids.
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36 559 Thus, in the Lanterman Range, the high-pressure eclogitic rocks (GHMC), the
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38 560 medium pressure amphibolite stage (BMC/EMC pro parte) and the low-pressure amphibolite
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40 561 stage (EMC pro parte) shared most of the late exhumation path.
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2 569 and is available at the PNRA (Programma Nazionale Ricerche in Antartide) rock repository
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5 570 located at the Museo Nazionale dell'Antartide (Siena, Italy).
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939 FIGURE CAPTIONS

940 Figure 1 – Location of the Ross orogenic belt in the Transantarctic Mountains

941 (modified from Godard and Palmeri, 2013). A) Location map; B) geological sketch map of

942 Northern Victoria Land (RBT, Robertson Bay Terrane; BT, Bowers Terrane; WT, Wilson

943 Terrane; LR, Lanterman Range; SR, Salamander Range; MR, Mountaineer Range); C)

944 geological map of the Lanterman Range (after Ghiribelli, 2000) and location of the sampling

945 area.

947 Figure 2 – Microphotographs of sample GO5 under optical microscope, plane

948 polarized light (PPL). (A) Typical microstructure with muscovite and biotite growing along

949 the main foliation, abundant plagioclase and quartz, garnet of ca. 400 μm . Cluster of

950 inclusions (FI and PI) (B) along the rims of a large crystal of garnet, or (C) in the core of a

951 tiny crystal of garnet partly replaced by down-grade biotite.

953 Figure 3 –Representative Raman spectra of phases and volatile species in: A) fluid

954 inclusion; B) shrinkage bubble in remelted glass; C) remelted glass.

956 Figure 4 – Microstructures of MI under scanning electron microscope (BSE images)

957 after remelting experiments in piston cylinder apparatus at increasing temperature: 740°C -

958 partially homogenized MI with glass (yellow arrows), quartz, biotite, muscovite, albite and

959 ilmenite; 760°C – glassy MI with rounded quartz (= partial resorption); 780°C - glassy MI

960 with regular walls, negative crystal shape and crystals of Ilm; 810°C - homogenized MI with

961 local offshots (white arrows) and secondary acicular Bt; 850°C - homogenized MI with
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2 962 lobate-cusped walls, offshots (white arrows), and secondary Grt suggesting overheating and
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4 963 host interaction; 900°C - glassy MI with segmented straight walls after assimilation of host
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7 964 Grt (inset).
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12 966 Figure 5 – Harker diagrams showing the major element concentrations (wt%) of melt
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14 967 inclusions in crustal enclaves after remelting experiments at 740°C to 900°C (data in Table 3).
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16 968 (A) MnO vs. FeOtot. (B) FeO + MgO + TiO₂ vs. SiO₂ with mixing vectors between the
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18 969 mineral pairs Grt-Qz and Feld-Qz (see text for details). (C) CaO vs. SiO₂. (D) Na₂O vs. SiO₂.
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21 970 (E) Al₂O₃ vs. SiO₂ with mixing vectors between the mineral pairs Grt-Qz and Feld-Qz. (F)
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24 971 ASI vs. SiO₂.
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27 972

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29 973 Figure 6 – CIPW Qz-Ab-Or and An-Ab.Or diagrams showing the normative
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31 974 compositions of all analyzed melt inclusions. Blue and green circles are for glassy inclusions
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33 975 remelted at 760°C and 780°C respectively. Grey dotted lines show the eutectic points and
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35 976 cotectic lines for the subaluminous haplogranite system at $a_{\text{H}_2\text{O}} = 1$ and different pressures
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38 977 (Holtz and Johannes 1992; Luth et al. 1964; Huang and Wyllie 1975). Grey circles and lines
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40 978 represent the effect of different amounts of normative melt An at constant 100 MPa pressure
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42 979 (data from James and Hamilton, 1969). References to experimental glasses from Ojén
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44 980 (Bartoli et al., 2013c) and El Hoyazo (Acosta-Vigil et al., 2007) where “met.” refers to
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46 981 metatexites, and “diat.” to diatexite.
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51 982

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53 983 Figure 7 – Composition of analyzed melt inclusions in (A) modified alkali-lime index
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55 984 (MALI = Na₂O + K₂O – CaO) vs. SiO₂ (wt%) after Frost et al. (2001); (B) total alkali vs.
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57 985 SiO₂ diagram. Symbols as in Fig. 6.
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987 Figure 8 – H₂O and CO₂ concentration of re-homogenized melt inclusions in
988 experiments performed at 760°C (blu circles) and 780°C (green circles) at 0.8-1.0 GPa
989 measured by NanoSIMS. White circles in (B) and (C) are averages of analyses at 760°C and
990 780°C. Horizontal bars are one standard deviation on average values. (D) Comparison of H₂O
991 with expected values from Holtz et al. (2001), cross symbols are averages at 760°C (green)
992 and 780°C (blue).

994 Figure 9 – (A) P–T section for GO5 calculated in MnNCaKFMASHT system (bulk
995 composition from XRF analysis). (B) Isomodes of biotite, garnet and melt. The green shaded
996 area in B defines the calculated modal composition consistent with mass balance calculations.

997
998 Fig. 10 (Supplementary Material) – Compositional isopleths for X_{Pyrr}, X_{Alm} and X_{Spn} in
999 garnet, X_{An} in plagioclase, X_{Mg} and Ti (a.p.f.u.) in biotite calculated in MnNCaKFMASHT
1000 system (bulk composition of GO5 from XRF analysis).

1001
1002 Fig. 11. P–T trajectory path for prograde to retrograde path of eclogite and felsic rocks
1003 from Lanterman Range (modified from Di Vincenzo et al., 2016). Blue box represents the
1004 estimated P-T conditions for partial melting in the Edixon Complex.

TABLE CAPTIONS

1005
1006
1007 Table 1 (Supplementary Material) – Mean compositions (wt%) of minerals from GO5
1008 (standard deviations within brackets).

1009
1010 Table 2 – Experimental conditions of piston-cylinder remelting experiments.

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Table 3 - Major element composition (wt%) of glasses in remelted inclusions on

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anhydrous basis. Numbers in parentheses refer to 1 σ standard deviation.

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Table 4 (Supplementary Material)- H₂O and CO₂ concentrations measured in remelted

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inclusions by NanoSIMS.

