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The Variscan subduction inheritance in the Southern Alps Sub-Continental Lithospheric Mantle: clues from the Middle Triassic shoshonitic magmatism of the Dolomites (NE Italy)

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To the Editor of *Lithos*

Dear Editor,

Please find enclosed the revised version of the manuscript “**The Variscan subduction inheritance in the Southern Alps Sub-Continental Lithospheric Mantle: clues from the Middle Triassic shoshonitic magmatism of the Dolomites (NE Italy)**” by Federico Casetta, Ryan B. Ickert, Darren F. Mark, Pier Paolo Giacomoni, Costanza Bonadiman, Theodoros Ntaflos, Alberto Zanetti and Massimo Coltorti for potential publication in *Lithos*.

We took into account and accomplished all Your and Referee’s indications. You will find our revised work in the docx file “Casetta et al., Revision, Changes marked”, which includes the new and re-written parts with red typing. A clean version (black typing) of the revised manuscript is also provided (Casetta et al., Manuscript).

We hope this new version will be more suitable for publication in *Lithos*, and we remain at Your disposal for any further modification You retain useful.

For the authors,
Federico Casetta

Authors’ replies to Editor’s and Reviewer’s comments:

Editor (Michael Roden): *I agree with Dr. Marzoli that you should report the actual XRF analyses rather than those normalized to 100%. Also double check the references as she suggests.*

Reviewer #2 (Andrea Marzoli): *The authors answered to all points raised by the previous reviews. I consider their answers as convincing (or at least reasonable). The only point I still disagree on is the whole-rock table; It is not ok to report in the table XRF recalculated to 100%, just like you do not recalculate mineral data to 100%. Please double check the reference list, it contains many small mistakes.*

Authors’ reply: Following Editor’s and Reviewer’s suggestions, the analyses normalized to 100% have been substituted with the actual XRF analyses (see revised version of Table 1).

The reference list has been also double checked and the occasional errors corrected throughout it. Authors apologize since we noticed that the reference about Boscaini et al. (2020) paper, recommended by the Reviewer during the previous revision round, was inserted in the text but not in the reference list. In the revised version of the manuscript, the reference has been now included (lines 762-764). Although such kind of discrepancies would have been noticed during the eventual proof correction stage, we acknowledge Reviewer #2 for his scrupulous check.

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Abstract

Although often speculated, the link between the Middle Triassic shoshonitic magmatism at the NE margin of the Adria plate and the subduction-related metasomatism of the Southern Alps Sub-Continental Lithospheric Mantle (SCLM) has never been constrained. In this paper, a detailed geochemical and petrological characterization of the lavas, dykes and ultramafic cumulates belonging to the shoshonitic magmatic event that shaped the Dolomites (Southern Alps) was used to model the composition and evolution of the underlying SCLM in the time comprised between the Variscan subduction and the opening of the Alpine Tethys. Geochemical models and numerical simulations enabled us to define that 5-7% partial melting of an amphibole + phlogopite-bearing spinel lherzolite, similar to the Finero phlogopite peridotite, can account for the composition of the primitive Mid-Triassic SiO₂-saturated to -undersaturated melts with shoshonitic affinity ($^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7032\text{-}0.7058$; $^{143}\text{Nd}/^{144}\text{Nd}_i = 0.51219\text{-}0.51235$; Mg #~ 70; ~1.1 wt% H₂O). By taking into account the H₂O content documented in mineral phases from the Finero phlogopite peridotite, it is suggested that the Mid-Triassic SCLM source was able to preserve a significant enrichment and volatile content (600-800 ppm H₂O) for more than 50 Ma, i.e. since the slab-related metasomatism connected to the Variscan subduction. The partial melting of a Finero-like SCLM represents the exhaustion of the subduction-related signature in the Southern Alps lithosphere that predated the Late Triassic-Early Jurassic asthenospheric upwelling related to the opening of the Alpine Tethys.

HIGHLIGHTS

- Middle Triassic shoshonitic magmatism in the Dolomites, Southern Alps
- Linking the metasomatized Finero phlogopite peridotite to the shoshonitic magmatism
- Exhaustion of the slab-derived signature inherited from the Variscan subduction

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52 **Keywords**

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54 Middle Triassic shoshonitic magmatism; Finero phlogopite peridotite

55

56 **1. Introduction**

57 The genesis of shoshonitic magmas is controversial, as it can be variably ascribed to syn- to
58 post-collisional tectonic settings or to subduction-related geodynamic contexts (e.g., Conticelli
59 et al., 2009; Peccerillo and Frezzotti, 2015). The typical Large-Ion Lithophile Elements (LILE),
60 Light Rare Earth Elements (LREE) enrichments and High Field Strength Elements (HFSE)
61 depletion of shoshonitic rocks, coupled with the recorded radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ and unradiogenic
62 $^{143}\text{Nd}/^{144}\text{Nd}$ ratios, always reflects enrichment of the underlying Sub-Continental Lithospheric
63 Mantle (SCLM) domains (Conceição and Green, 2004; Zheng, 2019). Shoshonitic melts are
64 thought to be generated by SCLM portions modified by the interaction with slab-derived fluids
65 or melts, or crustal recycled lithophile components (Plank, 2014; Zheng, 2019), and can be
66 powerful tools to track the composition and evolution of the lithosphere during major
67 geodynamic cycles.

68 In this regard, of paramount interest is the Middle Triassic (242-237 Ma) occurrence of
69 shoshonitic magmatism in the Dolomites (NE Italy; Fig. 1) and throughout the Southern Alps
70 (see Lustrino et al., 2019). The Mid-Triassic magmatism of the Southern Alps is particularly
71 interesting because of its timing, that is in between two major geodynamic episodes, i.e. the
72 Variscan subduction (430-340/300 Ma) and the Late Triassic/Early Jurassic opening of the
73 Alpine Tethys (Stampfli et al., 2013; Kroner and Romer, 2014). This last rifting phase was
74 marked by a strong asthenospheric upwelling that induced the production of highly alkaline
75 anorogenic magmas at 225-215 Ma throughout the Southern Alps, i.e. 20 Ma after the
76 termination of the shoshonitic magmatic cycle (Stahle et al., 2001; Casetta et al., 2019).

77 Notwithstanding its concomitance with dominant strike-slip to extensional tectonics (e.g.,
78 Doglioni, 2007), the Mid-Triassic magmatic event that shaped the Dolomites was historically

79 defined as “shoshonitic” due to the potassic affinity/orogenic-type of the magmatic products
80 (Sloman, 1989; Casetta et al., 2018a; 2018b; Lustrino et al., 2019). Despite its geodynamic
81 implications, this term was adopted throughout the text to indicate the magmatic affinity, in line
82 with previous studies, although multiple and articulated geodynamic/tectonic models for the
83 Southern Alps magmatism have been proposed so far. They include: i) aborted continental
84 rifting (Bernoulli and Lemoine, 1980); ii) sinistral strike-slip tectonics (Doglioni, 2007); iii)
85 arc/back-arc systems located at the Paleotethys NW limb (Zanetti et al., 2013); iv) subduction
86 of a Palaeozoic oceanic basin located between Austroalpine and Southern Alps (Bianchini et
87 al., 2018); v) reactivation, during an anorogenic rifting phases, of old SCLM domains
88 metasomatized during the Variscan subduction (Sloman, 1989; Bonadiman et al., 1994; De Min
89 et al., 2020).

90 Although the impact of the Variscan subduction on the Southern Alps/Austroalpine SCLM
91 portions is widely documented, as in case of the Finero and Ulten peridotitic massifs (Zanetti
92 et al., 1999; Tumiatì et al., 2003; Ionov et al., 2017; Giovanardi et al., 2020), a direct link
93 between the metasomatic event and the subsequent genesis of magmas with orogenic-like
94 affinity is still missing, feeding relevant questions:

- 95 - What was the composition of the Southern Alps SCLM during Middle Triassic?
- 96 - How did this mantle source evolve from the Variscan subduction to the Alpine Tethys
97 opening?
- 98 - In which way is Mid-Triassic shoshonitic magmatism related to the geochemical features
99 of the Southern Alps SCLM?

