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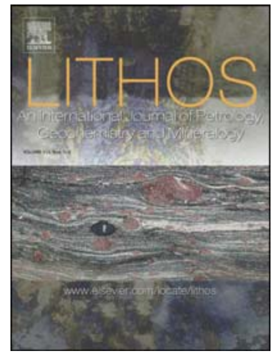

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Origin and age of zircon-bearing chromitite layers from the Finero Phologopite Peridotite (Ivrea-Verbano Zone, western Alps) and geodynamic consequences

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Abstract

An investigation has been performed on three chromitite layers segregated in dunite bodies of the Phlogopite Peridotite mantle unit in the Finero Complex (FPP, Ivrea–Verbano Zone, Southern Alps) aimed at providing new constraints to their origin and evolution.

Field relationships, the sub-chondritic Hf isotopic composition of the zircons ($\varepsilon Hf_{(188)}$ as low as -5.4), the heavy O isotopic composition of zircons and pyroxenes ($\delta^{18}O$ up to 6.9‰), the strict similarity of the trace element composition between the clinopyroxenes and amphiboles from the 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 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 FPP. A LA-ICP-HRMS survey of chromitite zircon grains has provided Early Jurassic U-Pb ages 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, 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 textures, but the same sub-chondritic Hf isotopic composition, which define an overall U-Pb age span from ~290 Ma to 180. The segregation of the chromitite layers and the main pervasive metasomatism likely occurred in the Early Permian (in a post-collisional, transtensional setting) or before (possibly, in a subduction-related setting). The rejuvenation of the zircon ages was accompanied by a progressive disapperance of the internal zoning, interpreted as the result of a prolonged residence at mantle depths with progressive re-equilibration of the U-Pb system due to 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 event, possibly due to lithospheric hyperextension. The evolution of the FPP appears completely 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 Variscan orogeny.

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

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 structural, petrochemical and age anomalies with respect to such a model. For instance, there is now abundant evidence of the occurrence in the Finero Mafic Complex (northernmost IVZ) of 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 (e.g. zircon-bearing diorites and nepheline diorite pegmatites) mostly showing Late Triassic to Early Jurassic ages (Oppizzi and Schaltegger, 1999; Grieco et al., 2001; Klötzli et al., 2007, 2009; Schaltegger et al., 2015, and references therein). Late Triassic intrusions of zircon-bearing diorites, associated to hornblendites (Stähle et al., 1990, 2001; Grieco et al., 2001), are also recorded by the associate mantle unit, the Finero Phlogopite Peridotite (FPP), which also shows peculiar late, discordant swarms of apatite-calcite-bearing gabbroic veins characterised by the presence of magmatic sapphirine (Giovanardi et al., 2013). A 225 Ma age has been found in a recrystallised rim of a zircon from a metasedimentary septum included in the intrusive External Gabbro unit and in a monazite from the adjacent Kinzigite Formation (Langone and Tiepolo, 2015).

These observations suggest the occurrence of two kinds of IVZ, the “Val Sesia”-Type (central IVZ) and the “Finero”-Type (northern IVZ), assumed to have had different tectono-magmatic evolutions (Rivalenti and Mazzucchelli, 2000; Zanetti et al., 2013, 2014; Mazzucchelli et al., 2014). Further main differences are the petrochemical features of the mantle bodies. The Val Sesia-Type mantle 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). Conversely, the Finero-Type mantle bodies are enriched in phlogopite and amphibole due to diffuse modal metasomatism (Rivalenti and Mazzucchelli, 2000; Zanetti et al., 2013; Mazzucchelli et al., 2014). According to the distribution of the different kinds of mantle peridotites, Zanetti et al. (2013) have speculatively proposed that their boundary may corresponds to the Anzola-Val Grande high temperature shear zone (HTSM in Fig. 1).

The FPP is the biggest mantle body of the northern IVZ. It shows a virtually complete recrystallization due to pervasive to channelled melt migrations (Zanetti et al., 1999). The pervasive metasomatism formed a main lithologic association constituted by phlogopite harzburgites associated to phlogopite pyroxenites (mainly websterites and orthopyroxenites), which do not show significant chemical gradients among them (Zanetti et al., 1999). The channelled migration stages formed dunite bodies, often containing stratiform to podiform chromites and, more rarely, 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 al., 2006; Selverstone and Sharp, 2011; Giovanardi, 2012; Mazzucchelli et al., 2014). Relatively late melt migration events also formed peridotite and pyroxenite veins and bands (Zanetti et al., 1999, Grieco et al., 2001; Morishita et al., 2003, 2008; Matsumoto et al. 2005; Raffone et al., 2006). These are often characterised by the presence of apatite and carbonates, and usually exhibit marked modal and chemical gradients with respect to the host phlogopite harzburgite. Examples of these lithologies are the apatite-dolomite-bearing wehrlites (Zanetti et al., 1999; Raffone et al., 2006), as well as the apatite-bearing orthopyroxenites described by Morishita et al. (2003, 2008), and Matsumoto et al. (2005), these latter displaying Triassic ages.

Thanks to such a unique lithological association, the FPP is one of the most studied mantle sequence in the world (Fig. 1). Several papers have dealt with the age of the petrochemical processes and the geochemical affinity of the melts that migrated through it. These melts only marginally match the magmatic record of the associated crustal rocks of the Finero Mafic Complex

(Giovanardi et al., 2014). Despite this, the geodynamic evolution of the FPP is still controversial and the geochemical affinity of the metasomatic melts strongly debated. In particular, geochronological surveys reported in the literature give a very large age interval, spanning from the Early Permian to the Early Jurassic, which is difficult to reconcile with the evidence that the mantle bodies of the central IVZ (i.e. the Balmuccia body) were emplaced within the continental crust, as part of accretionary prisms, at or before the end of the Variscan orogeny.

A special opportunity to place geochronological constraints on the evolution of the FPP is represented by the occurrence of large amounts of primary zircon in the chromitites (Ferrario and 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 parent melts, ii) the age of the metasomatism and iii) the geodynamic evolution of the mantle sequences in the Finero-Type IVZ.

Geological setting and selected samples

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

(AP) and iii) the External Gabbro (EG). To the N-NW, the Finero Complex is in contact across the Insubric line with an accretionary prism of the Alpine orogeny, namely the Sesia-Lanzo Zone, belonging to the Austroalpine domain (Fig. 1). To the S-SE, it is instead bounded by the metapelites and metavolcanics, from granulite-to-amphibolite-facies, of the Kinzigite Formation, i.e. the polymetamorphic basement of the Adria plate (Fig. 1).

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. (1997), Grieco et al. (2001, 2004) and Zaccarini et al. (2004), with the peculiar presence of zircons being firstly recognised by Ferrario and Garuti (1990). The chromitites locally contain abundant Fe-Ni-Cu-sulphides and Platinum Group Elements (PGE) minerals, but also zirconolite, baddeleyite, thorianite, uraninite, thorite or huttonite (Grieco et al., 2004; Zaccarini et al., 2004).

The dunite bodies were produced by stages of channelled melt migration (Grieco et al., 2001, 2004; Zaccarini et al., 2004). Most of them are elongated parallel to the mantle foliation and show sharp contacts with the host phlogopite-harzburgite and pyroxenite association. The abrupt change of the 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., 2004). In particular, a gradational transition has been observed for peculiar dunite bodies having elongations highly discordant to the mantle foliation (Giovanardi, 2012).

Field observations indicate that large concordant dunites (up to 20 m across) i) may be virtually free from late magmatic segregations or layers; ii) contain chromitite layers (from a few mm to dm); iii) possess chromitites layers associate to the late intrusion of magmatic sheets, (Giovanardi, 2012).

Field relationships indicate that the chromitite segregation preceded formation of the other magmatic layers inside the dunite bodies, consistent with the observation reported by Grieco et al. (2001, 2004).

The chromitite swarms here studied formed in dunite bodies outcropping in different sectors of the FPP unit. At outcrop scale, no late phlogopite-amphibole-bearing magmatic layers are associated to these chromitites. The samples selected are representative of two different chromitite end-members (Fig. 2). In particular, one sample (MR01CR) records only a large segregation of chromite with modest recovery of the previous texture dominated by deformed dunite olivine (hereafter Ol-chromitite; where Ol means olivine, Fig. 2c), while two samples document the complete recovery of the dunite texture, which is totally replaced by secondary chromite and orthopyroxene (hereafter 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 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-grained matrix. The latter basically consists of olivine, with subordinate, anhedral orthopyroxene, in part clearly overgrowing olivine. Very small clinopyroxene and amphibole grains have been detected by SEM-EDS inspection, while phlogopite is missing. Thorite was recognized during SEM-EDS inspection (Supplementary Material C). The two Opx-chromitites (FI09C04 and 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 thickness, locally merge in pockets up to 20 cm large (Fig. 2b). These chromitites show allotriomorphic texture and are mainly composed by chromite and orthopyroxene (chromite 75-55% by Vol.; orthopyroxene 35-15% by Vol.: Fig. 2d, e), with subordinate clinopyroxene, apparently in textural equilibrium with the other minerals. In these samples, olivine rarely occurs as very small, 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 sample with similar modal composition to those here studied was documented by Grieco et al. (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 mm^2 by Zaccarini et al. (2004).

Analytical methods

Separation of orthopyroxene, clinopyroxene and zircon was performed with magnetic and chromatographic methods at the IGG-CNR, Pisa. The rocks were first grinded in two different granulometries: 0.250 to 0.125 mm and less than 0.125 mm. Minerals were then concentrated and purified by hand picking under a binocular microscope. Sixty-one zircons were separated from MR01CR, 11 from FI09C34 and 54 from FI09C04 and mounted in resin. Zircon internal structure was characterized with cathode-luminescence (CL) imaging by means of a SEM (Jeol JXA 840A model) at IGG-CNR, Pavia. Geochronological data were obtained with ELA-ICP-HRMS at the IGG-CNR, Pavia. The instrument couples an ArF excimer laser microprobe of 193 nm (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 mode (Tiepolo, 2003). The laser was operated at a repetition rate of 5 Hz with a pulse-energy of about 12 J/cm^2; Instrumental and laser-induced U/Pb fractionations were simultaneously corrected using as external standard the 1065 Ma 91500 reference zircon (Widenbeck et al., 1995). The same integration intervals and spot size were used on both the external standard and unknowns. During 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 mm and laser fluency to 12J/cm^2. Data reduction was carried out using the “Glitter” software package (van Achterbergh et al., 2001) setting at 1% the error of the external standard. During each analytical run the reproducibility on the standards was propagated to all determinations according to 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 ^{204}Pb at background level therefore no common Pb correction was carried out. The reader would however consider that the relatively high background of Hg hampers the detection of low signals for ^{204}Pb. Ages were calculated for $^{207}Pb/^{206}Pb$, $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$ ratios with 2σ error (Tab. 1) using Isoplot software (Ludwig, 2003). Concordia ages were determined and concordia plots were constructed using the same software. All errors in the text are given at 2s level.

Mineral major element analyses were conducted with the electron microprobe JEOL 8200 Super Probe housed at the University of Milano (data are reported in Supplementary Material Tab. A) on 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 for the background.

Trace element concentrations in minerals have been determined with a LA-ICP-MS housed at IGG-C.N.R., Pavia (Supplementary Material Tab. B) consisting of a PerkinElmer SCIEX ELAN DCR-e quadrupole ICP-MS coupled with a Q-switched Nd:YAG laser source, model Brilliant (Quantel), whose fundamental emission (1064 nm) is converted to 266 nm by two harmonic generators. Spot diameter was typically 50-60 μm. Data reduction was done with the GLITTER software, using the reference synthetic glass NISTSRM 610 as external standard. Si was used as internal standards for zircons, Ca for clinopyroxene. Precision and accuracy were assessed via repeated analysis of BCR-2g reference material, resulting better than ±10% at ppm concentration level. More analytical details are reported in Miller et al. (2012).

