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Origin and age of zircon-bearing chromitite layers from the Finero Phologopite Peridotite (Ivrea-Verbano Zone, western Alps) and geodynamic consequences

Alberto Zanetti, Tommaso Giovanardi, Antonio Langone, Massimo Tiepolo, Fu-Yuan Wu, Luigi Dallai, Maurizio Mazzucchelli

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- 1 Origin and age of zircon-bearing chromitite layers from the Finero Phologopite Peridotite
- 2 (Ivrea-Verbano Zone, western Alps) and geodynamic consequences
- 3 Alberto Zanetti¹, <u>Tommaso Giovanardi²</u>, Antonio Langone¹, Massimo Tiepolo³, Fu-Yuan Wu⁴,
- 4 Luigi Dallai⁵, Maurizio Mazzucchelli^{6,1}.
- 5
- 6 1Istituto di Geoscienze e Georisorse CNR, Unità di Pavia, Via Ferrata, 1, I-27100 Pavia, Italy;
- 7 2Instituto de Geociências, Universidade de São Paulo, Rua do Lago, 562, Cidade Universitária,
- 8 05508-900 São Paulo, Brazil;
- 9 3Dipartimento di Scienze della Terra 'Ardito Desio', Università degli Studi di Milano, Via
- 10 Mangiagalli/Botticelli, 32/23, I-20133 Milano, Italy;
- 11 4State Key Laboratory of Lithospheric Evolution, Institute of Geology and Geophysics, Chinese
- 12 Academy of Sciences, P. O. Box 9825, 100029 Beijing, China;
- 13 5Istituto di Geoscienze e Georisorse CNR, Sede di Pisa, Via Giuseppe Moruzzi, 1, I-56124 Pisa,
 14 Italy;
- 15 6Dipartimento di Scienze Chimiche e Geologiche, Università di Modena e Reggio Emilia, Via
- 16 Campi, 103, I-41125 Modena, Italy.
- 17
- 18 <u>Corresponding Author</u>: Tommaso Giovanardi; e-mail: tommaso.giovanardi@gmail.com
 19
- 20 Abstract
- 21 An investigation has been performed on three chromitite layers segregated in dunite bodies of the
- 22 Phlogopite Peridotite mantle unit in the Finero Complex (FPP, Ivrea–Verbano Zone, Southern
- 23 Alps) aimed at providing new constraints to their origin and evolution.
- 24 Field relationships, the sub-chondritic Hf isotopic composition of the zircons (ϵ Hf₍₁₈₈₎ as low as -
- 25 5.4), the heavy O isotopic composition of zircons and pyroxenes (δ^{18} O up to 6.9‰), the strict
- 26 similarity of the trace element composition between the clinopyroxenes and amphiboles from the

27 chromitites and those from the phlogopite harzburgites and pyroxenites forming the typical FPP association, as well as the REE composition of zircons, which approaches equilibrium with the 28 29 associate clinopyroxene, suggest that the studied chromitites were segregated from melts, highly contaminated from continental crust, during the pervasive cycle of metasomatism recorded by the 30 31 FPP. A LA-ICP-HRMS survey of chromitite zircon grains has provided Early Jurassic U-Pb ages 32 mostly between 199 ± 3 Ma and 178 ± 2 Ma, with a pronounced peak at 187 Ma. Relevant exceptions are inherited domains of two grains giving Triassic ages of 242 ± 7 Ma and 229 ± 7 Ma, 33 34 and a third homogeneous zircon giving 208 ± 3 Ma. Our geochronological data and those reported in the literature show that the FPP chromitites have zircon populations with different internal CL 35 textures, but the same sub-chondritic Hf isotopic composition, which define an overall U-Pb age 36 37 span from ~290 Ma to 180. The segregation of the chromitite layers and the main pervasive 38 metasomatism likely occurred in the Early Permian (in a post-collisional, transtensional setting) or 39 before (possibly, in a subduction-related setting). The rejuvenation of the zircon ages was 40 accompanied by a progressive disapperance of the internal zoning, interpreted as the result of a 41 prolonged residence at mantle depths with progressive re-equilibration of the U-Pb system due to 42 thermal perturbations. The age peak at ~187 Ma is argued to constrain the timing of FPP exhumation at shallower, crustal levels. This process was characterised by an important reheating 43 event, possibly due to lithospheric hyperextension. The evolution of the FPP appears completely 44 45 different than that of mantle bodies of the central IVZ (i.e. the Val Sesia-Type bodies), which were emplaced within the continental crust, as part of accretionary prisms, at or before the end of the 46 47 Variscan orogeny.

48

49 Keywords: Finero; zircon; mantle; chromitite; metasomatism.

50

51 Introduction

The Ivrea-Verbano Zone (IVZ, Southern Alps, Italy-Switzerland) consists of a worldwide famous section of lower continental crust. The reference geodynamic model developed after detailed field, geochronological and geochemical investigation of its central part (i.e. Sesia and Sessera Valleys) involves Early Permian under- and intra-plating of mafic melts, which interacted with granulite to amphibolite-facies metamorphic basement and evolved towards acid compositions producing granite intrusions and rhyolite volcanism (Quick et al., 1995, 2009; Sinigoi et al., 1996, 2011; Mazzucchelli et al., 2014).

Zanetti et al. (2013, 2014) pointed out that the northern sector of the IVZ records a number of 59 structural, petrochemical and age anomalies with respect to such a model. For instance, there is now 60 abundant evidence of the occurrence in the Finero Mafic Complex (northernmost IVZ) of 61 62 widespread events of Triassic magmatism (Gebauer, 1993; Lu et al., 1997a,b; Hingerl et al., 2008; Zanetti et al., 2013, 2014; Klötzli, Personal Communication), with late alkaline veins and pockets 63 64 (e.g. zircon-bearing diorites and nepheline diorite pegmatites) mostly showing Late Triassic to 65 Early Jurassic ages (Oppizzi and Schaltegger, 1999; Grieco et al., 2001; Klötzli et al., 2007, 2009; 66 Schaltegger et al., 2015, and references therein). Late Triassic intrusions of zircon-bearing diorites, 67 associated to hornblendites (Stähle et al., 1990, 2001; Grieco et al., 2001), are also recorded by the 68 associate mantle unit, the Finero Phlogopite Peridotite (FPP), which also shows peculiar late, 69 discordant swarms of apatite-calcite-bearing gabbroic veins characterised by the presence of 70 magmatic sapphirine (Giovanardi et al., 2013). A 225 Ma age has been found in a recrystallised rim 71 of a zircon from a metasedimentary septum included in the intrusive External Gabbro unit and in a 72 monazite from the adjacent Kinzigite Formation (Langone and Tiepolo, 2015). 73 These observations suggest the occurrence of two kinds of IVZ, the "Val Sesia"-Type (central IVZ) 74 and the "Finero"-Type (northern IVZ), assumed to have had different tectono-magmatic evolutions 75 (Rivalenti and Mazzucchelli, 2000; Zanetti et al., 2013, 2014; Mazzucchelli et al., 2014). Further 76 main differences are the petrochemical features of the mantle bodies. The Val Sesia-Type mantle 77 bodies are constituted by refractory spinel peridotites, virtually free from metasomatism away from

pyroxenites and dunite channels (Rivalenti et al., 1981, 1995; Mazzucchelli et al., 2009).

79 Conversely, the Finero-Type mantle bodies are enriched in phlogopite and amphibole due to diffuse

80 modal metasomatism (Rivalenti and Mazzucchelli, 2000; Zanetti et al., 2013; Mazzucchelli et al.,

81 2014). According to the distribution of the different kinds of mantle peridotites, Zanetti et al. (2013)

82 have speculatively proposed that their boundary may corresponds to the Anzola-Val Grande high

83 temperature shear zone (HTSM in Fig. 1).

84 The FPP is the biggest mantle body of the northern IVZ. It shows a virtually complete

85 recrystallization due to pervasive to channelled melt migrations (Zanetti et al., 1999). The pervasive 86 metasomatism formed a main lithologic association constituted by phlogopite harzburgites associated to phlogopite pyroxenites (mainly websterites and orthopyroxenites), which do not show 87 88 significant chemical gradients among them (Zanetti et al., 1999). The channelled migration stages 89 formed dunite bodies, often containing stratiform to podiform chromites and, more rarely, 90 pyroxenite and hornblendite layers (Cumming et al., 1987; Hartmann and Wedepohl, 1993; Zanetti et al., 1999; Seitz and Woodland, 2000; Grieco et al., 2001, 2004; Zaccarini et al., 2004; Raffone et 91 al., 2006; Selverstone and Sharp, 2011; Giovanardi, 2012; Mazzucchelli et al., 2014). Relatively 92 93 late melt migration events also formed peridotite and pyroxenite veins and bands (Zanetti et al., 94 1999, Grieco et al., 2001; Morishita et al., 2003, 2008; Matsumoto et al. 2005; Raffone et al., 2006). 95 These are often characterised by the presence of apatite and carbonates, and usually exhibit marked 96 modal and chemical gradients with respect to the host phlogopite harzburgite. Examples of these 97 lithologies are the apatite-dolomite-bearing wehrlites (Zanetti et al., 1999; Raffone et al., 2006), as 98 well as the apatite-bearing orthopyroxenites described by Morishita et al. (2003, 2008), and 99 Matsumoto et al. (2005), these latter displaying Triassic ages.

100 Thanks to such a unique lithological association, the FPP is one of the most studied mantle

101 sequence in the world (Fig. 1). Several papers have dealt with the age of the petrochemical

102 processes and the geochemical affinity of the melts that migrated through it. These melts only

103 marginally match the magmatic record of the associated crustal rocks of the Finero Mafic Complex

104 (Giovanardi et al., 2014). Despite this, the geodynamic evolution of the FPP is still controversial

105 and the geochemical affinity of the metasomatic melts strongly debated. In particular,

106 geochronological surveys reported in the literature give a very large age interval, spanning from the

107 Early Permian to the Early Jurassic, which is difficult to reconcile with the evidence that the mantle

108 bodies of the central IVZ (i.e. the Balmuccia body) were emplaced within the continental crust, as

109 part of accretionary prisms, at or before the end of the Variscan orogeny.

110 A special opportunity to place geochronological constraints on the evolution of the FPP is

111 represented by the occurrence of large amounts of primary zircon in the chromitites (Ferrario and

112 Garuti, 1990; Grieco et al., 2001; Zaccarini et al., 2004). Thus, three chromitite swarms segregated

in dunite bodies of the FPP were investigated. Zircon and associate minerals have been subjected to

a detailed petrochemical investigation in order to shed light on i) the geochemical affinity of their

115 parent melts, ii) the age of the metasomatism and iii) the geodynamic evolution of the mantle

116 sequences in the Finero-Type IVZ.

117

118 Geological setting and selected samples

119 The IVZ represents the westernmost sector of the Southern Alps, which form the inner part of the Alpine orogen. The Southern Alps escaped Alpine subduction, thus preserving their lithospheric 120 mantle roots. In the IVZ, the lithologies were tilted about of 90° at the end of the Middle Miocene 121 122 (Wolff et al., 2012), as a consequence of a series of rotations started with the opening of the Jurassic 123 Tethys and culminated with the Alpine collision and relaxation of the Alpine orogen (Rutter et al., 124 2007; Wolff et al., 2012; Beltrando et al., 2015). The Finero Complex outcrops as an antiform in the 125 northernmost part of IVZ. The antiform core is composed by the Finero Phlogopite Peridotite 126 mantle unit (FPP; Cawthorn, 1975), which is wrapped by an intercalation of mafic-ultramafic 127 lithologies interpreted as pristine intrusive bodies (Cawthorn, 1975; Coltorti and Siena, 1984; Lu et 128 al., 1997a, b), and referred to as the Finero Mafic Complex (Fig. 1). The latter consists (from the 129 contact with the FPP outwards) of: i) the Layered Internal Zone (LIZ), ii) the Amphibole Peridotite

130 (AP) and iii) the External Gabbro (EG). To the N-NW, the Finero Complex is in contact across the

131 Insubric line with an accretionary prism of the Alpine orogeny, namely the Sesia-Lanzo Zone,

132 belonging to the Austroalpine domain (Fig. 1). To the S-SE, it is instead bounded by the metapelites

133 and metavolcanics, from granulite-to-amphibolite-facies, of the Kinzigite Formation, i.e. the

134 polymetamorphic basement of the Adria plate (Fig. 1).