100 To contribute in answering these questions, a geochemical and petrological investigation of
101 mafic to intermediate lava flows, dykes and entrained ultramafic cumulates cropping out in the
102 Dolomites was put forward. These products, located close to the coeval Predazzo Intrusive
103 Complex (Fig. 1; Casetta et al., 2018a; 2018b), belong to the acme of the Mid-Triassic
104 magmatism (from 239.04 ± 0.04 to 237.58 ± 0.04 ; Storck et al., 2020), which was typified by the

105 emission of effusive products recording a mantle-derived signature, with limited crustal
106 contamination (Storck et al., 2020). These volcanic sequences are ideal objects for investigating
107 the evolution of the Southern Alps SCLM, and modelling its possible links to the main
108 geodynamic processes recorded in the neighbouring orogenic peridotites (Zanetti et al., 1999;
109 Tumiati et al., 2003). Mineral phase, whole-rock major and trace element analyses and Sr-Nd
110 isotopes were used to constrain the *T-P* path of shoshonitic magmas during ascent and
111 differentiation, as well as to model the nature of the Southern Alps SCLM during Mid-Triassic.
112 Numerical models were performed to: i) reconstruct the composition of the primitive magmas
113 in equilibrium with potential mantle sources; ii) trace the evolution of the SCLM beneath Adria
114 during Mid-Triassic, in the large context of what recorded in the Finero, Balmuccia, Baldissero
115 and Ulten peridotitic massifs; iii) determine the geometry of the feeding system and the main
116 processes that led to ascent/differentiation of the shoshonitic magmas in the crust.

117

118 **2. Geological setting**

119 The Mid-Triassic magmatic event that took place throughout the Southern Alps between ~243
120 and 227 ± 6 Ma was characterized by the emission of significant amounts of igneous rocks with
121 calc-alkaline to shoshonitic affinity (Lustrino et al., 2019; Storck et al., 2020 and references
122 therein). In the central and eastern Southern Alps, an earlier volcanoclastic phase, with
123 discontinuous emission of acidic products was followed by a dominantly effusive stage,
124 characterized by the eruption of mafic lavas and emplacement of shallow intrusive bodies
125 (Bonadiman et al., 1994; Casetta et al., 2018a; Storck et al., 2020). The oldest magmatic
126 manifestations, i.e. tuff layers embedded in the sedimentary Buchenstein Fm. in the Brescian
127 Alps, record zircon U-Pb ages of 242.65 ± 0.04 Ma, whereas the youngest episodes, represented
128 by tuff layers inside the Wengen Fm. in the Dolomites, yield an age of 237.58 ± 0.04 Ma
129 (Wotzlav et al., 2018; Storck et al., 2020).

130 The most productive phase of the Mid-Triassic magmatic event is best preserved in the thick
131 sequences of mafic to mildly evolved volcanics of the Dolomites (Fig. 1), emplaced after an
132 earlier (242.01 ± 0.05 Ma) phreatomagmatic episode recorded by lapilli tuffs and tuff deposits.
133 The effusive phase lasted < 0.9 Ma (Storck et al., 2020), and was characterized by the production
134 of submarine pillow lavas, pillow breccias, lava breccias, hyaloclastites, and subaerial lava
135 flows, rarely intercalated with volcanoclastic sandstones. The production of volcanites was
136 associated to the emplacement of articulated dyke swarms and plutonic bodies, which are in
137 magmatic contact with the overlying lava flows and/or cut by dykes with the same composition
138 of the volcanites (Casetta et al., 2018b).

139

140 **3. Materials and methods**

141 The volcanic products (lava flows and lava breccias) and dykes crop out in the Dolomites, and
142 are spatially associated to the Predazzo Intrusive Complex (Fig. 1). Due to the widespread
143 hydrothermal alteration affecting the magmatic products in this area, an initial collection of
144 more than 200 samples was reduced to ~50 specimens for the purposes of this study, after
145 detailed screening based on thin section observations. The least altered samples were selected
146 for detailed whole-rock and in-situ chemical determinations. Whole-rock major and trace
147 element chemistry was determined by means of an ARL Advant-XP automated X-Ray
148 Fluorescence (XRF) spectrometer hosted at the Department of Physics and Earth Sciences of
149 the University of Ferrara. Full matrix correction procedure and intensities were completed
150 following Traill and Lachance (1966). Accuracy and precision were better than 2-5% for major
151 elements and 5-10% for trace elements. Detection limits were 0.01 wt% and 1-3 ppm for most
152 of the major and trace element concentrations, respectively. Rb, Sr, Y, Zr, Nb, Hf, Ta, Th, U,
153 and REE analyses were carried out at the Department of Physics and Earth Sciences of the
154 University of Ferrara by means of a Thermo Series X inductively coupled plasma-mass
155 spectrometer (ICP-MS). Precision and accuracy were better than 10% for all elements, well

156 above the detection limit. Mineral phase major element chemistry was determined by using a
157 CAMECA SX100 electron microprobe equipped with four WD and one ED spectrometers
158 hosted at the Department of Lithospheric Research of the University of Wien. The operating
159 conditions were as follows: 15 kV accelerating voltage, 20 nA beam current, and 20 s counting
160 time on peak position. For feldspars, a 5 μm defocused beam and 10 s counting time on peak
161 position for Na and K were used. Natural and synthetic standards were used for calibration, and
162 PAP corrections were applied to the intensity data (Pouchou and Pichoir, 1991). Trace element
163 concentrations of clinopyroxene were measured at the CNR - Istituto di Georisorse of Pavia by
164 laser ablation microprobe-inductively coupled plasma-mass spectrometry (LAM-ICP-MS)
165 following the protocols described by Tiepolo et al. (2003). NIST 610 and NIST 612 standard
166 glasses were used to calibrate relative element sensitivity. Precision and accuracy for trace
167 element analyses were assessed by standard sample BCR-2 (reference values from USGS
168 Geochemical Reference Materials Database). Each analysis was corrected with internal
169 standards using CaO. The detection limit, function of the ablation volume and counting time
170 were calculated for each analysis. Since ablation volume greatly depends on laser type and
171 sample (matrix), the detection limit decreased with decreasing spot size, beam power and cell
172 gas flow. A 40-100 μm beam diameter and 20 $\mu\text{m/s}$ scanning rate were used. The theoretical
173 detection limit ranges from 10 to 20 ppb for REE, Ba, Th, U, Zr and are about 2 ppm for Ti.
174 Whole-rock $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ analyses were made at the Scottish Universities
175 Environmental Research Centre (SUERC) by thermal ionization mass spectrometry (TIMS)
176 following procedures described by Casetta et al. (2018a). Eight measurements of SRM-987 and
177 twelve of JNdi-1 made during the course of this analytical programme yielded mean values of
178 0.710244 ± 0.000016 for $^{87}\text{Sr}/^{86}\text{Sr}$ and 0.512079 ± 0.000018 (2 SD) for $^{143}\text{Nd}/^{144}\text{Nd}$, consistent
179 with the consensus values of ~ 0.71025 and ~ 0.51210 .