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 (Beijing, China). Zircons were analyzed using a Geolas-193 laser ablation coupled with a Neptune Multi-Collector Inductively Coupled-Plasma Mass-Spectrometer (MC-ICP-MS) as described in Wu et al. (2006). During analyses, isobaric interference of ^{176}Lu on ^{176}Hf was corrected assuming $^{175}Lu/^{176}Lu = 0.02655$; and the isobaric interference of ^{176}Yb on ^{176}Hf was corrected using the average fractionation index measured from the individual analysis proposed by Iizuka and Hirata (2005). Reference zircon 91500 ($^{176}Hf/^{177}Hf = 0.282305$, Wu et al., 2006) was used as a primary 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 FI09C34 samples), clinopyroxene (FI09C04 and FI09C34 samples) and zircon (only from sample MR01CR) have been analyzed at the I.G.G.-C.N.R., Pisa by conventional laser fluorination (Sharp, 1995) coupled with a Finnigan Delta Plus mass spectrometer. Analyses were performed following methods described by Perinelli et al. (2011). Results are reported in Tab. 3.

Results

U-Pb zircon ages

Separated zircons are anhedral to subhedral, inclusion-free, but locally fractured. Zircons from the Ol-chromitite are slightly pinkish, while those from the Opx-chromitites are colourless. All the separated zircons show low cathodoluminescence (CL). No internal zoning is shown by the Opx-chromitite zircons (Fig. 3) and by most of those from the Ol-chromitite. However, some Ol-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 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 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 $^{206}Pb/^{238}U$ ages in the same interval defined by the concordant ages, with exception of zircon 33 showing a Late Triassic age (208 ± 4 Ma; Fig. 4).

Similarly, ten zircon grains from the FI09C34 sample yielded eight concordant ages ranging from 185 ± 6 to 193 ± 7 Ma, with a concordia ages of 187 ± 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).

Major and Trace Mineral chemistry

Chromites from the investigated samples cover the same compositional field defined by those from other FPP chromitite layers documented by Grieco et al. (2001, 2004; Fig. 6). Chromites from the Opx-chromitites also approach the compositions of those from the phlogopite harzburgites-pyroxenites association (Siena and Coltorti, 1989; Zanetti et al., 1999; Grieco et al., 2001), whereas the Ol-chromites from Sasso Rosso show higher Cr#. The same behaviour is shown by the pyroxenes composition, with those from the Opx-chromitites possessing larger Al_2O_3 contents (up to 1.3 wt.%), which approach those in the phlogopite harzburgites-pyroxenites association, while Ol-chromitite pyroxenes have a very Al-poor composition (0.6-0.1 wt.% Al_2O_3). The increase of Al content in pyroxenes is accompanied by a significant decrease of Mg# of pyroxenes and olivine. As a whole, Mg# of pyroxenes and olivines in chromitite layers are distinctly higher than in the phlogopite harzburgites-pyroxenites association.

The clinopyroxene from the Opx-chromitites is strongly enriched in LREE with respect to MREE 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 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 PM-normalised 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.

Zircons from Opx-chromitites show very similar trace elements composition. REE patterns are typically HREE-enriched, with La, Pr and Nd between 1-2 xCI (normalized to Chondrite I, CI: 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 (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.

Zircons from Ol-chromitite display REE patterns with the same La and Lu concentration of the Opx-chromitite ones (Fig. 8), but with slightly higher contents from Ce to Yb. Ce and Eu still determine positive and negative anomalies, respectively. Th and U are higher (~480 and ~670 ppm, 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.

Hf and O isotope composition

The zircons from the three studied samples share similar Hf isotopic ratios. In particular, $^{176}Hf/^{177}Hf$ is between 0.282486±18-0.282582±16 and 0.282492±18-0.282587±19 for FI09C04 and FI09C34, respectively, with the weighted average values identical within uncertainty (0.282542±11, MSWD

8.9, 95% conf. for FI09C04; 0.282535±22, MSWD 9.7, 95% conf., for FI09C34). MR01CR zircons have $^{176}Hf/^{177}Hf$ between 0.282550±12-0.282610±13 for, with a weighted average slightly higher thant that the Opx-chromitite zircons (0.282580±8, MSWD 7.9, 95% conf.). The calculated $\varepsilon Hf_{(188)}$ span from -5.9 to -1.6 (Fig. 9), with weighted average values of -3.9 ± 0.4 for FI09C04 (MSWD 7.6, 95% conf.), -4.2 ± 0.8 for FI09C34 (MSWD 9.7, 95% conf.), and -2.7 ± 0.3 for MR01CR (MSWD 7.9, 95% conf.). Badanina et al. (2013), for zircons from FPP chromitite layers reported $^{176}Hf/^{177}Hf$ values similar to those obtained in this study ($^{176}Hf/^{177}Hf$ between 0.282533-0282652 for ~90% of zircons).

Pyroxenes from Opx-Chromitites show very uniform O isotopic compositions, with positive fractionation in orthopyroxene ($\delta^{18}O$ is 6.5-6.7‰ for clinopyroxene, 6.8-6.9‰ for orthopyroxene; Fig. 10). Zircons from Ol-chromitite exhibit comparable $\delta^{18}O$ (6.8‰), but the associated orthopyroxene has significantly lighter O isotopic composition (5.4‰). The $\delta^{18}O$ of the pyroxenes from Opx-chromitites and of the zircons from the Ol-chromitite are significantly higher than the typical range defined by mantle lithologies and mantle-derived melts (5.5-5.9‰ and 5.8-6.2‰, respectively; see Bindeman, 2008 and references therein), as well as of the values reported by Selverstone and Sharp (2011) for a series of lithologies from the FPP, but they more closely match the compositions found in FPP amphiboles and phlogopites by Hartmann and Wedephol (1993).

Discussion

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 both series 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 scenario, phlogopite crystallisation was a successive event related to the late intrusion of clinopyroxenites, which induced K-metasomatism.

Zaccarini et al. (2004) concluded that chromitites and phlogopite metasomatism were the result of the interaction of uprising alkaline-carbonatitic fluids with the ambient harzburgite in the framework of mantle diapirism at the base of the continental crust induced by extensional tectonics. Our data place further constraints on the geochemical affinity of the chromitite parent melts, as well as on its compositional relationships with the metasomatic agent producing the phlogopite-harzburgite and pyroxenite association. The possible effects of the late melt migrations recorded by the FPP have to be evaluated, in particular in terms of zircon and pyroxenes crystallisation/recrystallisation. It has been now widely documented that mantle chromitites after their formation are particularly stable over a very large range of P-T-X conditions, and that they can record the migration of different melts/fluids (Howell et al., 2015), sometimes associated to the precipitation of zircons at mantle depths, over a very large time interval. This issue is relevant for the interpretation of the geochemical evolution of FPP chromitites, because they show some zoning of the mineral chemistry, and variations in the modal content of PGE minerals, which suggest possible interactions with late fluids/melts (Grieco et al., 2001, 2004).

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-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 by the whole rock data reported by Grieco et al. (2001). The segregation of the chromitite zircons from the same parent melt is suggested by their REE composition. Compared to the magmatic zircons from the external gabbro and nepheline diorite pegmatites, they result enriched in LREE and markedly depleted in HREE, similar to zircons segregated from mantle-derived kimberlites to carbonatites (see Fig. 4 in Hoskin and Schaltegger, 2003). Even more stringent is the match of the peculiar, high $\delta^{18}O$ of the chromitite pyroxenes with those of amphiboles and phlogopites from the phlogopite-harzburgites and pyroxenites association reported by Hartmann and Wedephol (1993). The identical O isotopic composition of MR01CR zircons and Opx-chromitite pyroxenes suggests precipitation from a common parent melt. Although the $\delta^{18}O$ partition coefficent between zircon and mafic phases is presently unconstrained, it has been widely documented that zircons segregated by mantle-derived melts show a very small $\delta^{18}O$ interval at 5.3 ± 0.4 (Valley et al., 2005; Bindeman, 2008; Tribuzio et al., 2014). A further valuable insight into the geochemical signature of 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 $^{176}Hf/^{177}Hf$ ratios, which are much lower than the depleted mantle array. The $\varepsilon Hf_{(188)}$ values are sub-chondritic, at -6.1 ± 0.6 to -1.6 ± 0.5, consistent with the data by Badanina et al. (2013). This observation excludes any genetic relationship with the nepheline diorite pegmatites, whose zircons have $\varepsilon Hf_{(t)}$ between +6 to +9.8, 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 $\delta^{18}O$ values, 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; Cumming et al., 1987; Hartmann and Wedepohl, 1993; Obermiller, 1994; Seitz and Woodland, 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 et al., 1999; Grieco et al., 2001, 2004; Morishita et al., 2003, 2008; Matsumoto et al., 2005), but alternatively also to extensional settings (Garuti et al., 2001; Zaccarini et al., 2004).

Negative εHf 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 continental crustal component in the parent melts of chromitite is also strongly supported by the high $\delta^{18}O$ 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 by the occurrence of thorianite, thorite and uraninite (Zaccarini et al., 2004; this study).

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 to heterogeneity of the melt source and/or changes in the proportion of melt sources through time.

FPP Chromitite segregation model

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-subduction zones, where chromite is basically associated to olivine. The formation of chromitite and surrounding dunite envelope is mainly explained as the result of the interaction between exotic melts and host harzburgite, in association with magma mixing (c.f. Zhou et al., 1994, 1996; Arai, 1997). The reference model assumes that in the first stage, an exotic SiO_2-undersaturated melt, introduced into the ambient peridotite at low pressure, may selectively dissolve pyroxenes, as well as hydrous minerals, and precipitate olivine producing a replacive dunite envelope. This process would form a relatively Si-rich melt, according to the following reaction: SiO_2-poor melt + pyroxenes + hydrous phases → olivine + SiO_2-rich melt. If the dunite channel is further supplied by the SiO_2-undersaturated primary melt, after mixing with the Si-rich melt, an over-saturation in spinel components (Cr+Al) takes place, leading to the isolated precipitation of spinel (c.f. Arai, 1997).

Grieco et al. (2001, 2004) proposed that the FPP chromitites and their dunite haloes formed by the interaction between basic melts and the ambient peridotites, where the segregation of phlogopite pyroxenites resulting from successive events of melt migration, unrelated to the chromitites.

Instead, Zaccarini et al. (2004) suggested that chromitite layers, dunite channels and phlogopite harzburgites were the result of migration of alkaline-carbonatitic melts.

By contrast, Zanetti et al. (1999) stressed that the widespread precipitation of newly-formed, magmatic orthopyroxene in both phlogopite harzburgites and pyroxenites pointed to a SiO_2-saturation of the metasomatic melts related to the pervasive recrystallisation of the FPP. As a consequence, the formation of dunite bodies evidences peculiar variations in melt composition, with pulses of SiO_2-undersaturated melts determining the virtually complete resorption of pyroxenes, amphibole and phlogopite in channels/bodies up to tens of meters across at relatively high, spinel-facies P conditions (see Mazzucchelli et al., 2009). The presence of dunite bodies both concordant to discordant with respect to the mantle foliation, as well as some geochemical changes shown by the magmatic minerals precipitated within dunites indicate that SiO_2-undersaturated melts occurred in different stages of the FPP metasomatic cycle (Giovanardi, 2012).

The porphyroclastic textures of the Ol-chromitite MR01CR suggests that the development of structural weaknesses into the dunite bodies may have driven the migration and mixing of the different melt components. The presence of SiO_2-saturated components in the parent melts is confirmed by the nearly ubiquitous presence of newly-formed orthopyroxene replacing olivine. In 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 documented by Ol-chromitite MR01CR. The SiO_2-saturation of the melt is more apparent in the Opx-chromitites, which record the complete recovery of the texture characterised by replacement of olivine by secondary orthopyroxene in textural equilibrium with chromite. These petrographic trends, along with the progressive chemical variation from the strongly refractory compositions of the Ol-chromitites to relatively Al-Fe-richer compositions in the Opx-chromitites, allow us to suggest that the Opx-chromitites are related to levels that experienced the largest time-integrated chromitite melt/dunite ratios.