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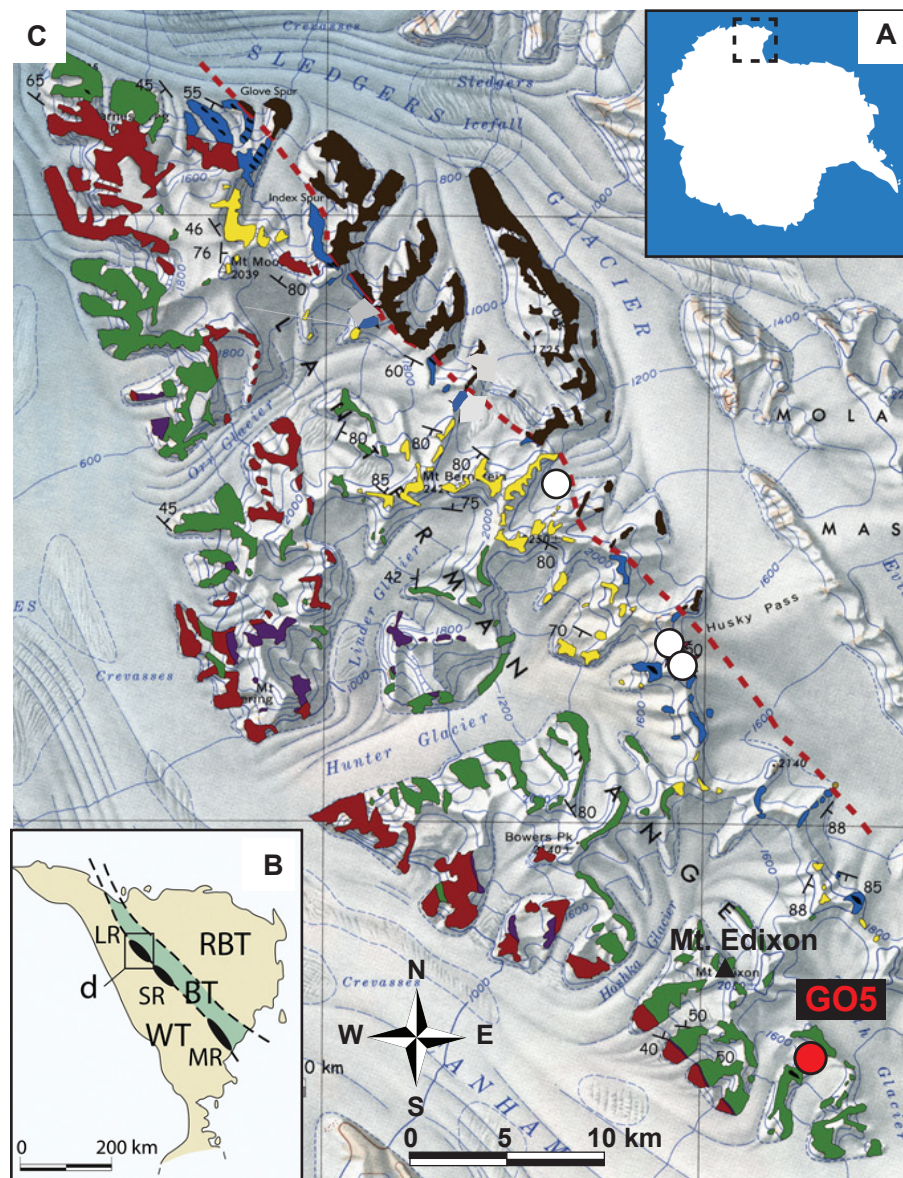
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Figure 1



■ BOWERS TERRANE (BT) ● **sampling area**

--- *Lanterman Fault*

WILSON TERRANE (WT)

- Gateway Hills Metamorphic Complex (GHMC)
eclogite facies rocks
- Bernstein Metamorphic Complex (BMC)
- Edixon Metamorphic Complex (EMC)
- Granite Harbour Intrusives (GHI)