180

181 **4. Petrography and mineral chemistry**

182 Lava flows with trachybasaltic to trachyandesitic composition crop out in thick sequences
183 emplaced over the Permo-Triassic volcano-sedimentary basement, and/or in direct contact with
184 the coeval plutonic counterparts. Multiple dykes, ranging in thickness from 10-20 cm up to 1-
185 2 m, cut both the basement and the volcano-plutonic bodies (Fig. 1). The crosscutting
186 relationships between the dykes and the plutonic bodies have been used to reconstruct the
187 temporal evolution of the multi-pulse Predazzo Intrusive Complex, which is composed of SiO₂-
188 saturated (SS) to SiO₂-undersaturated (SU) batches (Casetta et al., 2018b). Geochemical and
189 petrographic data (see below) support the discrimination between the two SS and SU suites for
190 the dykes studied in the present work, which mirror the geochemical features of the plutonic
191 rocks. They include hypersthene- to nepheline-normative trachybasalts, differentiating towards
192 quartz- and nepheline-normative trachytes, respectively. All lavas, on the other hand, are SiO₂-
193 saturated, highlighting how the occurrence of SU products was volumetrically limited.

194

195 ***4.1. Lava flows***

196 The texture of the SS trachybasaltic to trachyandesitic lavas is generally porphyritic to
197 glomeroporphyritic, with Porphyricity Index (P.I.) ranging from 5-10 up to 55-60%; these
198 values suggest near hypabyssal conditions of the last stages of their crystallization (Fig. 2). The
199 main phenocrysts are clinopyroxene (up to 3 mm), plagioclase (200 μm-2 mm) and spinel (50-
200 500 μm), together with some rare olivine (up to 1 mm), often altered in iddingsite or serpentine,
201 in the least differentiated samples (see Supplementary Material). Clinopyroxene ranges from
202 diopside to augite (Morimoto, 1988), with Mg# [calculated as Mg/(Mg+Fe_{TOT}) mol%] and TiO₂
203 contents of 70.0-74.1 and 0.5-1.3 wt%, respectively (Fig. 3). Its Al₂O₃ content varies between
204 2.5 and 4.7 wt%, whereas Cr₂O₃ is always < 0.1 wt%. Occasionally, glomerophyric aggregates
205 of clinopyroxene + spinel are observed. Spinel ranges in composition from magnetite to Ti-
206 magnetite (TiO₂ = 0.1-17.4 wt%). Plagioclase, ranging in composition from bytownitic to
207 labradoritic (An₅₈₋₈₇), is often sericitized in the less differentiated rocks, while it is quite fresh

208 in the trachyandesites. The groundmass of all lavas ranges from microcrystalline to
209 hypohyaline, constituted by the same mineral assemblage as the phenocrysts. Groundmass
210 plagioclase in the trachyandesitic lavas has andesinic composition (An_{37-44}).

211

212 **4.2. Dykes**

213 The P.I. of the trachybasaltic to trachyandesitic dykes varies from 5-10% up to 50-55%. SS
214 trachybasaltic to trachyandesites have paragenesis and texture comparable to those of the lavas,
215 except for the groundmass, which in some cases has coarser grain size and hosts some K-
216 feldspar crystals (Or_{79-90}). Plagioclase ranges in composition from bytownitic to labradoritic,
217 spinel is magnetite to Ti-magnetite, and clinopyroxene is similar in composition to the
218 phenocrysts in the SS lavas (Fig. 3) (see Supplementary Material).

219 Some trachybasaltic dykes, cropping out in proximity of the Latemar area (Fig. 1), have the
220 most primitive composition, and likely belong to the SU suite. They are constituted by
221 microcrystalline to hypohyaline groundmass, and clinopyroxene (400 μm -2 mm), plagioclase
222 (100-400 μm) and spinel (<200 μm) phenocrysts, together with fresh, zoned olivine phenocrysts
223 and xenocrysts (500 μm -2 mm; Fo_{85-92} cores to Fo_{78-81} rims, at NiO 0.07-0.15 wt%; Fig. 2). In
224 these dykes, clinopyroxene phenocrysts are Al-Cr-diopsides, with Mg# between 78.1 and 85.0,
225 TiO_2 contents up to 2.0 wt% and Cr_2O_3 concentrations ranging from 0.1 to 1.0 wt% (Fig. 3).
226 Some phenocrysts exhibit compositional zoning, with Cr-diopsidic cores (Mg# up to 88.4; TiO_2
227 = 0.2 wt%; Cr_2O_3 = 0.5 wt%) graded in diopsidic-augitic rims (Mg# down to 67.6; TiO_2 up to
228 1.0 wt%; Cr_2O_3 ~0.02 wt%). Throughout SU dykes, plagioclase phenocrysts range in
229 composition from bytownite to labradorite, whereas spinel encompass the magnetite to Ti-
230 magnetite compositional range. Spinel composition is different, however, in the least
231 differentiated SU trachybasaltic dyke, and ranges from Ti-magnetite to Cr-spinel (TiO_2 = 0.8-
232 16.9 wt%; Cr_2O_3 = 4.0-28.3 wt%). Groundmass K-feldspar ranges in composition from Or_{79} to
233 Or_{90} .

234 The trachytic dykes, distinguishable in the field by their bright pink to grey (SS quartz-
235 trachytes) or light pink to greenish (SU nepheline-trachytes) colours, have low P.I. (usually
236 <25%). Quartz-trachytes are mainly constituted by K-feldspar (300 μm -4 mm), plagioclase (up
237 to 7 mm) and spinel (<200 μm) phenocrysts in a micro- to cryptocrystalline groundmass made
238 by apatite, feldspars, biotite, quartz and Fe-Ti oxides. Nepheline-trachytes are usually highly
239 altered, and composed of K-feldspar (1-5 mm), plagioclase (up to 1 mm), and rare nepheline
240 phenocrysts in a cryptocrystalline to hypohyaline groundmass. Plagioclase phenocrysts and
241 groundmass crystals in SS-SU trachytes have large compositional variations, ranging from
242 labradorite to albite. K-feldspar is quite homogeneous, ranging from Or₉₂ to Or₉₉.
243 Overall, the composition of clinopyroxene phenocrysts is the best proxy for discriminating
244 between the SS and SU suites: clinopyroxene in SS rocks is generally more depleted in TiO₂
245 with respect to those in SU samples, at comparable Mg# (Fig. 3). Such a feature further
246 evidences the more pronounced “alkaline” character of SU samples with respect to the SS ones,
247 fostering a discrimination between the two suites.

248

249 ***4.3. Ultramafic cumulates***

250 In the Latemar area, some SU trachybasaltic dykes contain ultramafic xenoliths (Figg. 1-2). The
251 least altered clinopyroxenites have cumulitic texture, with small (200-300 μm) to medium-sized
252 (2-3 mm) fresh clinopyroxene crystals in contact by means of triple junctions, and sometimes
253 enclosing small olivine grains (Fig 2). Clinopyroxene is diopside (Mg# = 85.5-87.6; Wo = 47-
254 48), with TiO₂ and Cr₂O₃ contents of ~0.2 wt% and 0.1-0.3 wt%, respectively (Supplementary
255 Material). At the edges of the xenolith, the contact with the host SU trachybasalt induced
256 marked TiO₂ (up to 2.3 wt%) and Cr₂O₃ (up to 1.2 wt%) enrichments, coupled with a general
257 decrease in the Mg# (from 84.0 down to 62.5). Beside these contact areas, clinopyroxene show
258 compositional analogies with the phenocrysts of the SS suites (Fig. 3). In chondrite-normalized
259 (Sun and McDonough, 1989) incompatible element patterns, these clinopyroxene have Rb-Ba

260 and Nb-Ta depletions, as well as marked Zr-Hf and Ti negative anomalies. Their REE patterns
261 have the convex-upward shape typical of alkaline magmas, with $(La/Yb)_N$ values of 1.91-2.00
262 (Fig. 3). Although preserving the significant HFSE paucity, clinopyroxene grains in proximity
263 of the contact with the host basalt are significantly REE-enriched, and have $(La/Yb)_N$ values of
264 ~1.6 (Fig. 3).