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 of the dissolution of pyroxenes and hydrous minerals of the, already metasomatised, ambient harzburgites-pyroxenites association upon interaction with uprising of mafic melts. This is presumably the general process, always present in any FPP dunite body. However, the occurrence in some dunites of late phlogopite pyroxenites, rich in orthopyroxenes, strictly similar with those 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 SiO_2-saturated melts bearing the continental crustal component. The mixing of these two components, at a new transition of the melt composition (i.e. from SiO_2-undersaturated to SiO_2-saturated), may have triggered the precipitation of some chromitites, followed by a segregation of pyroxenites within the dunite bodies.

Interpretation of the U-Pb ages

The age of the petrologic processes recorded by the FPP is still controversial due to a very large time span documented by geochronological investigations, from the Early Permian to the Early Jurassic (Voshage et al., 1987, 1988; Stähle et al., 1990, 2001; Hartmann and Wedephol, 1993; 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 intrusive or metasomatic events (i.e. those documented by Stähle et al., 1990, 2001; the alkaline veins of Grieco et al., 2001; Matsumoto et al., 2005; Morishita et al., 2008) mainly provide Middle Triassic to Early Jurassic ages (from 240 Ma to 195 Ma). The Triassic to Early Jurassic U-Pb ages shown by chromitite zircons can be interpred in two different ways: i) the record of Jurassic, channelled melt migration with preservation of some Triassic relicts, or ii) the result of the 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 Permian ages of chromitite zircons from the FPP documented by Grieco et al. (2001) (208 ± 2 Ma), Badanina and Malitch (2012) and Badanina et al. (2013) (288 ± 7 Ma; 249 ± 3 Ma; 209 ± 4 Ma); 2) the Triassic age of late alkaline bodies discondatly cutting the harzburgite-pyroxenite association (225 ± 13 Ma; Stähle et al., 2001); 3) the Depleted Mantle geochemical affinity of the intrusives of the associate Finero Mafic Complex showing analogously Triassic to Early Jurassic radiometric data (231 ± 23 Ma to 214 ± 17 Ma: Lu et al., 1997a,b; 232 ± 3 Ma to 214 ± 5 Ma: Zanetti et al., 2013; 212.5 and 190 Ma: Schaltegger et al. 2015). Morever, the petrochemical observations reported in the previous sections clearly indicate that the chromitite zircons were segregated in the early metasomatic cycle producing the phlogopite harzburgites and pyroxenites association of FPP. Thus, it is here proposed that the different age clusters exhibited by the FPP chromitite zircons are 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 evidence of compositional homogenization due to a prolonged residence at high temperature in mantle conditions (Corfu et al., 2003).

The re-equilibration of the U-Pb system could most easily have occurred in fluid-assisted conditions. Currently no mineralogical or geochemical data support this hypothesis which, however, 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

Complex of the Val Sesia-Type IVZ, with the emplacement of large volumes of mantle-derived tholeiitic melts at the bottom of the Adria crust (Zanetti et al., 2013 and references therein).

The peculiar composition of the metasomatic melts recorded by the FPP requires the concomitant mobilisation of deep-seated reservoirs containing continental crust component. It was possibly related to Variscan subduction of continental crust and metasomatism of the overlying mantle wedge by crustal-derived melts / fluids at ~330 Ma (e.g. Ulten Area, Eastern Alps: Tumiati et al., 2003; Sapienza et al., 2009; Langone et al., 2011).

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.

A pronounced re-equilibration of the U-Pb zircon system similar to that shown by chromitite 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 common in the gabbroic rocks of the Mafic Complex documenting processes down to 25 km depth (i.e. ~0.8 GPa; Demarchi et al., 1998). Detailed inspections of magmatic zircons from the Mafic Complex performed by Peressini et al. (2007) evidenced the dominant presence of Early Permian ages, with only one Mesozoic age (180 Ma) given by a single-spot on recrystallised “white pest” rim. Consistently, up-to-date reconstructions of the thermal evolution of the polymetamorphic Kinzigite Formation of the IVZ do not provide evidence that the rifting of the Adriatic margin during the Early Jurassic induced conditions capable to reset the U-Pb system in zircon and monazite placed at crustal levels (Handy et al., 1999; Smye and Stockli, 2014; Ewing et al., 2015).

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.

Thus, it is here speculated that the prolonged re-equilibration of the U-Pb system displayed by chromitite zircons must be associated to peculiar P-T conditions affecting the FPP, such as permanence at great (mantle) depths, possible till the Early Jurassic, and/or a reheating phase due to a later (Early Jurassic) tectono-magmatic activity.

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 structures of the FPP. The precipitation of magmatic sapphirine in gabbroic rocks is consistent with pressures above than 1.1 GPa (Giovanardi et al., 2013). In particular, the composition of sapphirine-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 GPa), have been also estimated for the Finero Mafic Complex by Siena and Coltorti (1989) at ~1.0 GPa (at ~1000°C). Accordingly, Sills et al. (1983) and Christy (1989) estimated 0.9-1.1 GPa (at T of 800-950°C) for the subsolidus reaction involving formation of metamorphic sapphirine in the gabbroic lithologies of LIZ.

The development of regional thermal positive perturbations in the northern part of the IVZ, possibly associated to asthenosphere upwelling, can be inferred from the Triassic to the Early Jurassic cycles of magmatic activity segregating zircons. This hypothesis is consistent with the high temperature conditions (granulite-facies) argued for the lower IVZ by Brodie and Rutter (1987) in proximity of the Anzola-Val Grande shear zone during the Middle to Late Triassic. It is also indirectly supported by the change of the metamorphic conditions along the Pogallo Line, governed by brittle deformation to the south of Val d'Ossola, and by ductile deformation to the north (Handy, 1987). 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).

Constraints on the Mesozoic geodynamic evolution of Finero-Type IVZ

Geochronological data supports multiple melt injections throughout the Triassic to Early Jurassic in the Finero-Type IVZ (Zanetti et al., 2013 and references therein). The reappraisal of all data 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), 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) porphyroclastic intercalations in the pelagic succession of the western Southern Alps (Mundil et al., 1996). A second main stage was characterised by the intrusion of nepheline diorite pegmatites at 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 hornblendite dykes at 225-220 Ma and probably also apatite-carbonate-bearing orthopyroxenite veins (240 ± 41 Ma, phlogopite Ar-Ar (Matsumoto et al. 2005) and 213 ± 35 Ma, apatite U-Pb (Morishita et al., 2008).

The occurrence of anomalous heating processes at a regional scale has been confirmed by several papers dealing with cooling ages of the IVZ and adjacent area (Wolff et al. 2012; Smye and Stockli, 2014; Ewing et al., 2013, 2015; Beltrando et al., 2015). In particular, Beltrando et al. (2015) document the progressive westward rejuvenation of (U-Th)/(He) ages (hereafter ZHe ages), from 280-240 Ma in the Lombardian basin to 215-200 Ma near the Sostegno and Fenera basins, indicating that anomalously high thermal gradients were established in the Late Triassic towards the 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 anomaly, established at 215-210 Ma, followed by thermal decay at 200-190 Ma (Ewing et al., 2013, 2015; Beltrando et al., 2015).

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 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 considered to have spanned the period between 230-180 Ma. It also matches the final stages of 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 stages of the IVZ is still debated (see Zanetti et al., 2013), there is a wide consensus that the Late 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 23° of tilting; Wolff et al., 2012). It is thus proposed that the age interval shown by colourless 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.

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, possibly due to hyperextension of the Adriatic lithosphere. An Early Jurassic heating has been also invoked to explain the resetting of the ZHe thermochronometer in the Baveno granite (Wolff et al., 2012). Subsequent crust-wide extension led to breakup of continental crust and mantle exhumation. 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 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).

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 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 Early Permian, being exhumed at shallower, crustal levels only during Early Jurassic. The youngest event appears to have been connected to initiation of continental rifting and mantle exhumation, precursor events of the opening of the Alpine Tethys. Our data, along with those of Grieco et al. (2001), Badanina and Malitch (2012) and Badanina et al. (2013), support that lithosphere rifting and exhumation were affected by two strong thermal perturbations at 208 Ma and 187 Ma.

In our model for the northern IVZ, the pervasive metasomatism of the FPP occurred ~290 Ma and/or before. However, the possibility that the actual age of pervasive metasomatism of FPP was older and related to the Variscan orogenic cycle cannot be excluded.

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Supplementary Material Captions

Table A: Major-element composition of mineral phases as wt.% and a.p.f.u. Formulae.

Table B: Trace-element compositions of zircons, clinopyroxenes and amphiboles in ppm and single-analysis elements detection limits.

Supplemetary Material C: SEM images and EDS analysis of thorite in sample MR01CR.

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Figure Captions

Figure 1 – (a) Sketch map of the Ivrea-Verbano Zone, which represents the westernmost part of the Southern Alps. The rectangle indicates the location of the map of the Finero Complex in (b). CL, 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 modified after Steck and Tièche (1976). The empty stars document the locations of the zircon-bearing chromitites here studied.

Figure 2 – (a) The thickest layer of the FI09C34 chromitite swarm; (b) Chromitite pocket in sample FI09C34; (c) Texture of MR01CR chromitite, which is characterised by segregation of anhedral large chromite overgrowing a matrix formed by porphyroclastic dunite; (d) Allotriomorphic texture of chromite FI09C04 formed by chromite and orthopyroxene, where rounded olivine relicts rarely

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

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

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

Figure 5 - Concordia ages calculated with $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$ 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)..

Figure 7 – (a) CI-normalised (Lyubetskaya and Korenaga, 2007) REE patterns and (b) Pyrolite-normalised (McDonough and Sun, 1995) extended trace element diagrams of clinopyroxenes from 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 Morishita et al. (2008) (a); clinopyroxene from the harzburgite-pyroxenite association (Harz-Py Cpx) and apatite-bearing wehrlite (Ap-Wehrl Cpx) from Zanetti et al. (1999) (b); clinopyroxene from harzburgite (Harz Cpx) and chromitite veins (Chromitite Cpx) from Grieco et al. (2001) (c).

Figure 8 – (a) CI-normalised (Lyubetskaya and Korenaga, 2007) REE patterns and (b) Pyrolite-normalised (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.

Figure 9 – $\varepsilon Hf_{(t)}$ 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 $^{176}Hf/^{177}Hf$ ratio of 0.28325 from Nowell et al., 1998, and $^{176}Lu/^{177}Hf$ ratio of 0.0384 from Griffin et al., 2000). CHUR values are from Blichert-Toft and Albarede (1997).