135 In the FPP, stratiform to podiform chromitites mainly occur in dunite bodies. The petrogenesis of

the FPP chromitites has previously been discussed by Ferrario and Garuti (1990), Garuti et al.

137 (1997), Grieco et al. (2001, 2004) and Zaccarini et al. (2004), with the peculiar presence of zircons

138 being firstly recognised by Ferrario and Garuti (1990). The chromitites locally contain abundant Fe-

139 Ni-Cu-sulphides and Platinum Group Elements (PGE) minerals, but also zirconolite, baddeleyite,

140 thorianite, uraninite, thorite or huttonite (Grieco et al., 2004; Zaccarini et al., 2004).

141 The dunite bodies were produced by stages of channelled melt migration (Grieco et al., 2001, 2004;

142 Zaccarini et al., 2004). Most of them are elongated parallel to the mantle foliation and show sharp

143 contacts with the host phlogopite-harzburgite and pyroxenite association. The abrupt change of the

144 mineralogy is apparently the result of tectonic reactivation of the lithologic discontinuity. However,

some gradational transitions between dunites and host harzburgites are preserved (Zaccarini et al.,

146 2004). In particular, a gradational transition has been observed for peculiar dunite bodies having

147 elongations highly discordant to the mantle foliation (Giovanardi, 2012).

148 Field observations indicate that large concordant dunites (up to 20 m across) i) may be virtually free

149 from late magmatic segregations or layers; ii) contain chromitite layers (from a few mm to dm); iii)

150 possess chromitites layers associate to the late intrusion of magmatic sheets, (Giovanardi, 2012).

151 Field relationships indicate that the chromitite segregation preceded formation of the other

152 magmatic layers inside the dunite bodies, consistent with the observation reported by Grieco et al.

153 (2001, 2004).

154 The chromitite swarms here studied formed in dunite bodies outcropping in different sectors of the 155 FPP unit. At outcrop scale, no late phlogopite-amphibole-bearing magmatic layers are associated to

156 these chromitites. The samples selected are representative of two different chromitite end-members 157 (Fig. 2). In particular, one sample (MR01CR) records only a large segregation of chromite with 158 modest recovery of the previous texture dominated by deformed dunite olivine (hereafter Olchromitite; where Ol means olivine, Fig. 2c), while two samples document the complete recovery of 159 160 the dunite texture, which is totally replaced by secondary chromite and orthopyroxene (hereafter 161 Opx-chromitite, where Opx means orthopyroxene). The Ol-chromitite swarm, consisting of 1-cm thick parallel layers, outcrops in the area of Mount Sasso Rosso (MR01CR; Fig. 1). The chromitite 162 163 layers are made of large, anhedral, locally round chromite grains crystallised in a strongly-deformed porphyroclastic dunite characterised by the presence of olivine porphyroclasts embedded in a fine-164 grained matrix. The latter basically consists of olivine, with subordinate, anhedral orthopyroxene, in 165 part clearly overgrowing olivine. Very small clinopyroxene and amphibole grains have been 166 detected by SEM-EDS inspection, while phlogopite is missing. Thorite was recognized during 167 168 SEM-EDS inspection (Supplementary Material C). The two Opx-chromitites (FI09C04 and 169 FI09C34; Fig. 2a, b) outcrop along the Cannobino river to the south of Finero, close to the bridge towards Provola (Fig. 1). The single chromitite layers of the Cannobino occurrences, up to 6 cm in 170 171 thickness, locally merge in pockets up to 20 cm large (Fig. 2b). These chromitites show 172 allotriomorphic texture and are mainly composed by chromite and orthopyroxene (chromite 75-55% 173 by Vol.; orthopyroxene 35-15% by Vol.: Fig. 2d, e), with subordinate clinopyroxene, apparently in 174 textural equilibrium with the other minerals. In these samples, olivine rarely occurs as very small, 175 round relicts embedded in large orthopyroxenes. Phlogopite is missing, whereas rare amphibole grains, few tens of µm large, have been detected by SEM-EDS inspection. An Opx-chromitite 176 177 sample with similar modal composition to those here studied was documented by Grieco et al. 178 (2004).

According to Ferrario and Garuti (1990), Grieco et al. (2001) and Zaccarini et al. (2004), zircons are
up to 600 µm long. Euhedral zircons occur within chromite and sometimes in olivine. Most

- frequently, they are anhedral to subhedral, in interstitial position between chromite and olivine and/or orthopyroxene. Up to 25 zircons were observed in 50 mm² by Zaccarini et al. (2004).
- 183

184 Analytical methods

Separation of orthopyroxene, clinopyroxene and zircon was performed with magnetic and 185 chromatographic methods at the IGG-CNR, Pisa. The rocks were first grinded in two different 186 granulometries: 0.250 to 0.125 mm and less than 0.125 mm. Minerals were then concentrated and 187 purified by hand picking under a binocular microscope. Sixty-one zircons were separated from 188 MR01CR, 11 from FI09C34 and 54 from FI09C04 and mounted in resin. Zircon internal structure 189 was characterized with cathode-luminescence (CL) imaging by means of a SEM (Jeol JXA 840A 190 191 model) at IGG-CNR, Pavia. Geochronological data were obtained with ELA-ICP-HRMS at the 192 IGG-CNR, Pavia. The instrument couples an ArF excimer laser microprobe of 193 nm 193 (Geolas200Q-Microlas) with a ThermoFinnigan Element I ICP-HRMS. Mass signals 202 (Hg), 204 (Pb + Hg), 206 (Pb), 207 (Pb), 208 (Pb), 232 (Th) and 238 (U) were acquired in magnetic scan 194 mode (Tiepolo, 2003). The laser was operated at a repetition rate of 5 Hz with a pulse-energy of 195 about 12 J/cm²; Instrumental and laser-induced U/Pb fractionations were simultaneously corrected 196 197 using as external standard the 1065 Ma 91500 reference zircon (Widenbeck et al., 1995). The same 198 integration intervals and spot size were used on both the external standard and unknowns. During 199 each analytical run reference zircon 02123 (295 Ma; Ketchum et al., 2001) was analysed together with unknowns for quality control, accuracy resulted better than 99%. The spot size was set to 20 200 mm and laser fluency to 12J/cm². Data reduction was carried out using the "Glitter" software 201 202 package (van Achterbergh et al., 2001) setting at 1% the error of the external standard. During each 203 analytical run the reproducibility on the standards was propagated to all determinations according to 204 the equation in Horstwood et al. (2003). After this operation, analyses are considered accurate within quoted errors. All the analyses in the present work yield count rates for ²⁰⁴Pb at background 205 206 level therefore no common Pb correction was carried out. The reader would however consider that

207 the relatively high background of Hg hampers the detection of low signals for 204 Pb. Ages were 208 calculated for 207 Pb/ 206 Pb, 206 Pb/ 238 U and 207 Pb/ 235 U ratios with 2 σ error (Tab. 1) using Isoplot 209 software (Ludwig, 2003). Concordia ages were determined and concordia plots were constructed 210 using the same software. All errors in the text are given at 2s level.

211 Mineral major element analyses were conducted with the electron microprobe JEOL 8200 Super

212 Probe housed at the University of Milano (data are reported in Supplementary Material Tab. A) on

213 petrographic sections after carbon coating. Analytical conditions were 15 kV of acceleration

voltage, 15 nA of primary current beam, 10 s counting time for each element and 5 s counting time

215 for the background.

216 Trace element concentrations in minerals have been determined with a LA-ICP-MS housed at IGG-

217 C.N.R., Pavia (Supplementary Material Tab. B) consisting of a PerkinElmer SCIEX ELAN DCR-e

218 quadrupole ICP-MS coupled with a Q-switched Nd:YAG laser source, model Brilliant (Quantel),

219 whose fundamental emission (1064 nm) is converted to 266 nm by two harmonic generators. Spot

220 diameter was typically 50-60 µm. Data reduction was done with the GLITTER software, using the

221 reference synthetic glass NISTSRM 610 as external standard. Si was used as internal standards for

222 zircons, Ca for clinopyroxene. Precision and accuracy were assessed via repeated analysis of BCR-

223 2g reference material, resulting better than $\pm 10\%$ at ppm concentration level. More analytical

details are reported in Miller et al. (2012).

225 In-situ Hf isotopic compositions of zircon have been determined at the State Key Laboratory of Lithospheric Evolution, Institute of Geology and Geophysics of the Chinese Academy of Science 226 (Beijing, China). Zircons were analyzed using a Geolas-193 laser ablation coupled with a Neptune 227 228 Multi-Collector Inductively Coupled-Plasma Mass-Spectrometer (MC-ICP-MS) as described in Wu et al. (2006). During analyses, isobaric interference of ¹⁷⁶Lu on ¹⁷⁶Hf was corrected assuming 229 175 Lu/ 176 Lu = 0.02655; and the isobaric interference of 176 Yb on 176 Hf was corrected using the 230 average fractionation index measured from the individual analysis proposed by Iizuka and Hirata 231 (2005). Reference zircon 91500 (176 Hf/ 177 Hf = 0.282305, Wu et al., 2006) was used as a primary 232

standard for machine calibration, and also Mud Tank was used as a secondary reference material for data evaluation. During analyses, the obtained 176 Hf/ 177 Hf value from Mud Tank is 0.282521 ± 16 (2SD, n=12), which is consistent with the recommended value of 0.282507 ± 6 (2SD, n=5) within analytical uncertainty (Woodhead and Hergt, 2005). All of the Hf isotopic analyses were performed near the U-Pb spots, and the data are reported in Tab. 2. The calculation of ε_{Hf} and depleted mantle model age (T_{DM}) was done as in Wu et al. (2007). The O isotope composition of pure separates of orthopyroxene (from MR01CR, FI09C04 and

240 FI09C34 samples), clinopyroxene (FI09C04 and FI09C34 samples) and zircon (only from sample

241 MR01CR) have been analyzed at the I.G.G.-C.N.R., Pisa by conventional laser fluorination (Sharp,

242 1995) coupled with a Finnigan Delta Plus mass spectrometer. Analyses were performed following

243 methods described by Perinelli et al. (2011). Results are reported in Tab. 3.