Figure 1

Figure2

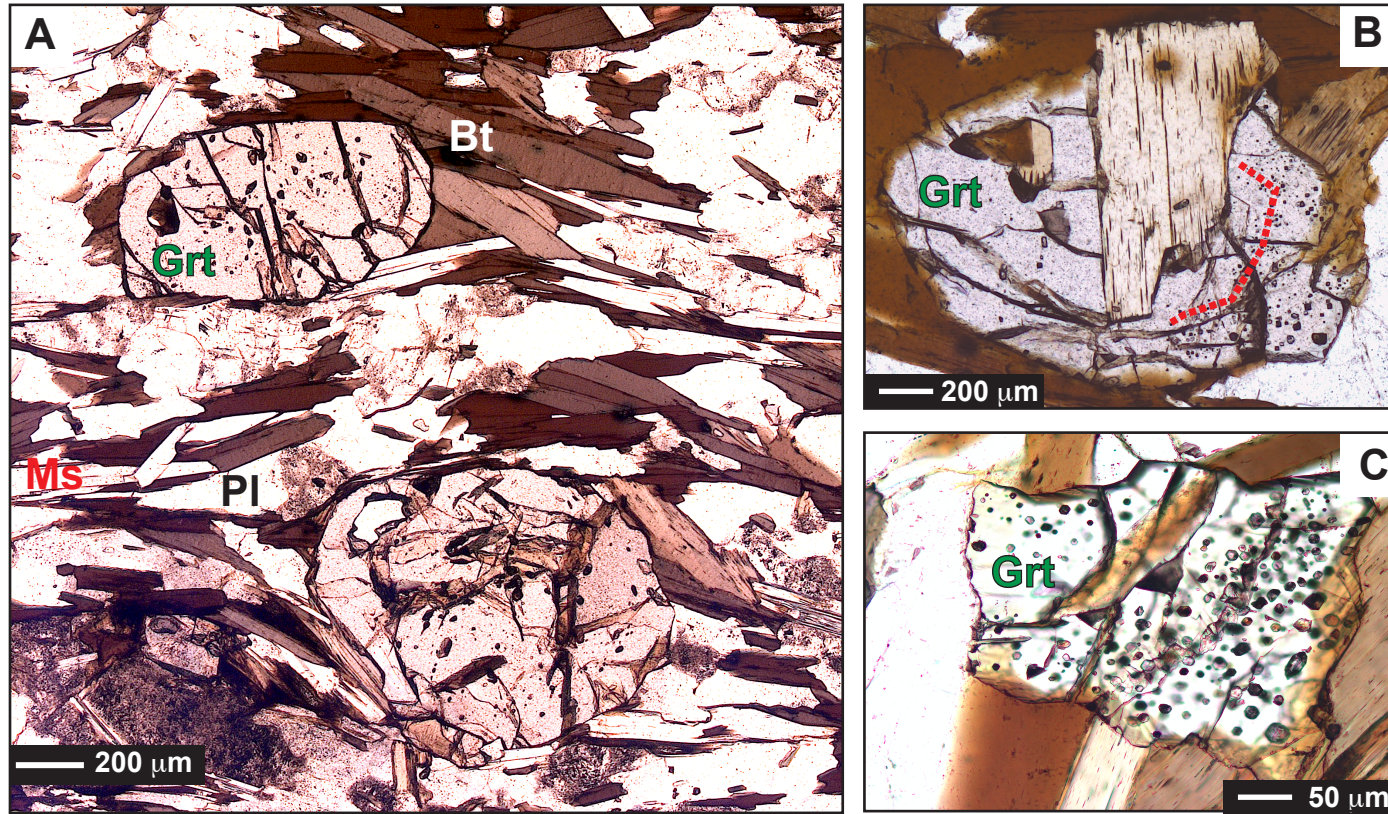


Figure 2

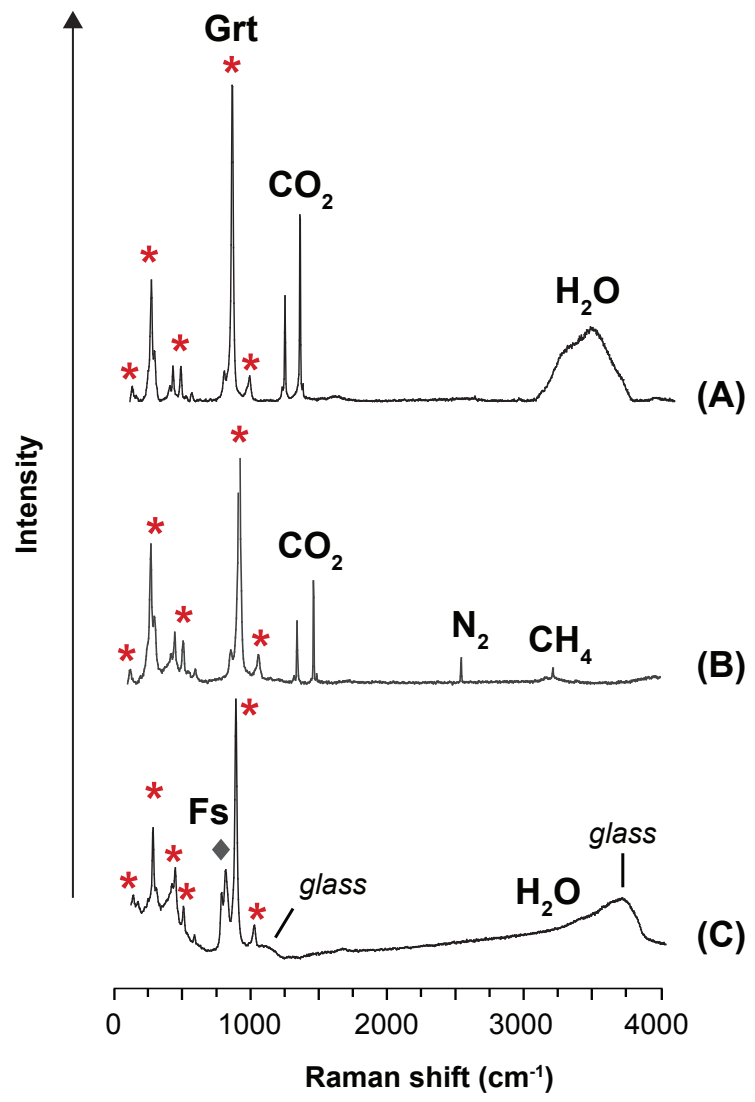


Figure 3

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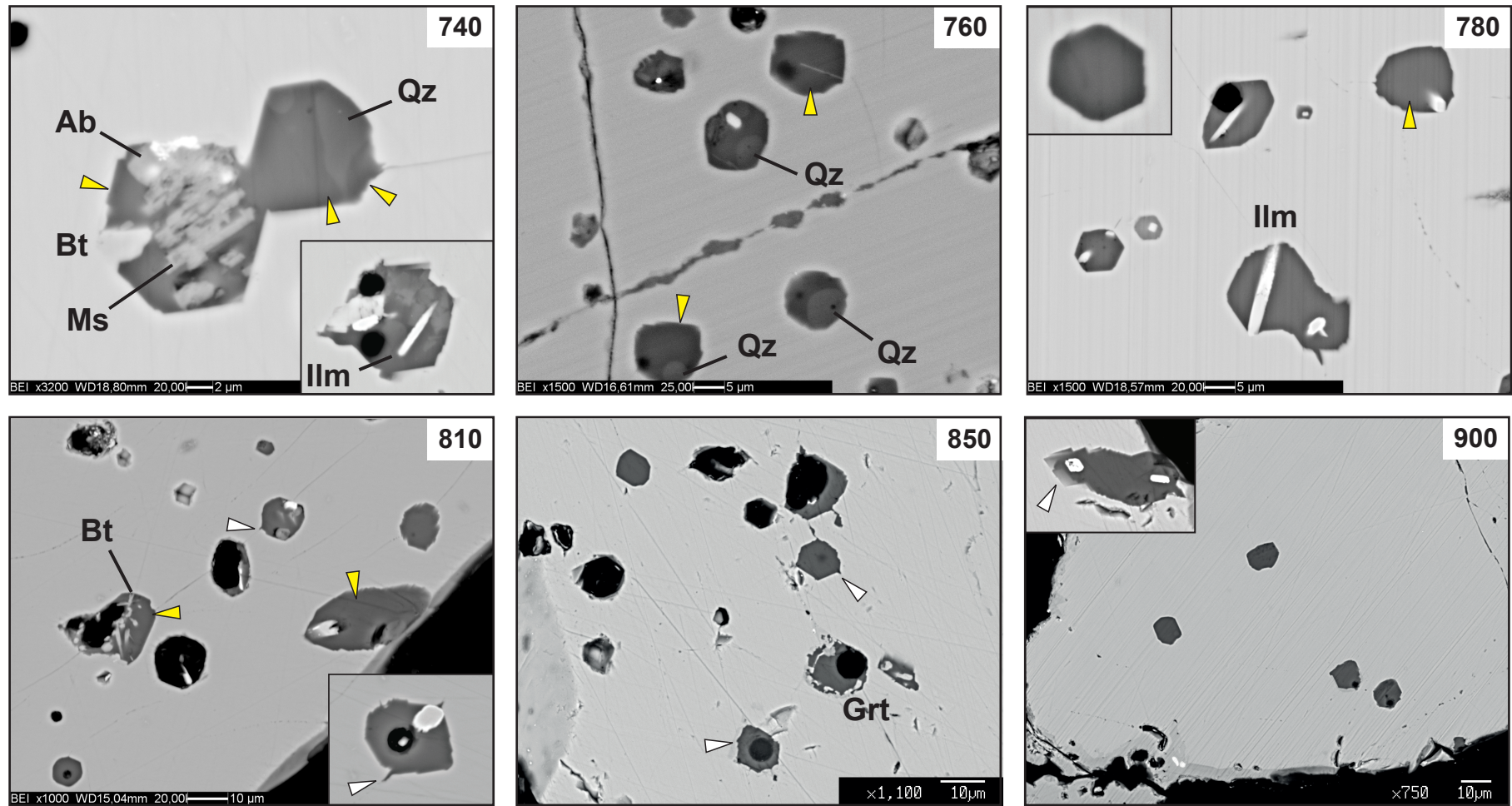


