



HAL
open science

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

Fabio Ferri, Bernardo Cesare, Omar Bartoli, Silvio Ferrero, Rosaria Palmeri, Laurent Remusat, Stefano Poli

► To cite this version:

Fabio Ferri, Bernardo Cesare, Omar Bartoli, Silvio Ferrero, Rosaria Palmeri, et al.. Melt inclusions at MT. Edixon (Antarctica): Chemistry, petrology and implications for the evolution of the Lanterman range. *Lithos*, 2020, 374-375, pp.105685. 10.1016/j.lithos.2020.105685 . hal-02959290

HAL Id: hal-02959290

<https://hal.science/hal-02959290>

Submitted on 24 Dec 2020

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

1 **MELT INCLUSIONS AT MT. EDIXON (ANTARCTICA):**
2
3
4 **CHEMISTRY, PETROLOGY AND IMPLICATIONS FOR THE**
5
6
7 **EVOLUTION OF THE LANTERMAN RANGE**
8
9

10
11
12
13 **Fabio Ferri**¹, Bernardo Cesare^{1*}, Omar Bartoli¹, Silvio Ferrero^{2,3}, Rosaria Palmeri⁴,
14
15
16 Laurent Remusat⁵ and Stefano Poli⁶
17

18
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

26
27 ⁵ Institut de Mineralogie, Physique des Matériaux et Cosmochimie Museum National
28 d'Histoire Naturelle, Paris, France

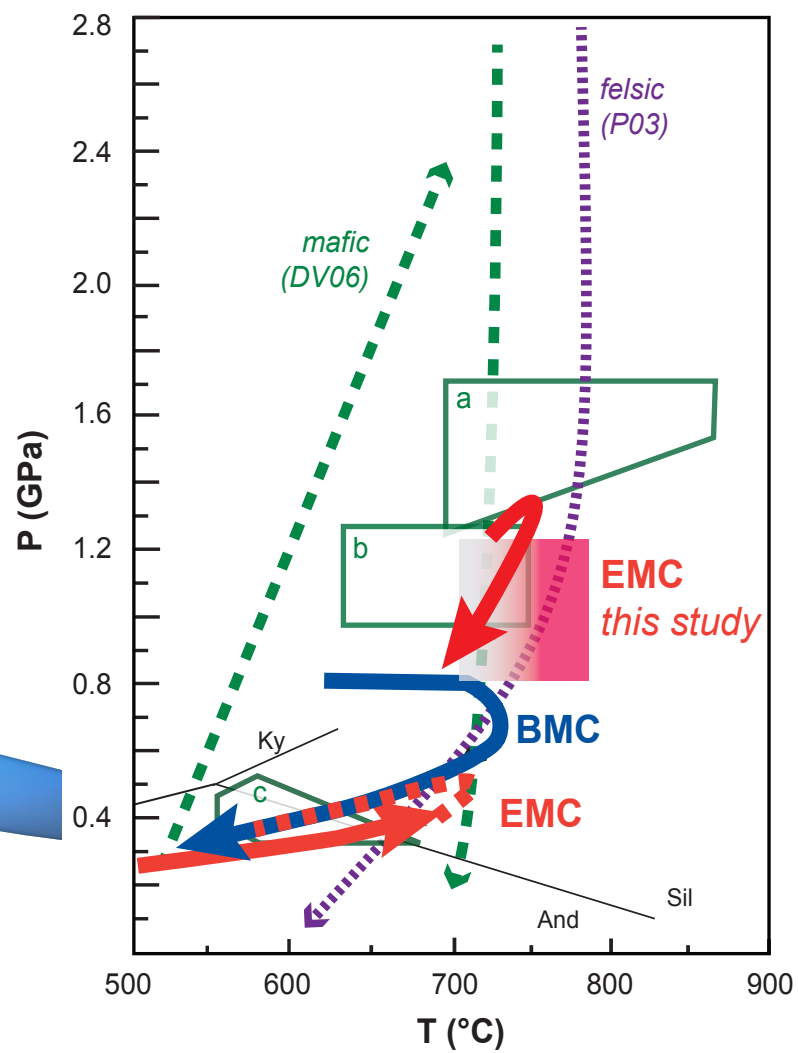
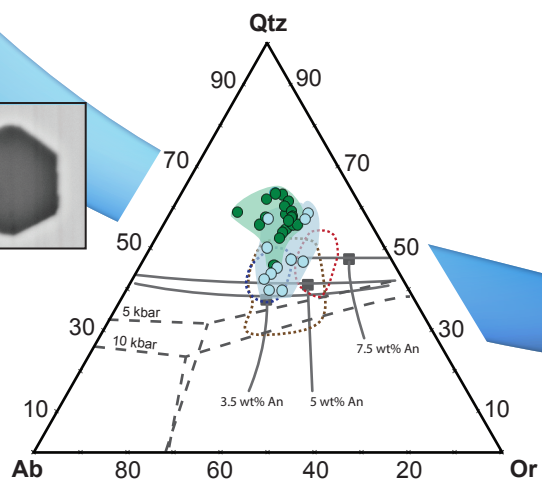
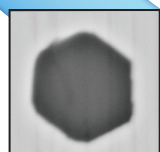
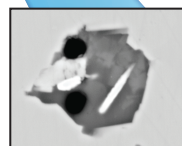
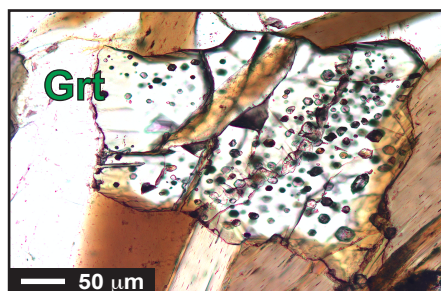
29
30
31 ⁶ Department of Earth Sciences, University of Milan, Italy
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49

20
21
22 * Corresponding author. Tel: +39-049-8279148; Fax: +39-049-8272010.
23

24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65
E-mail address: bernardo.cesare@unipd.it

*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

1
2
3
4 25 The recovery of melt inclusions in peritectic phases of metamorphic rocks brings the
5
6 26 opportunity to retrieve the chemical composition of primary melts and infer melt entrapment
7
8
9 27 conditions. Here we report the first recovery of such inclusions in a medium-grade muscovite-
10
11 28 biotite-plagioclase-K-feldspar-quartz gneiss from the Edixon Metamorphic Complex (EMC)
12
13
14 29 of the Lanterman Range in Northern Victoria Land, Antarctica.

15
16
17 30 The most common inclusions are polycrystalline (PI) and fluid (FI) types coexisting in the
18
19 31 same clusters and displaying frequent negative crystal shape and textures indicative of a
20
21
22 32 primary origin. The PI contain muscovite + quartz + H₂O ± biotite ± calcite ± siderite ± albite
23
24 33 while the FI are composed by CO₂ + H₂O + N₂ + CH₄ ± H₂S. The PI were remelted in a
25
26
27 34 piston cylinder apparatus at conditions of 740-900°C at 0.8-1.0 GPa with complete remelting
28
29 35 taking place at 760-780°C. The remelted glass is sub-alkaline, peraluminous (ASI ≈ 1.3-1.5)
30
31
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
33
34 37 400-1150 ppm of CO₂.

35
36
37
38 38 The stable mineral assemblage is represented by muscovite + plagioclase + quartz + K-
39
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-
46
47 42 730°C.

48
49
50 43 Our study confirms that the EMC at the Lanterman Range is not a homogeneous
51
52 44 metamorphic complex and that anatexis could have taken place at different crustal levels
53
54
55 45 during the evolution of the Ross Orogen. Our results indeed suggest that, in the Lanterman
56
57 46 Range, the high-pressure eclogitic rocks, the medium pressure amphibolite stage and the low-
58
59
60 47 pressure amphibolite stage shared most of the late exhumation path.

48

1

2 49

3

4 50

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

5

6 51 Edixon Metamorphic Complex

7

8

9

10

11

12

13

14

15

16

17

18

19

20

21

22

23

24

25

26

27

28

29

30

31

32

33

34

35

36

37

38

39

40

41

42

43

44

45

46

47

48

49

50

51

52

53

54

55

56

57

58

59

60

61

62

63

64

65

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
3
4 152 Peltier-cooled multichannel CCD detector and coupled with an Olympus BX41 petrographic
5
6 153 microscope. An air-cooled NdYAG laser was used for excitation ($\lambda = 532$ nm, laser power on
7
8 154 the sample of 2–3 mW) with a grating of 300 lines/mm, slit width and confocal hole set to
9
10 155 100 and 200 μm , respectively. The Raman spectra of the solid and fluid phases were acquired
11
12 156 with a 100 \times objective in the wavenumber range 100–4000 cm^{-1} , integrating 3 repetitions of
13
14 157 60 s with spectral resolution of 10 cm^{-1} .
15
16
17
18
19
20

21 159 The composition of mineral phases and glasses was determined using a Jeol JXA 8200
22
23 Superprobe at the Dipartimento di Scienze della Terra, Università di Milano (Italy) using
24 160 natural and synthetic silicates and oxides as standards. Analytical parameters for minerals
25
26 161 were 15 kV accelerating voltage, 5 nA current, counting time of 30 s on peak and 10 s on
27
28 162 background. For glass analyses, the beam diameter was reduced to 1 μm and the counting
29
30 163 time to 10 s on peak and 2 s on background. Sodium, K, Al, and Si were the first analyzed
31
32 164 elements to limit element loss. Concentrations were corrected against leucogranitic glass
33
34 165 standards with comparable H_2O contents (Morgan and London, 2005). Details concerning the
35
36 166 application of correction factors and the composition of the standard glasses are given by
37
38 167 Ferrero et al. (2012) and Bartoli et al. (2013a, 2013b).
39
40
41
42
43
44 168

45 169 The analysis of H and C in glasses was performed using the Cameca Nano Secondary
46
47 170 Ion Mass Spectrometry 50 (NanoSIMS) installed at Muséum National d'Histoire Naturelle in
48
49 171 Paris (France). Polished experimental capsules with MI exposed on the garnet surface and
50
51 172 standard glasses were mounted in Indium (Aubaud et al., 2007). MI were identified by
52
53 173 collecting secondary ion images of Si, K and Fe. For every analysis, we first performed a pre-
54
55 174 sputtering step on a $3 \times 3 \mu\text{m}^2$ surface area for 2 min with a 400 pA primary Cs^+ beam to
56
57
58
59
60
61
62
63
64
65

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.
1
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
3
4 201 (0.03 a.p.f.u.) is typical of amphibolite-facies subsolidus muscovite (e.g., Cesare 1999,
5
6
7 202 compare with Redler et al., 2011). The composition of the white mica is similar to that
8
9
10 203 reported by Di Vincenzo et al. (2001) in sample BT5 and BT2 collected from the Edixon and
11
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
17
18
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
20
21 208 biotite the titanium content is typical of amphibolite-facies subsolidus conditions (Tajčmanová
22
23
24 209 et al., 2009. Plagioclase composition has 15-23 mol. % anorthite, 76-84 mol.% albite and ca. 1
25
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
32
33
34 213 crystalline assemblage was completely melted to a single glassy phase, but melt + vapor
35
36 214 homogenization was not achieved; ‘rehomogenized’ are those inclusions that appear as a
37
38
39 215 homogeneous glass without bubbles (although former accidentally-trapped minerals may be
40
41 216 present). Experimental remelting of the melt inclusions was performed on hand-picked
42
43
44 217 separates of garnets obtained from crushed rock pieces sieved at < 500 μm size. The garnet
45
46 218 fragments were loaded into Au capsules filled with dry powder of silica or chemically pure
47
48
49 219 CaCO_3 . The capsules were loaded in a salt + MgO + graphite assembly and experiments were
50
51 220 performed in single stage piston cylinder apparatus at the Department of Geosciences of the
52
53
54 221 University of Milan at condition from 740°C to 900°C at 0.8 and 1.0 GPa, with duration of
55
56 222 21-68 hours (Table 2). The temperature was controlled by a K-type thermocouple with
57
58
59 223 uncertainty of $\pm 5^\circ\text{C}$. To avoid decrepitation of inclusions, quenching was performed at
60
61
62
63
64
65

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
1
2 248 CO₂ density of < 0.1 g cm⁻¹ (see Wang et al., 2011).
3

4 249 After the experiments, the inclusions were carefully inspected to infer the temperature
5
6
7 250 of entrapment from approach to remelting equilibrium (Fig. 4). At 740°C, the MI have well
8
9 251 defined negative crystal shapes and contain glass (yellow arrows in Fig. 4) together with
10
11 252 abundant unmelted crystals of quartz, biotite, muscovite, albite and ilmenite, indicating that
12
13 253 remelting was incomplete (i.e., they are partially-remelted inclusions). At 760°C, albite and
14
15 254 muscovite are completely resorbed, biotite is less frequent, and quartz is still common with
16
17 255 typical round shape, suggesting partial resorption. Few inclusions are completely remelted to
18
19 256 a homogeneous glass with presence of a shrinkage bubble. At 780°C experiments show the
20
21 257 best approach to equilibrium with > 40-50% remelted inclusions, displaying regular walls,
22
23 258 negative crystal shape and absence of offshoots. The presence of resorbed quartz is rare. At
24
25 259 810°C, completely or partly remelted inclusions display offshots (white arrows in Fig. 4) and
26
27 260 the presence of secondary minerals (i.e., formed during experiments or upon quenching) as
28
29 261 aggregates of tiny acicular Bt. Remelted inclusions represent > 70% of the total. At 850°C
30
31 262 and 900°C almost all inclusions are remelted or rehomogenized but show evidence of
32
33 263 overheating (see Cesare et al., 2015; Bartoli et al., 2013c; Acosta-Vigil et al., 2016) as
34
35 264 indicated by i) presence of lobate-cusate walls at the contact with the host garnet, ii) frequent
36
37 265 offshoots, iii) growth of secondary garnet at the inclusion walls, iv) segmented straight walls
38
39 266 after assimilation of the Grt (at 900°C). Ilmenite and rutile as trapped phases become less
40
41 267 common at T > 810°C and almost absent at 900°C. In the Raman spectrum shown in Figure
42
43 268 3C, the remelted glass has a broad band at ≈ 3400 cm⁻¹ related to the presence of H₂O, while
44
45 269 the shrinkage bubble within the glass contains a fluid mixture of CO₂ + N₂ + CH₄ (Fig. 3B).
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