265

266 **5. Whole-rock geochemistry**

267 ***5.1. Major and trace element chemistry***

268 Lavas and dykes belong to both the alkaline and transitional series with K-affinity (Fig. 4). On
269 the Total Alkali vs. Silica (TAS) diagram (Fig. 4), lavas plot between the basaltic and the
270 trachyandesitic fields ($SiO_2 = 49.2-59.9$ wt%; $Na_2O+K_2O = 3.1-8.5$ wt%), whereas dykes show
271 wider compositional variations, from basaltic to trachytic and phonolitic ($SiO_2 = 48.1-63.7$
272 wt%; $Na_2O+K_2O = 3.4-12.2$ wt%) (Fig. 4; Table 1). The Mg# [calculated as $MgO/(MgO+FeO)$
273 mol%, considering a Fe_2O_3/FeO ratio of 0.15, in agreement with fO_2 around +1 ΔFMQ (Kress
274 and Carmichael, 1991; Casetta et al., 2018a)] of the lavas ranges from 64.3 in trachybasaltic
275 samples down to 41.5 in trachyandesites, whereas dykes have Mg# of 65.4 (trachybasalts) to
276 23.7 (trachytes). The TiO_2 and FeO_{TOT} contents, as well as the CaO/Al_2O_3 ratio of the samples
277 are inversely correlated with the silica content, and decrease from 1.5 to 0.2 wt%, from 10.5 to
278 2.0 wt% and from 0.78 to 0.03, respectively (Fig. 4). Similarly, Ni decreases from 89 to 1 ppm,
279 while Zr, Nb and Rb concentrations increase from 77 to 779 ppm, from 5 to 51 ppm and from
280 12 to 344 ppm, respectively, with increasing SiO_2 (Fig. 4). As for the corresponding intrusive
281 suites (Casetta et al., 2018a; 2018b), SU and SS rocks can be easily discriminated by their FeO
282 contents, CaO/Al_2O_3 and Na_2O/K_2O ratios, as well as by their Rb, Zr and Nb concentrations at
283 comparable silica content (Fig. 4).

284 In the Primitive Mantle- (PM; Sun and McDonough, 1989) normalized incompatible element
285 diagram, all lavas and dykes are enriched in LILE, Th-U and LREE, and display the Nb-Ta and

286 Ti negative anomalies typical of magmatic suites generated by subduction-modified mantle
287 source (Fig. 5). Moving from trachybasaltic to trachytes, the incompatible element patterns are
288 increasingly enriched in LILE and depleted in Ti. All samples have flat Middle (M-) to Heavy
289 (H-) REE profiles in chondrite-normalized (Sun and McDonough, 1989) diagrams, and
290 moderate enrichment in LREE, with $(La/Yb)_N$ values comprised between 8.5 and 9.5 for the
291 least differentiated rocks of both suites (Fig. 5). Despite the similar incompatible element
292 concentrations with respect to the most primitive SS lavas and dykes, SU trachybasalts have
293 slightly higher Zr-Hf abundances, and less pronounced Nb-Ta-Ti anomalies. Significant
294 differences can be envisaged by comparing the most differentiated trachytic products of the two
295 suites: SU trachytes have extremely higher Th-U, Zr and LREE concentrations with respect to
296 SS ones, this latter parameter resulting in a marked difference in their $(La/Yb)_N$ ratios (Fig. 5;
297 Table 1). On the Th_N vs. Nb_N tectonic discrimination diagram (Fig. 6a), the SS/SU volcanic
298 series fall within the calc-alkaline, (i.e. orogenic) field. Their HFSE distribution is well
299 consistent with magmatic suites from active continental margins, as also testified by the
300 distribution of the least differentiated SS-SU rocks on the Th-Hf-Ta diagram (Fig. 6c). In both
301 diagrams, the marked difference between Mid-Triassic SS/SU rocks and Late Triassic
302 lamprophyres from the same area (Casetta et al., 2019) supports the hypothesis that different
303 mantle sources were activated within a period of ~20 Ma to generate the distinct magmatic
304 events. This feature is also highlighted by the Nb/La vs. La/Yb ratios of the least differentiated
305 SS/SU rocks (Fig. 6b), which suggest that the contribution of lithospheric mantle in their
306 genesis was more pronounced than for lamprophyres, which bear witness of a marked
307 asthenospheric involvement.

308

309 **5.2. $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$**

310 Whole-rock $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ isotopic ratios were measured on representative samples
311 of lavas and dykes from both the SS and SU suites (Table 1). Initial isotopic ratios, respectively

312 named $^{87}\text{Sr}/^{86}\text{Sr}_i$ and $^{143}\text{Nd}/^{144}\text{Nd}_i$, were corrected to an age of 238 Ma, in accordance with the
313 recently published U-Pb datings (from 238.19 ± 0.05 to 237.58 ± 0.04 Ma; Storck et al., 2020).
314 Lavas and dykes from the SS suite have $^{87}\text{Sr}/^{86}\text{Sr}_i$ values ranging from 0.7043 to 0.7057, and
315 $^{143}\text{Nd}/^{144}\text{Nd}_i$ from 0.51219 to 0.51235 (Fig. 7). Dykes belonging to the SU suite are
316 characterized by slightly higher $^{87}\text{Sr}/^{86}\text{Sr}_i$ (0.7055-0.7058) with respect to SS rocks, at
317 $^{143}\text{Nd}/^{144}\text{Nd}_i$ of 0.51226-0.51232. On the whole, Sr-Nd isotopic data for the SS and SU lavas
318 and dykes are similar to those of the corresponding intrusive rocks, lying in proximity of the
319 enriched mantle (EM I) end-member, in line with most of the Southern Alps Mid-Triassic
320 magmatic rocks (Fig. 7; Lustrino et al., 2019). They are also well distinct from the Late Triassic
321 lamprophyres, which plot close to the depleted MORB mantle (DMM) end-member (Fig. 7;
322 Workman and Hart, 2005). The only available analysis for cumulitic clinopyroxenite xenoliths
323 yields an $^{87}\text{Sr}/^{86}\text{Sr}_i$ value of 0.7044, i.e. similar to those of the SS rocks, but slightly lower with
324 respect to the initial Sr ratios typical of SU dykes.

325

326 **6. The primary shoshonitic magmas and their SCLM source**

327 ***6.1. Reconstruction of primary shoshonitic melts***

328 The variable extent of differentiation that affected the Mid-Triassic shoshonitic rocks makes it
329 difficult to understand the compositional features of the primary melts, and from there to
330 understand the nature of the underlying SCLM. The least evolved lavas and dykes have Mg#
331 always <65, and are often characterized by moderate to high crystal cargoes (P.I. up to 50-55),
332 which result in “doped” bulk rock analyses, where the high Mg# and the low REE abundances
333 are function of the hypabyssal nature of the rocks rather than of their primitivity. No melt
334 inclusions were found in the early-forming phases (olivine or clinopyroxene), being the
335 composition of the most Mg-rich crystals inside lavas, dykes and cumulitic xenoliths (Fig. 3)
336 the only petrological evidence for the existence of primitive melts. For these reasons, the most
337 Mg-rich and least porphyritic dykes from the two magmatic suites (SS-SU) were chosen and