Figure 4

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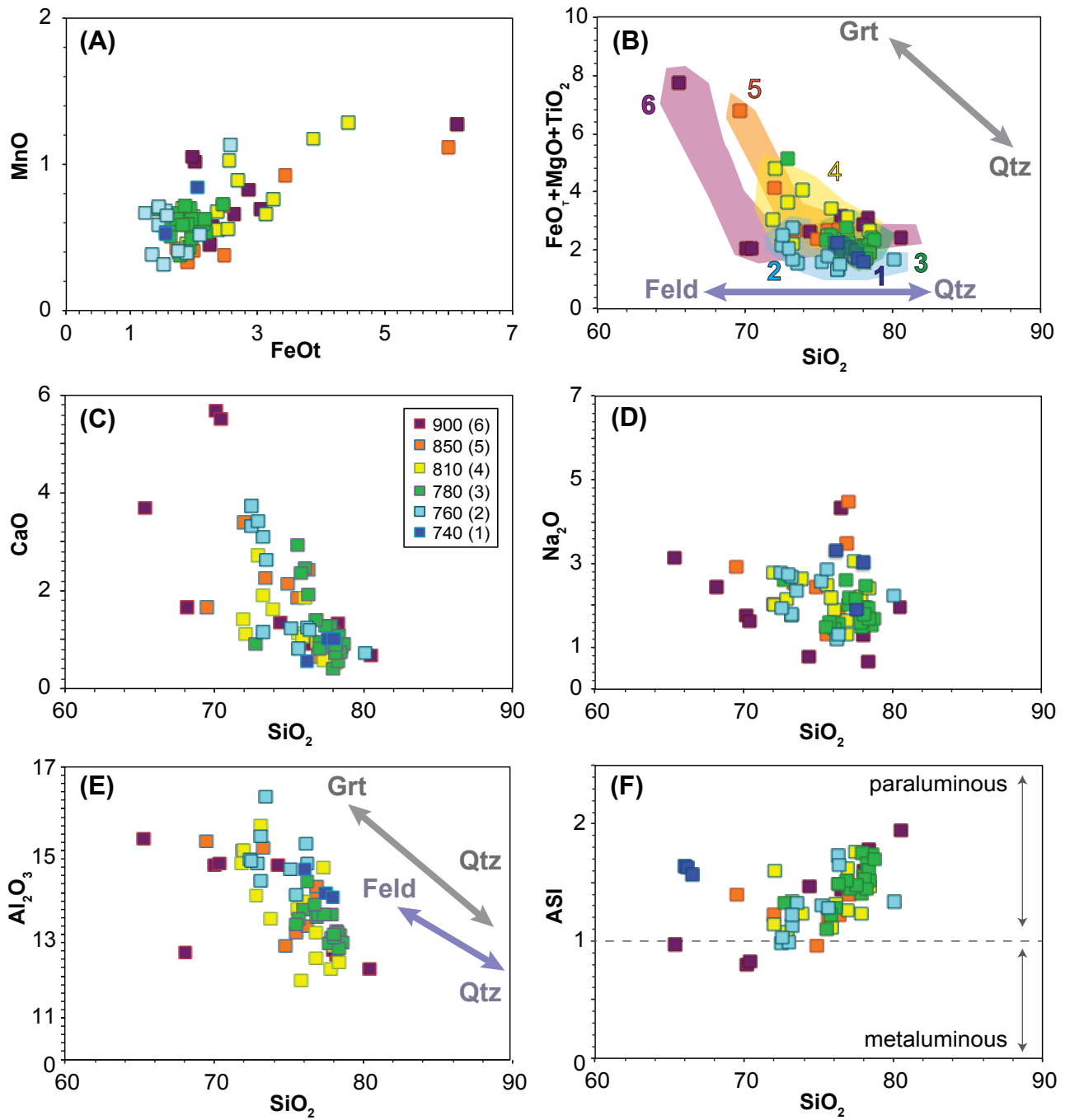


Figure 5

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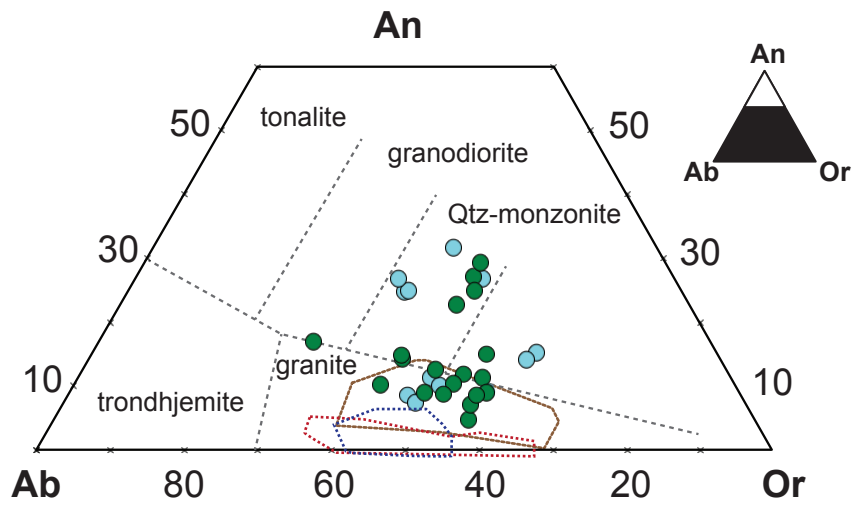
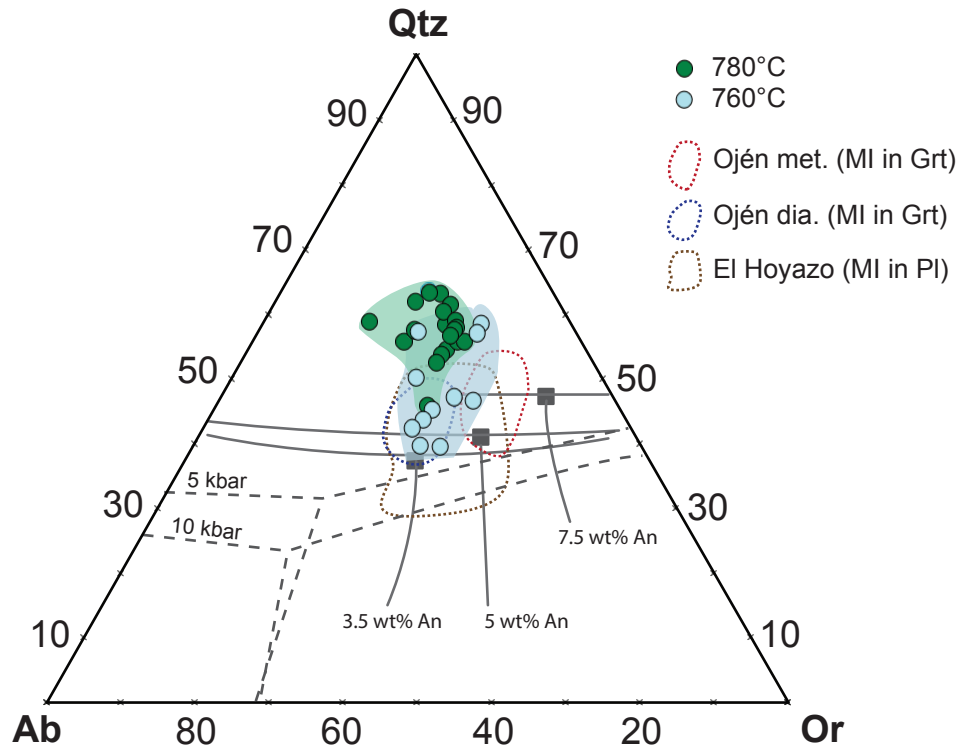


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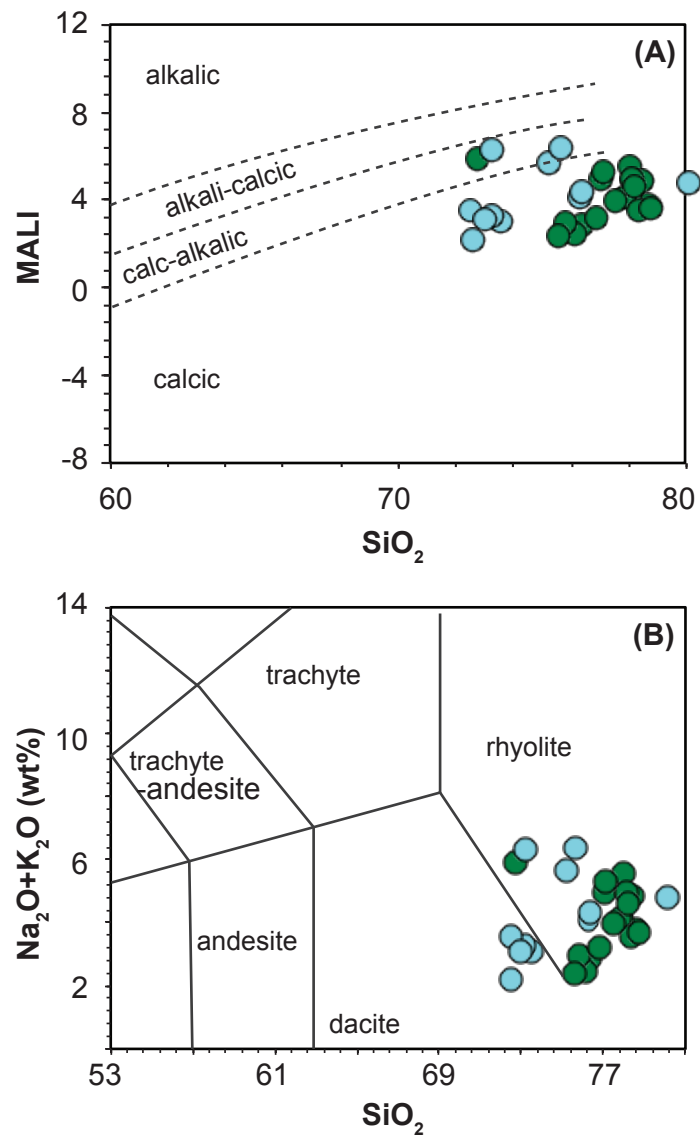