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

Table Captions

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

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

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

Fig. 1

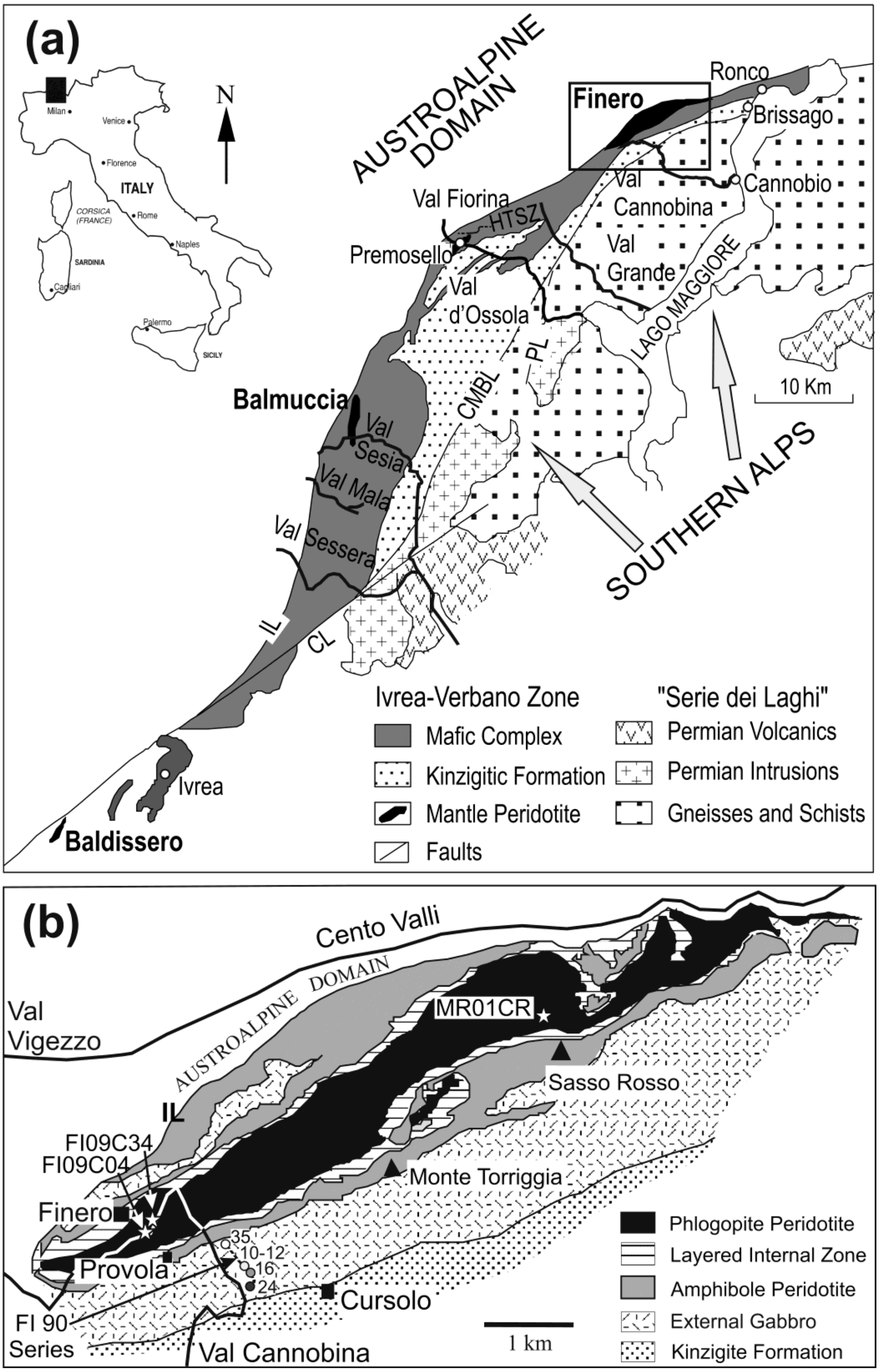

(a)
Milan
Venice
Florence
ITALY
CORSICA (FRANCE)
Rome
Naples
SARDINIA
Cagliari
Palermo
SICILY
N
AUSTROALPINE DOMAIN
Finero
Ronco
Brissago
Cannobio
Val Cannobina
Val Fiorina
HTSZ
Premosello
Val Grande
Val d'Ossola
LAGO MAGGIORE
PL
CMBL
10 Km
Balmuccia
Val Sesia
Val Mala
Val Sessera
SOUTHERN ALPS
IL
CL
Ivrea
Baldissero
Ivrea-Verbano Zone
Mafic Complex
Kinzigitic Formation
Mantle Peridotite
Faults
"Serie dei Laghi"
Permian Volcanics
Permian Intrusions
Gneisses and Schists
(b)
Cento Valli
Val Vigezzo
AUSTROALPINE DOMAIN
MR01CR
Sasso Rosso
IL
FI09C34
FI09C04
Monte Torriggia
Finero
Provola
35
10-12
16
24
Cursolo
FI 90 Series
Val Cannobina
1 km
Phlogopite Peridotite
Layered Internal Zone
Amphibole Peridotite
External Gabbro
Kinzigite Formation

Fig. 2

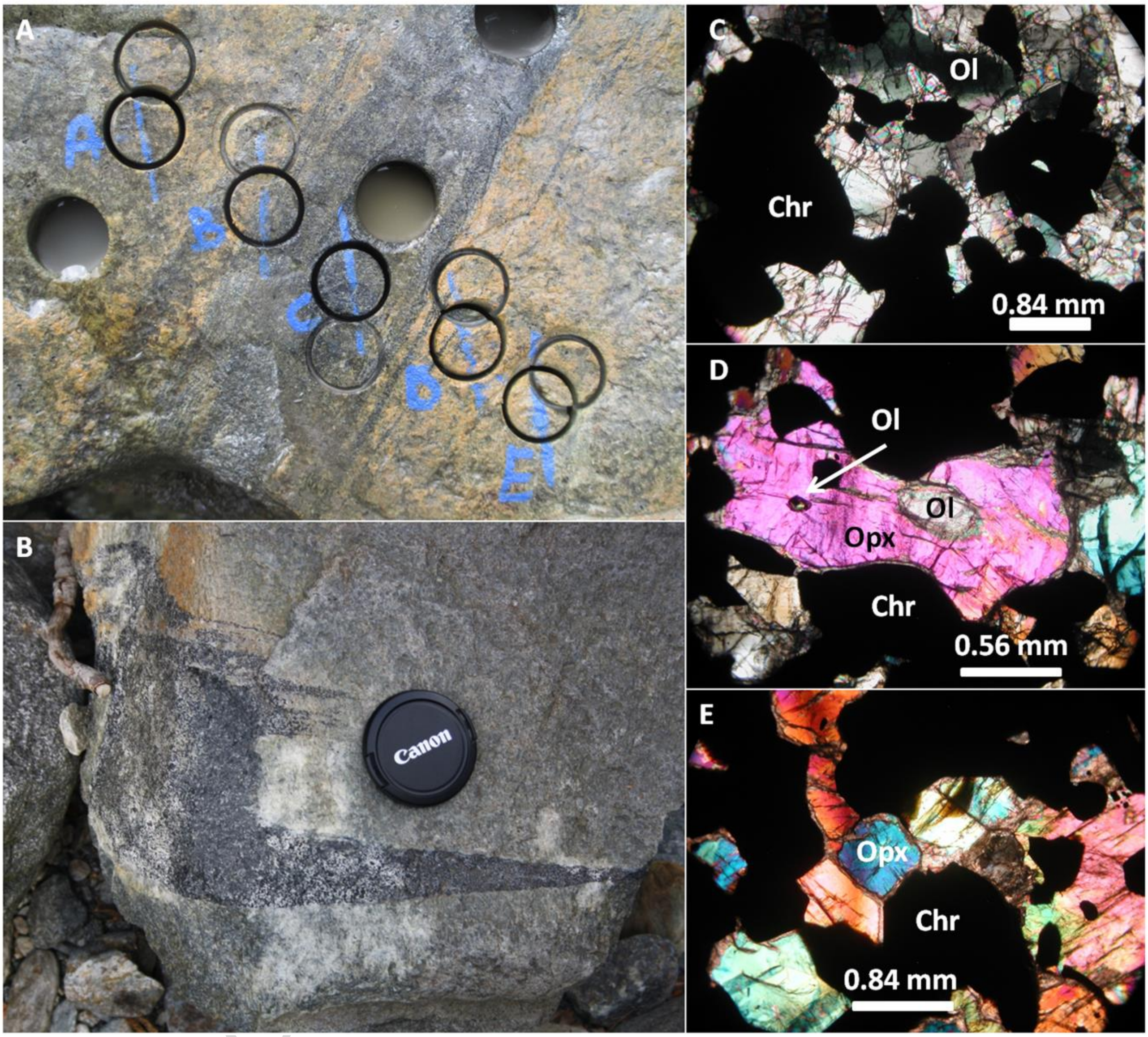
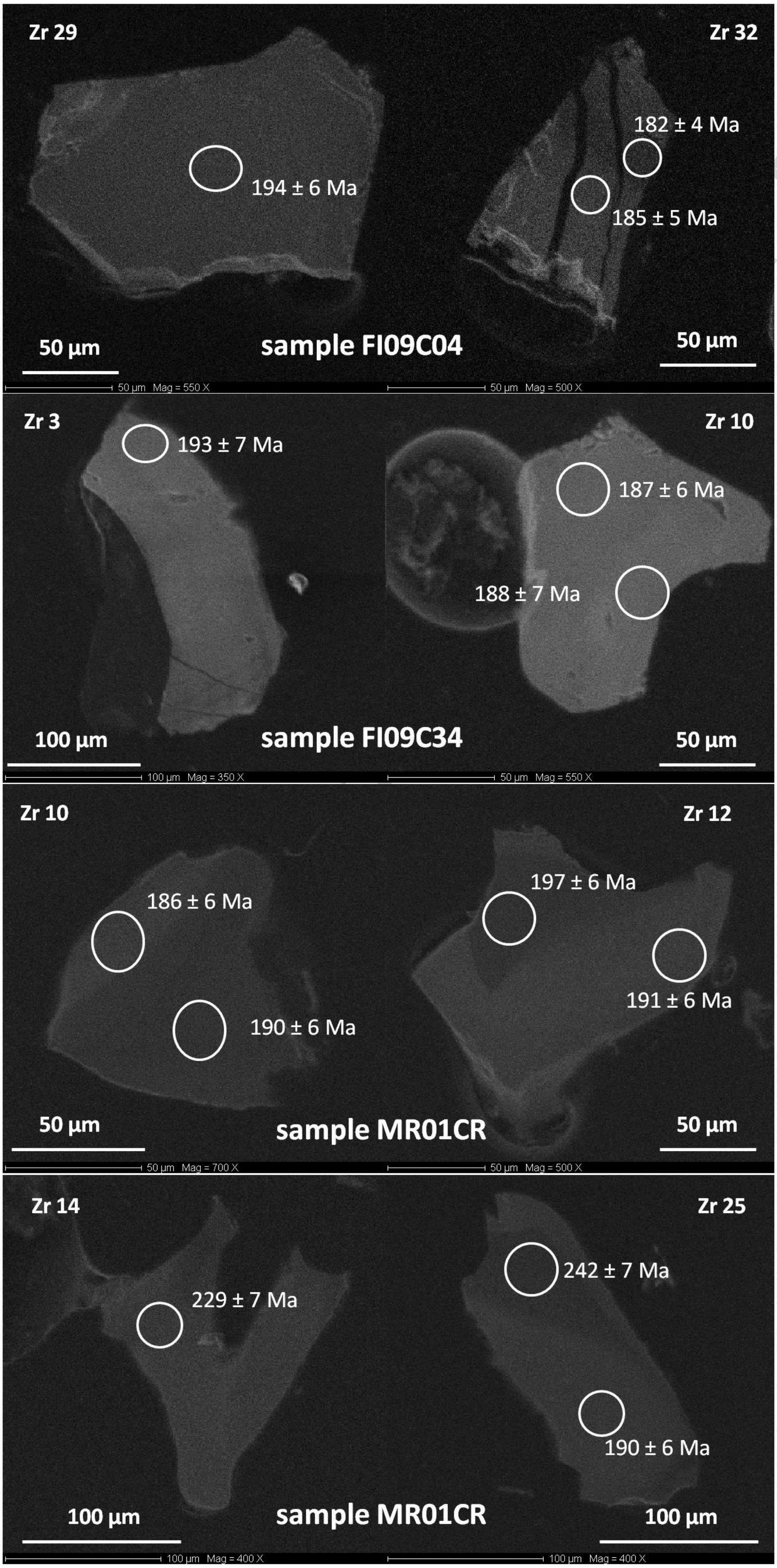
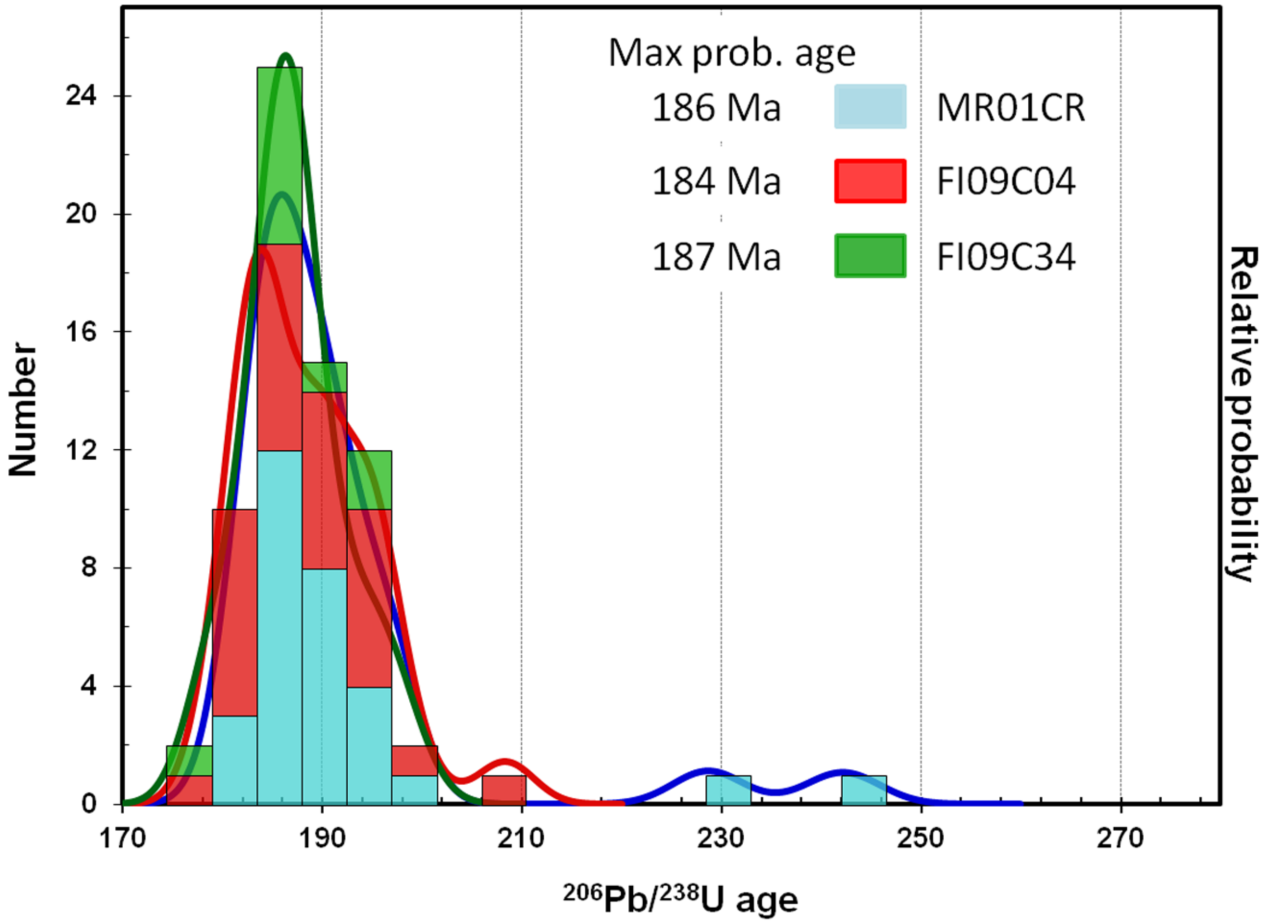