- 244
- 245 **Results**

246

247 *U-Pb zircon ages*

248 Separated zircons are anhedral to subhedral, inclusion-free, but locally fractured. Zircons from the 249 Ol-chromitite are slightly pinkish, while those from the Opx-chromitites are colourless. All the 250 separated zircons show low cathodoluminescence (CL). No internal zoning is shown by the Opx-251 chromitite zircons (Fig. 3) and by most of those from the Ol-chromitite. However, some Ol-252 chromitite zircons display two different, broad internal domains, the core being slightly darker (Fig. 3). Most of the U-Pb LA-ICP-HRMS analyses of the zircons from Opx-chromitites provide Early 253 254 Jurassic concordant U-Pb ages. The twenty-nine analyses from twenty-seven FI09C04 zircons provided twenty concordant ages varying from 178 ± 5 to 199 ± 6 Ma (Fig. 4), with a concordia 255 256 ages of 187 ± 2 Ma (95% confidence level error, MSWD=1.8; Fig. 5). The other single-spots show slightly discordant U-Pb data, but with ²⁰⁶Pb/²³⁸U ages in the same interval defined by the 257 concordant ages, with exception of zircon 33 showing a Late Triassic age (208 ± 4 Ma; Fig. 4). 258

- Similarly, ten zircon grains from the FI09C34 sample yielded eight concordant ages ranging from 185 \pm 6 to 193 \pm 7 Ma, with a concordia ages of 187 \pm 1 Ma (95% confidence level error, MSWD = 3.2).
- Most of the analyses of the MR01CR zircons (twenty-eight out of thirty) also give Lower Jurassic concordant ages ranging mainly from 181 ± 6 to 197 ± 6 Ma and providing a concordia age of 188 ± 1 Ma (95% confidence level error, MSWD = 2.0; Fig. 5). However, the darker internal domain of some zircons provide older ages, two of them giving Triassic (Anisian-Carnian) concordant ages at 242 ± 7 Ma and 229 ± 7 Ma (Tab. 1; Fig. 3, 4).
- 267

268 Major and Trace Mineral chemistry

Chromites from the investigated samples cover the same compositional field defined by those from 269 270 other FPP chromitite layers documented by Grieco et al. (2001, 2004; Fig. 6). Chromites from the 271 Opx-chromitites also approach the compositions of those from the phlogopite harzburgites-272 pyroxenites association (Siena and Coltorti, 1989; Zanetti et al., 1999; Grieco et al., 2001), whereas 273 the Ol-chromites from Sasso Rosso show higher Cr#. The same behaviour is shown by the 274 pyroxenes composition, with those from the Opx-chromitites possessing larger Al₂O₃ contents (up 275 to 1.3 wt.%), which approach those in the phlogopite harzburgites-pyroxenites association, while 276 Ol-chromitite pyroxenes have a very Al-poor composition (0.6-0.1 wt.% Al₂O₃). The increase of Al 277 content in pyroxenes is accompanied by a significant decrease of Mg# of pyroxenes and olivine. As 278 a whole, Mg# of pyroxenes and olivines in chromitite layers are distinctly higher than in the 279 phlogopite harzburgites-pyroxenites association. 280 The clinopyroxene from the Opx-chromitites is strongly enriched in LREE with respect to MREE 281 and HREE (Fig. 7; La_N/Sm_N between 3.09-5.26 and La_N/Yb_N between 21.43-42.47). The REE

282 patterns are comparable with those of the clinopyroxene from the phlogopite harzburgite-pyroxenite

- association (SIMS analysis: Zanetti et al., 1999), but significantly different from those in late
- dolomite-apatite-wehrlites and apatite-orthopyroxenites and their host harzburgites (Zanetti et al.,

1999; Morishita et al., 2008). These relationships are also apparent by inspection of the PMnormalised spider diagrams (Primitive Mantle values from McDonough and Sun, 1995), in which
the patterns of Opx-chromitite clinopyroxenes match considerably those of the phlogopite
harzburgites-pyroxenites association, in particular sharing the very low Nb/LREE ratio, the positive
Sr anomaly and the very large Sc/HREE ratio. The trace element concentration of amphibole in
sample FI09C34 is strictly similar to those found in the country phlogopite harzburgites, as in the
associated clinopyroxenites.

292 Zircons from Opx-chromitites show very similar trace elements composition. REE patterns are

293 typically HREE-enriched, with La, Pr and Nd between 1-2 xCI (normalized to Chondrite I, CI:

294 Lyubetskaya and Korenaga, 2007), a strong positive Ce anomaly (~10 xCI), and a steady

enrichment from Sm to Lu (with maximum at Lu_N from 36 to 53) with a slight negative Eu anomaly

296 (Fig. 8). Th and U concentrations are both ~200 ppm: as a consequence the Th/U is ~1. Pb and Ti

are ~6 and ~19 ppm, respectively.

298 Zircons from Ol-chromitite display REE patterns with the same La and Lu concentration of the

299 Opx-chromitite ones (Fig. 8), but with slightly higher contents from Ce to Yb. Ce and Eu still

300 determine positive and negative anomalies, respectively. Th and U are higher (~480 and ~670 ppm,

301 respectively), with Th/U of ~0.7. Pb and Ti are 7-16 and 7-12 ppm, respectively.

In all the chromitite zircons, HREE and Y are distinctly lower than in the magmatic ones from the
External Gabbro (Zanetti et al., 2013) and the nepheline diorite pegmatites (Schaltegger et al., 2015)
of the Finero Complex.

305

306 *Hf and O isotope composition*

307 The zircons from the three studied samples share similar Hf isotopic ratios. In particular, ¹⁷⁶Hf/¹⁷⁷Hf

308 is between 0.282486±18-0.282582±16 and 0.282492±18-0.282587±19 for FI09C04 and FI09C34,

309 respectively, with the weighted average values identical within uncertainty (0.282542±11, MSWD

310	8.9, 95% conf. for FI09C04; 0.282535±22, MSWD 9.7, 95% conf., for FI09C34). MR01CR zircons
311	have 176 Hf/ 177 Hf between 0.282550±12-0.282610±13 for, with a weighted average slightly higher
312	thant that the Opx-chromitite zircons (0.282580 \pm 8, MSWD 7.9, 95% conf.). The calculated ϵ Hf ₍₁₈₈₎
313	span from -5.9 to -1.6 (Fig. 9), with weighted average values of -3.9 \pm 0.4 for FI09C04 (MSWD
314	7.6, 95% conf.), -4.2 \pm 0.8 for FI09C34 (MSWD 9.7, 95% conf.), and -2.7 \pm 0.3 for MR01CR
315	(MSWD 7.9, 95% conf.). Badanina et al. (2013), for zircons from FPP chromitite layers reported
316	176 Hf/ 177 Hf values similar to those obtained in this study (176 Hf/ 177 Hf between 0.282533-0282652
317	for ~90% of zircons).
318	Pyroxenes from Opx-Chromitites show very uniform O isotopic compositions, with positive
319	fractionation in orthopyroxene (δ^{18} O is 6.5-6.7‰ for clinopyroxene, 6.8-6.9‰ for orthopyroxene;
320	Fig. 10). Zircons from Ol-chromitite exhibit comparable δ^{18} O (6.8‰), but the associated
321	orthopyroxene has significantly lighter O isotopic composition (5.4‰). The δ^{18} O of the pyroxenes
322	from Opx-chromitites and of the zircons from the Ol-chromitite are significantly higher than the
323	typical range defined by mantle lithologies and mantle-derived melts (5.5-5.9‰ and 5.8-6.2‰,
324	respectively; see Bindeman, 2008 and references therein), as well as of the values reported by
325	Selverstone and Sharp (2011) for a series of lithologies from the FPP, but they more closely match
326	the compositions found in FPP amphiboles and phlogopites by Hartmann and Wedephol (1993).
327	X

328 **Discussion**

329 *Concepts on the origin of chromitites*

Ferrario and Garuti (1990), Garuti et al. (1997), Grieco et al. (2001, 2004), Zaccarini et al. (2004)
propose models in which dunites formation and chromitites segregation were linked to the

pervasive metasomatism experienced by the FPP, but with significant differences in terms of bothseries of processes and melt compositions.

In particular, Grieco et al. (2001) proposed that chromite layers and their dunite haloes formed by

interaction between basic melts and the ambient harzburgite. The residual melts of this process

invaded the country rock harzburgites, with precipitation of clinopyroxene and amphibole. In this

337 scenario, phlogopite crystallisation was a successive event related to the late intrusion of

338 clinopyroxenites, which induced K-metasomatism.

339 Zaccarini et al. (2004) concluded that chromitites and phlogopite metasomatism were the result of

340 the interaction of uprising alkaline-carbonatitic fluids with the ambient harzburgite in the

341 framework of mantle diapirism at the base of the continental crust induced by extensional tectonics.

342 Our data place further constraints on the geochemical affinity of the chromitite parent melts, as well

343 as on its compositional relationships with the metasomatic agent producing the phlogopite-

344 harzburgite and pyroxenite association. The possible effects of the late melt migrations recorded by

345 the FPP have to be evaluated, in particular in terms of zircon and pyroxenes

346 crystallisation/recrystallisation. It has been now widely documented that mantle chromitites after

347 their formation are particularly stable over a very large range of P-T-X conditions, and that they can

348 record the migration of different melts/fluids (Howell et al., 2015), sometimes associated to the

349 precipitation of zircons at mantle depths, over a very large time interval. This issue is relevant for

350 the interpretation of the geochemical evolution of FPP chromitites, because they show some zoning

351 of the mineral chemistry, and variations in the modal content of PGE minerals, which suggest

352 possible interactions with late fluids/melts (Grieco et al., 2001, 2004).

353

354 Geochemical constraints on chromitites and zircons origin

Several lines of evidence point to a strict geochemical affinity of the parent melts of the chromitite minerals with the metasomatic agents provoking the main metasomatic event of the FPP. First of all, this consideration is supported by the similarity of the major element composition of pyroxenes and spinels (in particular, in terms of very high Mg# and Cr/Al values; Fig. 6) and the evident consistency of the peculiar trace element compositions (i.e. enriched in Th, U and LREE, strongly depleted in HREE) of clinopyroxene and amphibole in the chromitites and in the phlogopite-

361 harzburgites and pyroxenites association (Fig. 7). The similar geochemical affinity of chromitites

and harzburgites and pyroxenites is also supported by the similar trace element fractionation shown 362 by the whole rock data reported by Grieco et al. (2001). The segregation of the chromitite zircons 363 from the same parent melt is suggested by their REE composition. Compared to the magmatic 364 zircons from the external gabbro and nepheline diorite pegmatites, they result enriched in LREE 365 and markedly depleted in HREE, similar to zircons segregated from mantle-derived kimberlites to 366 367 carbonatites (see Fig. 4 in Hoskin and Schaltegger, 2003). Even more stringent is the match of the peculiar, high δ^{18} O of the chromitite pyroxenes with those of amphiboles and phlogopites from the 368 369 phlogopite-harzburgites and pyroxenites association reported by Hartmann and Wedephol (1993). The identical O isotopic composition of MR01CR zircons and Opx-chromitite pyroxenes suggests 370 precipitation from a common parent melt. Although the δ^{18} O partition coefficient between zircon 371 and mafic phases is presently unconstrained, it has been widely documented that zircons segregated 372 by mantle-derived melts show a very small δ^{18} O interval at 5.3 ± 0.4 (Valley et al., 2005; 373 Bindeman, 2008; Tribuzio et al., 2014). A further valuable insight into the geochemical signature of 374 375 the parent melts of the chromitite zircons and on the possible relationships with late melts migrating through the Finero Complex is provided by the zircon ¹⁷⁶Hf/¹⁷⁷Hf ratios, which are much lower than 376 377 the depleted mantle array. The ε Hf₍₁₈₈₎ values are sub-chondritic, at -6.1 ± 0.6 to -1.6 ± 0.5, 378 consistent with the data by Badanina et al. (2013). This observation excludes any genetic relationship with the nepheline diorite pegmatites, whose zircons have $\varepsilon Hf_{(1)}$ between +6 to +9.8, 379 380 evidence of segregation from mantle-derived melts (Schaltegger et al., 2015). The melts involved in the pervasive metasomatic event of the FPP, besides having high δ^{18} O values. 381 382 were characterised by isotopic composition of Nd, Sr, Pb, H, S, Cl and noble gases indicating the presence of "crustal" components (Hunziker and Zingg, 1982; Voshage et al., 1987, 1988; 383 384 Cumming et al., 1987; Hartmann and Wedepohl, 1993; Obermiller, 1994; Seitz and Woodland, 385 2000; Downes, 2001; Matsumoto et al., 2005; Selverstone and Sharp, 2011). The melt migration processes have been mainly attributed to supra-subduction environments (see among others Zanetti 386