338 traced back to the ideal condition of equilibrium with a fertile mantle assemblage, in line with
339 what proposed by Casetta et al. (2020) for Mt. Etna magmas. Olivine and clinopyroxene plus
340 small amounts of Cr-bearing spinel were re-added to the melt in a backward fractionation
341 process, being aware of respecting the Fe-Mg equilibrium conditions between the progressively
342 reconstructed melt and each added mineral phases (i.e. $^{Ol-Melt}Kd_{Fe-Mg} = 0.27-0.33$ and $^{Cpx-}$
343 $^{Melt}Kd_{Fe-Mg} = 0.24-0.30$; Roeder and Emslie, 1970; Putirka, 2008). The occurrence of high-Fo
344 olivine xenocrysts and olivine inclusions in cumulitic clinopyroxene suggests that olivine
345 dominated the earlier crystallization stages, probably in a restricted *T-P* interval, followed in a
346 rapid sequence by clinopyroxene in the possible liquid line of descent of these magmas.
347 Therefore, olivine + clinopyroxene (diopside-augite) assemblages, together with small amounts
348 of Cr-bearing spinel, were progressively re-added to the chosen starting SS and SU rocks until
349 their major element composition resulted in equilibrium with a fertile mantle source (i.e. Mg#
350 68-70, in equilibrium with a Fo₈₈₋₉₀ olivine; see McDonough, 1990). The trace element
351 composition of the ideal primitive SS and SU melts was modelled by means of the Rayleigh
352 fractionation equation (Shaw, 1979), following Casetta et al. (2020). The adopted mineral/melt
353 trace element partition coefficients are listed in the Supplementary Materials. The primary
354 transitional to alkali basaltic SS and SU melts were obtained after 10.5-11.4% backward
355 fractionation of olivine (6.6-6.9 vol.%), clinopyroxene (2.4-3.0 vol%) and spinel (1.5 vol.%)
356 (Table 2). The modelled primitive melts have Mg# ~70 and preserve the above-mentioned
357 differences between the SS and SU suites, in terms of both major and trace element features
358 (Fig. 4). Their PM-normalized incompatible element patterns are characterized by LILE-
359 enrichment as well as Nb-Ta-Ti negative anomalies, these latter less marked in the primitive
360 SU melt, which preserve the slightly higher Zr-Hf abundance with respect to the SS one (Fig.
361 5). When compared to the Late Triassic lamprophyres, SS and SU rocks can be easily
362 discriminated on the basis of their HFSE abundances (Fig. 5).

363

364 **6.2. The composition of the mantle source**

365 The shoshonitic affinity and incompatible element distribution of the studied magmas indicate
366 their derivation from an enriched SCLM source modified by subduction-related components.
367 In fact, Sr-Nd, Pb and Hf isotopic signatures point towards an enriched mantle-derived
368 fingerprint, and rule out the effect of large (>10%) crustal assimilation during the peak of
369 effusive magmatism at 238 Ma (Casetta et al., 2018a; De Min et al., 2020; Storck et al., 2020).
370 Several authors hypothesized a decoupling between the timing of the shoshonitic magmatism
371 and the metasomatism of the lithospheric mantle source, relating this latter process to the
372 Variscan subduction (Bonadiman et al., 1994; Lustrino et al., 2019). Direct evidence of the
373 impact of the Variscan subduction on the Southern Alps SCLM is visible in the Ulten and Finero
374 peridotitic massifs, where modal (i.e. presence of amphibole, phlogopite, Cl-rich apatite) and/or
375 cryptic subduction-related metasomatism is documented (e.g., Zanetti et al., 1999; 2016;
376 Tumiatì et al., 2003; Ionov et al., 2017; Tommasi et al., 2017).
377 To simulate the slab-derived modification of the Southern Alps SCLM, Sr-Nd isotopic mixing
378 models (Faure, 1986; see also Melchiorre et al., 2020) between DMM and selected subduction-
379 related components were performed. The average composition of the global subducting
380 sediments (GLOSS/GLOSS-II; Plank, 2014) and the average composition of the Southern Alps
381 crust (Voshage et al., 1990) were chosen to represent possible end-members of subducting
382 crustal components, for modelling the Sr-Nd isotopic variability caused by slab-derived
383 influxes (see Kessel et al., 2005; Plank, 2014). The SS rocks lie on a mixing trend between
384 DMM and GLOSS-II, whereas the relative Sr-Nd isotopic enrichment of SU magmas
385 approaches the mixing curve between DMM and GLOSS, confirming that two different mantle
386 sources are associated to the genesis of SS and SU magmas. By contrast, the Finero
387 amphibole+phlogopite-bearing peridotite samples plot towards the GLOSS-II end-member,
388 resulting slightly more enriched in radiogenic Sr with respect to SS magmas (Fig. 7). The Ulten
389 peridotites span a large Sr-Nd isotopic range, as result of complex interactions between mantle

390 and metasomatising fluids with crustal signature (Fig. 7; see Tumiati et al., 2003). Metasomatic
391 amphibole and phlogopite in the Finero phlogopite peridotite, recording post-Variscan ages as
392 old as 310 Ma (Zanetti et al., 2016; Malitch et al., 2017, and references therein), mostly plot
393 along the mixing trend between DMM and the average crust (Fig. 7), enhancing the role played
394 by subduction-related melts/fluids in metasomatising the Southern Alps SCLM prior to 290 Ma
395 (Hartmann and Wedepohl, 1993; Zanetti et al., 1999; 2016; Selverstone and Sharp, 2011;
396 Giovanardi et al., 2020). Following the Sr-Nd isotopic mixing models, the ideal trace element
397 composition of the enriched SCLM domains that generated the SS/SU primary melts has been
398 firstly approached by mixing DMM with 4% GLOSS-II (SS magmas) and 2% GLOSS (SU
399 magmas) components. The inferred mantle source of SS magmas has LILE-enrichment, Nb-
400 Ta-Ti negative troughs, slight LREE-enrichment and PM-like MREE-HREE contents (Fig. 8).
401 The calculated mantle source of SU magmas is relatively depleted in all trace elements with
402 respect to the SS one, being typified by PM-like LREE concentrations and DMM-like MREE-
403 HREE profiles (Fig. 8).

404 Subsequently, the mineral phases of highly fertile composition from the Southern Alps
405 peridotitic massifs (Mazzucchelli et al., 2010; Giovanardi et al., 2020) were used to model the
406 trace element content and modal composition of the ideal “SS” and “SU” sources *via* mass
407 balance calculations. Results showed that the trace element concentrations of the ideal “SS”
408 and “SU” sources are matched by considering two amphibole+phlogopite-bearing spinel-
409 lherzolites with the following modal compositions: 55.4% olivine + 23.0% orthopyroxene +
410 17.7% clinopyroxene + 1.1% spinel + 2.0% amphibole + 0.8% phlogopite (SS source); 57.5%
411 olivine + 24.3% orthopyroxene + 15.1% clinopyroxene + 1.1% spinel + 1.4% amphibole +
412 0.6% phlogopite (SU source) (Table 3). It should be highlighted that all the considered phases
413 belong to the Finero phlogopite peridotite, except clinopyroxene, which is from the Baldissero
414 peridotite. Clinopyroxene in the Finero phlogopite peridotite in fact is highly depleted,
415 recording up to ~20% melt extraction (Coltorti and Siena, 1984; Hartmann and Wedepohl,

416 1993), being thus unsuitable for simulating an enriched mantle source. By considering the
417 maximum H₂O contents measured in the mineral phases of the Finero peridotite (3.2 ppm
418 olivine; 111 ppm orthopyroxene; 267 ppm clinopyroxene; 1.93 wt% amphibole; 4.3 wt%
419 phlogopite; Tommasi et al., 2017), the amounts of H₂O retained in the modelled SS and SU
420 mantle sources would result of 784 and 598 ppm, respectively (Table 3). These values are well
421 within the range proposed for other worldwide SCLM domains, such as the Subei Basin in
422 China (328-1440 ppm H₂O; Hao et al., 2019) or the Norther Victoria Land in Antarctica (910-
423 1486 ppm H₂O; Giacomoni et al., 2020). In line with the trace element and Sr-Nd isotopes
424 mixing results, the compositional discrimination between the modelled SS and SU sources is
425 mirrored by different incompatible elements and H₂O concentrations, in turn related to the
426 lower modal amounts of clinopyroxene, amphibole and phlogopite (Fig. 8).