Fig. 3

1007 
Zr 29
Zr 32
194 ± 6 Ma
182 ± 4 Ma
185 ± 5 Ma
50 µm
sample FI09C04
50 µm
Zr 3
193 ± 7 Ma
Zr 10
187 ± 6 Ma
188 ± 7 Ma
100 µm
sample FI09C34
50 µm
Zr 10
Zr 12
186 ± 6 Ma
197 ± 6 Ma
191 ± 6 Ma
190 ± 6 Ma
50 µm
sample MR01CR
50 µm
Zr 14
Zr 25
242 ± 7 Ma
229 ± 7 Ma
190 ± 6 Ma
100 µm
sample MR01CR
100 µm

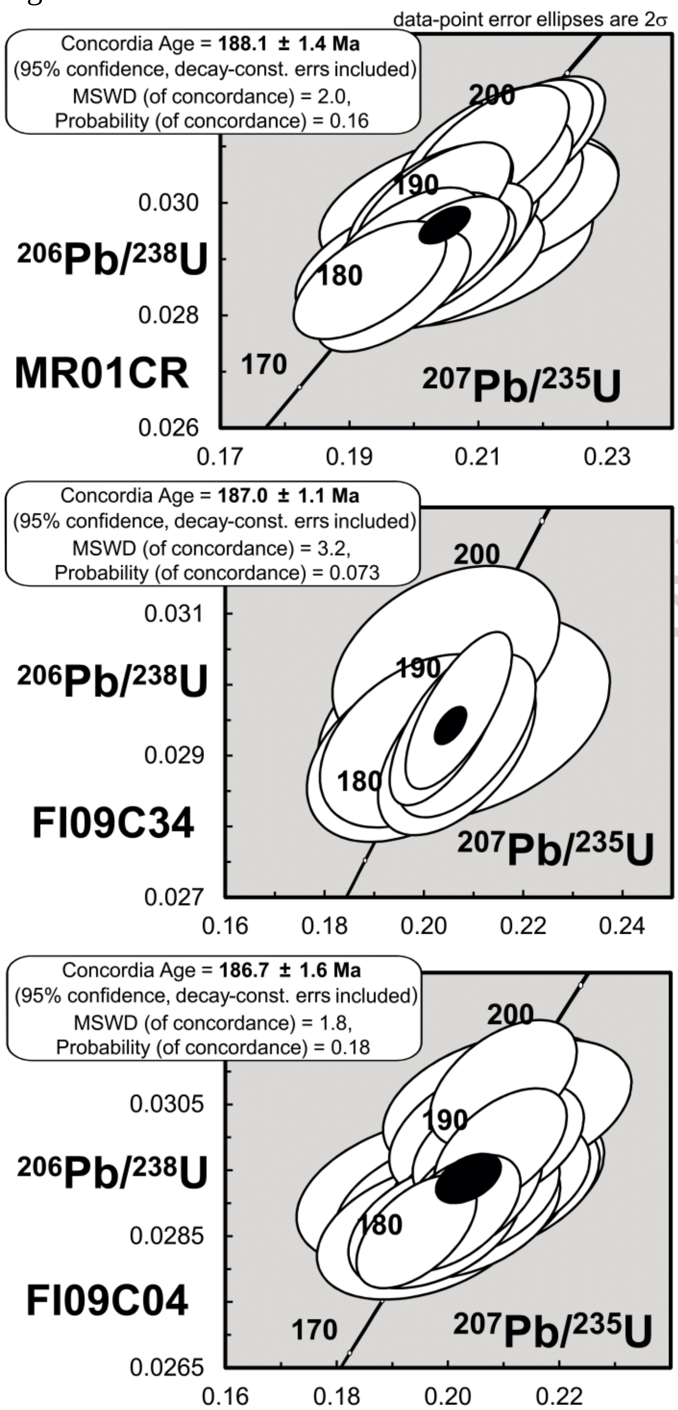
1009 Fig. 4



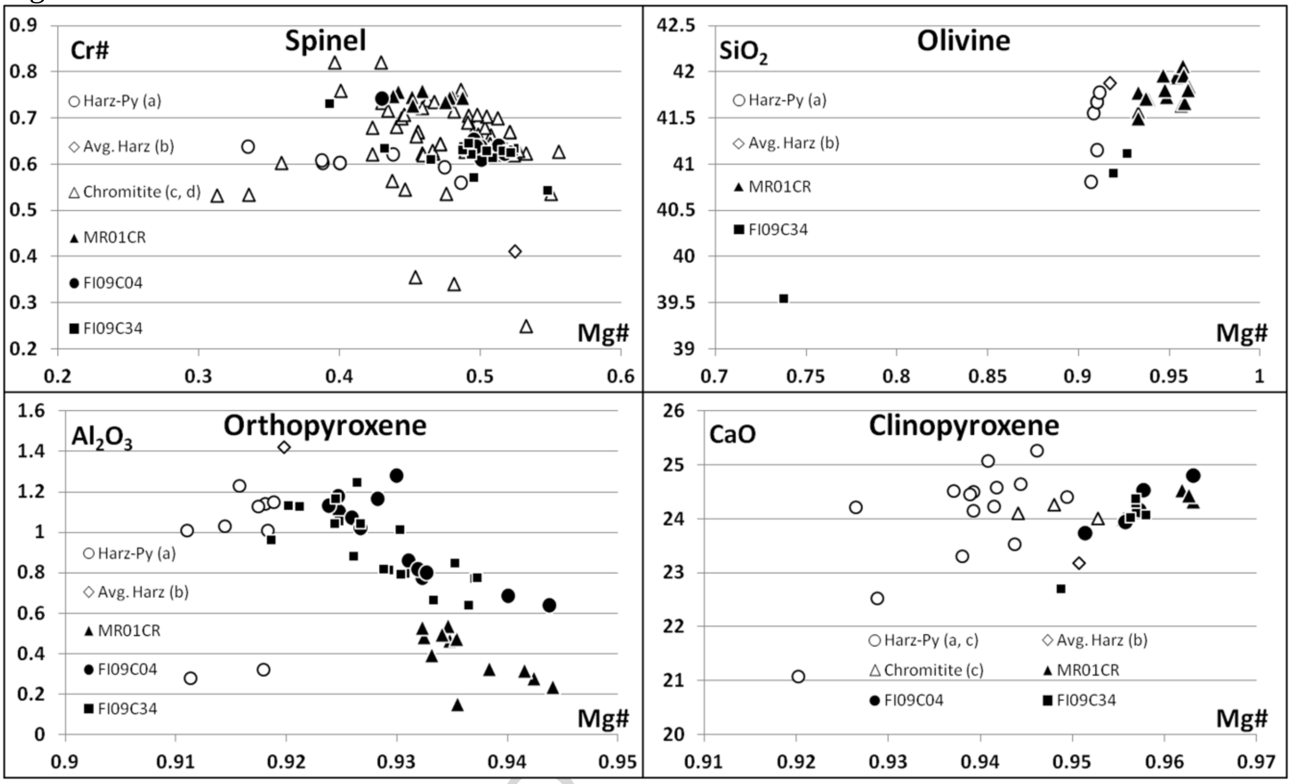
1010

1011

Fig. 5


data-point error ellipses are 2σ
Concordia Age = 188.1 ± 1.4 Ma
(95% confidence, decay-const. errs included)
MSWD (of concordance) = 2.0,
Probability (of concordance) = 0.16
200
190
180
170
0.030
0.028
0.026
206Pb/238U
MR01CR
207Pb/235U
0.17
0.19
0.21
0.23
Concordia Age = 187.0 ± 1.1 Ma
(95% confidence, decay-const. errs included)
MSWD (of concordance) = 3.2,
Probability (of concordance) = 0.073
200
190
180
0.031
0.029
0.027
206Pb/238U
FI09C34
207Pb/235U
0.16
0.18
0.20
0.22
0.24
Concordia Age = 186.7 ± 1.6 Ma
(95% confidence, decay-const. errs included)
MSWD (of concordance) = 1.8,
Probability (of concordance) = 0.18
200
190
180
170
0.0305
0.0285
0.0265
206Pb/238U
FI09C04
207Pb/235U
0.16
0.18
0.20
0.22