387 et al., 1999; Grieco et al., 2001, 2004; Morishita et al., 2003, 2008; Matsumoto et al., 2005), but 388 alternatively also to extensional settings (Garuti et al., 2001; Zaccarini et al., 2004). 389 Negative EHf values are interpreted in the literature as the result of continental crust recycling (e.g. Belusova et al., 2004; Scherer et al., 2007; Lee et al., 2007; Wu et al. 2007). The presence of 390 391 continental crustal component in the parent melts of chromitite is also strongly supported by the 392 high δ^{18} O zircon and by the large content in U and Th estimated for the parent melts by clinopyroxene composition and relevant clinopyroxene/liquid partition coefficients, but in particular 393 394 by the occurrence of thorianite, thorite and uraninite (Zaccarini et al., 2004; this study). 395 The occurrence of contrasting geochemical signatures in part suggesting oceanic crust derivation (Cumming et al., 1987; Selverstone and Sharp, 2011) may be tentatively interpreted as being related 396 397 to heterogeneity of the melt source and/or changes in the proportion of melt sources through time. 398

399 FPP Chromitite segregation model

400 Chromitites in dunite bodies are interpreted as late crystallization events of melts migrating into the dunite (Arai and Yurimoto, 1994; Arai, 1997). Such layers are common in dunites from supra-401 402 subduction zones, where chromite is basically associated to olivine. The formation of chromitite 403 and surrounding dunite envelope is mainly explained as the result of the interaction between exotic 404 melts and host harzburgite, in association with magma mixing (c.f. Zhou et al., 1994, 1996; Arai, 405 1997). The reference model assumes that in the first stage, an exotic SiO₂-undersaturated melt, 406 introduced into the ambient peridotite at low pressure, may selectively dissolve pyroxenes, as well 407 as hydrous minerals, and precipitate olivine producing a replacive dunite envelope. This process 408 would form a relatively Si-rich melt, according to the following reaction: SiO₂-poor melt + 409 pyroxenes + hydrous phases \rightarrow olivine + SiO₂-rich melt. If the dunite channel is further supplied by 410 the SiO₂-undersaturated primary melt, after mixing with the Si-rich melt, an over-saturation in 411 spinel components (Cr+Al) takes place, leading to the isolated precipitation of spinel (c.f. Arai, 412 1997).

413 Grieco et al. (2001, 2004) proposed that the FPP chromitites and their dunite haloes formed by the 414 interaction between basic melts and the ambient peridotites, where the segregation of phlogopite 415 pyroxenites resulting from successive events of melt migration, unrelated to the chromitites. 416 Instead, Zaccarini et al. (2004) suggested that chromitite layers, dunite channels and phlogopite harzburgites were the result of migration of alkaline-carbonatitic melts. 417 By contrast, Zanetti et al. (1999) stressed that the widespread precipitation of newly-formed, 418 magmatic orthopyroxene in both phlogopite harzburgites and pyroxenites pointed to a SiO₂-419 420 saturation of the metasomatic melts related to the pervasive recrystallisation of the FPP. As a 421 consequence, the formation of dunite bodies evidences peculiar variations in melt composition, with pulses of SiO₂-undersaturated melts determining the virtually complete resorption of pyroxenes, 422 423 amphibole and phlogopite in channels/bodies up to tens of meters across at relatively high, spinelfacies P conditions (see Mazzucchelli et al., 2009). The presence of dunite bodies both concordant 424 425 to discordant with respect to the mantle foliation, as well as some geochemical changes shown by the magmatic minerals precipitated within dunites indicate that SiO₂-undersaturated melts occurred 426 in different stages of the FPP metasomatic cycle (Giovanardi, 2012). 427 428 The porphyroclastic textures of the Ol-chromitite MR01CR suggests that the development of 429 structural weaknesses into the dunite bodies may have driven the migration and mixing of the 430 different melt components. The presence of SiO₂-saturated components in the parent melts is 431 confirmed by the nearly ubiquitous presence of newly-formed orthopyroxene replacing olivine. In 432 fact, this feature is present even in the first stages of chromitites formation solely characterised by chromite precipitation (in association to zircon and thorite as accessory mineral phases), as 433 434 documented by Ol-chromitite MR01CR. The SiO₂-saturation of the melt is more apparent in the Opx-chromitites, which record the complete recovery of the texture characterised by replacement of 435 436 olivine by secondary orthopyroxene in textural equilibrium with chromite. These petrographic 437 trends, along with the progressive chemical variation from the strongly refractory compositions of 438 the Ol-chromitites to relatively Al-Fe-richer compositions in the Opx-chromitites, allow us to

439 suggest that the Opx-chromitites are related to levels that experienced the largest time-integrated440 chromitite melt/dunite ratios.

441 Two different processes may be envisaged to explain the strong presence of such a metasomatic component in the parent melts of chromitites. The first, according to the Arai's model, is the result 442 of the dissolution of pyroxenes and hydrous minerals of the, already metasomatised, ambient 443 harzburgites-pyroxenites association upon interaction with uprising of mafic melts. This is 444 presumably the general process, always present in any FPP dunite body. However, the occurrence 445 in some dunites of late phlogopite pyroxenites, rich in orthopyroxenes, strictly similar with those 446 447 forming the main sequence (Grieco et al., 2001; Giovanardi, 2012) suggests that, at least locally, there might be mixing between mafic melts present in dunite channels and new upcoming SiO₂-448 449 saturated melts bearing the continental crustal component. The mixing of these two components, at 450 a new transition of the melt composition (i.e. from SiO₂-undersaturated to SiO₂-saturated), may 451 have triggered the precipitation of some chromitites, followed by a segregation of pyroxenites 452 within the dunite bodies.

453

454 Interpretation of the U-Pb ages

455 The age of the petrologic processes recorded by the FPP is still controversial due to a very large 456 time span documented by geochronological investigations, from the Early Permian to the Early 457 Jurassic (Voshage et al., 1987, 1988; Stähle et al., 1990, 2001; Hartmann and Wedephol, 1993; 458 Friedrichsen as cited by Hartmann and Wedephol, 1993; von Quadt et al., 1993; Grieco et al., 2001; Matsumoto et al., 2005; Morishita et al., 2008; Badanina et al., 2013; this work). Apparently late 459 460 intrusive or metasomatic events (i.e. those documented by Stähle et al., 1990, 2001; the alkaline 461 veins of Grieco et al., 2001; Matsumoto et al., 2005; Morishita et al., 2008) mainly provide Middle 462 Triassic to Early Jurassic ages (from 240 Ma to 195 Ma). The Triassic to Early Jurassic U-Pb ages 463 shown by chromitite zircons can be interpred in two different ways: i) the record of Jurassic,

- 464 channelled melt migration with preservation of some Triassic relicts, or ii) the result of the
 465 perturbation of the U-Pb zircon systems at Early Jurassic.
- The first scenario has some serious drawbacks, among which: 1) the abundance of Triassic to 466 Permian ages of chromitite zircons from the FPP documented by Grieco et al. (2001) (208 ± 2 Ma), 467 Badanina and Malitch (2012) and Badanina et al. (2013) (288 ± 7 Ma; 249 ± 3 Ma; 209 ± 4 Ma); 2) 468 the Triassic age of late alkaline bodies discondatly cutting the harzburgite-pyroxenite association 469 $(225 \pm 13 \text{ Ma}; \text{Stähle et al., } 2001); 3)$ the Depleted Mantle geochemical affinity of the intrusives of 470 471 the associate Finero Mafic Complex showing analogously Triassic to Early Jurassic radiometric data (231 \pm 23 Ma to 214 \pm 17 Ma: Lu et al., 1997a,b; 232 \pm 3 Ma to 214 \pm 5 Ma: Zanetti et al., 472 2013; 212.5 and 190 Ma: Schaltegger et al. 2015). Morever, the petrochemical observations 473 474 reported in the previous sections clearly indicate that the chromitite zircons were segregated in the 475 early metasomatic cycle producing the phlogopite harzburgites and pyroxenites association of FPP. 476 Thus, it is here proposed that the different age clusters exhibited by the FPP chromitite zircons are
 - 477 the result of progressive re-equilibration stages of the U-Pb system at subsolidus condition.
- This is consistent with the absence of CL zoning structures in most of the analyzed zircons of this
 study showing Early Jurassic ages. It is a common observation for mantle zircons, interpreted as the
- 480 evidence of compositional homogenization due to a prolonged residence at high temperature in
- 481 mantle conditions (Corfu et al., 2003).
- 482 The re-equilibration of the U-Pb system could most easily have occurred in fluid-assisted
- 483 conditions. Currently no mineralogical or geochemical data support this hypothesis which, however,484 cannot be discarded.
- In the framework depicted above, the 288 ± 7 Ma age provided by the pinkish zircon population with internal oscillatory-zoning of Badanina et al. (2013) is a minimum age of the FPP pervasive metasomatism. Such an Early Permian age would relate the FPP pervasive metasomatism to the transtensional regime affecting the Variscan orogen, and associated to the formation of the Mafic

489	Complex of the Val Sesia-Type IVZ, with the emplacement of large volumes of mantle-derived
490	tholeiitic melts at the bottom of the Adria crust (Zanetti et al., 2013 and references therein).
491	The peculiar composition of the metasomatic melts recorded by the FPP requires the concomitant
492	mobilisation of deep-seated reservoirs containing continental crust component. It was possibly
493	related to Variscan subduction of continental crust and metasomatism of the overlying mantle
494	wedge by crustal-derived melts / fluids at ~330 Ma (e.g. Ulten Area, Eastern Alps: Tumiati et al.,
495	2003; Sapienza et al., 2009; Langone et al., 2011).
496	

- 497

498 Evidence for peculiar P-T conditions of the Finero-Type IVZ

According to the reference model of Quick et al. (1995), the mantle bodies of Val Sesia-Type IVZ
were already intercalated into the crustal basement at least by the end of the Variscan orogeny,
having been progressively incorporated in the cumulates of the underplated Mafic Complex during
the Early Permian.

503 A pronounced re-equilibration of the U-Pb zircon system similar to that shown by chromitite 504 zircons has so far not been documented in the deepest rocks of the Val Sesia-Type IVZ. In particular, although zircons have never been found in associated mantle lithologies, they are 505 506 common in the gabbroic rocks of the Mafic Complex documenting processes down to 25 km depth 507 (i.e. ~0.8 GPa; Demarchi et al., 1998). Detailed inspections of magmatic zircons from the Mafic 508 Complex performed by Peressini et al. (2007) evidenced the dominant presence of Early Permian 509 ages, with only one Mesozoic age (180 Ma) given by a single-spot on recrystallised "white pest" 510 rim. Consistently, up-to-date reconstructions of the thermal evolution of the polymetamorphic 511 Kinzigite Formation of the IVZ do not provide evidence that the rifting of the Adriatic margin 512 during the Early Jurassic induced conditions capable to reset the U-Pb system in zircon and 513 monazite placed at crustal levels (Handy et al., 1999; Smye and Stockli, 2014; Ewing et al., 2015). 514 Locally, fluid-assisted partial recrystallization of zircon domains at ~220-200 Ma characterises

some IVZ metapelites of the Kinzigite Formation in the transitional zone between Val Sesia-Type IVZ and Finero-Type IVZ (Vavra et al., 1999; Ewing et al., 2013), presumably as a consequence of documented Late Triassic magmatism and of the related fluid activity. Vavra and Schaltegger (1999) also observed that monazites from the Kinzigite Formation yield a subconcordant discordia line with a lower intercept age of 210 ± 14 Ma, interpreted as an episode of fluid-driven Pb loss associated with the influx of hydrothermal fluids.