270 4.3 Major elements composition of remelted glasses

1
2 271 In order to determine the composition of the melt produced during anatexis of the
3
4
5 272 metapsammite, and to infer the temperature of entrapment of inclusions, we conducted EMP
6
7 273 glass analyses on completely or partially remelted glasses to explore their potential interaction
8
9
10 274 with the host garnet as function of experimental temperature (Table 3). We excluded analyses
11
12 275 with clear signal of contamination either by minerals preserved in the inclusions or by host
13
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 $(\text{FeO}_t + \text{MgO} + \text{TiO}_2) > 7$
15
16 277 wt\% . The difference to 100% of the totals may reflect not only the fluid content but also
17
18
19 278 interaction of the microprobe beam with the shrinkage bubble if present below the polished
20
21
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
23
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
27
28
29 282 respectively 0.5 wt% and 1.6-1.7 wt% up to 780°C, and systematically increases from 810°C
30
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%
32
33
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 $\text{FeO}_t + \text{MgO} +$
35
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.
37
38
39 286 SiO_2 has similar behavior for most temperatures with the SiO_2 increasing from 760°C to
40
41 287 780°C and decreasing from 780°C to 810°C, 850°C and 900°C (Fig. 5C). No temperature
42
43
44 288 dependence is visible on the Na_2O vs. SiO_2 plot, except at 900°C where the SiO_2 is
45
46 289 systematically lower for comparable Na_2O contents. The Al_2O_3 is inversely proportional to
47
48
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
50
51 291 the exception of few inclusions remelted at 850-900°C. The systematic variations reported in
52
53 292 Fig. 5A-C and E may result from contamination of melt by mixtures of Grt-Qz and Feld-Qz
54
55
56 293 (see directions of exchange vectors reported in Fig. 5B and E) suggesting that host interaction

294 is not negligible at $T > 810^{\circ}\text{C}$. As a consequence, we consider the glass compositions obtained
1
2 295 at $760\text{-}780^{\circ}\text{C}$ as most representative of the anatectic melt trapped within garnets.
3

4 296 In Fig. 6 the composition of glasses remelted at $760\text{-}780^{\circ}\text{C}$ is plotted on the
5
6
7 297 pseudoternary normative diagrams Qz-Ab-Or and An-Ab-Or. Compositions at 760°C include
8
9 298 remelted glasses with and without daughter quartz. The MI cluster above the cotectic lines at
10
11 299 $0.5\text{-}1.0\text{ GPa}$, which is consistent with the displacement of cotectic lines towards Qz apex as
12
13 300 effect of increasing Al and Ca content in the granite model system (see Bartoli et al., 2016).
14
15 301 The composition of MI at 760°C differ by lower SiO_2 which is related to the widespread
16
17 302 presence of relict unmelted quartz in the inclusions. However, few glasses at 760°C overlap
18
19 303 with glasses at 780°C suggesting that equilibrium melt composition are presumably achieved
20
21 304 within the interval $760\text{-}780^{\circ}\text{C}$ and represent the closest approximation to the original
22
23 305 anatectic melt. Displacement of the MI composition from the minimum melt in the Qz-Ab-Or
24
25 306 model system has been explained by Wilke et al. (2017) as due to the presence of extra
26
27 307 components in natural systems, mostly Ca, Fe and Ti. Wilke et al. (2017) report a
28
29 308 displacement of the eutectic point to Or-enriched composition at similar Qz amount when
30
31 309 melts have $\approx 3.5\text{-}7\%$ normative An. Such An content corresponds to that of remelted glasses
32
33 310 clustering at $2\text{-}7\%$ with few analyses up to $11\text{-}16\%$. The MI show variation in the Qz/feldspar
34
35 311 ratio as indicated by their distribution along the directories $\text{Ab/Or} = 0.38$ and $\text{Ab/Or} = 0.5$
36
37 312 collinear with the Qz apex. Such trends differ from those of Bartoli et al. (2016) who reported
38
39 313 compositional alignments of remelted granitic glasses parallel to the Qz-feldspar join and
40
41 314 interpreted them as due to slow diffusion of Si and Al compared to alkalis in granitic melts
42
43 315 (Acosta-Vigil et al., 2006, 2017). In the An-Ab-Or projection, most glass compositions plot in
44
45 316 the granite field close to the boundary with Qz-monzonite, with few Qz-monzonite
46
47 317 compositions. Figure 6 also reports the composition of MI recovered in garnet of
48
49 318 metasediments from Ojén (Bartoli et al., 2013b, 2013c) and in plagioclase of Grt-Bt-Sil
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

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

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

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

361 The petrographic study shows that the stable mineral assemblage is represented by Ms
362 + Pl + Kf + Grt + Qz + Liq ± Rt coexisting with excess C-O-H fluid as indicated by the
363 presence of FI and PI within the same cluster. In the pseudosection of Fig. 9, such assemblage
364 is stable at P > 0.5 GPa within a triangular field limited by the fluid-present solidus curve and
365 by the Ms-out/Als-in curve to temperatures of 690°C < T < 820°C. The Fig. 10
366 (Supplementary Material) reports the calculated compositional isopleths for garnet,
367 plagioclase and biotite. In larger garnets from cores to rims, the X_{Pyp} slightly increases and
368 X_{Sps} slightly decreases suggesting a ≈ 50°C temperature increase during the garnet growth,

369 while smaller garnets are homogeneous with composition similar to that of large garnet rims.
1
2 370 The X_{Alm} in garnet has a slightly reversed tendency (higher temperature in the core and lower
3
4 371 at the rim). The combination of garnet isopleths with the X_{An} in plagioclase, delimit a P-T
5
6
7 372 space corresponding to approximately 680°C to 730°C at 1.0-1.3 GPa (Fig. 9). The low Ti
8
9 373 content in biotite (0.12 a.p.f.u.) is not compatible with such temperatures and supports the
10
11
12 374 petrographic observation of post peak recrystallization of biotite at $T < 600^\circ\text{C}$.

13
14 375 The calculated isomodes for the phases biotite, garnet and melt reported in Fig. 9B
15
16 376 and muscovite (nor reported for clarity) indicate that biotite and muscovite are partially
17
18
19 377 consumed while limited amount of garnet is formed during progressive metamorphism at P-T
20
21
22 378 evolving from condition 1 to conditions 2 in Fig. 9. Modelled muscovite decreases from 12 to
23
24 379 8 wt%, biotite from 22 to 20 wt% and garnet increases from 1.5% to 2-4 wt%. The melt
25
26 380 produced is limited to 2-4 wt%. The calculated isomodes are in agreement with the modal
27
28
29 381 phase proportions constrained by mass balance calculation and corresponding to 7.9 wt% of
30
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**

39
40 385 The present study is an example of successful remelting of inclusions that adds to
41
42
43 386 previous case studies of metasedimentary rocks undergoing anatexis (e.g., Cesare et al, 2015;
44
45 387 Bartoli et al., 2016; Tacchetto et al., 2019; Carvalho et al., 2018). Previous works report MI
46
47
48 388 entrapped under fluid present (Bartoli et al., 2013b) or fluid-absent melting of felsic (e.g.,
49
50 389 Cesare et al., 2015) or ultramafic rocks (Ferrero et al., 2018), and in a range of temperatures
51
52
53 390 from upper-amphibolite to UHT conditions. The present case study yields new data on the
54
55 391 onset of low-T anatexis of metasedimentary crust under possibly fluid-present conditions,
56
57
58 392 which are likely to be more common than previously thought (see Carvalho et al., 2018 and
59
60 393 discussion therein).

394 The P-T-fluid conditions of melt entrapment within the host mineral can be
1
2 395 reconstructed combining the microstructures produced during remelting experiments at
3
4 396 increasing temperature with the proportion of remelted inclusions, the compositional variation
5
6
7 397 of major elements in glass and the thermodynamic modelling. In remelting experiments,
8
9 398 microstructural changes include i) type of minerals coexisting with glass (daughter minerals
10
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
13
14 400 new phases (indicative overheating), v) appearance of zoning/growth structures (indicative of
15
16 401 intense host interaction). Since glasses remelted at 780° are mostly free of daughter minerals,
17
18 402 have no visible cracks/offshoots, and display limited or no chemical interaction with the host
19
20 403 garnet (e.g., low MnO + FeO_t in Fig. 5), they can be confidently interpreted as representative
21
22 404 of entrapment conditions. However, at 760°C, where most inclusions have quartz which has
23
24 405 not been re-homogenized and hence lower SiO₂, few remelted inclusions resemble in texture
25
26 406 and composition those typically recovered at 780°C. One may question about the effect of
27
28 407 kinetics on remelting conditions and the counter play of temperature and run duration on the
29
30 408 achievement of equilibrium for remelting. Preliminary experiments at 780°C performed at
31
32 409 different durations (hours to days, unpublished data) seem to confirm that run duration has an
33
34 410 effect on remelting and that the chemical signature of the glasses recovered at lower
35
36 411 experimental duration partly overlaps (though not completely) with the composition of
37
38 412 homogeneous glasses recovered at 760°C. Even though the kinetics dilemma is out of the
39
40 413 scope of this work, it is reasonable to consider the interval 760-780°C at 0.8-1.0 GPa as
41
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
45
46 416 the wet solidus of the rock locates at ≈690-710°C between 0.5 GPa and 1.5 GPa, and that the
47
48 417 stability field of the assemblage Ms + Pl + Kf + Grt + Qz + Liq ± Rt is between 700-780°C.
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

1
2 419 The temperature intervals indicated by compositional isopleths of Grt and Pl and the mineral
3
4 420 isomodes (Fig. 9) are more consistent with $\approx 700\text{-}730^\circ\text{C}$ and $\approx 730\text{-}760^\circ\text{C}$ respectively. The
5
6 421 discrepancy between calculated and experimental conditions is maximum $60\text{-}80^\circ\text{C}$ and can be
7
8 422 ascribed to diverse sources of uncertainty in phase equilibria calculations, such as the quality
9
10 423 of thermodynamic data and, in turn, the quality of mineral and melt solution models (White et
11
12 424 al., 2011). For example, the current melt model requires improvements to reproduce properly
13
14 425 natural processes (see discussion in Bartoli et al., 2013b), as already pointed out in previous
15
16 426 works (e.g., Bartoli et al., 2013a and 2016; Grant, 2009 and White et al., 2011).

19 427 **5.2 Fluid regime and volatile content of the primary melt**

21
22 428 There are few lines of evidence suggesting that partial melting of the investigated
23
24 429 rocks occurred in the presence of a fluid phase, including i) the coexistence of primary FI and
25
26 430 PI within the same cluster, and ii) the major element composition of primary melt.

28
29 431 The abundant FI coexisting with PI in the same cluster indicate the presence of a C-O-
30
31 432 H fluid dominated by H_2O and CO_2 . Graphite was not observed within the FI but carbonates
32
33 433 (calcite and siderite) are common within garnets. Recently Carvalho et al. (2018) discussed
34
35 434 the recovery of C-O-H bearing FI in metasedimentary rocks from Ivrea at amphibolite to
36
37 435 granulite conditions and argued that “a (graphitic) rock undergoing melting will always show
38
39 436 the presence of a free C-O-H fluid, in variable amounts and with variable $a_{\text{H}_2\text{O}}$ depending on
40
41 437 pressure, temperature and the bulk composition of the system”. According to Carvalho et al.
42
43 438 (2018), the fluid regime during anatexis might be characterized by CO_2 -dominated fluid in
44
45 439 processes dominated by dehydration of OH-bearing phases. In the Edixon Complex, where
46
47 440 the fluid was likely H_2O dominated, conditions of entrapment are compatible with the
48
49 441 beginning of partial melting and potentially with both muscovite and biotite fluid-present
50
51 442 conditions. The calculated isomodes suggest that muscovite and biotite are consumed at the
52
53 443 solidus to produce limited amounts of melt and peritectic garnet, always in the absence of
54
55
56
57
58
59
60
61
62
63
64
65

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
3
4 446 assemblage is more consistent with a muscovite + biotite fluid-present melting reaction
5
6
7 447 (Weinberg and Hasalova, 2015). Increasing $a_{\text{H}_2\text{O}}$ should reduce garnet stability (Conrad et al.,
8
9 448 1988; Gardien et al., 2000), but the experiments of Alonso-Perez et al. (2009) on Si-rich melts
10
11 449 indicated the opposite, i.e. that stability of peritectic garnet increases with increasing H_2O in
12
13
14 450 melt.