427

428 **7. The genesis of shoshonitic magmatism**

429 ***7.1. Mantle source melting model***

430 The partial melting conditions of the mantle sources required to produce the primary SS and
431 SU transitional to alkali basaltic melts were simulated by using the mineral phase modal
432 proportions and the eutectic composition of the peridotitic systems as function of the degree of
433 partial melting (see Casetta et al., 2020). A “melting+recrystallization” mass balance model
434 enabled us to simulate the contribution of the fusible components (e.g., TiO₂, Al₂O₃ and alkalis)
435 from each mineral phase of the lherzolitic source, and then reproduce the major element
436 composition of the produced magmas. In this process, the contribution during melting of each
437 phase with fertile composition is counterbalanced by recrystallization of a restitic residuum
438 (e.g., melting of Al-rich clinopyroxene + recrystallization of Al-poor clinopyroxene), in
439 accordance with the conditions of melting of a amphibole+phlogopite-bearing lherzolite in the
440 spinel stability field (Gaetani and Grove, 1998; Conceição and Green, 2004; Liu et al., 2006;

441 Condamine and Médard, 2014). The relative contribution of each mineral phase at the eutectic
442 was evaluated, for major elements (oxide, wt%) by least square method, following the equation:

443
$$S = \sum_{i=1}^n r_i^2 .$$

444 Here, the optimal parameter S is fitted by minimizing the sum of squared residuals r^2 between
445 the predicted values for each oxide “i” (e.g., SiO₂ wt% of the primitive melt) and the values
446 calculated from the melting model (e.g., SiO₂ wt% resulting from the
447 “melting+recrystallization” process). For our purposes, the model was considered accurate
448 when $S < 1$, i. e. when the difference between modelled and calculated parameters, expressed as
449 oxides wt%, is well within the analytical errors (i.e. XRF and/or EPMA determinations). The
450 partial melting degree (F) associated to the production of primitive melts was then calculated
451 as the difference between the melted fertile assemblage and the recrystallized restitic residuum,
452 checking the consistency with the experimental data on genesis of shoshonitic melts (e.g.,
453 Conceição and Green, 2004). Once fixed the mineral phase proportions required to produce the
454 primary SS/SU magmas at the eutectic, the non-modal batch melting equation of Shaw (1979)
455 was applied to model their trace element distribution. In this equation, expressed as:

456
$$C_L = \frac{C_0}{[D+F \times (1-P)]},$$

457 C_L represents the concentration of each trace element in the primary melt, C_0 its concentration
458 in the modelled source, D its partition coefficients weighted for the modal composition of the
459 source and P the partition coefficients weighted for the eutectic mineral proportions (see Table
460 2 and Supplementary Material for the partition coefficients and mineral melting proportions).

461 The application of this equation to elements with highly incompatible behaviour, i.e. with $D \rightarrow 0$
462 and $P \rightarrow 0$, approximated by the ratio:

463
$$F = \frac{C_0}{C_L},$$

464

465 was also used throughout the calculations to check the consistency of the partial melting
466 degrees.

467

468 ***7.2. Partial melting conditions and role of H₂O***

469 The partial melting models suggest that 7% partial melting of a fertile spinel-lherzolite bearing
470 small proportions of amphibole (1.98%) and phlogopite (0.76%) can produce the major/trace
471 element content of primary SS transitional basalt. Hydrous phases were totally consumed
472 during this melting stage, significantly contributing to the eutectic of the system (Table 4). A
473 negligible contribution of spinel accompanied by positive melting proportions of
474 orthopyroxene, clinopyroxene and by dominant recrystallization of olivine, is also required to
475 account for the major/trace element distribution of primary SS melts. On the other hand, the
476 production of the SU primary alkali basalt can be explained by 5% partial melting of a spinel-
477 lherzolite with slightly lower abundances of hydrous phases (1.36% amphibole + 0.62%
478 phlogopite). Again, hydrous phases were completely consumed, together with significant
479 proportions of orthopyroxene and clinopyroxene. Using experimental estimations of mineral-
480 melt water partitioning during peridotite melting, the estimated H₂O amounts of the produced
481 primary melts are in the range of 1.06-1.12 wt% (Table 4; Supplementary Material). In terms
482 of trace elements, the predicted and calculated SS/SU melts are very close, especially the LILE
483 and REE (Fig. 9). In both models, the main discrepancies are related to U, Zr-Hf and Sr, whose
484 concentrations in the sources resulted too low to match those of the primary melts (see
485 discussion below).

486

487 ***7.3. Peridotite melting and mantle source depth***

488 Conceição and Green (2004) showed that decompression melting of a metasomatised spinel-
489 peridotite bearing small amounts amphibole and phlogopite can produce shoshonitic magmas
490 at ~5% partial melting, $T = 1075^{\circ}\text{C}$ and $P = 1.0 \text{ GPa}$, by exhausting hydrous phases very close

491 to the solidus. Consistent with their findings, our model shows that 5-7% melting of an
492 amphibole+phlogopite-bearing spinel lherzolite can reproduce the composition of the primary
493 SS/SU melts, requiring the complete disappearance of both amphibole and phlogopite.

494 During the two melting stages, the contribution of the pyroxenes was opposite. Indeed, while
495 the SS melting stage was characterized by an orthopyroxene-dominated eutectic, in the SU
496 melting stage the maximum contribution at the eutectic was given by clinopyroxene (Table 4).
497 This behaviour can be explained by the role played by volatiles (H₂O, CO₂) and alkalis during
498 partial melting. Experimental studies showed that the amount of clinopyroxene and
499 orthopyroxene consumed during melting of spinel lherzolites at 1.1-1.2 GPa are negatively and
500 positively correlated to the amount of H₂O in the system, respectively (Gaetani and Grove,
501 1998; Condamine and Médard, 2014). Moreover, Liu et al. (2006) found that the effect of H₂O
502 is somewhat similar to that of K₂O, so that increasing the concentration of these elements in the
503 peridotitic source results in the genesis of SiO₂-saturated (to oversaturated) melts. Conversely,
504 decreasing H₂O and/or increasing Na₂O in the source system leads to the production of
505 nepheline-normative liquids (Liu et al., 2006). Accordingly, the slightly different H₂O contents
506 of the SS and SU sources are associated to the genesis of hypersthene- and nepheline-normative
507 melts, respectively.

508 According to Selverstone and Sharp (2011) results, the Finero amphibole-phlogopite peridotite
509 equilibrated at *P* of 1.2-1.6 GPa. These values are consistent with the experimental works on
510 the genesis of shoshonitic melts from peridotitic sources (e.g., Conceição and Green, 2004) and
511 with the experimental studies on amphibole/phlogopite stability in the SCLM (Condamine and
512 Médard, 2014; Mandler and Grove, 2016). In this framework, it is likely that the transitional to
513 alkali basaltic SS/SU melts were generated was located at 40-60 km of depth, i.e. well above
514 the spinel-garnet transition in SCLM domains (2-3 GPa, i.e. 60-90 km; Falloon and Green,
515 1988). This melting depth is consistent with the equilibration depth calculated by Carraro and
516 Visonà (2003) for the spinel-lherzolitic xenoliths sampled along the same SCLM column (i.e.

517 ~45 km), and quite shallower than the one from which alkaline lamprophyres were generated
518 during the subsequent Late Triassic cycle (70-80 km; Casetta et al., 2019).