Figure 7

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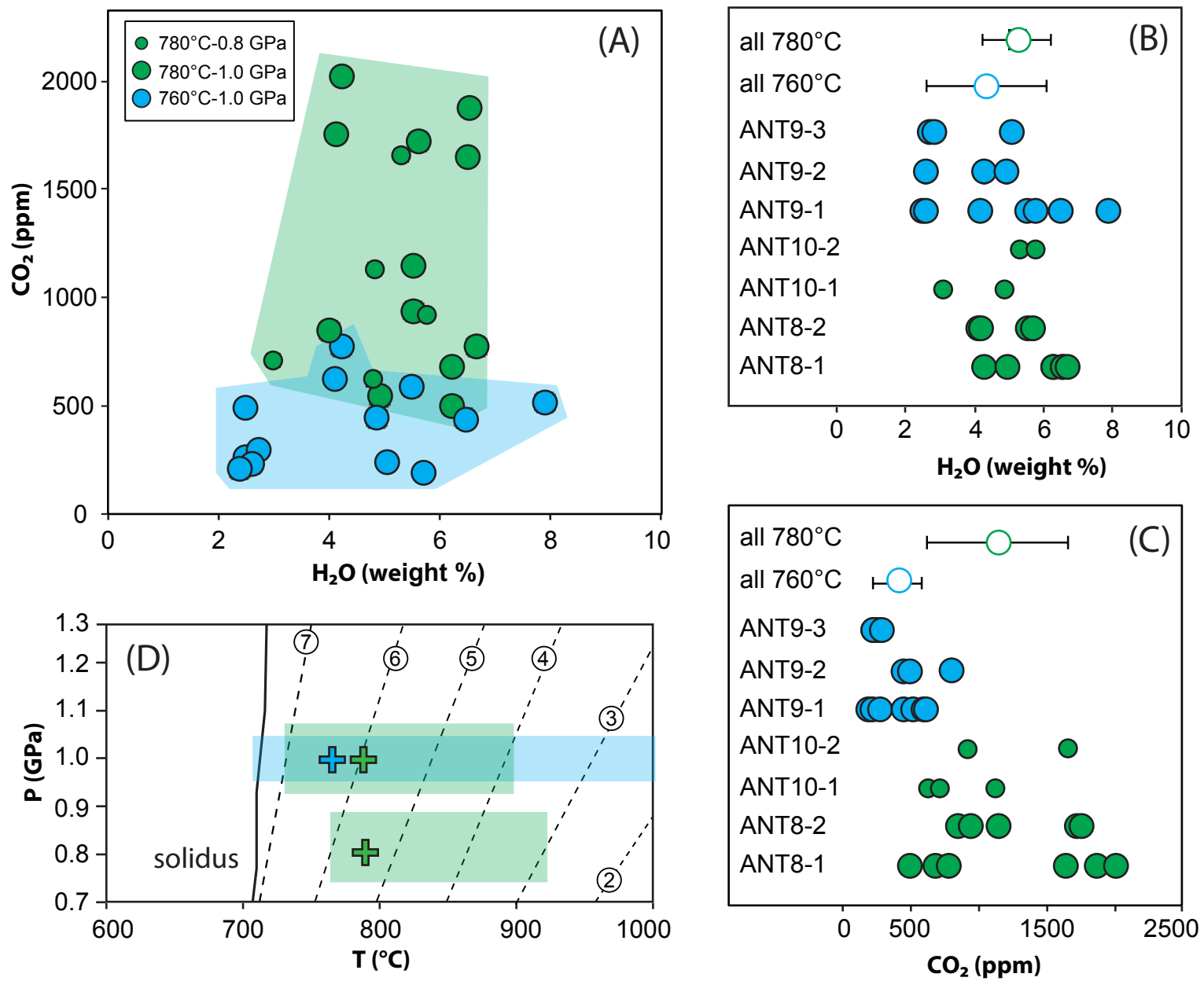


Figure 8

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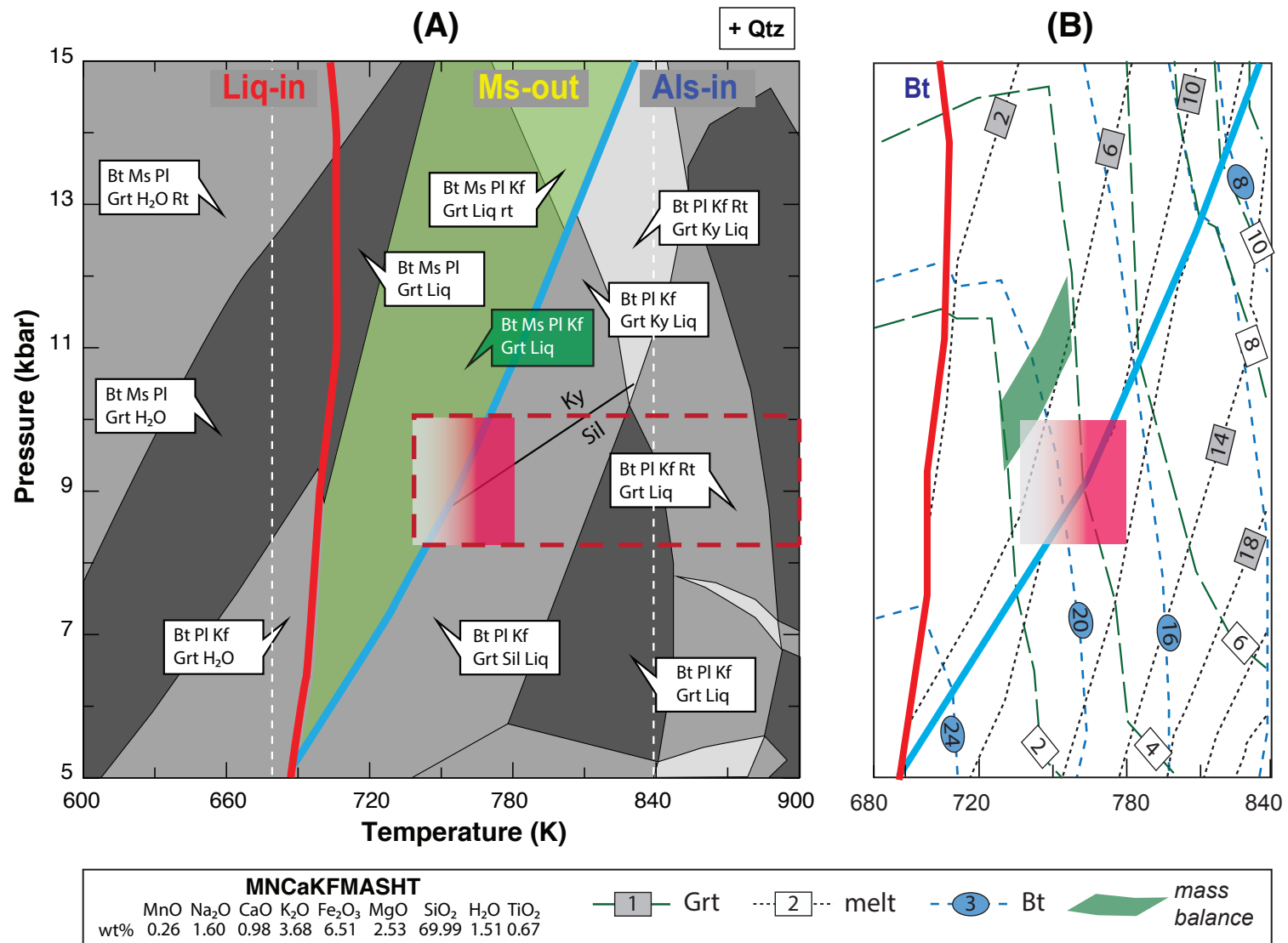