15
16
17 451 The H_2O content of glasses at 760°C is more scattered ($\approx 2\text{-}8$ wt%) than at 780°C ($\approx 4\text{-}$
18
19 452 7 wt%) but in both cases it is close to the values predicted for H_2O saturation at 0.8 GPa for
20
21 453 haplogranitic melts (≈ 7.6 wt%; Holtz et al., 2001, Fig. 8D). The lower H_2O contents of some
22
23
24 454 glasses may be explained as effect of heterogeneous distribution of the fluid (see discussion in
25
26 455 Bartoli et al., 2014) and/or by the presence of a diluting carbonic species in the COH fluid,
27
28
29 456 namely CO_2 . The H_2O content of glasses determined as difference to 100 from EMPA
30
31 457 analyses may yield similar contents, though Acosta-Vigil et al. (2016) report deviations up to
32
33
34 458 30%. In Table 3, the average difference to 100 total decreases with increasing experimental
35
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-
37
38
39 460 900°C . All values overestimate the measured H_2O in glasses. Possible explanations may
40
41 461 account for the presence of light and incompatible elements as B, which is abundant in first
42
43 462 anatexis products or micropores and would be consistent with the presence of tourmaline in
44
45
46 463 the rock. Concerning other volatiles, the Cl and F content of the glasses was below the
47
48
49 464 detection limits (Table 3). This inconsistency requires further investigation.

50
51 465 The maximum CO_2 content of the remelted glasses from NanoSIMS (with all C
52
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
54
55
56 467 difference may be explained considering that i) the CO_2 solubility in felsic melts slightly
57
58 468 increases with increasing T at least at 0.5 GPa (Tamic et al., 2001), ii) the CO_2 solubility and
59
60
61
62
63
64
65

469 speciation are affected by the glass composition, with total CO₂ increasing with increasing
1
2 470 SiO₂ and decreasing with increasing Na₂O, and molecular CO₂ prevailing on the (CO₃)²⁻
3
4 471 species in SiO₂ rich- and Na₂O poor- compositions (Brooker et al., 1999). At 780°C, glasses
5
6 472 have higher SiO₂-CaO and lower Na₂O with respect to glasses produced at 760°C, so the CO₂
7
8 473 speciation is expected to be dominated by molecular CO₂. Even though most of the analyzed
9
10 474 MI appeared homogeneous and bubble-free on polished surface, the presence of shrinkage
11
12 475 bubbles deep within the inclusion cannot be excluded. Since the fluid phase preserved in the
13
14 476 shrinkage bubble is mostly CO₂-rich (Fig. 3B) in agreement with Wallace et al. (2015), this
15
16 477 may account for an additional CO₂ reservoir. If (slightly) higher resorption of the shrinkage
17
18 478 bubbles takes place at increasing temperature, this may explain the higher CO₂ content of
19
20 479 glasses remelted at 780°C vs. 760°C.
21
22
23
24
25

26 480 Granitic melt produced by water-present melting of metasedimentary rocks are
27
28 481 expected to have higher ASI values than those produced by muscovite or biotite dehydration
29
30 482 melting (Acosta-Vigil et al., 2003), to be enriched in Na₂O and CaO (Weinberg and Hasalova,
31
32 483 2015), and to contain higher proportion of MgO and FeO at increasing water content or water
33
34 484 activity (e.g. Conrad et al., 1988; Holtz and Johannes, 1991) at constant P–T (Patiño Douce
35
36 485 and Beard, 1996). This is partly consistent with the glass composition at 780°C, with ASI
37
38 486 value of 1.57, MALI (= Na₂O + K₂O - CaO) value of $\approx 4.0 \pm 1.0$ wt%, CaO content of ≈ 1.27
39
40 487 ± 0.7 wt.% and FeO_{tot} + MgO > 2.1 wt%, supporting the hypothesis of water-saturated partial
41
42 488 melting.
43
44
45
46
47

48 489 **5.3 Anatexis and evolution of the Edixon Metamorphic Complex**

49
50 490 The tectonometamorphic history of the Edixon Metamorphic Complex and its role in
51
52 491 the evolution of the Wilson Terrane has been primarily constrained by investigations along
53
54 492 the western Lanterman Range but also southward within a wider belt including Salamander
55
56 493 Range and Niagara Icefalls-Mountaineer Range regions which are exposed, as the Lanterman
57
58
59
60
61
62
63
64
65

494 Range, close to the Wilson-Bowers tectonic boundary (e.g., Ghiribelli, 2000; Talarico et al.,
1
2 495 1998; 2004). In general, the EMC contains no record of an early HP metamorphic stage, but
3
4 496 preserves a broad metamorphic zonation from low/medium grade to high grade up to anatexis.
5
6
7 497 However, previous studies showed the presence of rocks with contrasting P-T paths in
8
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
12
13 500 Range, that the rocks of the EMC show close similarities in lithological assemblages and
14
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
18
19 503 Grt-Crd migmatites outcropping at the Deep Freeze Range (Palmeri et al. 1991, 1994;
20
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
24
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
30
31 509 event and, at the same time, promoted the juxtaposition of migmatites with contrasting P-T
32
33 510 paths likely formed during the same Ross orogenic cycle (Palmeri, 1997).
34
35
36
37
38
39
40

41 511 In other localities, as at the Mountaineer Range region, a different P-T path has been
42
43 512 proposed (Castelli et al., 2003; Talarico et al., 1998), involving an early decompression path
44
45 513 following a moderate thickening, magmatic intrusions between 500 to 480 Ma (the Granite
46
47 514 Harbour intrusives), and a final cooling-unloading path has been inferred (Castelli et al.,
48
49 515 2003).
50
51
52

53 516 In this paper, the MI preserved in peritectic garnet and phase equilibria modelling
54
55 517 demonstrate that the portion of the EMC close to Mt. Edixon experienced anatexis above
56
57 518 700°C and up to 760-780°C at ≈ 1.0 GPa, corresponding to the P-T conditions of entrapment.
58
59
60
61
62
63
64
65

1
2 519 In the NW part of the Lanterman Range (Carnes Crag) Grt-Crd migmatites yield T-P
3 520 conditions at $\approx 700^\circ\text{C}$ and $\approx 0.5\text{ GPa}$ (Ghiribelli, 2000). Different conditions, however, were
4
5 521 found in the southern portion of the EMC (Mountaineer Range), where the highest
6
7 522 metamorphic grade shows temperature at $\approx 700\text{-}750^\circ\text{C}$ and pressure from ≈ 0.5 up to 0.6 GPa
8
9 523 (Castelli et al., 2003). The spread in low to medium pressure rock types and contrasting P-T
10
11 524 paths at the boundary with the Wilson-Bowers terranes is not uncommon (Ricci et al., 1997;
12
13 525 Ricci and Tessensohn, 2003), and supports the interpretation that anatexis could have taken
14
15 526 place at different crustal levels during the evolution of the tectono-metamorphic stages of the
16
17 527 Ross Orogen.

18
19 528 The latter is in keeping with the multiple and diachronous generations of white micas
20
21 529 evidenced by Di Vincenzo et al., (2001) in the different metamorphic complexes of the
22
23 530 Lanterman Range. The different white micas have been interpreted as testifying different P-T
24
25 531 deformation stages developed during the evolution of the three metamorphic complexes (Di
26
27 532 Vincenzo et al., 2001).

28
29 533 The white micas in the investigated sample GO5 have no evidence of retrogression,
30
31 534 they are texturally and compositionally similar to type-2 white mica of Di Vincenzo et al.,
32
33 535 (2001), in fact they grow along the main amphibolite-facies foliation, and show a light
34
35 536 enrichment in celadonitic component ($\text{Si} = 3.1\text{ a.p.f.u.}$ and $\text{Ti} = 0.022\text{ a.p.f.u.}$). Moreover, data
36
37 537 collected in this paper indicate that white micas are compatible with fluid-present melting as
38
39 538 reconstructed by combination of thermodynamic modelling and MI remelting experiments.
40
41 539 Since the composition of white micas in the Wilson Terrane evolves with time and the type-2
42
43 540 mica dates to ca. $480\text{-}490\text{ Ma}$, we infer that the latter age could be referred to melting in the
44
45 541 Lanterman Range.

46
47 542 Even though it is not possible to reconstruct a proper P-T-relative time path based on
48
49 543 the single sample GO5, the equilibrium conditions reconstructed from the remelting
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

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
3
4 546 Lanterman Range is not an homogeneous metamorphic complex characterized by a low-
5
6
7 547 pressure metamorphic imprint as postulated previously (Ghiribelli, 2000) but, similar to the
8
9 548 HP-UHP belt (GHMC), it represents a mélange of different crustal slices; ii) the outcrop,
10
11 549 from which the study sample comes from could be the extension towards the southern
12
13 550 Lanterman Range of the intermediate-pressure belt of the BMC. Whatever the solution of the
14
15 551 latter problem is, combining (Fig.10) geochronological data collected at the Lanterman Range
16
17 552 (Di Vincenzo et al., 2016 and references therein), previous P-T paths (Di Vincenzo et al.,
18
19 553 2001; Ghiribelli 2000; Talarico et al., 2004) and the results for sample GO5, the docking
20
21 554 between the BMC and/or the different tectonic slices of the EMC might have been before 480
22
23 555 Ma. The latter age is widespread in the area and it was interpreted as the Early Ordovician
24
25 556 final “closure” of micas at the Lanterman Range (Di Vincenzo et al., 2001) and mainly
26
27 557 referred to the exhumation process of the different complexes, whereas in the inner Wilson
28
29 558 Terrane that age correspond to the youngest undeformed calc-alkaline granitoids.
30
31
32
33

34 559 Thus, in the Lanterman Range, the high-pressure eclogitic rocks (GHMC), the
35
36 560 medium pressure amphibolite stage (BMC/EMC pro parte) and the low-pressure amphibolite
37
38 561 stage (EMC pro parte) shared most of the late exhumation path.
39
40
41
42
43
44

45 562 **ACKNOWLEDGEMENTS**

46
47
48 563 This work was funded by the CaRiPaRo project Makearth (Cassa di Risparmio di
49
50 564 Padova e Rovigo) to FF, the grant SIR RBSI14Y7PF (Italian Ministry of Education,
51
52 565 University, Research) to OB, the PNRA (PdR 13/B2.07) to RP and by the German Federal
53
54 566 Ministry for Education and Research and the Deutsche Forschungsgemeinschaft (Project FE
55
56 567 1527/2-1 and FE 1527/2-2) to SF.
57
58
59
60
61
62
63
64
65

568 Rock sample used in this study was collected by Gianni Capponi e Giacomo Oggiano
1
2 569 and is available at the PNRA (Programma Nazionale Ricerche in Antartide) rock repository
3
4
5 570 located at the Museo Nazionale dell'Antartide (Siena, Italy).
6
7

8 571 REFERENCES

10
11 572 Acosta-Vigil, A., London, D., Morgan, G. B., VI, Dewers, T. A. (2003). Solubility of
12
13
14 573 excess alumina in hydrous granitic melts in equilibrium with peraluminous minerals at 700–
15
16 574 800°C and 200MPa, and applications of the aluminum saturation index. Contributions to
17
18
19 575 Mineralogy and Petrology, 146, 100–119.
20

21 576
22
23 577 Acosta-Vigil, A., London, D., Morgan, G.B. VI (2006). Experiments on the kinetics of
24
25
26 578 partial melting of a leucogranite at 200 MPa H₂O and 690–800°C: Compositional variability
27
28 579 of melts during the onset of H₂O-saturated crustal anatexis. Contributions to Mineralogy and
29
30
31 580 Petrology, 151, 539–557.
32

33 581
34
35 582 Acosta-Vigil, A., Cesare, B., London, D., Morgan, G. B., VI (2007). Microstructures
36
37
38 583 and composition of melt inclusions in a crustal anatexis environment, represented by
39
40
41 584 metapelitic enclaves within El Hoyazo dacites, SE Spain. Chemical Geology 235, 450-465.
42

43 585
44
45 586 Acosta-Vigil, A., Buick, I., Hermann, J., Cesare, B., Rubatto, D., London, D., Morgan,
46
47
48 587 G. B., VI (2010). Mechanisms of crustal anatexis: a geochemical study of partially melted
49
50
51 588 metapelitic enclaves and host dacite, SE Spain. Journal of Petrology 51, 785-821.
52

53 589
54
55 590 Acosta-Vigil, A., Barich, A., Bartoli, O. et al. (2016). The composition of
56
57
58 591 nanogranitoids in migmatites overlying the Ronda peridotites (Betic Cordillera, S Spain): the
59
60
61
62
63
64
65

592 anatectic history of a polymetamorphic basement. *Contributions to Mineralogy and Petrology*,
1
2 593 171, 24. <https://doi.org/10.1007/s00410-016-1230-3>
3
4
5 594
6
7 595 Acosta-Vigil, A., London, D., Morgan, G. B., VI, Cesare, B., Buick, I., Hermann, J.,
8
9
10 596 Bartoli, O. (2017). Primary crustal melt compositions: Insights into the controls, mechanisms
11
12 597 and timing of generation from kinetics experiments and melt inclusions. *Lithos*, 286-287,
13
14 598 454-479. doi:10.1016/j.lithos.2017.05.020
15
16
17 599
18
19 600 Alonso-Perez, R., Müntener, O., Ulmer, P. (2008). Igneous garnet and amphibole
20
21 601 fractionation in the roots of island arcs: experimental constraints on andesitic liquids.
22
23
24 602 *Contributions to Mineralogy and Petrology*, 157, 541-558. 10.1007/s00410-008-0351-8.
25
26
27 603
28
29 604 Aubaud, C., Withers, A. C., Hirschmann, M. M., Guan, Y., Leshin, L. A., Mackwell,
30
31 605 S. J., Bell, D. R. (2007). Intercalibration of FTIR and SIMS for hydrogen measurements in
32
33
34 606 glasses and nominally anhydrous minerals. *American Mineralogist*, 92, 811-828.
35
36 607
37
38
39 608 Bartoli, O., Cesare, B., Poli, S., Acosta-Vigil, A., Esposito, R., Turina, A., Bodnar,
40
41 609 R.J., Angel, R.J., Hunter, J. (2013a). Nanogranite inclusions in migmatitic garnet: behavior
42
43
44 610 during piston cylinder re-melting experiments. *Geofluids* 13, 405–420.
45
46 611
47
48
49 612 Bartoli, O., Cesare, B., Poli, S., Bodnar, R.J., Acosta-Vigil, A., Frezzotti, M.L., Meli,
50
51 613 S. (2013b). Recovering the composition of melt and the fluid regime at the onset of crustal
52
53
54 614 anatexis and S-type granite formation. *Geology*, 41, 115–118. CONTROLLARE 2013 abc
55
56 615
57
58
59
60
61
62
63
64
65