519

520 **8. The ascent and emplacement of the Middle Triassic melts**

521 ***8.1. Equilibrium conditions and adopted thermobarometric/hygrometric models***

522 The *T-P-H₂O* paths followed by SS and SU melts in the interval comprised between segregation
523 from the mantle source and their emplacement/differentiation in shallow crust were tracked by
524 means of thermobarometric and hygrometric equations based on olivine- and clinopyroxene-
525 melt equilibria. To reconstruct the *T* conditions of crystallization of the least differentiated SU
526 magmas, the olivine- and clinopyroxene-melt thermometers of Putirka (2008; Equations 22 and
527 33) were applied to the phenocrysts yielding equilibrium with the whole-rock composition of
528 the host dykes (i.e. $^{Ol-Melt}Kd_{Fe-Mg} = 0.27-0.33$ and $^{Cpx-Melt}Kd_{Fe-Mg} = 0.24-0.30$; Fig. 10). Due to
529 the generally low crystal cargo of the SU trachybasaltic dyke (P.I. down to 5-10%; Table 1), its
530 whole-rock compositions was considered a good approximation of a melt composition. The
531 pressures of clinopyroxene crystallization were calculated by means of the Putirka (2008,
532 Equation 32b) H₂O-dependent barometer, in turn derived from the Putirka et al. (1996, Equation
533 T1) thermometer. Because of the sensitivity of the chosen thermobarometers to the H₂O content
534 of the melt, multiple calculations were performed using as input variable amounts of H₂O (from
535 1 to 5 wt%). These H₂O-dependent equations were then iterated with the *T-P*-dependant
536 clinopyroxene-based hygrometer of Perinelli et al. (2016) until matching results were obtained,
537 in a process that enable to skip circular assumptions of the *T-P-H₂O* conditions of the
538 crystallizing melts (see Casetta et al., 2018a; Lanzafame et al., 2020).

539 Due to the absence of unaltered olivine phenocrysts, the primary *T-P-H₂O* conditions of
540 crystallization of SS melts were modelled by applying thermobarometric and hygrometric
541 equations to clinopyroxene in early-derived clinopyroxenitic cumulates. Accordingly,
542 clinopyroxene grains inside clinopyroxenitic xenolith were traced back to the equilibrium

543 conditions with the least differentiated SS trachybasaltic dyke in terms of both major (Fig. 10a)
544 and trace elements (Fig. 10b; see Supplementary Material for the adopted mineral-melt partition
545 coefficients for trace elements). The most Mg-poor clinopyroxenes in trachybasaltic lavas
546 resulted in clear disequilibrium with their host rock (Fig. 10), likely because the high P.I. of the
547 studied lavas, crystallized in almost hypabyssal conditions, make them far from resembling a
548 melt in terms of whole-rock composition. Therefore, these crystals were traced back to
549 equilibrium conditions (i.e. $K_{\text{Fe-Mg}}^{\text{Cpx-Melt}} = 0.24\text{-}0.30$; Putirka, 2008) with less porphyritic
550 trachybasaltic and trachyandesitic rocks belonging to the same magmatic suite (see also Casetta
551 et al., 2018a). Afterwards, the chosen clinopyroxene-melt thermobarometers and hygrometers
552 (Putirka, 2008; Perinelli et al., 2016) were iteratively applied to determine the T - P - H_2O space
553 of clinopyroxene crystallization. Unfortunately, no reliable P - H_2O results were obtained by the
554 iterative procedures for the clinopyroxene crystals inside clinopyroxenitic xenoliths. Therefore,
555 to obtain a rough estimate of their crystallization P , the Putirka et al. (2003) H_2O -independent
556 barometer was used. Further details, together with a complete list of the obtained results and of
557 the errors associated to each adopted equation are listed in Table 5.

558

559 **8.2. T - P - H_2O paths**

560 Thermometric models showed that the phenocrysts of olivine (Fo_{85}) and clinopyroxene (Mg\#_{Cpx}
561 85) in equilibrium with the SU trachybasaltic dyke (Mg\#_{Liq} 64) record the most congruent
562 crystallization T (from 1195 to 1204-1207 °C), by considering a H_2O content in the melt of 1.5-
563 1.9 wt%. The corresponding crystallization P obtained from the clinopyroxene-melt barometer
564 is ~830 MPa (Fig. 10). The deepest conditions of crystallization of SS magmas ($P = 570\text{-}710$
565 MPa) are from clinopyroxene grains inside cumulitic xenoliths (Mg\#_{Cpx} 86-88) in equilibrium
566 with the SS trachybasaltic dyke (Mg\#_{Liq} 65), at T between 1100 and 1128 °C (Fig. 10). The
567 most Mg-poor clinopyroxene phenocrysts (Mg\#_{Cpx} 70-83) analysed in SS trachybasaltic to
568 trachyandesitic lavas and dykes (equilibrium melts = Mg\#_{Liq} 41-53) record shallower

569 conditions. Their crystallization occurred along a polybaric path at P comprised between 470
570 and 230-170 MPa, until the onset of a nearly isobaric cooling regime took place at P of 50-170
571 MPa (Fig. 10). Although plausible and in line with previous estimates for the same magmatic
572 suite (e.g., Casetta et al., 2018a; De Min et al., 2020), these T - P results have to be considered
573 together with the associated errors (i.e. ± 29 - 42°C ; ± 170 - 260 MPa; Table 5).

574

575 **9. Discussion**

576 ***9.1. The Southern Alps SCLM and the genesis of shoshonitic melts***

577 A possible relationship between the metasomatized orogenic peridotites and the Mid-Triassic
578 orogenic-like magmatism in the Southern Alps has been often hypothesized (Zanetti et al.,
579 1999; 2013; Mazzucchelli et al., 2010; De Min et al., 2020). As shown in Fig. 7, the shoshonitic
580 magmas of the Dolomites share a common enriched (EM I-like) Sr-Nd isotopic signature
581 (Bonadiman et al., 1994; Casetta et al., 2018a; Lustrino et al., 2019), mirroring the fingerprint
582 of the underlying mantle source. The existence of K-H₂O-rich mantle-derived melts in the
583 Southern Alps SCLM during Mid-Triassic is also supported by the intrusion of gabbroic dykes
584 with similar Sr-Nd isotopic signature to the SS/SU magmas inside the Finero peridotite
585 (Giovanardi et al., 2020). According to many authors, the slab-derived enrichment of the
586 Southern Alps SCLM has to be ascribed to the Variscan subduction (Bonadiman et al., 1994;
587 Lustrino et al., 2019), the effect of which are also preserved in the metasomatised Finero and
588 Ulten peridotites (Zanetti et al., 1999; Ionov et al., 2017; Giovanardi et al., 2020). In the Finero
589 peridotite, several generations of hydrous phases (amphibole and phlogopite) formed after
590 pervasive and prolonged metasomatism of depleted harzburgites (Voshage et al., 1987;
591 Hartmann and Wedepohl, 1993; Zanetti et al., 1999; 2013; 2016; Tommasi et al., 2017;
592 Giovanardi et al., 2020). The infiltration of slab-related metasomatic agents in the Finero main
593 peridotitic body is also suggested by the Nb-Ta-Ti negative anomalies and enrichment in
594 radiogenic Sr of amphibole grains, which lie on the mixing trend between DMM and the

595 average crustal (and/or GLOSS/GLOSS-II) components (Fig. 7; see also Giovanardi et al.,
596 2020). Although partially overprinted by the complex *T-P-t* trajectories followed from
597 depletion to the subsequent Palaeozoic metasomatism and the Mesozoic (~180 Ma) exhumation
598 at shallower levels (Hartmann and Wedepohl, 1993; Zanetti et al., 2016; Langone et al., 2017),
599 the Finero peridotite is the best approximation of the Southern Alps SCLM during
600 MiddleTriassic, i.e. at the time of the shoshonitic magmatism. Accordingly, our models enabled
601 us to set out that an assemblage composed of phases with most fertile composition from the
602 Finero (and Baldissero) peridotite is able to approach, in terms of major/trace elements, a bulk-
603 rock source composition similar to that resulting from a mixing between DMM (98-96%) and
604 GLOSS/GLOSS-II (2-4%) components (Fig. 9; Table 3). Besides confirming the role played
605 by subductive crustal components in the modification of the Southern Alps SCLM, the
606 modelled fertile mantle source perfectly encompasses the Sr-Nd isotopic variations of the Mid-
607 Triassic shoshonitic magmas (Fig. 7; Table 1). The modelled sources are consistently able to
608 account for the major/trace element composition of the primary SS/SU magmas at 5-7% partial
609 melting. These melting degrees are higher than those hypothesized by Lustrino et al. (2019),
610 who suggested that a supra-subductive peridotite interacting with slab-derived components is
611 able to generate the Mid-Triassic shoshonitic magmatism at 2% slab + 1.1% peridotite melting.
612 According to these authors, although reliable in terms of geodynamic interpretation, such a
613 model is limited by the use of a non-primary target melts (Mg# 61), resulting in excessively
614 low degrees of partial melting.