521 Thus, it is here speculated that the prolonged re-equilibration of the U-Pb system displayed by

522 chromitite zircons must be associated to peculiar P-T conditions affecting the FPP, such as

523 permanence at great (mantle) depths, possible till the Early Jurassic, and/or a reheating phase due to

a later (Early Jurassic) tectono-magmatic activity.

525 The residence of the FPP at relatively high pressure up to Mesozoic time is supported by the

presence of magmatic sapphirine in one of the late, if not the last, magmatic intrusions represented
by apatite-calcite-bearing gabbroic dyke swarms, discordantly cutting all the other rocks and

528 structures of the FPP. The precipitation of magmatic sapphirine in gabbroic rocks is consistent with

529 pressures above than 1.1 GPa (Giovanardi et al., 2013). In particular, the composition of sapphirine-

530 saturated melts corresponds to basalt to andesite at pressures of 1.1–1.5 GPa, and the stability field

of the magmatic sapphirine extends to P >3 GPa (Milholland and Presnall, 1999). Equilibrium

pressures exceeding those at the bottom of the Mafic Complex of the Val Sesia-Type IVZ (i.e. 0.8

533 GPa), have been also estimated for the Finero Mafic Complex by Siena and Coltorti (1989) at ~1.0

534 GPa (at ~1000°C). Accordingly, Sills et al. (1983) and Christy (1989) estimated 0.9-1.1 GPa (at T

535 of 800-950°C) for the subsolidus reaction involving formation of metamorphic sapphirine in the

536 gabbroic lithologies of LIZ.

537 The development of regional thermal positive perturbations in the northern part of the IVZ, possibly 538 associated to asthenosphere upwelling, can be inferred from the Triassic to the Early Jurassic cycles 539 of magmatic activity segregating zircons. This hypothesis is consistent with the high temperature 540 conditions (granulite-facies) argued for the lower IVZ by Brodie and Rutter (1987) in proximity of

541 the Anzola-Val Grande shear zone during the Middle to Late Triassic. It is also indirectly supported

542 by the change of the metamorphic conditions along the Pogallo Line, governed by brittle

543 deformation to the south of Val d'Ossola, and by ductile deformation to the north (Handy, 1987).

544 The temperatures recorded by the cooling paths along the Pogallo Line are consistently higher in

the northern than in the southern sector of the IVZ (Wolff et al., 2012).

546

547 Constraints on the Mesozoic geodynamic evolution of Finero-Type IVZ

548 Geochronological data supports multiple melt injections throughout the Triassic to Early Jurassic in
549 the Finero-Type IVZ (Zanetti et al., 2013 and references therein). The reappraisal of all data

550 available suggests that a first magmatic stage was dominated by segregation of gabbroic to

anorthositic rocks from ~240 to 230 Ma (Gebauer, 1993; Hingerl et al., 2008; Zanetti et al., 2013),

552 possibly associated to the emplacement of anatectic granites in the Kinzigite Formation adjacent to

the Finero Complex at 242 ± 3 Ma (Vignola et al., 2008), matching the oldest age found in the core

of zircons from MR01CR (242 ± 7 Ma). Volcanic activity also formed (241-238 Ma)

555 porphyroclastic intercalations in the pelagic succession of the western Southern Alps (Mundil et al.,

556 1996). A second main stage was characterised by the intrusion of nepheline diorite pegmatites at

557 225-190 Ma (Klötzli et al., 2007, 2009; Schaltegger et al., 2015). A Late Triassic magmatic event in

the FPP is recorded by the emplacement of apatite-carbonate-bearing alkaline diorite and

bornblendite dykes at 225-220 Ma and probably also apatite-carbonate-bearing orthopyroxenite

veins (240 \pm 41 Ma, phlogopite Ar-Ar (Matsumoto et al. 2005) and 213 \pm 35 Ma, apatite U-Pb

561 (Morishita et al., 2008).

562 The occurrence of anomalous heating processes at a regional scale has been confirmed by several

563 papers dealing with cooling ages of the IVZ and adjacent area (Wolff et al. 2012; Smye and Stockli,

564 2014; Ewing et al., 2013, 2015; Beltrando et al., 2015). In particular, Beltrando et al. (2015)

565 document the progressive westward rejuvenation of (U-Th)/(He) ages (hereafter ZHe ages), from

566 280-240 Ma in the Lombardian basin to 215-200 Ma near the Sostegno and Fenera basins,

567 indicating that anomalously high thermal gradients were established in the Late Triassic towards the 568 area where the actual rifting of Alpine Tethys was later localized. This suggests that rift localization along the western margin of the Adriatic plate was probably favoured by a lithospheric thermal 569 570 anomaly, established at 215-210 Ma, followed by thermal decay at 200-190 Ma (Ewing et al., 2013, 2015; Beltrando et al., 2015). 571 572 The Early Jurassic (~200-180 Ma) age interval provided by most of the zircons from the FPP chromitites broadly corresponds to the final stages of extensional faulting as recorded in the IVZ by 573 574 the Pogallo Line, which was active between 210 and 170 Ma (Zingg et al., 1990), and the Anzola-Val Grande high-T shear zone (Brodie and Rutter, 1987; Brodie et al., 1989), whose movement is 575 considered to have spanned the period between 230-180 Ma. It also matches the final stages of 576 577 development of the Lombardian Basin at ~220-180 Ma (Bertotti et al., 1999 and references threin), located just east of the IVZ. Although the geodynamic setting of the Middle Triassic deformation 578 579 stages of the IVZ is still debated (see Zanetti et al., 2013), there is a wide consensus that the Late 580 Triassic-Early Jurassic deformation phases was a precursor events of the opening of the Alpine Tethys, which involved crustal thinning, mantle exhumation and a partial rotation of the IVZ (15° to 581 582 23° of tilting; Wolff et al., 2012). It is thus proposed that the age interval shown by colourless 583 smoky chromitite zircons and, in particular, the closure of the U-Pb system of the mantle zircons at ~180 Ma, document the exhumation stage of the FPP. 584 585 Smye and Stockli (2014) evidenced that the IVZ underwent a reheating event of sufficient duration and T to reset the U-Pb system of rutile in granulites of the Kinzigite Formation at ~180-190 Ma, 586

587 possibly due to hyperextension of the Adriatic lithosphere. An Early Jurassic heating has been also

588 invoked to explain the resetting of the ZHe thermochronometer in the Baveno granite (Wolff et al.,

589 2012). Subsequent crust-wide extension led to breakup of continental crust and mantle exhumation.

590 ZHe ages in detrital zircons from syn-tectonic sandstone constrain the onset of normal faulting in

the axial zone at 185-180 Ma (Beltrando et al., 2015). It is thus concluded that the ages shown by

592 FPP chromitite zircons record thermal perturbations in the Triassic-Early Jurassic time span, the

youngest one reflecting hyperextension of the Adriatic lithosphere (Smye and Stockli, 2014), and
regional Early Jurassic magmatic activity (Mazzucchelli et al., 2010; Schaltegger et al., 2015). The
absence of evidence for partial melting in the FPP suggests that such thermal perturbations never
exceeded 965°C, which is the solidus temperature extimated for the Finero phlogopite harzburgites
(Giovanardi et al., 2013).

598

599 Concluding Remarks

Field relationships, the major element composition of spinel and pyroxenes, the trace element composition of clinopyroxene and zircon, the O isotopic composition of zircon and pyroxenes, and the Hf isotopic composition of zircon converge in indicating that the chromitite layers here studied were segregated from hybrid melts derived from the mantle but stronly contaminated by continental crust.

The FPP chromitite zircons yield ages spanning the Early Permian to the Early Jurassic, interpreted 605 606 as indicating primary crystallization in the Early Permian and resetting during thermal disturbances in the Jurassic. The chromitites zircon data indicate that the FPP was at mantle depths since the 607 608 Early Permian, being exhumed at shallower, crustal levels only during Early Jurassic. The youngest 609 event appears to have been connected to initiation of continental rifting and mantle exhumation, 610 precursor events of the opening of the Alpine Tethys. Our data, along with those of Grieco et al. 611 (2001), Badanina and Malitch (2012) and Badanina et al. (2013), support that lithosphere rifting and 612 exhumation were affected by two strong thermal perturbations at 208 Ma and 187 Ma. 613 In our model for the northern IVZ, the pervasive metasomatism of the FPP occurred ~290 Ma 614 and/or before. However, the possibility that the actual age of pervasive metasomatism of FPP was 615 older and related to the Variscan orogenic cycle cannot be excluded. 616

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- 627
- 628 Supplementary Material Captions
- 629 Table A: Major-element composition of mineral phases as wt.% and a.p.f.u. Formulae.
- 630 Table B: Trace-element compositions of zircons, clinopyroxenes and amphiboles in ppm and631 single-analysis elements detection limits.
- 632 Supplemetary Material C: SEM images and EDS analysis of thorite in sample MR01CR.
- 633

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- 928

929 Figure Captions

Figure 1 - (a) Sketch map of the Ivrea-Verbano Zone, which represents the westernmost part of the

- 931 Southern Alps. The rectangle indicates the location of the map of the Finero Complex in (b). CL,
- 932 Cremosina Line; IL, Insubric Line; CMBL, Cossato-Mergozzo-Brrissago Line; PL, Pogallo Line;
- HTSZ, high-temperature shear zone of the Anzola-Val Grande area (Brodie and Rutter, 1987;
- Rivalenti and Mazzucchelli, 2000; Rutter et al., 2007). (b) Sketch map of the Finero Complex
- 935 modified after Steck and Tièche (1976). The empty stars document the locations of the zircon-
- 936 bearing chromitites here studied.
- 937
- Figure 2 (a) The thickest layer of the FI09C34 chromitite swarm; (b) Chromitite pocket in sample
- 939 FI09C34; (c) Texture of MR01CR chromitite, which is characterised by segregation of anhedral
- 940 large chromite overgrowing a matrix formed by porphyroclastic dunite; (d) Allotriomorphic texture
- 941 of chromite FI09C04 formed by chromite and orthopyroxene, where rounded olivine relicts rarely

942 occur within large orthopyroxene; (e) Allotriomorphic texture of chromitite FI09C34, with presence943 of serpentinisation along the grain boundary.

944

- 945 Figure 3 Cathodoluminescence images of zircons from FI09C04, FI09C34 and MR01CR.
- 946 FI09C04 and FI09C34 zircons are virtually free from internal structures, while some of those from
- 947 MR01CR show broad darker areas, usually returning relatively older ages than the lighter ones.
- 948 Spot analyses are reported toghether with single-spot concordant ages.

949

- 950 Figure 4 Age histogram and relative probability diagram of chromitite zircons of this study.
- 951 Maximum probability age is reported for each sample.

952

Figure 5 - Concordia ages calculated with ²⁰⁶Pb/²³⁸U and ²⁰⁷Pb/²³⁵U ratios for chromitite zircons
from MR01CR, FI09C34 and FI09C04 samples.