616 Bartoli, O., Tajčmanová, L., Cesare, B., Acosta-Vigil, A., (2013c). Phase equilibria
1
2 617 constraints on melting of stromatic migmatites from Ronda (S. Spain): insights on the
3
4 618 formation of peritectic garnet. *Journal of Metamorphic Geology*, 31, 775–789.
5
6
7 619
8
9 620 Bartoli, O., Acosta-Vigil, A., Ferrero, S., and Cesare, B. (2016). Granitoid magmas
10
11 621 preserved as melt inclusions in high-grade metamorphic rock. *American Mineralogist*, 101,
12
13 622 1543-1559.
14
15
16 623
17
18
19 624 Bartoli, O., Cesare, B., Remusat, L., Acosta-Vigil, A., Poli, S. (2014). The H₂O
20
21 625 content of granite embryos. *Earth and Planetary Science Letters*, 395, 281-290.
22
23
24 626 doi:10.1016/j.epsl.2014.03.031
25
26 627
27
28
29 628 Bartoli, O. (2017). Phase equilibria modelling of residual migmatites and granulites:
30
31 629 an valuation of the melt-reintegration approach. *Journal of Metamorphic Geology*, 35, 919–
32
33 630 942.
34
35
36 631
37
38
39 632 Behrens, H., Jantos, N. (2001) The effect of anhydrous composition on water
40
41 633 solubility in granitic melts *American Mineralogist*, 86, 14-20.
42
43 634
44
45
46 635 Bomparola, R.M., Belousova, E., Ghezzi, C., Griffin, W.L., O'Reilly, Y.O. (2006).
47
48 636 Resetting of the U–Pb zircon system in Cambro-Ordovician intrusives of the Deep Freeze
49
50 637 Range, northern Victoria Land, Antarctica. *Journal of Petrology*, 48, 327–364.
51
52
53 638
54
55
56 639 Borg, S. G., De Paolo, D. J., and Smith, B. M. (1990). Isotopic structure and tectonics
57
58 640 of the central Transantarctic Mountains. *Journal of Geophysical Research*, 95, 6647-6667.
59
60
61
62
63
64
65

641

642 Bradshaw, J. D., Laird, M.G. (1983). The pre-beacon geology of Northern Victoria

643 Land. In: Oliver, R.L., James, P.R., Jago, J.B. (Eds.), Antarctic Earth Science. Australian

644 Academy of Science, Canberra, pp. 98–101.

645

646 Brooker, R. A., Kohn, S. C., Holloway, J. R., McMillan, P. F., Carroll, M. R. (1999).

647 Solubility, speciation and dissolution mechanisms for CO₂ in melts on the NaAlO₂–SiO₂ join.

648 *Geochimica et Cosmochimica Acta*, 63, 21, 3549–3565.

649

650 Bureau, H., Trocellier, P., Shaw, C., Khodja, H., Bolfan-Casanova, N., and

651 Demouchy, S., (2003). Determination of the concentration of water dissolved in glasses and

652 minerals using nuclear microprobe. *Nucl. Instrum. Methods Phys. Res., Sect. B* 210, 449–

653 454.

654

655 Capponi, G., Crispini, L., Meccheri, M. (1999). Structural history and tectonic

656 evolution of the boundary between the Wilson and Bowers Terranes, Lanterman Range,

657 Northern Victoria Land, Antarctica. *Tectonophysics*, 312, 249–266. [http://dx.doi.org/10.](http://dx.doi.org/10.1016/S0040-1951(99)00174-2)

658 [1016/S0040-1951\(99\)00174-2](http://dx.doi.org/10.1016/S0040-1951(99)00174-2).

659

660 Capponi G., Castorina F., Di Pisa A., Meccheri M., Petrini R., Villa, I. M.,

661 (2002). The meta-igneous rocks of the Barber Glacierarea (northern Victoria Land,

662 Antarctica): a clue to the enigmatic Borchgrevink Orogeny? In: Gamble J., Skinner D.N.B.

663 and Henrys S. (eds.), *Antarctica at the close of a millennium*. Royal Society of New

664 Zealand Bulletin, 35, 99-104.

665

666 Carvalho, B. B., Bartoli, O., Ferri, F., Cesare, B., Ferrero, S., Remusat, L., Capizzi,
1
2 667 L.S., Poli, S. (2019). Anatexis and fluid regime of the deep continental crust: New clues from
3
4 668 melt and fluid inclusions in metapelitic migmatites from Ivrea zone (NW Italy). *Journal of*
5
6
7 669 *Metamorphic Geology*, 37, 951-975. doi:10.1111/jmg.12463
8

9 670
10
11
12 671 Cawood, P. (2005). Terra Australis Orogen: Rodinia breakup and development of the
13
14 672 Pacific and Iapetus margins of Gondwana during the Neoproterozoic and Paleozoic. *Earth-*
15
16
17 673 *Science Reviews*, 69, 249–279.
18

19 674
20
21
22 675 Cesare, B. (1999). Multi-stage pseudomorph replacement of garnet during
23
24 676 polymetamorphism: 1. microstructures and their interpretation. *Journal of Metamorphic*
25
26
27 677 *Geology*, 17, 723-734. doi:10.1046/j.1525-1314.1999.00229.x
28

29 678
30
31
32 679 Cesare, B., Acosta-Vigil, A., Bartoli, O., Ferrero, S. (2015). What can we learn from
33
34 680 melt inclusions in migmatites and granulites? *Lithos*, 239, 186–216.
35

36 681
37
38
39 682 Castelli, D., Oggiano, G., Talarico, F., Belluso, E., and Colombo, F. (2003). Mineral
40
41 683 chemistry and petrology of the Wilson Terrane Metamorphics from Retreat Hills to Lady
42
43
44 684 Newnes Bay, northern Victoria Land, Antarctica. In: Ricci C.A. and Tessensohn F. (Eds.),
45
46 685 *Aspects of a suture zone, Marnier Glacier area Antarctica. Geologisches Jahrbuch B*, 85, 133-
47
48
49 686 172.
50

51 687
52
53 688 Coggon, R., Holland, T. J. B. (2002). Mixing properties of phengitic micas and revised
54
55
56 689 garnet–phengite thermobarometers. *Journal of Metamorphic Geology*, 20, 683–696.
57

58 690
59
60
61
62
63
64
65

691 Connolly, J. A. D. (2009). The geodynamic equation of state: What and how.

692 Geochemistry, Geophysics, Geosystems, 10, Q10014.

693

694 Connolly, J. A. D., Cesare, B. (1993). C-O-H-S fluid composition and oxygen

695 fugacity in graphitic metapelites. *Journal of Metamorphic Geology*, 11, 368-378.

696

697 Conrad, W. K., Nicholls, I. A., and Wall, V. J. (1988). Water-saturated and -

698 undersaturated melting of metaluminous and peraluminous crustal compositions at 10 kb:

699 Evidence for the origin of silicic magmas in the Taupo volcanic zone, New Zealand, and other

700 occurrences. *Journal of Petrology*, 29, 765-803. doi:10.1093/petrology/29.4.765

701

702 Crispini, L., Di Vincenzo, G., Palmeri, R. (2007). Petrology and ⁴⁰Ar–³⁹Ar dating of

703 shear zones in the Lanterman Range (northern Victoria Land, Antarctica): implications for the

704 metamorphic and temporal evolution at terrane boundaries. *Mineralogy and Petrology*, 89,

705 217–249. <http://dx.doi.org/10.1007/s00710-006-0164-2>.

706

707 Di Vincenzo, G., Palmeri, R., Talarico, F., Andriessen, P. A. M., Ricci, C. A. (1997).

708 Petrology and geochronology of eclogites from the Lanterman Range. Antarctica. *Journal of*

709 *Petrology*, 38, 1391–1417.

710

711 Di Vincenzo, G., Ghiribelli, B., Giorgetti, G., Palmeri, R. (2001). Evidence of a close

712 link between petrology and isotope records: constraints from SEM, EMP, TEM and in situ

713 ⁴⁰Ar–³⁹Ar laser analyses on multiple generations of white micas (Lanterman Range,

714 Antarctica). *Earth and Planetary Science Letters*, 192, 389–405.

715

716 Di Vincenzo, G., Grande, A., Rossetti, F. (2014). Paleozoic siliciclastic rocks from
1
2 717 northern Victoria Land (Antarctica): provenance, timing of deformation and implications for
3
4 718 the Antarctica/Australia connection. *Geological Society of America Bulletin* 126, 1416–1438.
5
6

7 719
8
9 720 Di Vincenzo, G., Horton F., Palmeri, R. (2016). Protracted (~30 Ma) eclogite-facies
10
11 721 metamorphism in northern Victoria Land (Antarctica): Implications for the geodynamics of
12
13 722 the Ross/Delamerian Orogen. *Gondwana Research*, 40, 91-106.
14
15

16 723
17
18
19 724 Federico, L., Capponi, G., Crispini, L. (2006). The Ross orogeny of the transantarctic
20
21 725 mountains: a northern Victoria Land perspective. *International Journal of Earth Sciences*
22
23 726 (*Geol Rundsch*) 95, 759. <https://doi.org/10.1007/s00531-005-0063-5>.
24
25

26 727
27
28
29 728 Federico, L., Crispini, L., Capponi, G. (2010). Fault-slip analysis and transpressional
30
31 729 tectonics: a study of Paleozoic structures in northern Victoria Land, Antarctica. *Journal of*
32
33 730 *Structural Geology*, 32, 667–684.
34
35

36 731
37
38
39 732 Ferrero, S., Bartoli, O., Cesare, B., Salvioli, Mariani, E., Acosta-Vigil, A., Cavallo, A.,
40
41 733 Groppo, C., Battiston, S. (2012). Microstructures of melt inclusions in anatectic
42
43 734 metasedimentary rocks. *Journal of Metamorphic Geology*, 30, 303–322.
44
45

46 735
47
48 736 Ferrero, S., Wunder, B., Walczak, K., O'Brien, P. J., Ziemann, M. A. (2015).
49
50 737 Preserved near ultrahigh-pressure melt from continental crust subducted to mantle depths.
51
52 738 *Geology*, 43, 447–450 doi:10.1130/G36534.1.
53
54

55 739
56
57
58
59
60
61
62
63
64
65

740 Ferrero, S., Godard, G., Palmeri, R., Wunder, B., and Cesare B. (2018). Partial
1
2 741 melting of ultramafic granulites from Dronning Maud Land, Antarctica: Constraints from
3
4 742 melt inclusions and thermodynamic modelling. *American Mineralogist*, 103, 610-622
5
6
7 743 <https://doi.org/10.2138/am-2018-6214>.
8
9 744
10
11 745 Frost, B.R., Barnes, C.G., Collins, W.J., Arculus, R.J., Ellis, D.J., and Frost, C.D.
12
13
14 746 (2001). A geochemical classification for granitic rocks. *Journal of Petrology*, 42, 2033–2048.
15
16 747
17
18 748 Gardien, V., Thompson, A. B., Ulmer, P. (2000). Melting of biotite + plagioclase +
19
20
21 749 quartz gneisses: The role of H₂O in the stability of amphibole. *Journal of Petrology*, 41, 651-
22
23
24 750 666. doi:10.1093/petrology/41.5.651
25
26 751
27
28 752 Ghezzi, C., Baldelli, C., Biagini, R., Carmignani, L., Di Vincenzo, G., Gosso, G.,
29
30
31 753 Lelli, A., Lombardo, B., Montrasio, A., Pertusati, P. C., Salvini, F. (1989). Granitoids from
32
33
34 754 the David Glacier ± Aviator Glacier segment of the Transantarctic Mountains (Victoria Land
35
36 755 Antarctica). In: Ricci CA (ed.) *Proceedings of the Meeting: Geosciences in Victoria Land,*
37
38
39 756 *Antarctica, Siena, September 2-3 1987. Mem Soc Geol Ital: 143-159.*
40
41 757
42
43 758 Ghiribelli, B. (2000). *Evoluzione tettonica e metamorfica del margine attivo del*
44
45
46 759 *Gondwana: Lanterman e Salamander Ranges (Antartide) Unpublished Ph.D. Thesis Siena*
47
48
49 760 *University, p. 127 pp.*
50
51 761
52
53 762 Ghiribelli, B., Frezzotti, M.L., Palmeri, R. (2002). Coesite in eclogites of the
54
55
56 763 Lanterman Range (Antarctica): evidence from textural and raman spectroscopy studies.
57
58 764 *European Journal of Mineralogy*, 14, 355–360.
59
60
61
62
63
64
65

765

1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

766 Giacomini, F., Tiepolo, M., Dallai, L., Ghezzi, C. (2007). On the onset and evolution
767 of the Ross-orogeny magmatism in North Victoria Land—Antarctica. *Chemical Geology*,
768 240, 103–128.

769
770 Godard, G., and Palmeri, R. (2013). High-pressure metamorphism in Antarctica from
771 the Proterozoic to the Cenozoic: A review and geodynamic implications. *Gondwana*
772 *Research*, 23, 844-864, <http://dx.doi.org/10.1016/j.gr.2012.07.012>

773
774 Grant, J. A. (2009). Thermocalc and experimental modelling of melting of pelite,
775 morton pass, Wyoming. *Journal of Metamorphic Geology*, 27, 571-578. doi:10.1111/j.1525-
776 1314.2009.00846.x

777
778 Grew, E.S., Kleinschmidt, G., Schubert, W. (1984). Contrasting metamorphic belts in
779 north Victoria Land, Antarctica. *Geologisches Jahrbuch B60*, 253–263.