1015 Fig. 6



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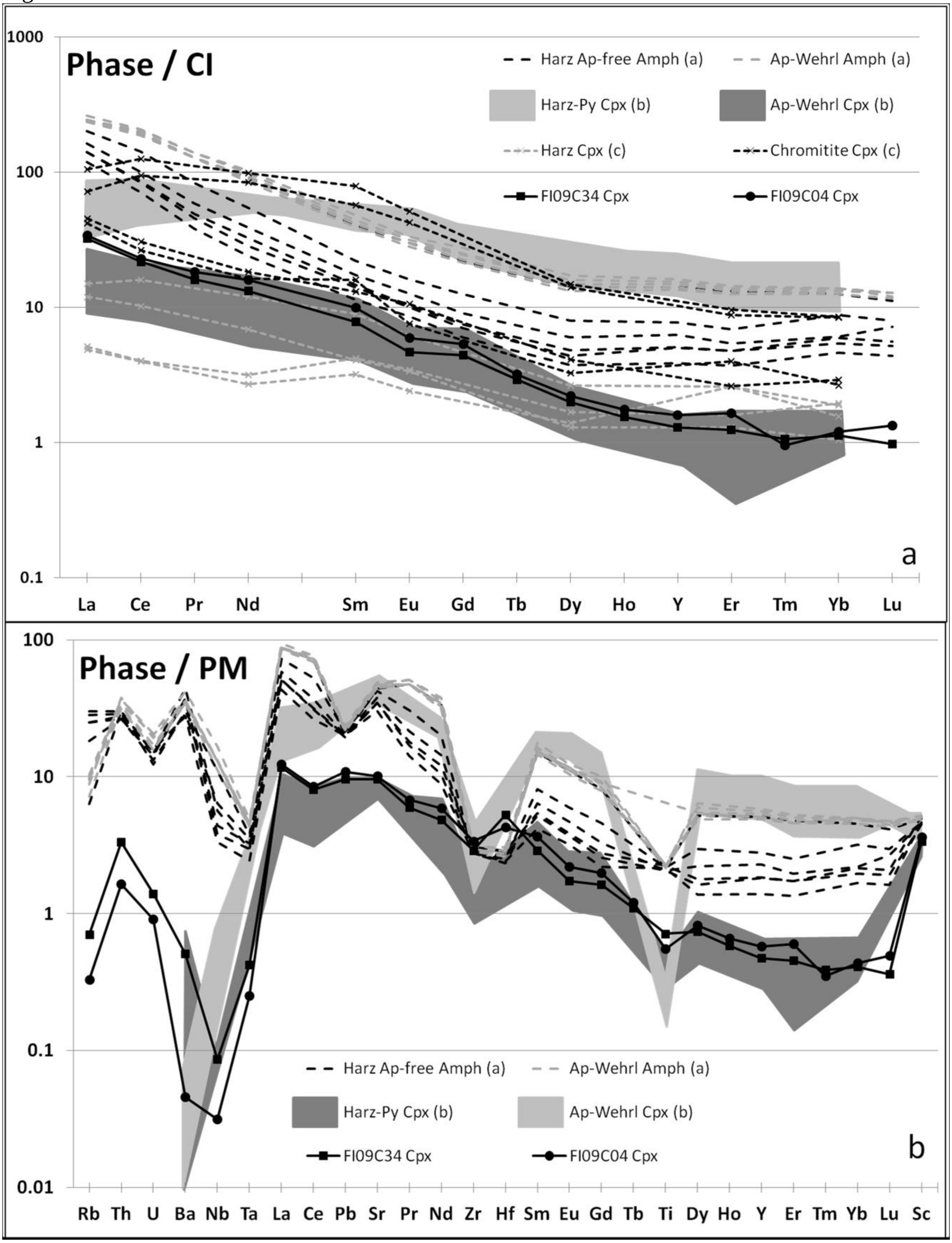
1017

1018 Fig. 7



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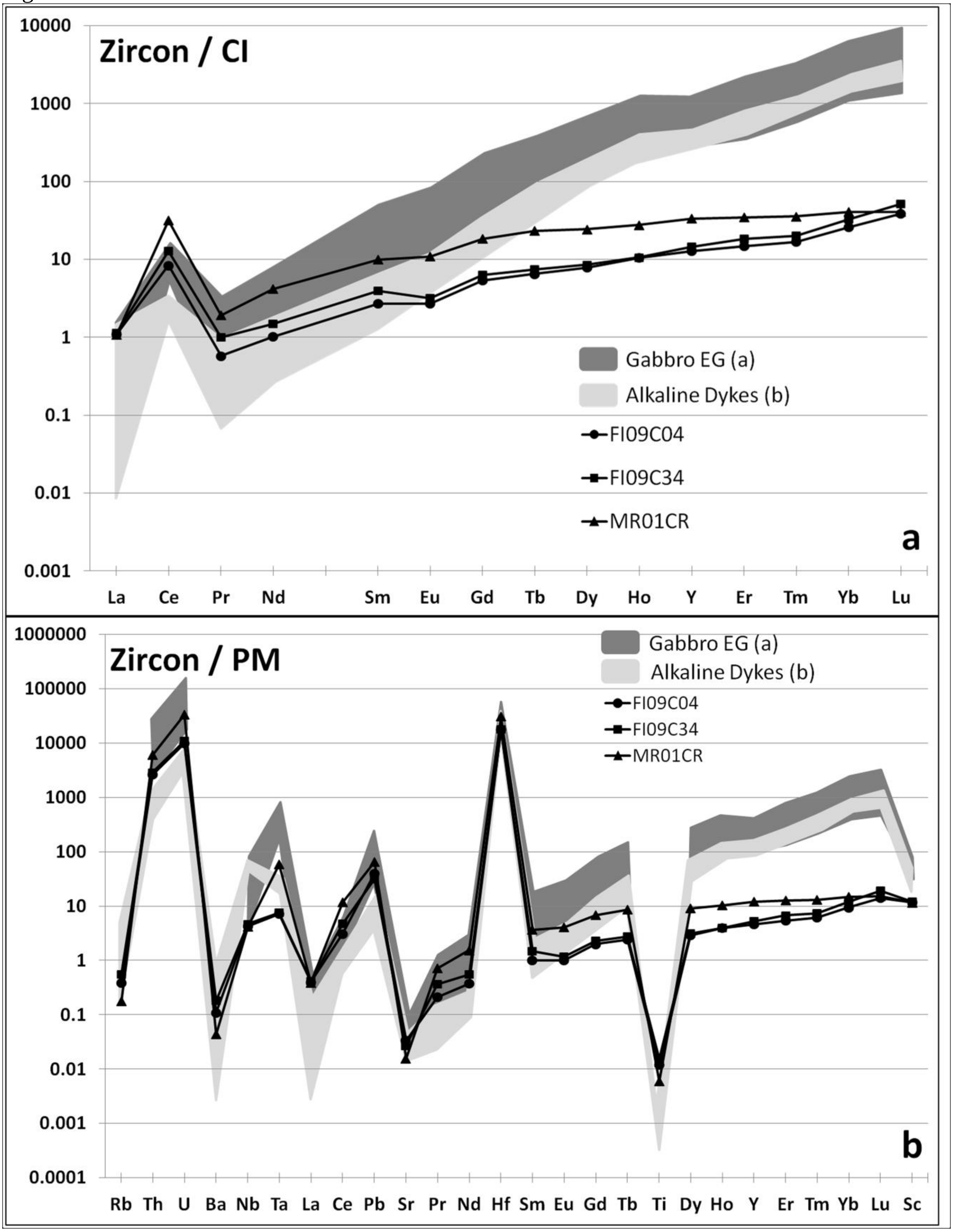
1020

1021 Fig. 8



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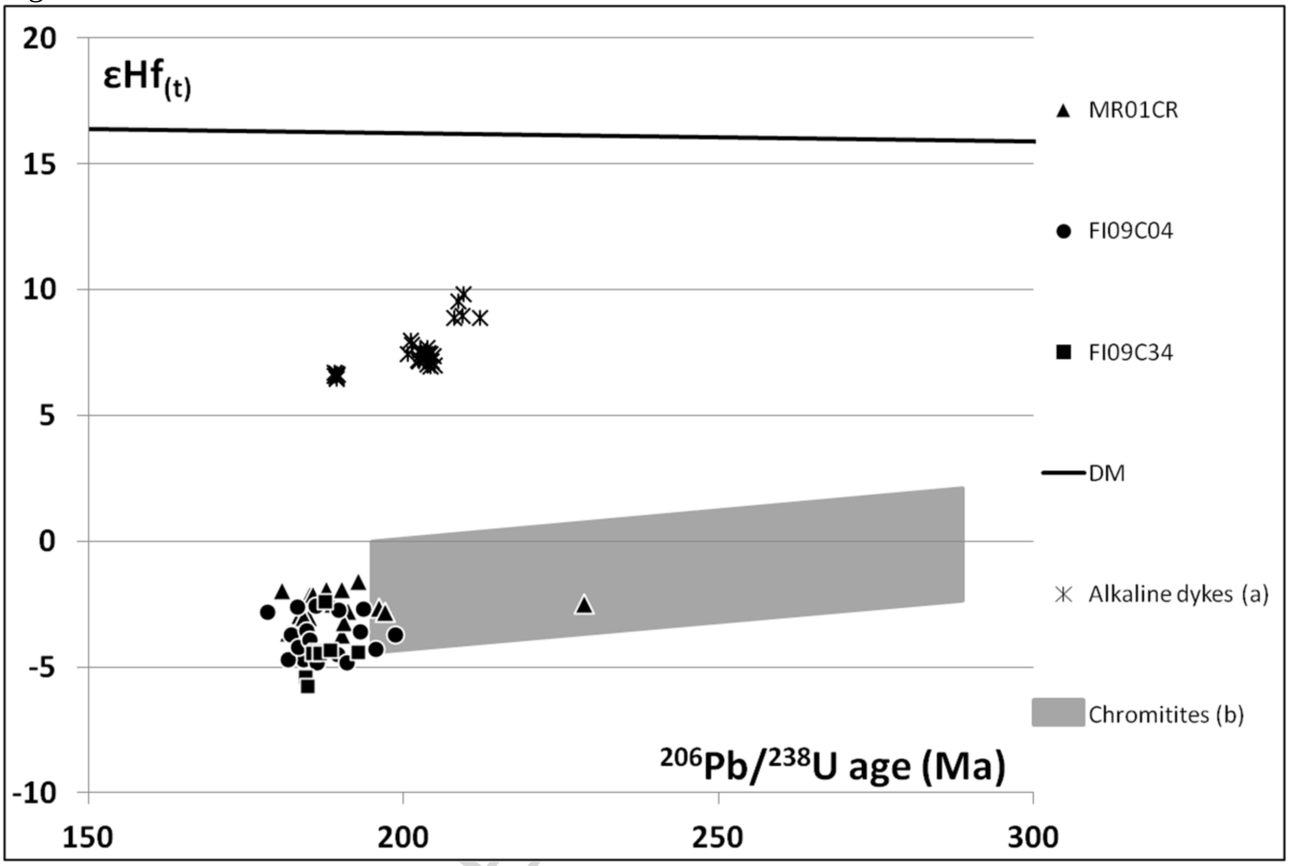
1024 Fig. 9