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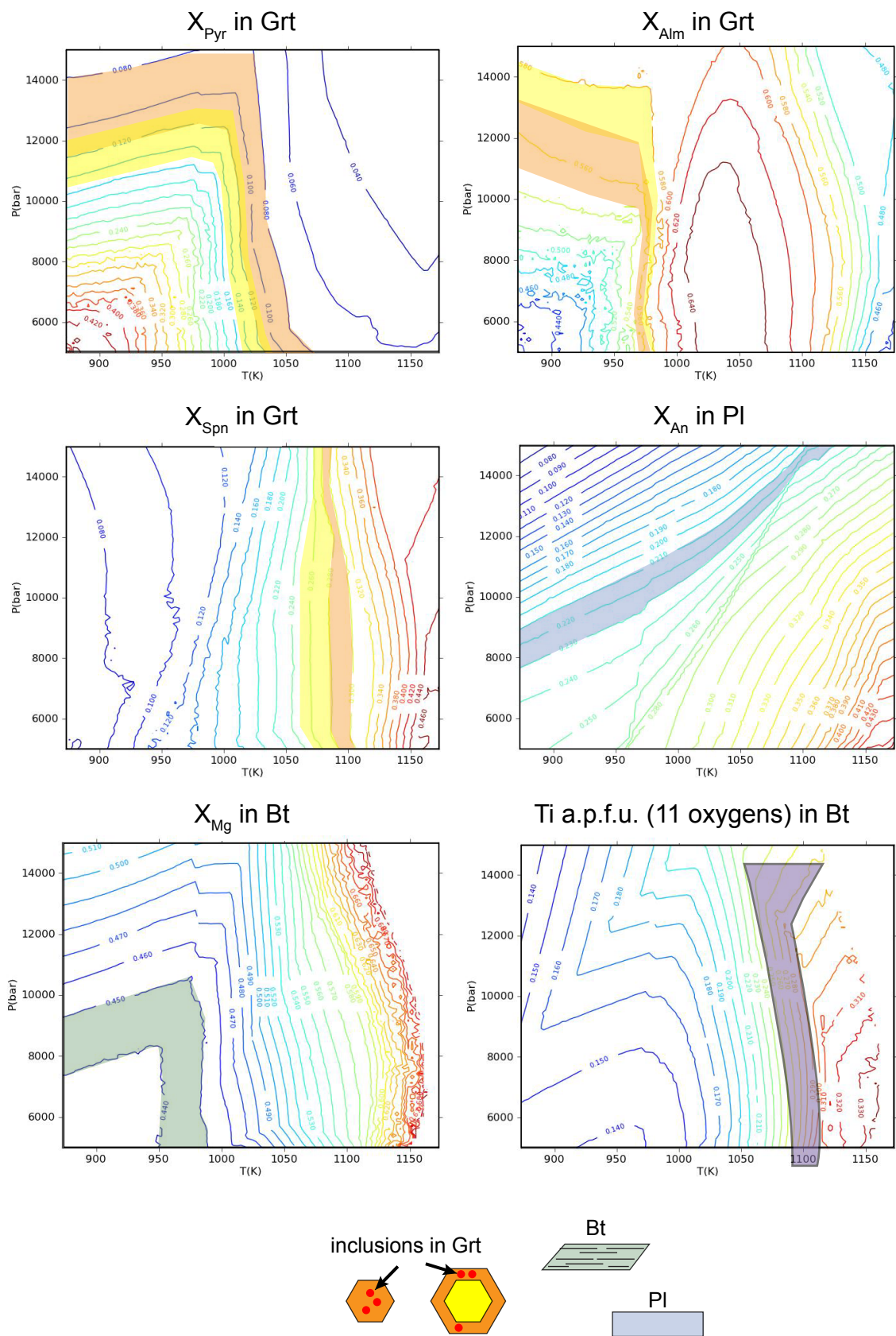


Figure 10 (Supplementary Material)