780
781 Hiroi, Y., Hokada, T., Kato, M., et al. (2019). Felsite–nanogranite inclusions and three
782 Al₂SiO₅ polymorphs in the same garnet in ultrahigh–temperature granulites from
783 Rundvågshetta, Lützow–Holm Complex, East Antarctica. *Journal of Mineralogical and*
784 *Petrological Sciences*, 114, 60-78, doi 10.2465/jmps.181118.

785
786 Holland, T. J. B., and Powell, R. (1998). An internally consistent ther- modynamic
787 data set for phases of petrological interest. *Journal of Metamorphic Geology*, 16, 309–343.

788

789 Holland, T. J. B., Powell, R. (2003). Activity-composition relations for phases in
1
2 790 petrological calculations: An asymmetric multicomponent formulation. Contributions to
3
4 791 Mineralogy and Petrology, 145, 492–501.
5
6
7 792
8
9
10 793 Holness, M. B., Cesare, B., Sawyer, E.W. (2011). Melted rocks under the microscope:
11
12 794 microstructures and their interpretation. Elements, 7, 247-252.
13
14 795
15
16
17 796 Holtz, F., Johannes, W., Tamic, N., Behrens, H. (2001). Maximum and minimum
18
19 797 water contents of granitic melts generated in the crust: A reevaluation and implications.
20
21 798 Lithos, 56, 1–14.
22
23
24 799
25
26 800 Holtz, F., Johannes, W. (1991). Genesis of peraluminous granites I. experimental
27
28 801 investigation of melt compositions at 3 and 5 kb and various H₂O activities. Journal of
29
30 802 Petrology, 32, 935-958. doi:10.1093/petrology/32.5.935
31
32
33 803
34
35
36 804 Holtz, F., Johannes, W. (1992). Peraluminous granites: The effect of alumina on melt
37
38 805 composition and coexisting minerals. Transactions of the Royal Society of Edinburgh: Earth
39
40 806 Sciences, 83, 409-416. doi:10.1017/S0263593300008075
41
42
43 807
44
45
46 808 Huang, W. L., Wyllie, P. J. (1975). Melting reactions in the system NaAlSi₃O₈-
47
48 809 KAlSi₃O₈-SiO₂ to 35 kilobars, dry and with excess water. Journal of Geology, 83, 737–748.
49
50
51 810
52
53 811 Kleinschmidt, G., Tessensohn, F. (1987). Early Paleozoic westward directed
54
55 812 subduction at the Pacific continental margin of Antarctica, sixth Gondwana symposium.
56
57 813 American Geophysical Union, Geophysical Monograph 40, 89–105.
58
59
60
61
62
63
64
65

814

1

2 815 Luth, W.C., Jahns, R.H., Tuttle, O.F. (1964) The granite system at pressure of 4 to 10

3

4 816 kilobars. *Journal of Geophysical Research*, 69, 759–773.

6

7 817

8

9 818 Morgan VI, G.B., London, D. (2005). The effect of current density on the electron

11

12 819 microprobe analysis of alkali aluminosilicate glasses. *American Mineralogist*, 90, 1131–

13

14 820 1138.

15

16 821

18

19 822 Newton, R. C., Charlu, T. V., Kleppa, O. J. (1980). Thermochemistry of high

20

21 823 structural state plagioclases. *Geochimica et Cosmochimica Acta*, 44, 933–941.

23

24 824

25

26 825 Palmeri, R., Talarico, F., Meccheri, M., Oggiano, G., Pertusati, P.C., Rastelli, N.,

28

29 826 Ricci, C.A. (1991). Progressive deformation and low pressure/high temperature

30

31 827 metamorphism in the Deep Freeze Range, Wilson Terrane, northern Victoria Land,

32

33 828 Antarctica. *Memorie della Società Geologica Italiana*, 46, 179–195.

35

36 829

37

38 830 Palmeri, R., Pertusati, P.C., Ricci, C.A., Talarico, F. (1994). Late Proterozoic(?)–early

40

41 831 Paleozoic of the active Pacific margin of Gondwana: evidence from the southern Wilson

42

43 832 terrane (northern Victoria Land, Antarctica). *Terra Antarctica*, 1, 5–9.

45

46 833

47

48 834 Palmeri, R. (1997). P-T paths and migmatite formation: An example from Deep

49

50 835 Freeze Range, northern Victoria Land, Antarctica. *Lithos*, 42, 47–66.

52

53 836

54

55 837 Palmeri, R., Ghiribelli, B., Talarico, F., Ricci, C.A. (2003). Ultra-high-pressure

57

58 838 metamorphism in felsic rocks: the garnet-phengite gneisses and quartzites from the

59

60

61

62

63

64

65

839 Lanterman Range. Antarctica: European Journal of Mineralogy, 15, 513–525.

840 [http://dx.doi.org/ 10.1127/0935-1221/2003/0015-0513](http://dx.doi.org/10.1127/0935-1221/2003/0015-0513).

841

842 Palmeri, R., Ghiribelli, B., Ranalli, G., Talarico, F., Ricci, C.A. (2007). Ultrahigh-

843 pressure meta- morphism and exhumation of garnet-bearing ultramafic rocks from the

844 Lanterman Range (northern Victoria Land, Antarctica). Journal of Metamorphic Geology, 25,

845 225–243. <http://dx.doi.org/10.1111/j.1525-1314.2006.00686.x>.

846

847 Palmeri, R., Frezzotti, M.L., Godard, G., Davies, J. (2009). Pressure-induced incipient

848 amorphization of α -quartz and transition to coesite in an eclogite from Antarctica: a first

849 record and some consequences. Journal of Metamorphic Geology, 27, 685–705.

850

851 Palmeri, R., Talarico, F.M., Ricci, C.A. (2011). Ultrahigh-pressure metamorphism at

852 the Lanterman Range (northern Victoria Land, Antarctica). Geological Journal, 46, 126–136.

853

854 Patiño Douce, A.E., Beard, J.S. (1996). Effects of P, f(O₂) and Mg/Fe ratio on

855 dehydration-melting of model metagraywackes. Journal of Petrology, 37, 999–1024.

856

857 Redler, C., Johnson, T. E., White, R. W., and Kunz, B. E. (2012). Phase equilibrium

858 constraints on a deep crustal metamorphic field gradient: Metapelitic rocks from the Ivrea

859 Zone (NW Italy). Journal of Metamorphic Geology, 30, 235–254, [https://doi.org/10.](https://doi.org/10.1111/j.1525-1314.2011.00965.x)

860 [1111/j.1525-1314.2011.00965.x](https://doi.org/10.1111/j.1525-1314.2011.00965.x)

861

862 Ricci, C.A., Talarico, F., Palmeri, R., Di Vincenzo, G., and Pertusati, P.C. (1996).
1
2 863 Eclogite at the Antarctic palaeo-Pacific active margin of Gondwana (Lanterman Range,
3
4 864 northern Victoria Land, Antarctica). *Antarctic Science*, 8, 277–280.
5
6
7 865
8
9 866 Ricci, C.A., Talarico, F., and Palmeri, R. (1997). Tectonothermal evolution of the
10
11 867 Antarctic paleo-pacific active margin of Gondwana: a northern Victoria Land perspective. In:
12
13 868 Ricci, C.A. (Ed.), *The Antarctic Region: Geological Evolution and Processes*. Terra Antartica
14
15 869 Publications, Siena, 213–218.
16
17
18
19 870
20
21 871 Ricci, C.A., and Tessensohn, F. (2003). The Lanterman-Mariner suture: Antarctic
22
23 872 evidence for active margin tectonics in Paleozoic Gondwana. *Geologisches Jahrbuch B85*,
24
25 873 303–332.
26
27
28
29 874
30
31 875 Rocchi, S., Di Vincenzo, G., Ghezzi, C. (2004). The Terra Nova Intrusive Complex
32
33 876 (Victoria Land, Antarctica), with 1:50000 geopetrographic map. *Terra Antartica Reports* 10,
34
35 877 51.
36
37
38
39 878
40
41 879 Roedder, E. (1984). Fluid inclusions. Mineralogical Society of America, *Reviews in*
42
43 880 *Mineralogy*, 12, 644 pp.
44
45
46 881
47
48 882 Stump, E. (1995). *The Ross Orogen of the transantarctic mountains*. Cambridge
49
50 883 University Press, Cambridge. 284 pp.
51
52
53 884
54
55 885 Tacchetto, T., Bartoli, O., Cesare, B., Berkesi, M., Aradi, L. E., Dumond, G., Szabó,
56
57 886 C. (2019). Multiphase inclusions in peritectic garnet from granulites of the Athabasca
58
59
60
61
62
63
64
65

1 887 granulite terrane (Canada): Evidence of carbon recycling during Neoproterozoic crustal melting.

2 888 *Chemical Geology*, 508, 197-209, <https://doi.org/10.1016/j.chemgeo.2018.05.043>.

3
4 889

5
6
7 890 Tajčmanová, L., Connolly, J. A. D., Cesare, B. (2009). A thermodynamic model for
8
9 891 titanium and ferric iron solution in biotite. *Journal of Metamorphic Geology*, 27, 153-165.

10
11 892

12
13
14 893 Talarico, F., Borsi, L., Lombardo, B. (1995). Relict granulites in the Ross Orogen of
15
16 894 northern Victoria Land (Antarctica), II. Geochemistry and palaeo-tectonic implications.
17
18 895 *Precambrian Research*, 75, 157–174. [http://dx.doi.org/10.1016/0301-9268\(95\)80004-2](http://dx.doi.org/10.1016/0301-9268(95)80004-2).

19
20
21 896

22
23
24 897 Talarico, F., Ghiribelli, B., Smith, S.C., Palmeri, R., Ricci, C.A. (1998). The northern
25
26 898 Victoria Land segment of the Antarctic paleo-Pacific margin of eastern Gondwana: new
27
28 899 constraints from the Lanterman and Mountaineer Ranges. *Terra Antarctica*, 5, 245–252.

29
30
31 900

32
33
34 901 Talarico, F., Palmeri, R., and Ricci, C.A. (2004). Regional metamorphism and P-T
35
36 902 evolution of the Ross orogen in northern Victoria Land (Antarctica): A review. *Periodico*
37
38 903 *Mineralogia*, 73, 185-196.

39
40
41 904

42
43 905 Tamic, N., Behrens, H., Holtz, F. (2001). The solubility of H₂O and CO₂ in rhyolitic
44
45 906 melts in equilibrium with a mixed CO₂-H₂O fluid phase. *Chemical Geology*, 174, 333-347.

46
47
48 907

49
50
51 908 Thomen, A., Robert, F., Remusat, L. (2014). Determination of the nitrogen abundance
52
53 909 in organic materials by NanoSIMS quantitative imaging. *J. Anal. At. Spectrom.* 29, 512-519.

54
55
56 910

57

58

59

60

61

62

63

64

65

1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

911 Thompson, J. B., Hovis, G. L. (1979). Entropy of mixing in sanidine. American
912 Mineralogist, 64, 57–65.
913
914 Wallace, P. J., Kamenetsky, V. S., and Cervantes, P. (2015). Melt inclusion CO₂
915 contents, pressures of olivine crystallization, and the problem of shrinkage bubbles. American
916 Mineralogist, 100, 787–794, 003-004X/15/0004–87\$05.00/DOI:[http://dx.doi.org/10.2138/am-](http://dx.doi.org/10.2138/am-2015-5029)
917 2015-5029
918
919 Wang, X., Chou, I. M., Hu, W., Burruss, R. C., Sun, Q., Song, Y. (2011). Raman
920 spectroscopic measurements of CO₂ density: Experimental calibration with high-pressure optical cell
921 (HPOC) and fused silica capillary capsule (FSCC) with application to fluid inclusion observations.
922 Geochimica et Cosmochimica Acta, 75, 4080-4093.
923
924 Weinberg, R. F., Hasalová, P. (2015). Water-fluxed melting of the continental crust: A
925 review. Lithos, 212, 158-188.
926
927 White, R. W., Powell, R., Holland, T. J. B. (2007). Progress relating to calculation of
928 partial melting equilibria for metapelites. Journal of Metamorphic Geology, 25, 511–527.
929
930 White, R. W., Stevens, G., Johnson, T. E. (2011) Is the crucible reproducible?
931 Reconciling melting experiments with thermodynamic calculations. Elements, 7, 241–
932 246.
933
934 Wilke, S., Holtz, F., Neave, D. A., Almeev, R. (2017). The Effect of Anorthite
935 Content and Water on Quartz–Feldspar Cotectic Compositions in the Rhyolitic System and

936 Implications for Geobarometry. *Journal of Petrology*, 2017, Vol. 58, No. 4, 789–818 doi:

937 10.1093/petrology/egx034

938

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.

952 **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
1
2 962 lobate-cusped walls, offshots (white arrows), and secondary Grt suggesting overheating and
3
4 963 host interaction; 900°C - glassy MI with segmented straight walls after assimilation of host
5
6
7 964 Grt (inset).
8

9
10 965

11
12 966 Figure 5 – Harker diagrams showing the major element concentrations (wt%) of melt

13
14 967 inclusions in crustal enclaves after remelting experiments at 740°C to 900°C (data in Table 3).

15
16 968 (A) MnO vs. FeO_{tot}. (B) FeO + MgO + TiO₂ vs. SiO₂ with mixing vectors between the

17
18
19 969 mineral pairs Grt-Qz and Feld-Qz (see text for details). (C) CaO vs. SiO₂. (D) Na₂O vs. SiO₂.