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1027 Fig. 10

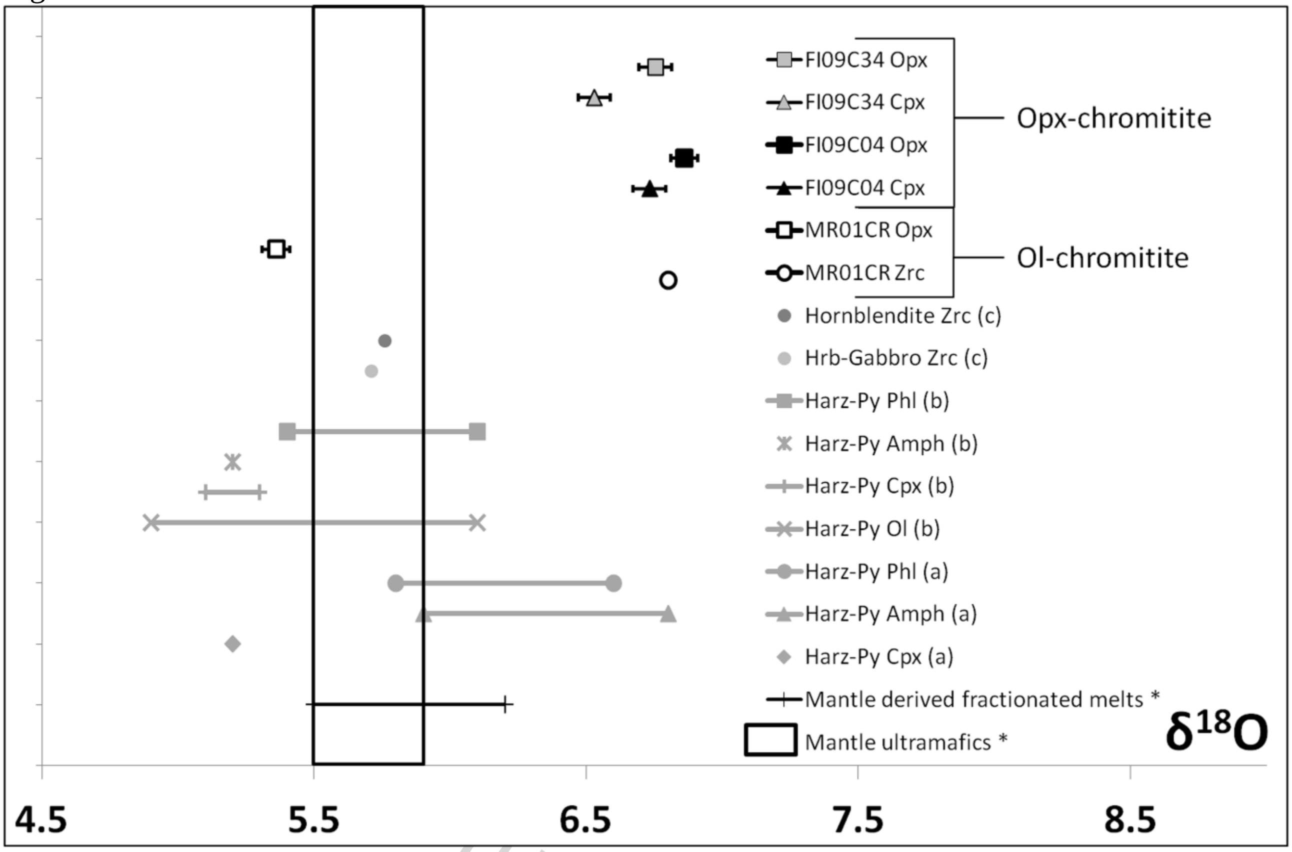

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Table 1: summary of ELA-ICP-HRMS U-Pb zircon analysis from chromitite layers from FPP.

Sample: Zircon	MR01 CR Position	Ratio $^{207}Pb/^{206}Pb$	1σ	$^{206}Pb/^{238}U$	1σ	$^{207}Pb/^{235}U$	1σ	Ages $^{207}Pb/^{206}Pb$	2σ	$^{206}Pb/^{238}U$	2σ	$^{207}Pb/^{235}U$	2σ	Concordia	2σ	% of discordance
7	core	0.0498	0.0011	0.0292	0.0005	0.2005	0.0049	187.1	8.3	185.4	5.9	185.6	9.1			
9	core	0.0497	0.0011	0.0309	0.0005	0.2115	0.0051	178.6	7.8	196.2	6.2	194.8	9.4	196.1	6.1	0.69%
9	rim	0.0507	0.0011	0.0290	0.0005	0.2027	0.0049	227.7	9.9	184.3	5.8	187.4	9.0	184.4	5.7	-1.70%
10	core	0.0492	0.0011	0.0299	0.0005	0.2029	0.0050	157.4	7.0	190.0	6.0	187.6	9.2	189.9	5.9	1.31%
10	rim	0.0511	0.0012	0.0292	0.0005	0.2057	0.0051	244.0	11.2	185.7	5.9	189.9	9.5	185.8	5.8	-2.29%
12	core	0.0503	0.0011	0.0310	0.0005	0.2148	0.0053	209.8	9.5	196.9	6.1	197.5	9.7	196.9	6.0	-0.34%
12	rim	0.0511	0.0014	0.0301	0.0005	0.2111	0.0062	245.3	13.5	191.2	6.2	194.4	11.4	191.3	6.1	-1.68%
14	core	0.0509	0.0012	0.0361	0.0006	0.2531	0.0062	238.1	10.8	228.7	7.1	229.1	11.2	228.7	7.0	-0.19%
16	core	0.0494	0.0011	0.0292	0.0005	0.1981	0.0049	165.0	7.6	185.3	5.8	183.5	9.2	185.2	5.7	0.95%
16	rim	0.0516	0.0020	0.0291	0.0005	0.2079	0.0081	269.1	20.4	184.8	6.6	191.8	14.9	184.9	6.5	-3.80%
17	core	0.0506	0.0012	0.0311	0.0005	0.2163	0.0054	222.2	10.2	197.1	6.2	198.9	9.9	197.2	6.1	-0.88%
18	core	0.0498	0.0013	0.0309	0.0005	0.2116	0.0058	186.1	9.6	196.2	6.2	194.9	10.7	196.1	6.1	0.67%
20	core	0.0506	0.0011	0.0304	0.0005	0.2118	0.0052	224.0	10.0	192.8	6.1	195.1	9.6	192.9	6.0	-1.17%
21	core	0.0492	0.0012	0.0299	0.0005	0.2025	0.0052	157.8	7.6	190.0	5.9	187.2	9.7	189.9	5.8	1.48%
23	core	0.0506	0.0015	0.0291	0.0005	0.2018	0.0062	224.0	13.2	185.0	6.0	186.7	11.5	185.1	5.9	-0.88%
24	core	0.0502	0.0012	0.0297	0.0005	0.2054	0.0054	204.3	10.0	188.7	6.1	189.7	10.0	188.7	6.0	-0.56%
25	core	0.0513	0.0011	0.0383	0.0006	0.2697	0.0066	253.0	11.3	242.2	7.4	242.4	11.8	242.2	7.3	-0.11%
25	rim	0.0519	0.0019	0.0299	0.0005	0.2125	0.0078	282.3	20.4	190.1	6.3	195.7	14.4	190.2	6.3	-2.92%
26	core	0.0491	0.0011	0.0286	0.0004	0.1931	0.0048	150.7	7.0	182.0	5.5	179.3	8.9	181.8	5.4	1.48%
27	core	0.0507	0.0014	0.0292	0.0005	0.2038	0.0057	227.7	12.2	185.7	5.8	188.3	10.6	185.7	5.7	-1.43%
28	core	0.0507	0.0012	0.0295	0.0005	0.2055	0.0054	225.8	11.1	187.7	5.9	189.8	10.0	187.8	5.8	-1.14%
28	rim	0.0508	0.0013	0.0295	0.0005	0.2055	0.0057	232.2	12.1	187.2	6.0	189.7	10.5	187.3	5.9	-1.37%
30	core	0.0504	0.0011	0.0284	0.0004	0.1968	0.0049	212.1	9.7	180.7	5.6	182.4	9.0	180.8	5.5	-0.95%
30	rim	0.0499	0.0016	0.0289	0.0005	0.1977	0.0066	188.9	12.1	183.8	6.0	183.1	12.1	183.8	5.9	0.38%
33	core	0.0504	0.0023	0.0300	0.0006	0.2085	0.0095	211.6	19.1	190.4	7.0	192.3	17.4	190.5	6.9	-0.98%
34	core	0.0502	0.0012	0.0291	0.0004	0.2012	0.0051	204.3	9.6	184.9	5.7	186.1	9.4	185.0	5.6	-0.64%
35	core	0.0495	0.0013	0.0300	0.0005	0.2048	0.0057	173.5	9.1	190.4	6.2	189.2	10.5	190.3	6.1	0.61%
35	rim	0.0497	0.0014	0.0301	0.0005	0.2071	0.0062	182.4	10.4	191.4	6.2	191.1	11.4	191.4	6.1	0.14%
38	core	0.0516	0.0015	0.0289	0.0005	0.2047	0.0063	265.9	15.8	183.5	5.8	189.1	11.6	183.6	5.7	-3.08%
39	core	0.0510	0.0012	0.0290	0.0004	0.2031	0.0052	239.9	11.4	184.1	5.6	187.8	9.5	184.2	5.6	-1.99%
Std02123		0.0514	0.0018	0.0466	0.0008	0.3309	0.0116	257.0	17.7	293.5	9.5	290.2	20.3	293.4	9.3	1.10%

Sample: Zircon	FI09C04 Position	Ratio						Ages						U-Pb		
		$^{207}Pb/^{206}Pb$	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	Concordia	2σ	% of discordance
26	core	0.0588	0.0019	0.0297	0.0004	0.2412	0.0078	557.8	35.4	188.9	5.4	219.4	14.2			-16.12%
26	rim	0.0546	0.0011	0.0298	0.0004	0.2243	0.0048	394.6	15.7	189.3	4.6	205.4	8.8			-8.53%
27	core	0.0499	0.0016	0.0287	0.0004	0.1975	0.0063	192.2	12.0	182.2	5.0	183.0	11.7	182.2	4.9	-0.44%
28	core	0.0513	0.0016	0.0288	0.0004	0.2039	0.0066	255.7	16.2	183.2	5.0	188.4	12.2	183.2	4.9	-2.87%
29	core	0.0496	0.0020	0.0305	0.0004	0.2088	0.0084	174.9	14.0	193.7	5.6	192.5	15.5	193.7	5.5	0.62%
30	core	0.0494	0.0012	0.0308	0.0004	0.2099	0.0054	165.9	8.1	195.7	5.0	193.5	9.9	195.6	4.9	1.13%
31	core	0.0482	0.0022	0.0292	0.0005	0.1949	0.0091	110.1	10.3	185.4	5.9	180.8	16.8	185.3	5.8	2.47%
32	core	0.0510	0.0016	0.0291	0.0004	0.2042	0.0065	239.5	14.9	184.7	4.9	188.6	12.0	184.7	4.8	-2.15%
32	rim	0.0493	0.0011	0.0286	0.0004	0.1943	0.0044	162.6	6.9	181.6	4.5	180.3	8.2	181.5	4.4	0.72%
33	core	0.0584	0.0017	0.0328	0.0005	0.2645	0.0080	544.8	32.0	208.3	5.7	238.3	14.4			-14.41%
34	core	0.0509	0.0011	0.0298	0.0004	0.2093	0.0049	235.4	10.4	189.5	4.7	193.0	9.0	189.6	4.7	-1.85%
35	core	0.0586	0.0014	0.0310	0.0004	0.2507	0.0065	553.4	27.4	196.9	5.1	227.2	11.7			-15.39%
36	core	0.0521	0.0012	0.0288	0.0004	0.2070	0.0049	287.6	13.0	183.3	4.5	191.0	9.1			-4.22%
37	core	0.0513	0.0018	0.0293	0.0004	0.2072	0.0074	252.1	17.9	186.0	5.2	191.2	13.7	186.1	5.2	-2.79%
38	core	0.0580	0.0013	0.0308	0.0004	0.2461	0.0058	528.6	23.6	195.4	4.8	223.4	10.5			-14.33%
39	core	0.0672	0.0014	0.0306	0.0004	0.2832	0.0064	842.7	36.1	194.2	4.8	253.2	11.5			-30.40%
40	core	0.0501	0.0022	0.0290	0.0004	0.2008	0.0087	201.5	17.4	184.2	5.7	185.8	16.1	184.2	5.6	-0.88%
41	core	0.0495	0.0020	0.0286	0.0004	0.1957	0.0079	170.6	13.8	181.7	5.4	181.5	14.6	181.7	5.4	0.08%
42	core	0.0509	0.0019	0.0304	0.0004	0.2135	0.0079	234.9	17.3	193.2	5.5	196.5	14.6	193.2	5.4	-1.70%
43	core	0.0592	0.0013	0.0283	0.0004	0.2313	0.0054	573.7	25.4	180.2	4.5	211.3	9.9			-17.27%
44	core	0.0506	0.0012	0.0289	0.0004	0.2011	0.0049	220.4	10.0	183.3	4.6	186.0	9.0	183.4	4.5	-1.45%
46	core	0.0508	0.0015	0.0281	0.0004	0.1969	0.0058	232.7	13.4	178.3	4.8	182.5	10.8	178.4	4.7	-2.32%
47	core	0.0602	0.0014	0.0302	0.0004	0.2508	0.0060	609.7	27.5	191.9	4.9	227.2	10.8			-18.38%
48	core	0.0494	0.0015	0.0301	0.0004	0.2048	0.0063	164.5	9.9	191.2	5.2	189.2	11.6	191.1	5.1	1.04%
50	core	0.0508	0.0015	0.0299	0.0004	0.2089	0.0063	230.4	13.5	189.8	5.0	192.6	11.6	189.8	4.9	-1.50%
51	core	0.0502	0.0024	0.0293	0.0005	0.2045	0.0096	202.4	19.1	186.4	6.1	188.9	17.7	186.4	6.0	-1.34%
52	core	0.0509	0.0012	0.0291	0.0004	0.2042	0.0052	235.4	11.3	185.0	4.8	188.7	9.6	185.1	4.7	-1.96%
53	core	0.0499	0.0021	0.0313	0.0004	0.2150	0.0090	189.4	15.9	198.8	5.7	197.7	16.5	198.8	5.6	0.54%
54	core	0.0510	0.0022	0.0293	0.0005	0.2056	0.0088	240.4	20.6	185.9	5.9	189.9	16.3	185.9	5.8	-2.15%
Std02123		0.05295	0.0022	0.04717	0.0007	0.34439	0.0143	326.7	27.2	297.1	8.9	300.5	25.0	297.2	8.7	-1.10%

Table 1: continue.