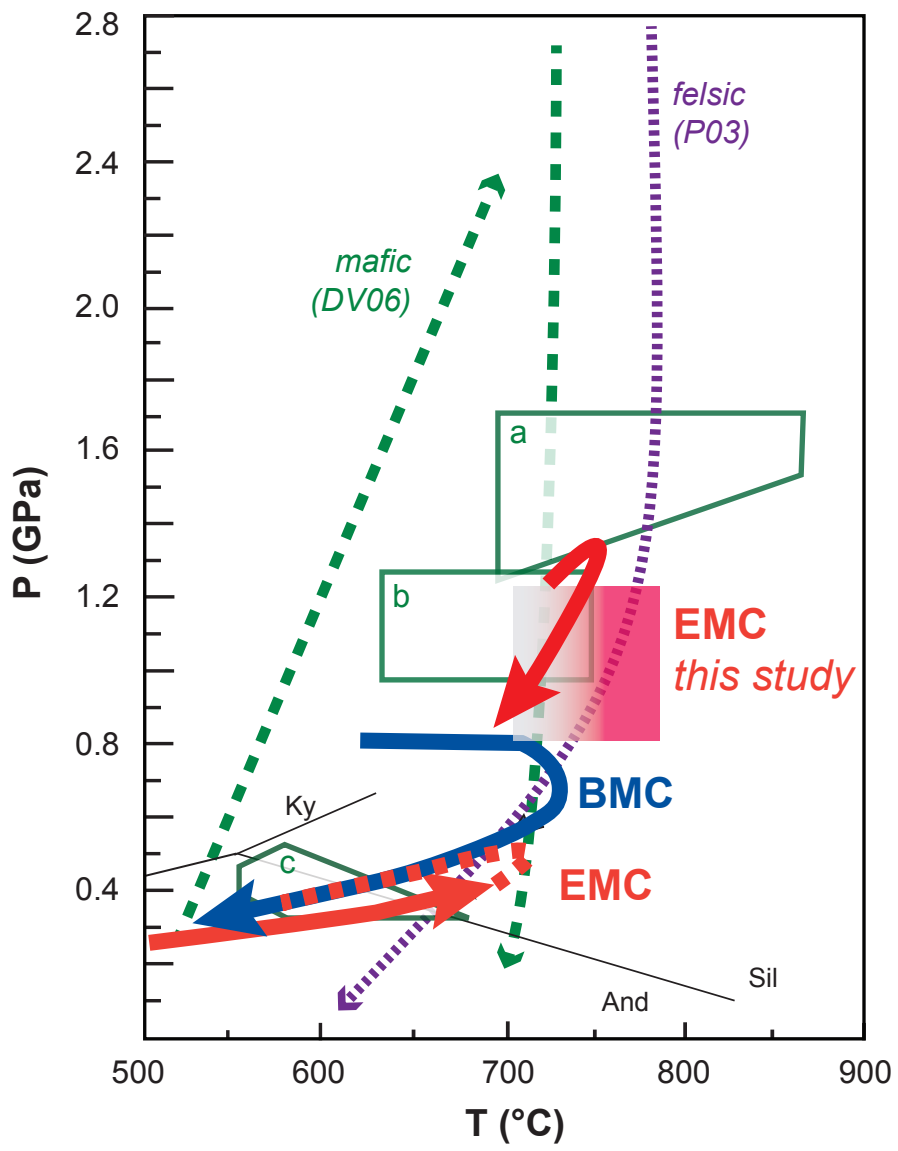


Figure 11

Table1

| <i>(n. analyses)</i> | <i>Garnet (6)</i> | | | | <i>Muscovite (8)</i> | | <i>Biotite (9)</i> | | <i>Plagioclase (8)</i> | | <i>Kfeldspar (7)</i> | |
|--------------------------------|--------------------------------|--------------------------------|--------------------------------|--------------------------------|--|---------------------------------|---|---------------------------------|---|---------------------------------|---|---------------------------------|
| | core | | rim | | mean | STD | mean | STD | mean | STD | mean | STD |
| | mean | STD | mean | STD | | | | | | | | |
| SiO ₂ | SiO ₃ | SiO ₄ | SiO ₅ | SiO ₆ | SiC SiO ₈ | SiO ₉ | SiC SiO ₁₁ | SiO ₁₂ | SiC SiO ₁₄ | SiO ₁₅ | SiC SiO ₁₇ | SiO ₁₈ |
| TiO ₂ | TiO ₃ | TiO ₄ | TiO ₅ | TiO ₆ | TiC TiO ₈ | TiO ₉ | TiC TiO ₁₁ | TiO ₁₂ | TiC TiO ₁₄ | TiO ₁₅ | TiC TiO ₁₇ | TiO ₁₈ |
| Al ₂ O ₃ | Al ₂ O ₄ | Al ₂ O ₅ | Al ₂ O ₆ | Al ₂ O ₇ | Al ₂ Al ₂ O ₉ | Al ₂ O ₁₀ | Al ₂ Al ₂ O ₁₂ | Al ₂ O ₁₃ | Al ₂ Al ₂ O ₁₅ | Al ₂ O ₁₆ | Al ₂ Al ₂ O ₁₈ | Al ₂ O ₁₉ |
| FeO | FeO | FeO | FeO | FeO | FeC FeO | FeO | FeC FeO | FeO | FeC FeO | FeO | FeC FeO | FeO |
| MnO | MnO | MnO | MnO | MnO | Mn MnO | MnO | Mn MnO | MnO | Mn MnO | MnO | Mn MnO | MnO |
| MgO | MgO | MgO | MgO | MgO | Mg MgO | MgO | Mg MgO | MgO | Mg MgO | MgO | Mg MgO | MgO |
| CaO | CaO | CaO | CaO | CaO | Ca(CaO | CaO | Ca(CaO | CaO | Ca(CaO | CaO | Ca(CaO | CaO |
| Na ₂ O | Na ₂ O | Na ₂ O | Na ₂ O | Na ₂ O | Na ₂ Na ₂ O | Na ₂ O | Na ₂ Na ₂ O | Na ₂ O | Na ₂ Na ₂ O | Na ₂ O | Na ₂ Na ₂ O | Na ₂ O |
| K ₂ O | K ₂ O | K ₂ O | K ₂ O | K ₂ O | K ₂ (K ₂ O | K ₂ O | K ₂ (K ₂ O | K ₂ O | K ₂ (K ₂ O | K ₂ O | K ₂ (K ₂ O | K ₂ O |
| Total | 0.00 | (0.64) | 0.00 | (0.63) | 0.00 | (0.40) | 0.00 | (0.25) | 0.00 | (0.46) | 0.00 | (0.46) |
| Si | 2.98 | (0.01) | 2.99 | (0.01) | 3.04 | (0.05) | 2.74 | (0.03) | 2.80 | (0.03) | 3.01 | (0.02) |
| Ti | 0.00 | (0.00) | 0.00 | (0.00) | 0.03 | (0.01) | 0.12 | (0.01) | 0.00 | (0.00) | 0.00 | (0.00) |
| Al | 2.00 | (0.01) | 1.99 | (0.02) | 2.71 | (0.07) | 1.63 | (0.04) | 1.21 | (0.03) | 1.00 | (0.02) |
| Fe ³⁺ | 0.06 | (0.02) | 0.05 | (0.02) | 0.00 | (0.00) | 0.00 | (0.01) | | | | |
| Fe ²⁺ | 1.69 | (0.09) | 1.67 | (0.03) | 0.12 | (0.01) | 1.30 | (0.02) | 0.00 | (0.00) | 0.00 | (0.01) |
| Mn | 0.76 | (0.04) | 0.83 | (0.04) | 0.00 | (0.00) | 0.04 | (0.00) | 0.00 | (0.00) | 0.00 | (0.00) |
| Mg | 0.35 | (0.09) | 0.29 | (0.04) | 0.10 | (0.02) | 1.04 | (0.01) | 0.00 | (0.00) | 0.00 | (0.00) |
| Ca | 0.17 | (0.01) | 0.16 | (0.03) | 0.00 | (0.00) | 0.00 | (0.00) | 0.21 | (0.02) | 0.00 | (0.00) |
| Na | 0.00 | (0.00) | 0.00 | (0.00) | 0.06 | (0.01) | 0.01 | (0.00) | 0.77 | (0.03) | 0.01 | (0.01) |
| K | 0.00 | (0.00) | 0.00 | (0.00) | 0.87 | (0.01) | 0.92 | (0.02) | 0.01 | (0.00) | 0.97 | (0.01) |
| xMg(Fetot) | 0.16 | (0.04) | 0.14 | (0.02) | 0.45 | (0.01) | 0.44 | (0.01) | | | | |
| Al(IV) | | | | | 0.96 | (0.05) | 1.26 | (0.02) | | | | |
| Al(VI) | | | | | 1.75 | (0.03) | 0.37 | (0.03) | | | | |
| Grossular | 0.11 | (0.03) | 0.09 | (0.01) | | | | | | | | |
| Almandine | 0.54 | (0.03) | 0.55 | (0.01) | | | | | | | | |
| Pyrope | 0.05 | (0.00) | 0.05 | (0.01) | | | | | | | | |
| Spessartite | 0.25 | (0.01) | 0.27 | (0.02) | | | | | | | | |
| Anorthite | | | | | | | | | 0.21 | (0.02) | 0.00 | (0.00) |
| Albite | | | | | | | | | 0.78 | (0.03) | 0.01 | (0.01) |
| Orthoclase | | | | | | | | | 0.01 | (0.00) | 0.99 | (0.01) |

Garnet compositions normalized to 12 oxygens, muscovite and biotite compositions normalized to 11 oxygens, plagioclase and feldspar compositions normalized to 8 oxygens, STD standard deviation