20
21
22 970 (E) Al₂O₃ vs. SiO₂ with mixing vectors between the mineral pairs Grt-Qz and Feld-Qz. (F)

23
24 971 ASI vs. SiO₂.
25

26
27 972

28
29 973 Figure 6 – CIPW Qz-Ab-Or and An-Ab.Or diagrams showing the normative

30
31 974 compositions of all analyzed melt inclusions. Blue and green circles are for glassy inclusions

32
33
34 975 remelted at 760°C and 780°C respectively. Grey dotted lines show the eutectic points and

35
36 976 cotectic lines for the subaluminous haplogranite system at $a_{\text{H}_2\text{O}} = 1$ and different pressures

37
38
39 977 (Holtz and Johannes 1992; Luth et al. 1964; Huang and Wyllie 1975). Grey circles and lines

40
41 978 represent the effect of different amounts of normative melt An at constant 100 MPa pressure

42
43
44 979 (data from James and Hamilton, 1969). References to experimental glasses from Ojén

45
46 980 (Bartoli et al., 2013c) and El Hoyazo (Acosta-Vigil et al., 2007) where “met.” refers to

47
48
49 981 metatexites, and “diat.” to diatexite.
50

51 982

52
53 983 Figure 7 – Composition of analyzed melt inclusions in (A) modified alkali-lime index

54
55
56 984 (MALI = Na₂O + K₂O – CaO) vs. SiO₂ (wt%) after Frost et al. (2001); (B) total alkali vs.

57
58
59 985 SiO₂ diagram. Symbols as in Fig. 6.
60
61
62
63
64
65

986

1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

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.

1011

1

21012

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

3

41013

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

5

6

71014

8

91015

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

10

111016

inclusions by NanoSIMS.