Sample:	FI09C34	Ratio						Ages						U-P
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	Concordi
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.

Table 1: continue.

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

sample	zircon	position	Age (Ma)	$^{176}Yb/^{177}Hf$	2σ	$^{176}Lu/^{177}Hf$	2σ	$^{176}Hf/^{177}Hf$	2σ	$e_{Hf}(0)$	$e_{Hf}(t)$	2σ	T_{DM}	T_{DM}^{C}	$f_{Lu/Hf}$
MR01CR	9	core	188	0.001629	0.000039	0.000052	0.000001	0.282580	0.000015	-6.77	-2.65	0.53	927	1395	-1.00
MR01CR	10	core	188	0.000730	0.000003	0.000022	0.000000	0.282582	0.000012	-6.73	-2.60	0.44	925	1391	-1.00
MR01CR	12	rim	188	0.000479	0.000008	0.000015	0.000000	0.282577	0.000012	-6.89	-2.77	0.44	931	1402	-1.00
MR01CR	14	core	188	0.000533	0.000025	0.000017	0.000001	0.282584	0.000012	-6.64	-2.51	0.44	921	1386	-1.00
MR01CR	16	core	188	0.000603	0.000010	0.000017	0.000000	0.282596	0.000013	-6.24	-2.11	0.45	906	1360	-1.00
MR01CR	17	core	188	0.000620	0.000019	0.000017	0.000000	0.282575	0.000011	-6.96	-2.83	0.40	933	1406	-1.00
MR01CR	20	core	188	0.000661	0.000007	0.000019	0.000000	0.282610	0.000013	-5.73	-1.60	0.47	886	1328	-1.00
MR01CR	21	core	188	0.000742	0.000012	0.000024	0.000000	0.282600	0.000011	-6.09	-1.97	0.39	900	1351	-1.00
MR01CR	23	core	188	0.000550	0.000025	0.000017	0.000000	0.282553	0.000011	-7.74	-3.61	0.40	964	1455	-1.00
MR01CR	24	core	188	0.000573	0.000004	0.000017	0.000000	0.282585	0.000013	-6.61	-2.48	0.47	920	1384	-1.00
MR01CR	25	rim	188	0.000420	0.000006	0.000013	0.000000	0.282550	0.000012	-7.85	-3.73	0.41	968	1462	-1.00
MR01CR	26	core	188	0.000534	0.000006	0.000016	0.000000	0.282552	0.000013	-7.79	-3.66	0.45	966	1458	-1.00
MR01CR	27	core	188	0.000836	0.000026	0.000025	0.000001	0.282595	0.000012	-6.25	-2.12	0.41	906	1361	-1.00
MR01CR	28	core	188	0.000461	0.000007	0.000013	0.000000	0.282601	0.000013	-6.05	-1.92	0.48	898	1348	-1.00
MR01CR	30	core	188	0.000492	0.000005	0.000014	0.000000	0.282599	0.000013	-6.11	-1.98	0.47	900	1352	-1.00
MR01CR	30	rim	188	0.000990	0.000012	0.000033	0.000000	0.282556	0.000020	-7.63	-3.51	0.72	960	1449	-1.00
MR01CR	33	core	188	0.000382	0.000015	0.000012	0.000000	0.282564	0.000013	-7.34	-3.22	0.47	948	1430	-1.00
MR01CR	34	core	188	0.000404	0.000011	0.000012	0.000000	0.282570	0.000014	-7.15	-3.03	0.48	941	1418	-1.00
MR01CR	35	core	188	0.000783	0.000007	0.000024	0.000000	0.282601	0.000015	-6.05	-1.92	0.54	898	1349	-1.00
MR01CR	38	core	188	0.000403	0.000020	0.000013	0.000000	0.282572	0.000012	-7.06	-2.93	0.44	937	1412	-1.00
MR01CR	39	core	188	0.000409	0.000004	0.000012	0.000000	0.282570	0.000015	-7.16	-3.03	0.53	941	1418	-1.00
FI09C04	26	core	188	0.000531	0.000005	0.000021	0.000000	0.282527	0.000019	-8.68	-4.55	0.66	1000	1515	-1.00
FI09C04	26	rim	188	0.000872	0.000019	0.000033	0.000001	0.282566	0.000016	-7.27	-3.15	0.55	946	1426	-1.00
FI09C04	27	core	188	0.000569	0.000005	0.000022	0.000000	0.282550	0.000016	-7.85	-3.72	0.55	968	1462	-1.00
FI09C04	28	core	188	0.000576	0.000024	0.000024	0.000001	0.282582	0.000016	-6.74	-2.61	0.58	925	1392	-1.00
FI09C04	29	core	188	0.000577	0.000005	0.000022	0.000000	0.282579	0.000016	-6.82	-2.70	0.55	928	1397	-1.00
FI09C04	30	core	188	0.000442	0.000007	0.000017	0.000000	0.282534	0.000016	-8.41	-4.28	0.57	990	1498	-1.00
FI09C04	31	core	188	0.001212	0.000037	0.000047	0.000001	0.282523	0.000019	-8.81	-4.69	0.66	1006	1524	-1.00
FI09C04	32	core	188	0.000572	0.000009	0.000022	0.000000	0.282554	0.000020	-7.71	-3.58	0.70	963	1454	-1.00
FI09C04	33	core	188	0.000638	0.000009	0.000025	0.000000	0.282546	0.000018	-7.99	-3.86	0.65	973	1471	-1.00
FI09C04	34	core	188	0.000613	0.000007	0.000024	0.000000	0.282528	0.000017	-8.63	-4.51	0.58	998	1512	-1.00
FI09C04	35	core	188	0.000626	0.000008	0.000025	0.000000	0.282528	0.000021	-8.64	-4.52	0.73	999	1513	-1.00
FI09C04	36	core	188	0.000754	0.000009	0.000028	0.000000	0.282489	0.000019	-10.00	-5.88	0.66	1052	1599	-1.00

sample	zircon	position	Age (Ma)	$^{176}Yb/^{177}Hf$	2σ	$^{176}Lu/^{177}Hf$	2σ	$^{176}Hf/^{177}Hf$	2σ	$e_{Hf}(0)$	$e_{Hf}(t)$	2σ	T_{DM}	T_{DM}^{C}	$f_{Lu/Hf}$
FI09C04	37	core	188	0.000605	0.000011	0.000025	0.000000	0.282582	0.000021	-6.71	-2.59	0.75	924	1391	-1.00
FI09C04	38	core	188	0.000544	0.000011	0.000021	0.000000	0.282486	0.000018	-10.13	-6.01	0.64	1056	1607	-1.00
FI09C04	39	core	188	0.000619	0.000011	0.000026	0.000000	0.282524	0.000021	-8.76	-4.64	0.74	1003	1520	-1.00
FI09C04	40	core	188	0.000497	0.000018	0.000020	0.000001	0.282522	0.000023	-8.84	-4.71	0.80	1006	1525	-1.00
FI09C04	41	core	188	0.000628	0.000008	0.000024	0.000000	0.282522	0.000022	-8.84	-4.72	0.79	1007	1525	-1.00
FI09C04	42	core	188	0.000493	0.000006	0.000020	0.000000	0.282553	0.000016	-7.73	-3.61	0.57	963	1455	-1.00
FI09C04	44	core	188	0.000690	0.000012	0.000025	0.000000	0.282536	0.000017	-8.36	-4.23	0.59	988	1495	-1.00
FI09C04	46	core	188	0.000976	0.000018	0.000037	0.000000	0.282576	0.000018	-6.94	-2.81	0.64	933	1405	-1.00
FI09C04	48	core	188	0.000558	0.000009	0.000022	0.000000	0.282518	0.000018	-8.97	-4.85	0.65	1012	1534	-1.00
FI09C04	50	core	188	0.000510	0.000018	0.000022	0.000001	0.282578	0.000017	-6.86	-2.74	0.62	930	1400	-1.00
FI09C04	51	core	188	0.000901	0.000005	0.000035	0.000000	0.282519	0.000017	-8.96	-4.84	0.61	1011	1533	-1.00
FI09C04	52	core	188	0.001005	0.000035	0.000041	0.000001	0.282544	0.000019	-8.05	-3.93	0.68	976	1476	-1.00
FI09C04	53	core	188	0.000646	0.000010	0.000026	0.000000	0.282550	0.000018	-7.84	-3.72	0.65	968	1462	-1.00
FI09C34	1	core	188	0.000808	0.000017	0.000031	0.000000	0.282503	0.000020	-9.52	-5.40	0.69	1033	1569	-1.00
FI09C34	2	core	188	0.000783	0.000020	0.000032	0.000001	0.282532	0.000019	-8.50	-4.38	0.69	994	1504	-1.00
FI09C34	3	core	188	0.000886	0.000049	0.000038	0.000002	0.282530	0.000022	-8.55	-4.43	0.76	996	1508	-1.00
FI09C34	4	core	188	0.000575	0.000014	0.000023	0.000000	0.282529	0.000018	-8.58	-4.45	0.64	996	1509	-1.00
FI09C34	5	core	188	0.000646	0.000012	0.000028	0.000000	0.282492	0.000018	-9.90	-5.78	0.63	1048	1593	-1.00
FI09C34	6	rim	188	0.000925	0.000036	0.000039	0.000001	0.282551	0.000032	-7.80	-3.68	1.13	967	1460	-1.00
FI09C34	8	core	188	0.000498	0.000004	0.000019	0.000000	0.282577	0.000018	-6.90	-2.78	0.62	931	1402	-1.00
FI09C34	10	core	188	0.000427	0.000008	0.000018	0.000000	0.282529	0.000020	-8.59	-4.46	0.72	997	1509	-1.00
FI09C34	10	rim	188	0.000651	0.000007	0.000029	0.000000	0.282587	0.000019	-6.55	-2.43	0.69	918	1381	-1.00
FI09C34	11	core	188	0.000881	0.000017	0.000039	0.000000	0.282533	0.000017	-8.46	-4.34	0.59	992	1502	-1.00

Table 2: continue.

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

Sample	Phase	$\delta^{18}O$ avg.	Std. Dev.
MR01CR	Zrc (1)	6.80	
	Opx (2)	5.36	0.05
FI09C04	Opx (2)	6.86	0.05
	Cpx (3)	6.73	0.06
FI09C34	Opx (2)	6.76	0.06
	Cpx (2)	6.53	0.06

Highlights:

zircon-bearing mantle chromitites formed along with the Finero phlogopite harzbugites

the Finero Phlogopite Peridotite was firstly metasomatised in Early Permian or before

younger ages record thermal perturbations related to tectono-magmatic events

the exhumation of Finero Phlogopite Peridotite occurred in Early Jurassic

the IVZ exhumation was accompanied by a thermal perturbation at ~187 Ma