- Figure 6 Major element mineral chemistry of chromitites of this study. Data from literature are
 reported for comparison: FPP harzburgite-pyroxenite association (Harz-Py) from Zanetti et al.
 (1999) (a) and Grieco et al. (2001) (c); average FPP harzburgite (Avg. Harz) from Siena and
 Coltorti (189) (b); chromitite veins (Chromitite) from Grieco et al. (2001, 2004) (c, d)..
- 961 Figure 7 (a) CI-normalised (Lyubetskaya and Korenaga, 2007) REE patterns and (b) Pyrolite-
- 962 normalised (McDonough and Sun, 1995) extended trace element diagrams of clinopyroxenes from
- 963 FI09C04 and FI09C34 chromitites. Literature data are reported for comparison: amphibole from
- apatite-bearing wehrlites (Ap-Wehrl Amph) and its host harzburgite (Harz Ap-free Amph) from
- 965 Morishita et al. (2008) (a); clinopyroxene from the harzburgite-pyroxenite association (Harz-Py
- 966 Cpx) and apatite-bearing wehrlite (Ap-Wehrl Cpx) from Zanetti et al. (1999) (b); clinopyroxene
- 967 from harzburgite (Harz Cpx) and chromitite veins (Chromitite Cpx) from Grieco et al. (2001) (c).

968

Figure 8 – (a) CI-normalised (Lyubetskaya and Korenaga, 2007) REE patterns and (b) Pyrolitenormalised (McDonough and Sun, 1995) extended trace element diagrams for zircons from the
chromitites of this study. Data from zircons from (a) gabbros of the EG (Zanetti et al., 2013) and (b)
from alkaline dykes within the Mafic Complex (Schaltegger et al., 2015) are reported for
comparison.

974

Figure 9 – ϵ Hf_(t) vs U-Pb age for the zircons from the chromitites of this study. Literature data of (a) zircons from miaskite-type nepheline pegmatites in the Finero Mafic Complex (Schaltegger et al., 2015) and (b) FPP chromitites (Badanina et al., 2013). The Depleted Mantle (DM) evolution line is calculated using the values of present-day ¹⁷⁶Hf/¹⁷⁷Hf ratio of 0.28325 from Nowell et al., 1998, and ¹⁷⁶Lu/¹⁷⁷Hf ratio of 0.0384 from Griffin et al., 2000). CHUR values are from Blichert-Toft and Albarede (1997).

981

Figure 10 – Isotopic oxygen composition (δ^{18} O mineral vs. SMOW‰) of orthopyroxene (Opx) and 982 983 clinopyroxene (Cpx) separates from FI09C04 and FI09C34 chromitites, and of orthopyroxene and 984 zircon (Zrc) separates from MR01CR. FPP data for minerals (Ol = olivine; Cpx = clinopyroxene; 985 Amph = amphibole; Phl = phlogopite) of the harzburgite-pyroxenite association (Harz-Py) from 986 Hartmann and Wedephol (1993) (a) and Selverstone and Sharp (2011) (b) are reported for 987 comparison, as well as the range of mantle peridotites and mantle-derived melts from Bindeman 988 (2008) (*) and zircon values from a hernblende-gabbro and a hornblendite from the Ligurian 989 ophiolites (Tribuzio et al., 2014) (c).

990

991 Table Captions

Table 1: summary of ELA-ICP-HRMS U-Pb zircon analysis from chromitite layers from FPP.993

994 Table 2: summary of MC-ICP-MS *in-situ* Hf isotopic compositions of zircon from FPP.

995

- 996 Table 3: Isotopic oxygen composition (δ^{18} O vs. SMOW‰) of mineral separates from the FPP
- 997 chromitites here studied . Numbers between parenthesis represent the number of replicates of the
- 998 measurements on different aliquots of the same sample.

999

Chilling and a second s

Fig. 1 1000 PINE (a) Ronco Ν Finero Brissago P/ Cannobio /a ITALY Val Fiorina Cannobina Grande G Val Premosello d'Ossola 10 Km SOUTHERNALPS Balmuccia Val Sesser "Serie dei Laghi" Ivrea-Verbano Zone Permian Volcanics Mafic Complex Permian Intrusions Kinzigitic Formation ++++ vrea Mantle Peridotite **Gneisses and Schists** Baldissero Faults Cento Valli (b) AUSTROM PINE DOMAIN Val MR01CI Vigezzo Sasso Rosso FI09C3 FI09C04 Monte Torriggia Finero Phlogopite Peridotite Layered Internal Zone rovola Cursolo Amphibole Peridotite 옷을 External Gabbro FI 90 1 km Series . Val Cannobina 1001 Kinzigite Formation

1003 Fig. 2



1006 Fig. 3

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K K









1031 Table 1: summary of ELA-ICP-HRMS U-Pb zircon analysis from chromitite layers from FPP.

Samp	MR01															
le:	CR			Ratio						Ages						
Zirco	Positi	²⁰⁷ Pb/ ²⁰		²⁰⁶ Pb/ ²³		²⁰⁷ Pb/ ²³		²⁰⁷ Pb/ ²⁰		²⁰⁶ Pb/ ²³	2	²⁰⁷ Pb/ ²³		Concor	2	% of
n	on	۴Pb	1σ	⁸ U	1σ	⁵U	1σ	۴Pb	2σ	⁸ U	σ	⁵U	2σ	dia	σ	discordance
			0.00		0.00		0.00				5.					
7	core	0.0498	11	0.0292	05	0.2005	49	187.1	8.3	185.4	9	185.6	9.1			
			0.00		0.00		0.00				6.				6.	
9	core	0.0497	11	0.0309	05	0.2115	51	178.6	7.8	196.2	2	194.8	9.4	196.1	1	0.69%
			0.00		0.00		0.00				5.				5.	
9	rim	0.0507	11	0.0290	05	0.2027	49	227.7	9.9	184.3	8	187.4	9.0	184.4	7	-1.70%
			0.00		0.00		0.00				6.				5.	
10	core	0.0492	11	0.0299	05	0.2029	50	157.4	7.0	190.0	0	187.6	9.2	189.9	9	1.31%
			0.00		0.00		0.00		11.		5.				5.	
10	rim	0.0511	12	0.0292	05	0.2057	51	244.0	2	185.7	9	189.9	9.5	185.8	8	-2.29%
			0.00		0.00		0.00				6.				6.	
12	core	0.0503	11	0.0310	05	0.2148	53	209.8	9.5	196.9	1	197.5	9.7	196.9	0	-0.34%
			0.00		0.00		0.00	-	13.		6.		11.		6.	
12	rim	0.0511	14	0.0301	05	0.2111	62	245.3	5	191.2	2	194.4	4	191.3	1	-1.68%
			0.00		0.00		0.00		10.		7.		11.		7.	
14	core	0.0509	12	0.0361	06	0.2531	62	238.1	8	228.7	1	229.1	2	228.7	0	-0.19%
			0.00		0.00		0.00				5.				5.	
16	core	0.0494	11	0.0292	05	0.1981	49	165.0	7.6	185.3	8	183.5	9.2	185.2	7	0.95%
			0.00		0.00		0.00		20.		6.		14.		6.	
16	rim	0.0516	20	0.0291	05	0.2079	81	269.1	4	184.8	6	191.8	9	184.9	5	-3.80%
47		0.0500	0.00		0.00	0.0460	0.00		10.		6.	400.0	~ ~	407.0	6.	0.000/
1/	core	0.0506	12	0.0311	05	0.2163	54	222.2	2	197.1	2	198.9	9.9	197.2	1	-0.88%
4.0			0.00		0.00		0.00			406.0	6.		10.	100.1	6.	0.670/
18	core	0.0498	13	0.0309	05	0.2116	58	186.1	9.6	196.2	2	194.9	/	196.1	1	0.67%
20		0.0500	0.00	0.0204	0.00	0.2440	0.00	224.0	10.	402.0	6.	405.4	0.0	402.0	6.	4 4 70/
20	core	0.0506	11	0.0304	05	0.2118	52	224.0	0	192.8	1	195.1	9.6	192.9	0	-1.1/%
24		0.0402	0.00	0.0200	0.00	0.2025	0.00	457.0	7.0	100.0	5.	407.2	07	400.0	5.	1 100/
21	core	0.0492	12	0.0299	05	0.2025	52	157.8	7.6	190.0	9	187.2	9.7	189.9	8	1.48%
22		0.0500	0.00	0.0204	0.00	0.2010	0.00	224.0	13.	405.0	6.	406 7	11.	405.4	5.	0.000/
23	core	0.0506	15	0.0291	05	0.2018	62	224.0	2	185.0	0	186.7	5	185.1	9	-0.88%
24		0.0503	0.00	0.0207	0.00	0 2054	0.00	204.2	10.	100 7	6. 1	100 7	10.	100 7	ь. О	0 5 6 9/
24	core	0.0502	12	0.0297	05	0.2054	54	204.3	11	188.7		189.7	11	188.7	-0	-0.56%
25		0.0513	0.00	0.0202	0.00	0 2007	0.00	252.0	11.	242.2	7.	242.4	11.	242.2	7.	0.110/
25	core	0.0513	0.00	0.0385	0.00	0.2697	00	253.0	30	242.2	4	242.4	11	242.2	5	-0.11%
25	rim	0.0510	0.00	0.0200	0.00	0 2125	0.00	2022	20.	100 1	0. ว	10E 7	14.	100.2	ס. כ	2 0.2%
25	r i i i i	0.0519	0.00	0.0299	0.00	0.2125	78	282.3	4	190.1	5	195.7	4	190.2	5	-2.92%
26	coro	0.0401	0.00	0.0296	0.00	0 1021	0.00	150.7	70	192.0	5.	170.2	80	101 0	э. И	1 / 2%
20	LUIE	0.0491	0.00	0.0280	0.00	0.1931	0.00	150.7	12	102.0	5	179.5	10	101.0	5	1.40%
77	coro	0.0507	0.00	0 0202	0.00	0 2028	57	7777	12.	195 7	٦. ٥	100 2	10. 6	195 7	J. 7	1 / 2%
27	LUIE	0.0307	0.00	0.0292	0.00	0.2038	0.00	227.7	11	105.7	5	100.5	10	105.7	5	-1.4376
28	core	0.0507	12	0 0295	0.00	0 2055	54	225.8	11.	187 7	٦. ۵	180.8	10.	187 8	٦. ۵	-1 1/1%
20	core	0.0507	0.00	0.0255	0.00	0.2055	0 00	225.0	12	107.7	6	105.0	10	107.0	5	1.1470
28	rim	0.0508	13	0.0295	0.00	0.2055	57	222.2	12. 1	187 2	0. N	189 7	±0. 5	187 २	9. 9	-1 37%
20		0.0000	0.00	0.0255	0 00	0.2000	0.00	252.2	-	107.2	5	100.7	5	107.5	5	1.5770
30	core	0.0504	11	0.0284	0.00	0,1968	49	212 1	9.7	180 7	у. 6	182 4	9.0	180 R	5	-0 95%
50		0.0004	0.00	0.0204	0.00	0.1500	0.00	<i><i><i>L</i> 1 <i>L</i> 1</i></i>	12	100.7	6	102.4	12	100.0	5	0.5570
30	rim	0 0499	16	0 0289	05	0 1977	66	188 9	1	183.8	0	183 1	1	183.8	9	0 38%
20		0.0 + 3 3	0.00	0.0205	0.00	0.1011	0.00	100.0	19	105.0	7.	100.1	17.	100.0	6	0.0070
33	core	0.0504	23	0.0300	06	0.2085	95	211.6		190.4	0	192.3	4	190.5	9	-0.98%
		0.000 +	0.00	0.0000	0.00	0.2000	0.00	211.0	-	200.7	5	102.0		150.5	5	0.50%
34	core	0.0502	12	0.0291	04	0.2012	51	204.3	9.6	184.9	7	186.1	9.4	185.0	6	-0.64%
0.	0010	0.0002	0.00	0.0201	0.00	0.2012	0.00	20110	5.0	10	6.	10011	10.	20010	6.	010 170
35	core	0.0495	13	0,0300	05	0.2048	57	173.5	9.1	190.4	2	189.2	-5	190.3	1	0.61%
		0.0100	0.00	2.0000	0.00		0.00	275.5	10	100.1	6	100.2	11	200.0	6	0.01/0
35	rim	0.0497	14	0.0301	05	0.2071	62	182.4	4	191.4	2	191.1	4	191.4	1	0.14%
		2.0.07	0.00		0.00		0.00	101.1	15		5.		11.		5.	0.1
38	core	0.0516	15	0.0289	05	0.2047	63	265.9	8	183.5	8	189.1	6	183.6	7	-3.08%
'			0.00		0.00		0.00		11.		5.		-		5.	/*
39	core	0.0510	12	0.0290	04	0.2031	52	239.9	4	184.1	6	187.8	9.5	184.2	6	-1.99%
		'	0.00		0.00		0.01		17.		9.	~	20.		9.	
Std021	.23	0.0514	18	0.0466	08	0.3309	16	257.0	7	293.5	5	290.2	3	293.4	3	1.10%
			-		'		-				-		-			