12

13

14

15

16

17

18

19

20

21

22

23

24

25

26

27

28

29

30

31

32

33

34

35

36

37

38

39

40

41

42

43

44

45

46

47

48

49

50

51

52

53

54

55

56

57

58

59

60

61

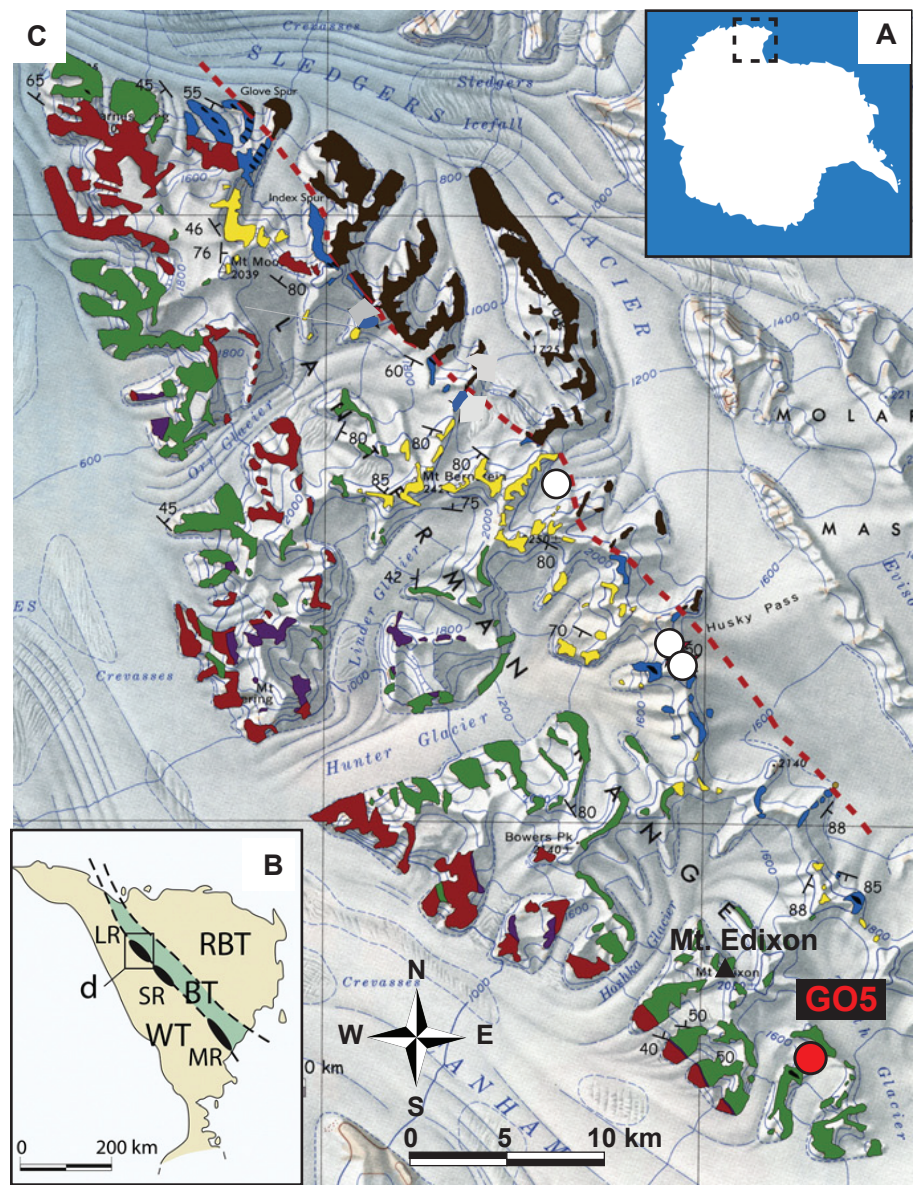
62

63

64

65

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

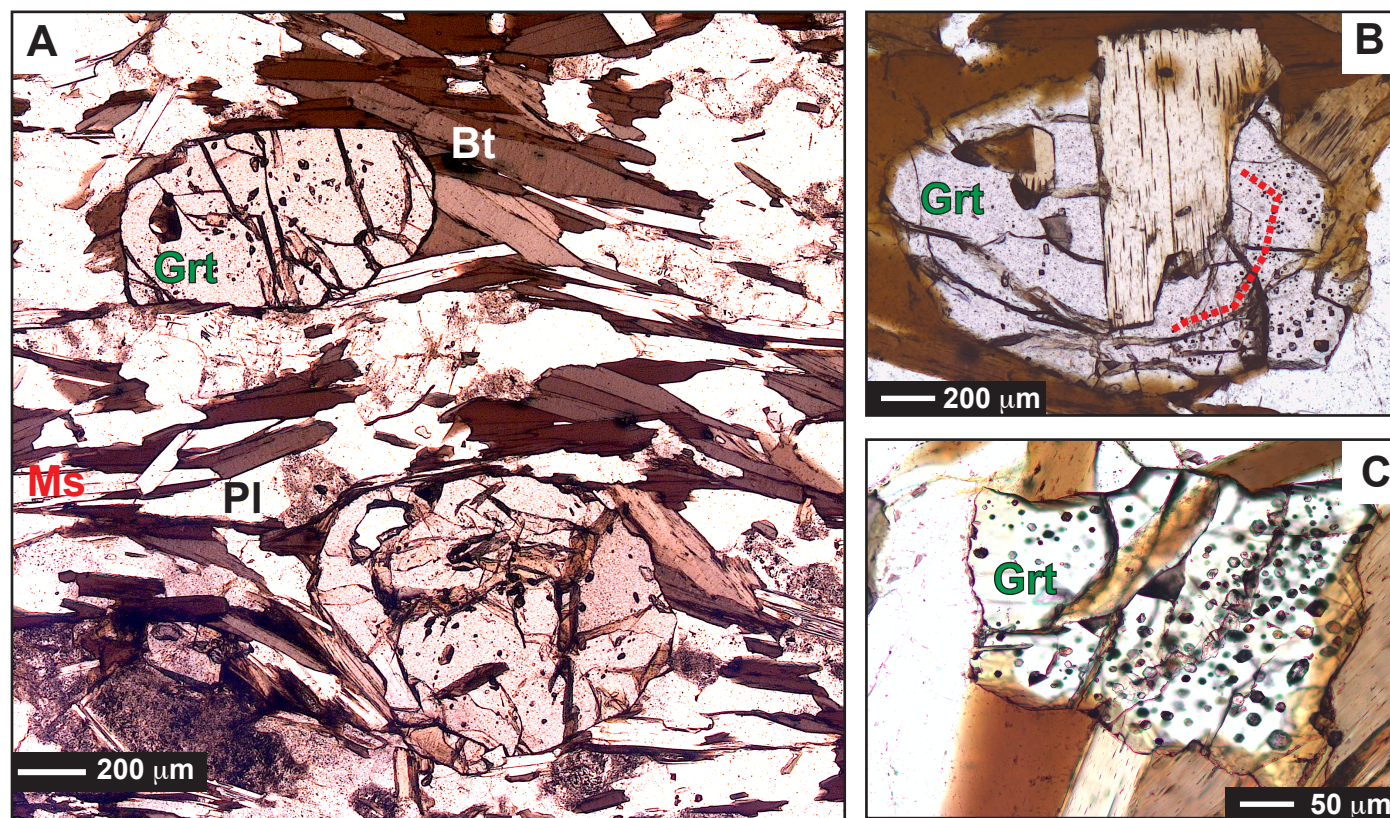


Figure 2

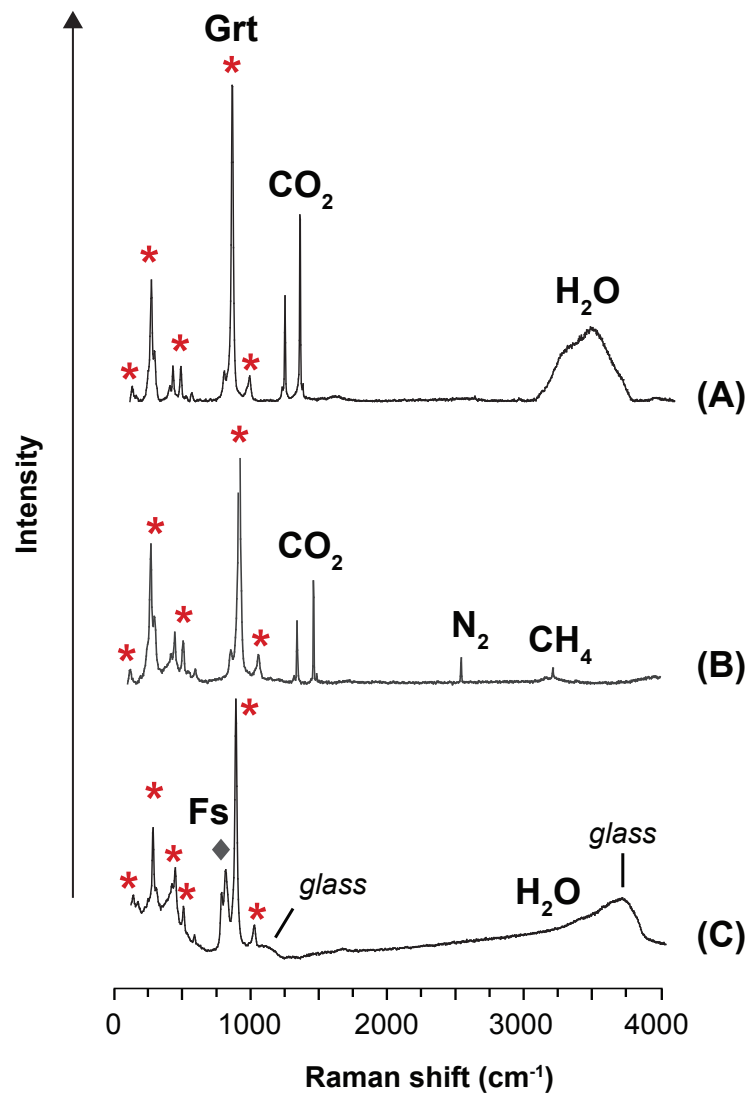


Figure 3

Figure4

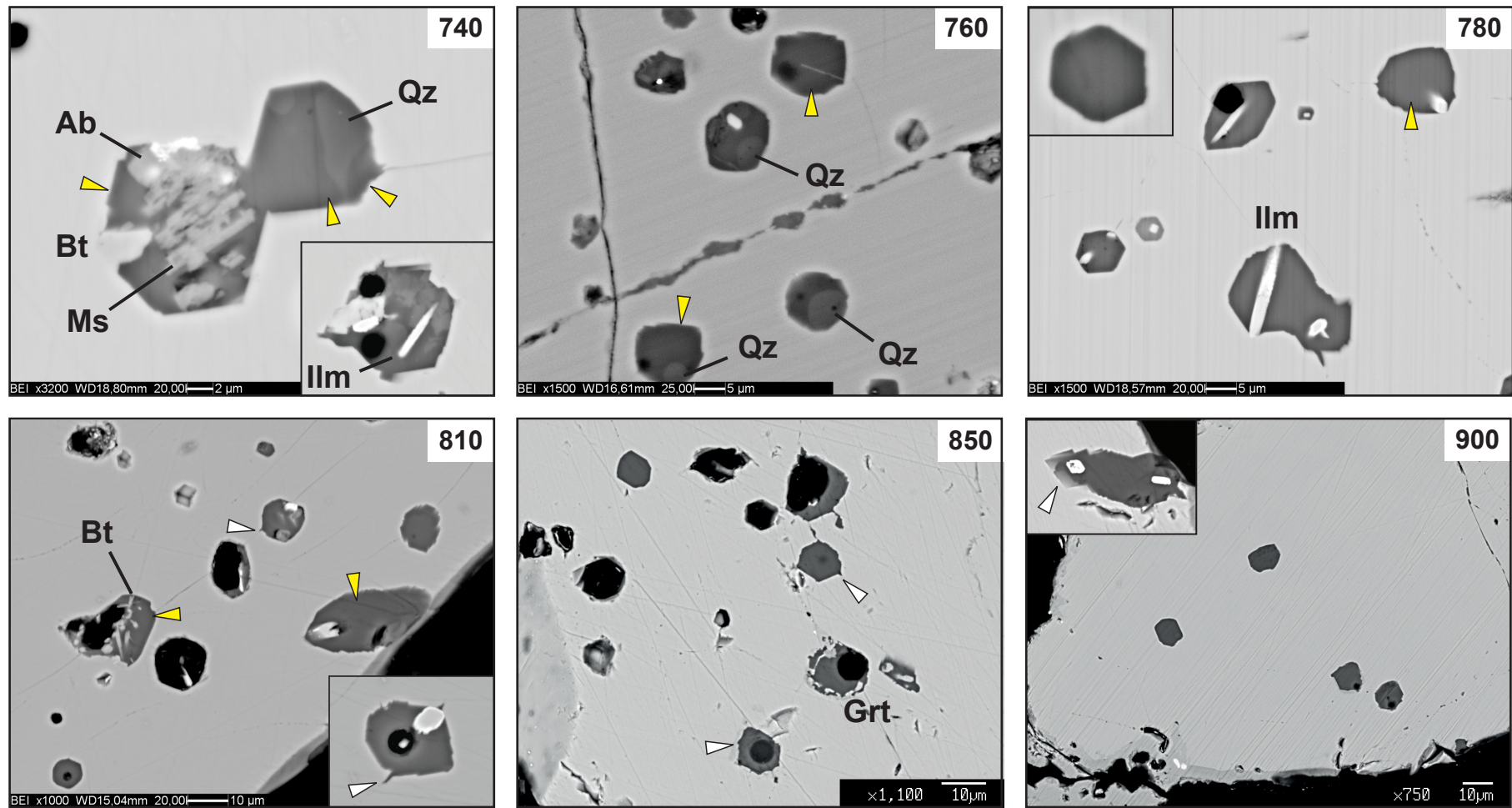


Figure 4

Figure 5

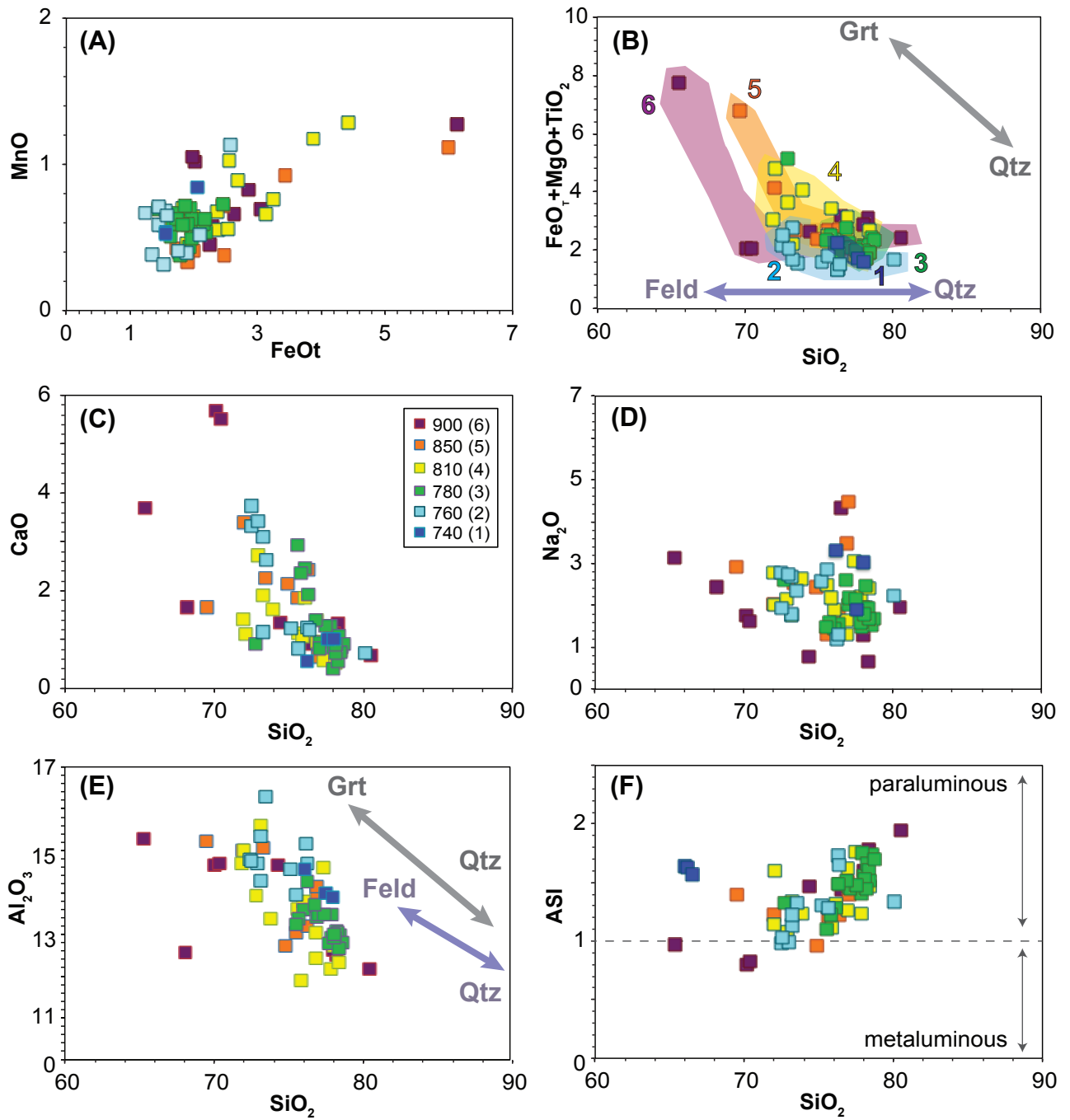


Figure 5

Figure6

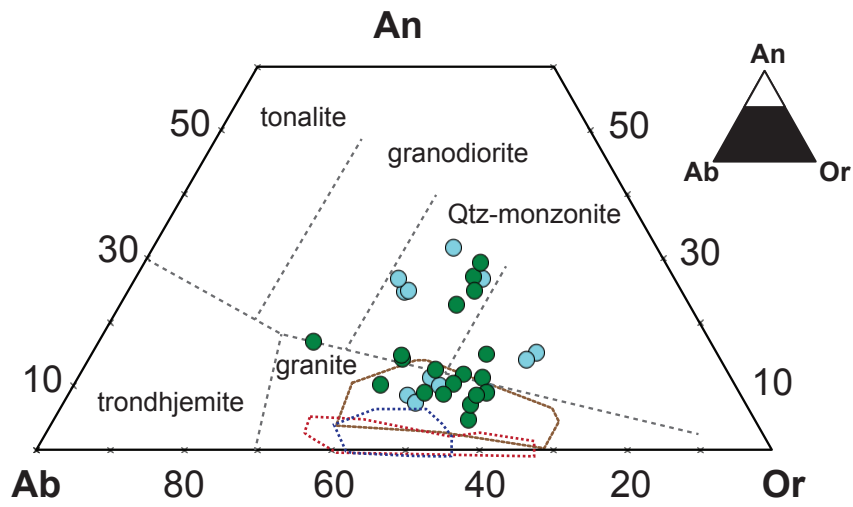
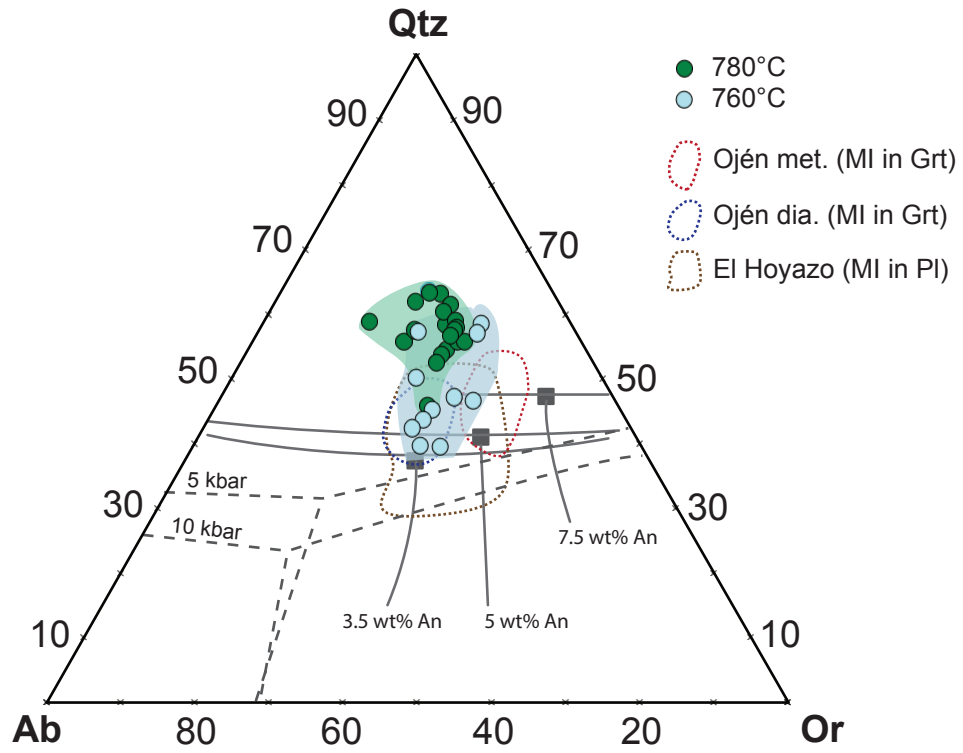