Table2

| # | matrix | T(°C) | P(GPa) | time (hrs) |
|----------|--|--------------|---------------|-------------------|
| ANT7 | CaCO ₃ | 740 | 0.8 | 46 |
| ANT9 | CaCO ₃ and SiO ₂ | 760 | 1.0 | 48 |
| ANT8 | SiO ₂ | 780 | 0.8 | 68 |
| ANT10 | SiO ₂ | 780 | 1.0 | 67 |
| ANT5 | CaCO ₃ | 810 | 0.8 | 23 |
| ANT3 | CaCO ₃ | 850 | 0.8 | 22 |
| ANT4 | CaCO ₃ | 900 | 0.8 | 21 |

Table3

| Temp. (No. Analyses) | 740°C | | 760°C | | 780°C | | 810°C | | 850°C | | 900°C | |
|-------------------------------------|--------|------------------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|
| | mean | ^c STD | mean | STD | mean | STD | mean | STD | mean | STD | mean | STD |
| SiO ₂ | 77.28 | (0.97) | 74.13 | (1.55) | 77.48 | (1.00) | 75.52 | (2.30) | 71.58 | (7.96) | 73.54 | (5.23) |
| TiO ₂ | 0.02 | (0.02) | 0.10 | (0.11) | 0.10 | (0.10) | 0.08 | (0.12) | 0.22 | (0.21) | 0.57 | (1.30) |
| Al ₂ O ₃ | 14.13 | (0.37) | 14.83 | (0.63) | 13.22 | (0.40) | 13.49 | (1.14) | 15.59 | (4.38) | 13.63 | (1.20) |
| FeO | 1.72 | (0.28) | 1.69 | (0.40) | 1.92 | (0.23) | 2.59 | (0.83) | 3.81 | (2.12) | 3.25 | (1.65) |
| MnO | 0.63 | (0.18) | 0.57 | (0.24) | 0.60 | (0.09) | 0.76 | (0.25) | 0.75 | (0.34) | 0.85 | (0.28) |
| MgO | 0.12 | (0.06) | 0.11 | (0.05) | 0.19 | (0.14) | 0.20 | (0.10) | 0.64 | (0.66) | 0.21 | (0.12) |
| CaO | 0.86 | (0.26) | 2.18 | (1.16) | 1.24 | (0.71) | 1.29 | (0.59) | 1.50 | (0.91) | 2.44 | (1.99) |
| Na ₂ O | 2.75 | (0.74) | 2.22 | (0.63) | 1.83 | (0.35) | 2.18 | (0.47) | 1.90 | (1.25) | 2.00 | (1.17) |
| K ₂ O | 2.48 | (0.70) | 4.15 | (0.40) | 3.42 | (0.54) | 3.89 | (0.87) | 4.00 | (2.38) | 3.51 | (1.57) |
| total | 100.00 | | 100.00 | | 100.00 | | 100.00 | | 100.00 | | 100.00 | |
| ^a total _{EMPA} | 85.70 | (0.81) | 87.86 | (2.20) | 88.97 | (1.10) | 90.06 | (1.60) | 86.53 | (1.33) | 89.29 | (2.14) |
| ^b H ₂ O(diff) | 14.30 | (0.99) | 12.14 | (1.87) | 11.03 | (1.18) | 9.94 | (1.71) | 13.47 | (5.81) | 10.71 | (2.25) |
| ASI | 1.61 | (0.04) | 1.27 | (0.26) | 1.50 | (0.17) | 1.35 | (0.21) | 1.59 | (0.51) | 1.33 | (0.42) |
| AI | 0.07 | (0.00) | 0.07 | (0.02) | 0.06 | (0.01) | 0.06 | (0.01) | 0.08 | (0.04) | 0.06 | (0.01) |
| MALI | 4.38 | (0.59) | 4.19 | (1.46) | 4.01 | (0.98) | 4.79 | (0.86) | 4.40 | (2.56) | 3.07 | (1.55) |

^a totals from EMPA, ^b H₂O by difference to 100%, ^c standard deviation

Table4

| sample | H ₂ O (wt%) | STD | CO ₂ (ppm) | STD |
|--|------------------------|--------|-----------------------|----------|
| <i>remelting T = 780°C; mean SiO₂ = 68.27 wt. %</i> | | | | |
| ANT8_1_1 | 6.66 | (0.09) | 774.39 | (19.19) |
| ANT8_1_2 | 4.90 | (0.06) | 541.17 | (19.04) |
| ANT8_1_3 | 4.22 | (0.05) | 2019.91 | (39.55) |
| ANT8_1_4 | 6.52 | (0.09) | 1878.96 | (36.74) |
| ANT8_1_5 | 6.48 | (0.09) | 1648.30 | (31.55) |
| ANT8_1_6 | 6.21 | (0.08) | 680.84 | (16.56) |
| ANT8_1_7 | 6.21 | (0.08) | 496.32 | (17.25) |
| mean | 5.89 | (0.94) | 1148.56 | (670.27) |
| ANT8_2_1 | 5.52 | (0.07) | 939.69 | (19.50) |
| ANT8_2_2 | 5.52 | (0.07) | 1148.56 | (22.02) |
| ANT8_2_3 | 4.01 | (0.05) | 849.99 | (18.09) |
| ANT8_2_4 | 4.11 | (0.05) | 1750.82 | (34.39) |
| ANT8_2_3 | 5.63 | (0.07) | 1725.19 | (37.19) |
| mean | 4.96 | (0.82) | 1282.85 | (429.48) |
| ANT10_1_1 | 4.83 | (0.06) | 1128.05 | (22.70) |
| ANT10_1_2 | 2.98 | (0.05) | 710.32 | (17.85) |
| ANT10_1_3 | 4.80 | (0.05) | 624.46 | (17.12) |
| mean | 4.20 | (1.06) | 820.94 | (269.41) |
| ANT10_2_1 | 5.73 | (0.07) | 916.62 | (20.10) |
| ANT10_2_2 | 5.28 | (0.06) | 1661.12 | (34.52) |
| mean | 5.51 | (0.32) | 1288.87 | (526.44) |
| <i>remelting T = 760°C; mean SiO₂ = 65.02 wt. %</i> | | | | |
| ANT9_1_1 | 2.47 | (0.05) | 265.23 | (18.28) |
| ANT9_1_2 | 2.40 | (0.05) | 209.09 | (18.51) |
| ANT9_1_3 | 6.44 | (0.09) | 443.40 | (17.03) |
| ANT9_1_4 | 7.88 | (0.13) | 512.97 | (17.69) |
| ANT9_1_5 | 5.49 | (0.07) | 588.63 | (18.11) |
| ANT9_1_6 | 4.11 | (0.05) | 613.04 | (18.78) |
| ANT9_1_7 | 5.72 | (0.08) | 188.34 | (19.18) |
| mean | 4.93 | (2.04) | 402.96 | (180.26) |
| ANT9_2_1 | 4.21 | (0.05) | 770.47 | (19.33) |
| ANT9_2_2 | 2.50 | (0.05) | 489.78 | (16.15) |
| ANT9_2_3 | 4.87 | (0.06) | 443.40 | (17.42) |
| mean | 3.86 | (1.22) | 567.89 | (176.97) |
| ANT9_3_1 | 2.71 | (0.05) | 296.96 | (17.88) |
| ANT9_3_2 | 2.61 | (0.05) | 228.61 | (18.55) |
| ANT9_3_3 | 5.03 | (0.06) | 248.14 | (19.26) |
| mean | 3.45 | (1.37) | 257.90 | (35.20) |

STD = standard deviation