Samp le:	FI09C 04			Ratio						Ages				U-Pb		
Zirco	Positi	²⁰⁷ Pb/ ²⁰		²⁰⁶ Pb/ ²³		²⁰⁷ Pb/ ²³		²⁰⁷ Pb/ ²⁰	_	²⁰⁶ Pb/ ²³	2	²⁰⁷ Pb/ ²³	_	Concor	2	% of
n	on	°Pb	1σ	°U	1σ	٠U	1σ	°Pb	2σ	°U	σ	<u> </u>	2σ	dia	σ	discordance
26	core	0.0588	0.00 19	0.0297	0.00	0.2412	0.00 78	557.8	35. 4	188.9	5. 4	219.4	14. 2			-16,12%
	00.0	0.0000	0.00	0.0207	0.00	0.2.12	0.00	00710	15.	20010	4.					10112/
26	rim	0.0546	11	0.0298	04	0.2243	48	394.6	7	189.3	6	205.4	8.8			-8.53%
27		0.0400	0.00	0.0207	0.00	0 4 0 7 5	0.00	402.2	12.	402.2	5.	102.0	11.	402.2	4.	0.440
27	core	0.0499	16	0.0287	0.00	0.1975	63 0.00	192.2	0 16	182.2	5	183.0	12	182.2	9 4	-0.44%
28	core	0.0513	16	0.0288	0.00	0.2039	66	255.7	2	183.2	0	188.4	2	183.2	 9	-2.87%
			0.00		0.00		0.00		14.	C	5.		15.		5.	
29	core	0.0496	20	0.0305	04	0.2088	84	174.9	0	193.7	6	192.5	5	193.7	5	0.62%
30	core	0 0494	0.00	0 0308	0.00	0 2000	0.00 54	165.9	8 1	195 7	5. 0	103 5	٩٩	105.6	4. 0	1 1 3 %
30	COLE	0.0494	0.00	0.0308	0.00	0.2099	0.00	105.5	10.	195.7	5.	193.5	9.9 16.	195.0	5.	1.1370
31	core	0.0482	22	0.0292	05	0.1949	91	110.1	3	185.4	9	180.8	8	185.3	8	2.47%
			0.00		0.00		0.00		14.		4.		12.		4.	
32	core	0.0510	16	0.0291	04	0.2042	65	239.5	9	184.7	9 1	188.6	0	184.7	8	-2.15%
32	rim	0.0493	0.00	0.0286	0.00	0.1943	0.00 44	162.6	6.9	181.6	4. 5	180.3	8.2	181.5	4. 4	0.72%
-		010100	0.00	0.0200	0.00	0.10	0.00		32.	101.0	5.	10010	14.	10110	•	017270
33	core	0.0584	17	0.0328	05	0.2645	80	544.8	0	208.3	7	238.3	4			-14.41%
~ .		0 0500	0.00		0.00		0.00		10.	400 5	4.	400.0		100.0	4.	1.05%
34	core	0.0509	11	0.0298	04	0.2093	49	235.4	4 27	189.5	5	193.0	9.0 11	189.6	/	-1.85%
35	core	0.0586	0.00 14	0.0310	0.00	0.2507	65	553.4	4	196.9	J. 1	227.2	7			-15.39%
			0.00		0.00		0.00		13.		4.					
36	core	0.0521	12	0.0288	04	0.2070	49	287.6	0	183.3	5	191.0	9.1		_	-4.22%
27	coro	0.0512	0.00	0 0202	0.00	0 2072	0.00	252.1	17.	196.0	5. ว	101.2	13.	106 1	5. ว	2 70%
57	LUIE	0.0515	0.00	0.0295	0.00	0.2072	0.00	252.1	23.	180.0	4.	191.2	10.	100.1	2	-2.79%
38	core	0.0580	13	0.0308	04	0.2461	58	528.6	6	195.4	8	223.4	5			-14.33%
			0.00		0.00		0.00		36.		4.		11.			
39	core	0.0672	14	0.0306	04	0.2832	64	842.7	1	194.2	8	253.2	5		-	-30.40%
40	core	0 0501	0.00	0 0290	0.00	0 2008	0.00 87	201 5	17. A	184.2	5. 7	185.8	16. 1	184.2	5. 6	-0 88%
-0	core	0.0501	0.00	0.0250	0.00	0.2000	0.00	201.5	13.	104.2	, 5.	105.0	14.	104.2	5.	0.0070
41	core	0.0495	20	0.0286	04	0.1957	79	170.6	8	181.7	4	181.5	6	181.7	4	0.08%
		0 0500	0.00	0.0004	0.00	0.0405	0.00		17.	400.0	5.	100 5	14.	400.0	5.	1 700/
42	core	0.0509	19	0.0304	04	0.2135	79 0.00	234.9	3 25	193.2	5 1	196.5	6	193.2	4	-1.70%
43	core	0.0592	13	0.0283	0.00	0.2313	0.00 54	573.7	2 <i>5</i> . 4	180.2	- . 5	211.3	9.9			-17.27%
		7	0.00		0.00		0.00		10.		4.				4.	
44	core	0.0506	12	0.0289	04	0.2011	49	220.4	0	183.3	6	186.0	9.0	183.4	5	-1.45%
16	coro		0.00	0 0 2 9 1	0.00	0 1060	0.00	ד רכר	13.	170 0	4. °	100 E	10.	170 /	4.	2 2 20/
40	LUIE	0.0508	0.00	0.0281	0.00	0.1909	0.00	252.7	27.	170.5	ہ 4.	102.5	10.	170.4	'	-2.52%
47	core	0.0602	14	0.0302	04	0.2508	60	609.7	5	191.9	9	227.2	8			-18.38%
			0.00		0.00		0.00				5.		11.		5.	
48	core	0.0494	15	0.0301	04	0.2048	63	164.5	9.9	191.2	2	189.2	6	191.1	1	1.04%
50	coro	0.0508	0.00	0 0200	0.00	0 2080	0.00	220 /	13.	190.9	5. 0	102.6	11. 6	190.9	4. 0	1 50%
50	COLE	0.0508	0.00	0.0299	0.00	0.2089	0.00	230.4	19.	109.0	6.	192.0	17.	105.0	6.	-1.50%
51	core	0.0502	24	0.0293	05	0.2045	96	202.4	1	186.4	1	188.9	7	186.4	0	-1.34%
			0.00		0.00		0.00	c	11.		4.				4.	
52	core	0.0509	12	0.0291	04	0.2042	52	235.4	3 15	185.0	8	188.7	9.6	185.1	7	-1.96%
53	core	0.0499	0.00 21	0.0313	0.00	0.2150	0.00 90	189.4	тэ. 9	198.8	э. 7	197.7	10. 5	198.8	э. 6	0.54%
			0.00		0.00	00	0.00		20.	5.0	5.		16.		5.	5.5 .70
54	core	0.0510	22	0.0293	05	0.2056	88	240.4	6	185.9	9	189.9	3	185.9	8	-2.15%
C+-10-2-1	20	0.05205	0.00	0.0471	0.00	0.3443	0.01	226 -	27.	2074	8.	200 5	25.	207.2	8.	4 4 6 6 1
Sta021	.23	0.05295	22	/	07	9	43	326.7	2	297.1	9	300.5	U	297.2	/	-1.10%

1036 Table 1: continue.

Sample:	FI09C34			Ratio						Ages				U-I
Zircon	Position	²⁰⁷ Pb/ ²⁰⁶ Pb	1σ	²⁰⁶ Pb/ ²³⁸ U	1σ	²⁰⁷ Pb/ ²³⁵ U	1σ	²⁰⁷ Pb/ ²⁰⁶ Pb	2σ	²⁰⁶ Pb/ ²³⁸ U	2σ	²⁰⁷ Pb/ ²³⁵ U	2σ	Concord
1	core	0.0494	0.0022	0.0290	0.0005	0.1977	0.0087	166.9	15.2	184.5	6.5	183.2	16.2	184
2	core	0.0517	0.0015	0.0293	0.0005	0.2084	0.0057	270.8	15.5	186.0	6.0	192.2	10.6	186
3	core	0.0488	0.0023	0.0304	0.0005	0.2044	0.0093	138.2	13.1	192.9	6.7	188.8	17.2	192
4	core	0.0492	0.0021	0.0292	0.0005	0.1987	0.0080	157.4	13.2	185.6	6.3	184.0	14.9	185
5	core	0.0516	0.0017	0.0291	0.0005	0.2066	0.0065	266.4	17.3	184.8	6.2	190.7	12.0	184
6	core	0.0599	0.0017	0.0270	0.0005	0.2230	0.0062							
6	rim	0.0564	0.0014	0.0282	0.0004	0.2185	0.0052	466.6	23.1	179.0	5.6	200.7	9.5	
7	core	3.8182	3.0059	0.0285	0.0228	15.0262	2.7782							
8	core	0.0569	0.0019	0.0310	0.0005	0.2424	0.0078	486.5	32.7	196.7	6.5	220.4	14.2	
10	core	0.0504	0.0012	0.0294	0.0004	0.2042	0.0046	213.0	10.0	186.7	5.6	188.7	8.4	186
10	rim	0.0513	0.0030	0.0295	0.0006	0.2080	0.0120	252.1	30.0	187.5	7.6	191.9	22.1	187
11	core	0.0506	0.0011	0.0296	0.0005	0.2070	0.0044	224.5	10.1	188.3	5.8	191.0	8.1	188
Std02123		0.0530	0.0021	0.0464	0.0008	0.3388	0.0128	329.2	25.9	292.1	9.9	296.3	22.4	292