Figure 6

Figure 7

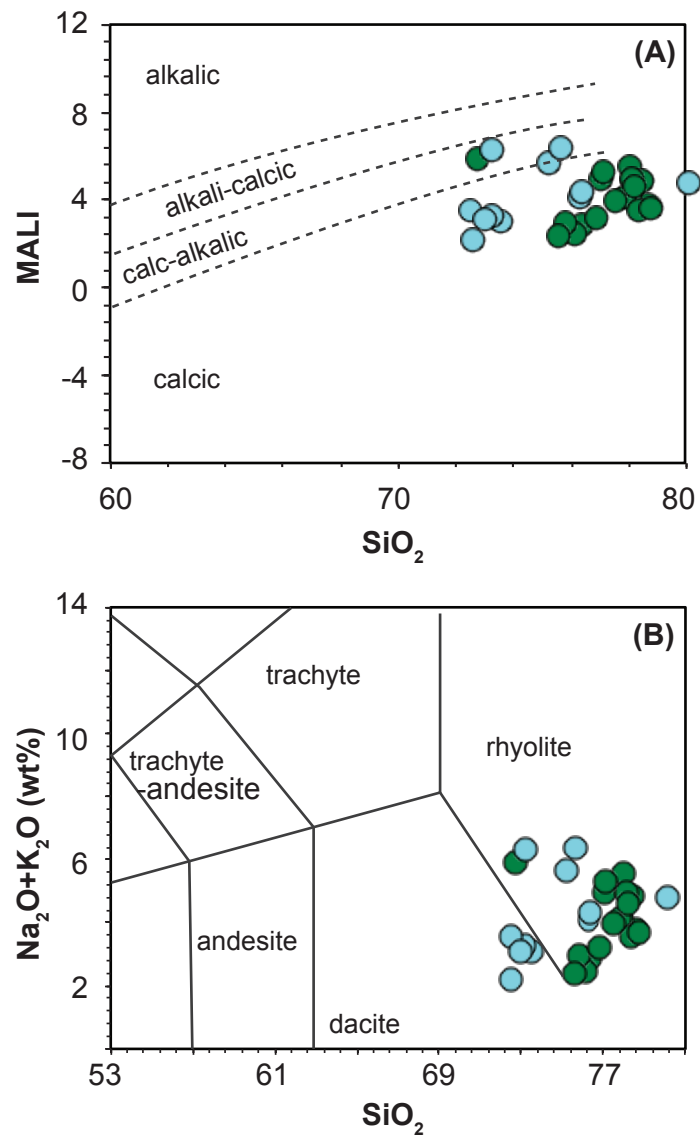


Figure 7

Figure8

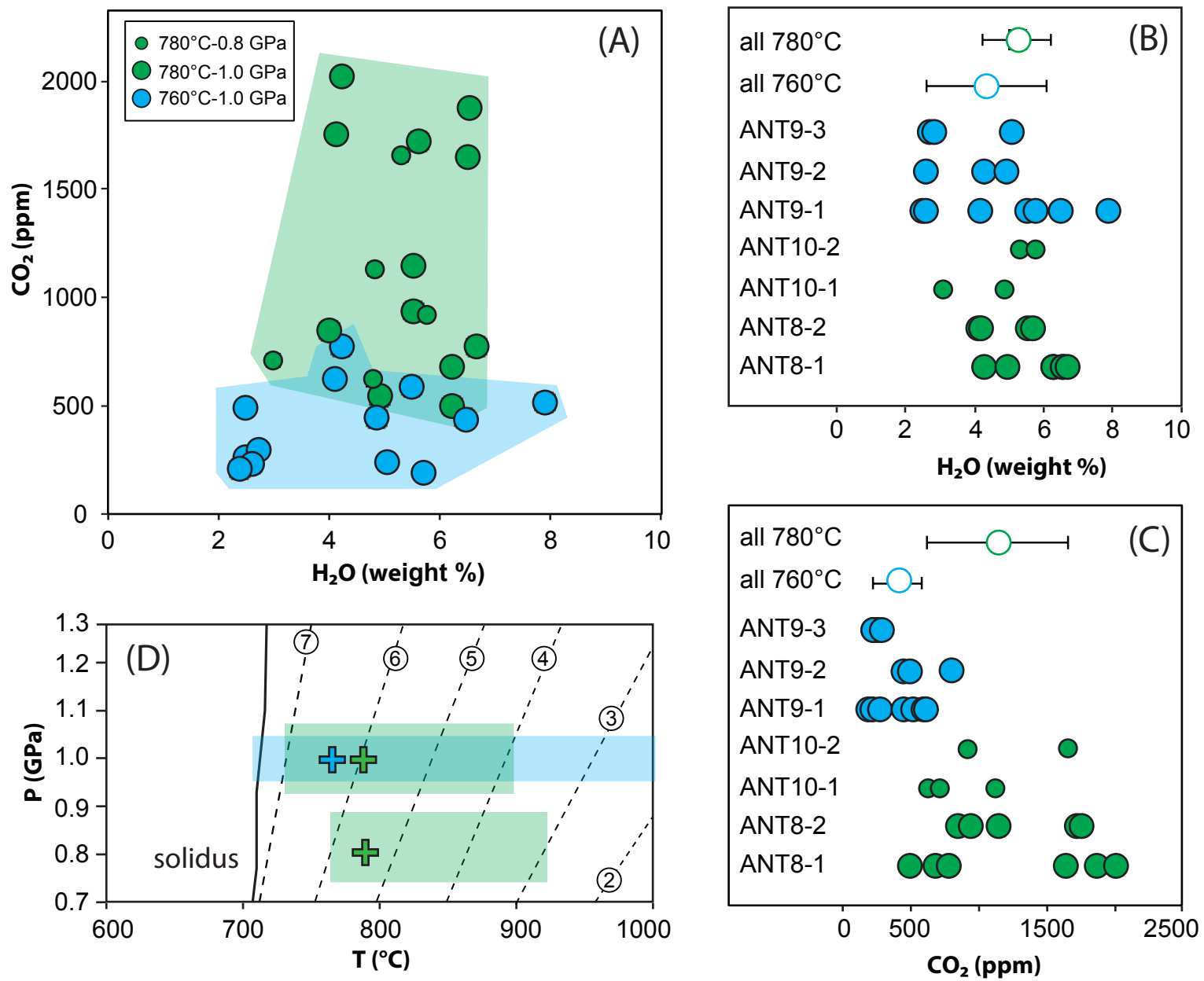


Figure 8

Figure9

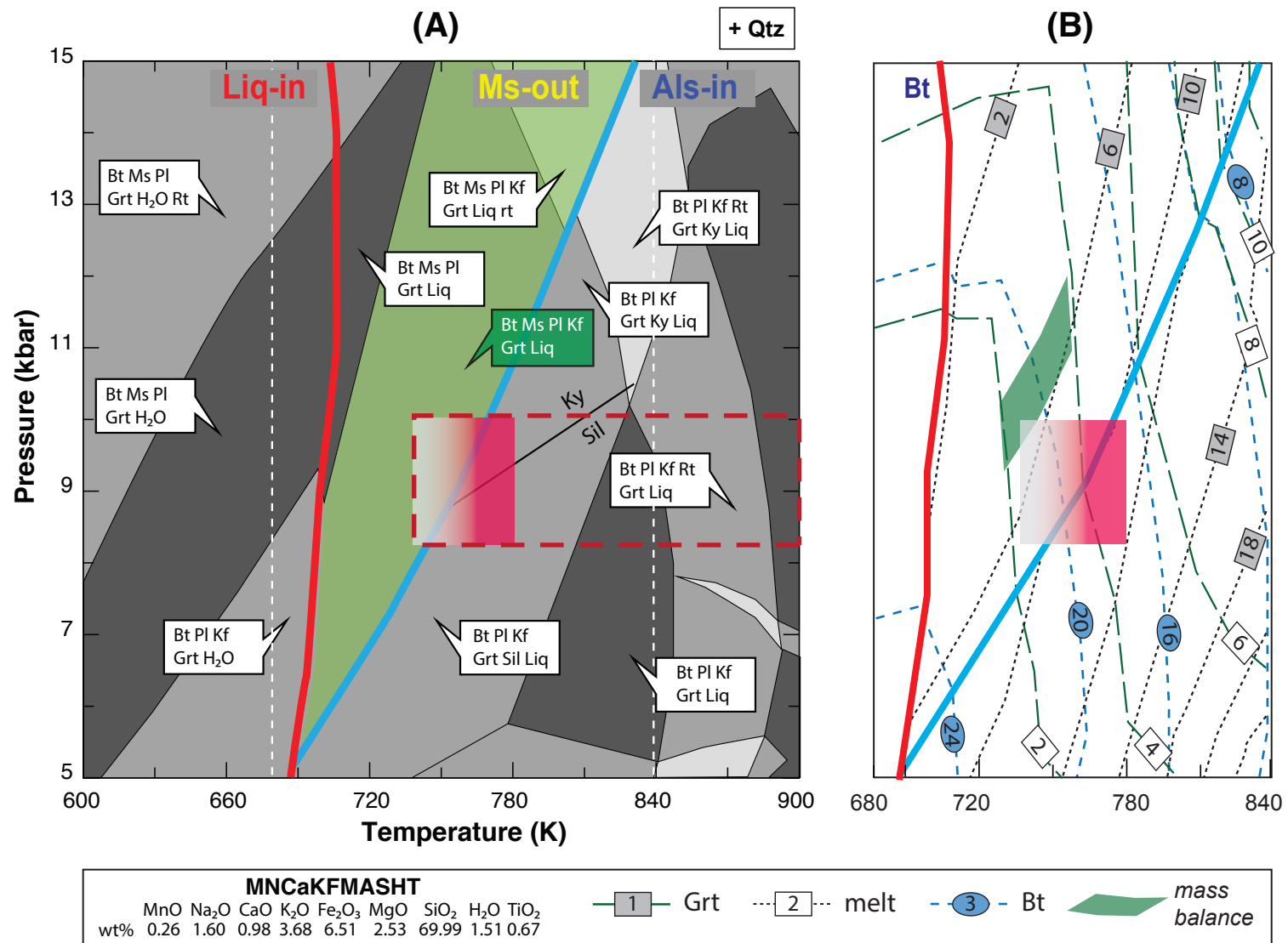


Figure 9

Figure10

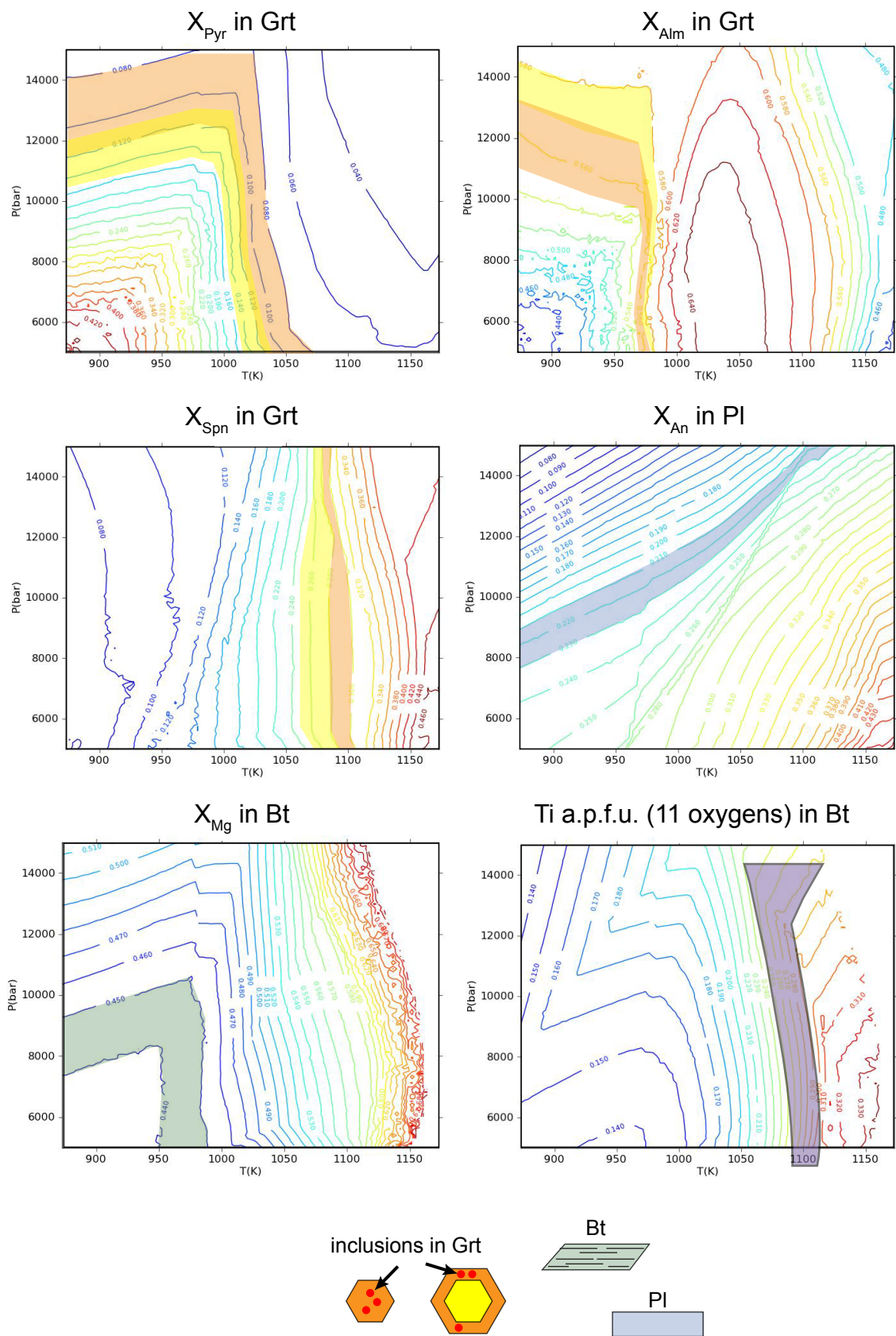


Figure 10 (Supplementary Material)

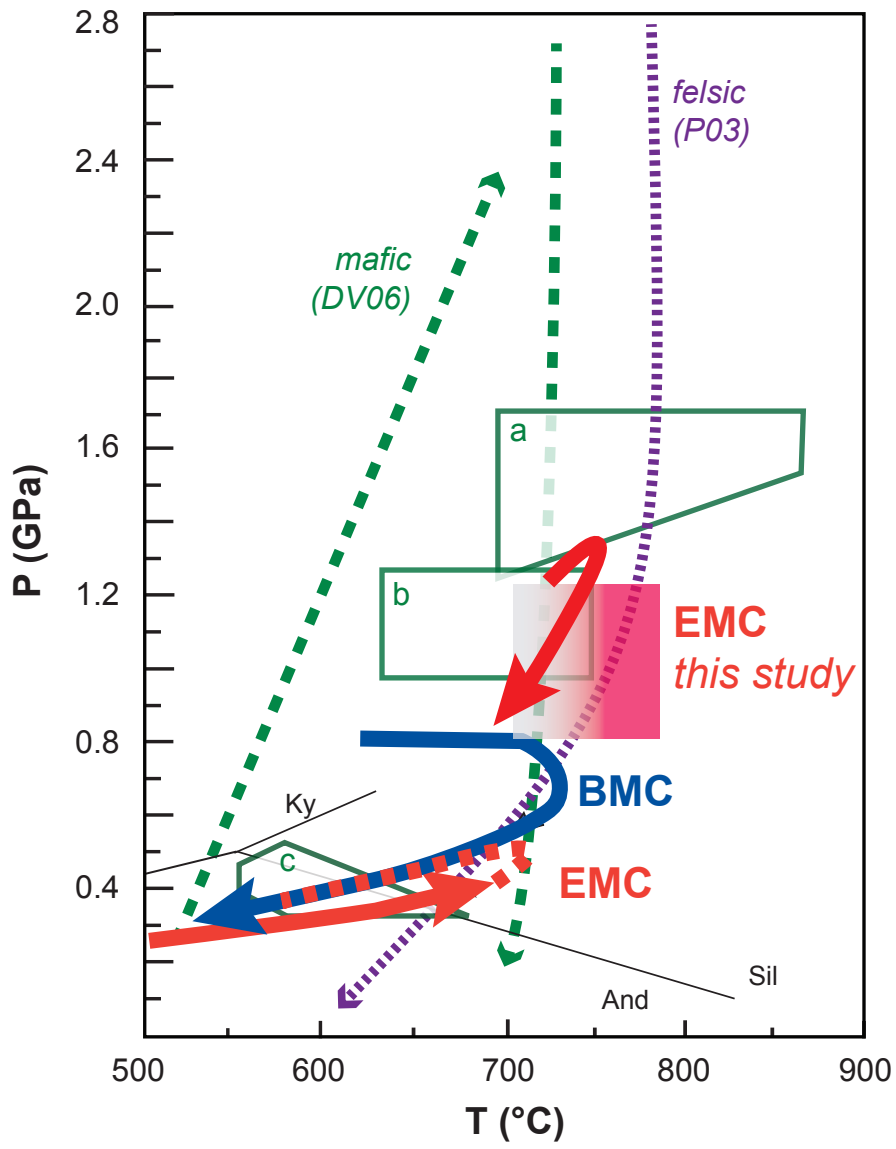


Figure 11

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	CaC CaO	CaO	CaC CaO	CaO	CaC CaO	CaO	CaC 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 C K ₂ O	K ₂ O	K ₂ C K ₂ O	K ₂ O	K ₂ C K ₂ O	K ₂ O	K ₂ C 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