1041 Table 1: continue.

043 Table 2: summary of MC-ICP-MS *in-situ* Hf isotopic compositions of zircon from FPP.

1043 1044

sample	zirco n	positio n	Age (Ma)	¹⁷⁶ Yb/ ¹⁷⁷ H f	2σ	¹⁷⁶ Lu/ ¹⁷⁷ H f	2σ	¹⁷⁶ Hf/ ¹⁷⁷ H f	2σ	e _{Hf} (0)	e _{нf} (t)	2σ	T_{DM}	T _{DM} ^C	$\mathbf{f}_{\mathrm{Lu/Hf}}$
MR01C R	9	core	188	0.001629	0.00003 9	0.000052	0.00000	0.282580	0.00001 5	-6.77	- 2.65	0.5 3	927	139 5	- 1.00
MR01C R	10	core	188	0.000730	0.00000	0.000022	0.00000 0	0.282582	0.00001 2	-6.73	- 2.60	0.4 4	925	139 1	- 1.00
MR01C R	12	rim	188	0.000479	0.00000	0.000015	0.00000	0.282577	0.00001	-6.89	2 77	0.4 4	931	140 2	-
MR01C	14	core	188	0.000533	0.00002	0.000017	0.00000	0.282584	0.00001	-6.64	- 2 51	0.4	921	138	-
MR01C	16	core	188	0.000603	0.00001	0.000017	0.00000	0.282596	0.00001	6.24	2.31	0.4	906	136 0	-
MR01C	17	core	188	0.000620	0.00001	0.000017	0.00000	0.282575	0.00001	-0.24 6.06	2.11 - 2.02	0.4	022	140 6	-
MR01C	20	core	188	0.000661	0.00000	0.000019	0.00000	0.282610	0.00001	-0.90	2.05	0.4	935	132	-
MR01C	21	core	188	0.000742	0.00001	0.000024	0.00000	0.282600	0.00001	-5.73	1.00	0.3	000	8 135	-
R MR01C	23	core	188	0.000550	0.00002	0.000017	0.00000	0.282553	0.00001	-6.09	1.97	9 0.4	900	1 145	1.00
R MR01C	24	core	188	0.000573	0.00000	0.000017	0.00000	0.282585	1 0.00001	-7.74	3.61	0.4	964	5 138	1.00
R MR01C	25	rim	188	0.000420	4 0.00000	0.000013	0.00000	0.282550	3 0.00001	-6.61	2.48	7 0.4	920	4 146	1.00
R MR01C	26	core	188	0.000534	6 0.00000	0.000016	0 0.00000	0.282552	2 0.00001	-7.85	3.73	1 0.4	968	2 145	1.00 -
R MR01C	27	core	188	0.000836	6 0.00002	0.000025	0 0.00000	0 282595	3 0.00001	-7.79	3.66 -	5 0.4	966	8 136	1.00 -
R MR01C	28	core	188	0.000461	6 0.00000	0.000013	1 0.00000	0 282601	2 0.00001	-6.25	2.12	1 0.4	906	1 134	1.00 -
R MR01C	20	coro	100	0.000401	7 0.00000	0.000013	0 0.00000	0.2825001	3 0.00001	-6.05	1.92 -	8 0.4	898	8 135	1.00 -
R MR01C	20	rim	100	0.000492	5 0.00001	0.000014	0 0.00000	0.202555	3 0.00002	-6.11	1.98	7 0.7	900	2 144	1.00 -
R MR01C	50		100	0.000990	2 0.00001	0.000055	0 0.00000	0.202550	0 0.00001	-7.63	3.51 -	2 0.4	960	9 143	1.00 -
R MR01C	33	core	188	0.000382	5 0.00001	0.000012	0 0.00000	0.282564	3 0.00001	-7.34	3.22	7 0.4	948	0 141	1.00 -
R MR01C	34	core	188	0.000404	1 0.00000	0.000012	0 0.00000	0.282570	4 0.00001	-7.15	3.03	8 0.5	941	8 134	1.00
R MR01C	35	core	188	0.000783	7	0.000024	0	0.282601	5	-6.05	1.92	4	898	9 141	1.00
R MR01C	38	core	188	0.000403	0	0.000013	0	0.282572	2	-7.06	2.93	4	937	2	1.00
R	39	core	188	0.000409	4	0.000012	0.00000	0.282570	5	-7.16	3.03	3	941	8	1.00
FI09C04	26	core	188	0.000531	0.00000	0.000021	0.00000	0.282527	0.00001	0.00	-	0.6	100	151	-
FI09C04	26	rim	188	0.000872	0.00001	0.000033	0.00000	0.282566	9 0.00001	-8.68	4.55	6 0.5	0	5 142	-
FI09C04	27	core	188	0.000569	9 0.00000	0.000022	1 0.00000	0.282550	6 0.00001	-7.27	3.15	5 0.5	946	6 146	1.00
FI09C04	28	core	188	0.000576	5 0.00002	0.000024	0 0.00000	0.282582	6 0.00001	-7.85	3.72	5 0.5	968	2 139	1.00 -
FI09C04	29	core	188	0.000577	4 0.00000	0.000022	1 0.00000	0 282579	6 0.00001	-6.74	2.61	8 0.5	925	2 139	1.00 -
FINGCOA	30	core	188	0.000377	5 0.00000	0.000017	0 0.00000	0.282534	6 0.00001	-6.82	2.70	5 0.5	928	7 149	1.00 -
FI09C04	21	coro	100	0.000442	7 0.00003	0.000017	0 0.00000	0.2025334	6 0.00001	-8.41	4.28 -	7 0.6	990 100	8 152	1.00 -
FI09C04	22	core	100	0.001212	7 0.00000	0.000047	1 0.00000	0.202525	9 0.00002	-8.81	4.69 -	6 0.7	6	4 145	1.00 -
FI09C04	32	core	188	0.000572	9 0.00000	0.000022	0 0.00000	0.282554	0 0.00001	-7.71	3.58 -	0 0.6	963	4 147	1.00 -
FI09C04	33	core	188	0.000638	9 0.00000	0.000025	0 0.00000	0.282546	8 0.00001	-7.99	3.86 -	5 0.5	973	1 151	1.00 -
FI09C04	34	core	188	0.000613	7	0.000024	0	0.282528	7	-8.63	4.51	8 0.7	998	2 151	1.00
FI09C04	35	core	188	0.000626	8	0.000025	0	0.282528	1	-8.64	4.52	3	999 105	3	1.00
FI09C04	36	core	188	0.000754	9.00000	0.000028	0.00000	0.282489	9	10.00	- 5.88	6	2	9	1.00

FI09C04	37	core	188	0.000605	0.00001	0.000025	0.00000	0.282582	0.00002	-6 71	- 2 59	0.7	924	139 1	- 1 00
F109C04	38	core	188	0.000544	0.00001	0.000021	0.00000	0.282486	0.00001	-0.71	2.59	0.6	105	160	-
FI09C04	39	core	188	0.000619	1 0.00001	0.000026	0.00000	0.282524	8 0.00002 1	10.13 8 76	6.01 -	4 0.7	6 100 2	/ 152	1.00
					T		0		T	-8.70	4.04	4	3	0	1.00
<u> </u>	zirco	positio	on Age	¹⁷⁶ Yb/ ¹⁷⁷ H		¹⁷⁶ Lu/ ¹⁷⁷ H		¹⁷⁶ Hf/ ¹⁷⁷ H		е _{нf} (0	e _{∺f} (t		_	- (
sample	n		(Ma)	f	2σ	f	2σ	ŕ	2σ))	2σ	Т _{DM}	Трм	t _{Lu/Hf}
FI09C0	40	core	188	0.000497	0.00001	0.000020	0.00000	0.282522	0.00002	0.04	-	0.8	100	152	-
4 FI09C0					8 0.00000		0.00000		0.00002	-8.84	4.71	0.7	ь 100	5 152	1.00
4	41	core	188	0.000628	8	0.000024	0	0.282522	2	-8.84	4.72	9	7	5	1.00
FI09C0	42	core	188	0.000493	0.00000	0.000020	0.00000	0.282553	0.00001		-	0.5		145	-
					6 0.00001		0		6	-7.73	3.61	7	963	5 1/10	1.00
4	44	core	188	0.000690	2	0.000025	0.00000	0.282536	0.00001	-8.36	4.23	9	988	5	1.00
FI09C0	46	core	188	0 000976	0.00001	0 000037	0.00000	0 282576	0.00001		-	0.6		140	-
4	10	core	100	0.000370	8	0.000037	0	0.202570	8	-6.94	2.81	4	933	5	1.00
F109C0 4	48	core	188	0.000558	0.00000	0.000022	0.00000	0.282518	0.00001	-8.97	- 4.85	0.6	101	153 4	-
FI09C0	50		100	0.000510	0.00001	0 000022	0.00000	0 202570	0.00001	0.57	-	0.6	-	140	-
4	50	core	188	0.000510	8	0.000022	1	0.282578	7	-6.86	2.74	2	930	0	1.00
FI09C0	51	core	188	0.000901	0.00000	0.000035	0.00000	0.282519	0.00001	0.00	-	0.6	101	153	-
4 FI09C0					0.00003		0.00000		0.00001	-8.90	4.84	0.6	1	3 147	-
4	52	core	188	0.001005	5	0.000041	1	0.282544	9	-8.05	3.93	8	976	6	1.00
FI09C0	53	core	188	0.000646	0.00001	0.000026	0.00000	0.282550	0.00001		-	0.6		146	-
4		00.0	100	01000010	0		0	0.202000	8	-7.84	3.72	5	968	2	1.00
					0.00001		0 00000		0 00002		_	0.6	103	156	_
4	1	core	188	0.000808	0.00001	0.000031	0.00000	0.282503	0.00002	-9.52	5.40	0.0 9	3	9	1.00
FI09C3	2	core	188	0 000783	0.00002	0.000032	0.00000	0 282532	0.00001		-	0.6		150	-
4	-	core	100	0.000705	0	0.000032	1	0.202332	9	-8.50	4.38	9	994	4	1.00
F109C3	3	core	188	0.000886	0.00004	0.000038	0.00000	0.282530	0.00002	-8.55	4.43	0.7	996	150	-
FI09C3	4		100	0.000575	0.00001	0 000022	0.00000	0 202520	0.00001	0.00	-	0.6	550	150	-
4	4	core	100	0.000575	4	0.000023	0	0.282529	8	-8.58	4.45	4	996	9	1.00
FI09C3	5	core	188	0.000646	0.00001	0.000028	0.00000	0.282492	0.00001	0.00	- E 70	0.6	104	159	-
4 FI09C3				\mathbf{V}	0.00003		0.00000		0.00003	-9.90	5.76	5 1.1	0	5 146	-
4	6	rim	188	0.000925	6	0.000039	1	0.282551	2	-7.80	3.68	3	967	0	1.00
FI09C3	8	core	188	0.000498	0.00000	0.000019	0.00000	0.282577	0.00001		-	0.6		140	-
4 ELOQC2	-				4		0		8 0 0000 0	-6.90	2.78	2	931	2 150	1.00
4	10	core	188	0.000427	8	0.000018	0.0000.0	0.282529	0.00002	-8.59	4.46	2	997	130 9	- 1.00
FI09C3	10	rim	199	0 000651	0.00000	0 000020	0.00000	0 282587	0.00001		-	0.6		138	-
4	10	1111	100	0.000031	7	0.000029	0	0.202307	9	-6.55	2.43	9	918	1	1.00
FI09C3	11	core	188	0.000881	0.00001	0.000039	0.0000.0 0	0.282533	0.00001 7	-8.46	-	U.5 a	992	150 2	-
-7					1		0		1	0.40	+.J+	J	552	4	1.00

1045

1046 Table 2: continue.

1048

- 1049 Table 3: Isotopic oxygen composition (δ18O vs. SMOW‰) of mineral separates from the FPP
- 1050 chromitites here studied . Numbers between parenthesis represent the number of replicates of the
- 1051 measurements on different aliquots of the same sample.
- 1052

Sample	Phase	δ ¹⁸ O avg.	Std. Dev.
MD01CD	Zrc (1)	6.80	
WIKUICK	Opx (2)	5.36	0.05
5100004	Opx (2)	6.86	0.05
F109C04	Срх (3)	6.73	0.06
F100C24	Opx (2)	6.76	0.06
F109C34	Cpx (2)	6.53	0.06

1053

6.53 0.06

1054 Highlights:

- 1055 zircon-bearing mantle chromitites formed along with the Finero phlogopite harzbugites
- 1056 the Finero Phlogopite Peridotite was firstly metasomatised in Early Permian or before
- 1057 younger ages record thermal perturbations related to tectono-magmatic events
- 1058 the exhumation of Finero Phlogopite Peridotite occurred in Early Jurassic
- 1059 the IVZ exhumation was accompanied by a thermal perturbation at ~187 Ma