615 The incompatible element patterns of the modelled SS/SU melts mirror those of the primary
616 magmas, except for some discrepancies in the Th-U, Sr and Zr-Hf distributions (Fig. 9), which
617 likely reflect the low concentrations of these elements in the modelled sources, i.e. in the
618 mineral phases in the Finero peridotite (see Giovanardi et al., 2020). The “ideal” mantles
619 resulting from the mixing between DMM and GLOSS/GLOSS-II components are enriched in
620 Th-U and Zr-Hf with respect to the modelled Finero-like sources (Fig. 9). The positive Sr spike

621 in the primary melts, a peculiar feature of magmas generated from subduction-modified sources
622 (Zheng, 2019), cannot be reproduced neither by the modelled sources nor by the “mixed”
623 DMM+GLOSS mantles. This discrepancy could be explained by the variable concentration of
624 Sr in worldwide sediments (see Plank, 2014) or, alternatively, by the extremely variable and T -
625 P -dependent mobility of Sr in metasomatic melts/fluids and/or supercritical fluids in
626 subduction-related settings (Kessel et al., 2005).

627

628 ***9.2. Ascent and emplacement of the Middle Triassic magmas***

629 The maximum P of crystallization of SS/SU magmas yielded by clinopyroxene-melt
630 barometers are comprised between 710 and 830 MPa (i.e. 24.5-28.6 km assuming a $\Delta P/\Delta z$ of
631 ~ 29 MPa/km; Fig. 10). During ascent, the crystallization processes continued in a polybaric
632 regime until 160-230 MPa, where the nearly isobaric cooling from T of 1060 °C down to T
633 < 1000 °C reflects the formation, in a shallow crustal environment, of discrete magma batches,
634 dominated by the onset of fractional crystallization processes (Fig. 10). Evidence of the
635 progressive alkali- and H₂O-enrichment during the shallow evolution of the melts are widely
636 documented in the Predazzo and Mt. Monzoni plutons, which record emplacement P of 40-165
637 MPa (Bonadiman et al., 1994; Casetta et al., 2018a). The hygrometric results show that
638 clinopyroxene crystallization in SS and SU melts took place at progressively increasing H₂O
639 amounts, from 1.7 wt% in trachybasalts at 710-830 MPa up to 3.9 wt% in trachyandesites at
640 380 MPa (Fig. 10; Table 5). Such H₂O concentrations are not far from what expected if a
641 completely incompatible behaviour of H₂O is assumed during progressive differentiation of
642 mantle-derived transitional to alkali basaltic melts with ~ 1.1 wt% H₂O. At shallower P , volatiles
643 exsolution promoted crystallization of clinopyroxene, feldspar and Fe-Ti oxides, followed by
644 amphibole and phlogopite in the case of magmas differentiating and cooling in
645 hypabyssal/plutonic conditions (Fig. 10).

646

647 ***9.3. Geodynamic implications: a mantle perspective***

648 Our study provides a robust link between the shoshonitic magmatism of the Dolomites and the
649 metasomatised Finero peridotite, enabling speculation about the evolution of the Southern Alps
650 SCLM. In fact, 7% partial melting of a Finero-like mantle containing ~800 ppm H₂O is able to
651 generate primary transitional basaltic SS melts resembling those emplaced during the most
652 productive phase of the Mid-Triassic (239-237 Ma) magmatism in the Dolomites. On the other
653 hand, 5% partial melting of a slightly different SCLM domain (H₂O ~600 ppm) produced the
654 SU primary alkali basaltic melts, which emplaced mostly in form of dykes and small intrusive
655 bodies, for which no age data have been reported so far. The progressive depletion of the SCLM
656 beneath the Southern Alps continued towards Late Triassic, when the asthenospheric influx
657 related to the Alpine Tethys opening phase led to the production of alkaline lamprophyres with
658 anorogenic signature at 219.22±0.73 Ma (Casetta et al., 2019). The extreme volatile-rich nature
659 of lamprophyres can be ascribed to small degree partial melts (1-0-2.5%) of their amphibole-
660 garnet-lherzolitic source (Casetta et al., 2019). Overall, the shifting from shoshonitic to alkaline
661 magmatism recorded in the Dolomites is a very effective way to track progressive depletion, in
662 terms of both major/trace elements and volatiles content, of the Southern Alps SCLM prior to
663 the rifting phase connected to the Alpine Tethys opening.

664 On the other hand, the retention of the modal/compositional enrichment in the lithosphere for a
665 time interval as long as 50 Ma, i.e. from the Variscan subduction to the Mid-Triassic
666 magmatism, can be matter of debate. In particular, it is complicated by the Permian (~294 to
667 274 Ma) onset of a large magmatic phase, during which mantle-derived anhydrous basaltic
668 melts with tholeiitic to subduction-related affinity were produced by the SCLM and/or the
669 asthenosphere (Sinigoi et al., 2016; Boscaini et al., 2020). These magmas underplated the thick
670 crustal piles and induced extensive anatexis, leading to the emplacement of mafic/ultramafic to
671 granitic/rhyolitic rocks throughout the Ivrea-Verbano Zone, particularly in Val Sesia (Voshage
672 et al., 1990; Sinigoi et al., 2016; Karakas et al., 2019) and the Athesina Volcanic District (Barth

673 et al., 1993; Marocchi et al., 2008). Furthermore, the gabbroic bodies of the Finero Mafic
674 Complex record the intrusion of multiple discrete tholeiitic/transitional magmatic batches
675 between Permian and Triassic, suggesting the existence of a multi-stage thermal evolution for
676 the Southern Alps lithosphere (Lu et al., 1997; Zanetti et al., 2013; Langone et al., 2017). In
677 this framework, the timing between the enrichment of Southern Alps SCLM and the Mid-
678 Triassic shoshonitic magmatism can belong to two different scenarios:

679 a) The existence of a Triassic arc/back-arc system at the NW Paleotethys limb (Zanetti et al.,
680 2013). In this view, the infiltration of slab-derived components into a depleted SCLM would
681 promote the formation of a Finero-like suprasubductive mantle domain during Mid-Triassic,
682 making the shoshonitic magmatism a simple expression of an active subduction-related setting.
683 This model implies that the enriched signature related to the Variscan subduction in the
684 Southern Alps SCLM was exhausted by the voluminous Permian magmatic event (see Storck
685 et al., 2020).

686 b) The preservation, in the Southern Alps lithospheric mantle, of an enrichment induced by the
687 Variscan subduction (Bonadiman et al., 1994; Zanetti et al., 1999; Lustrino et al., 2019). In this
688 view, the SCLM would have been capable of generating a transitional to alkaline magmatic
689 phase delayed with respect to the metasomatic events. The Sr-Nd isotopes support the
690 plausibility of such a scenario, showing how the hypothetical slab-derived components (i.e.
691 GLOSS/GLOSS-II), the Finero peridotite, the metasomatic amphiboles and the Mid-Triassic
692 magmatic rocks would lie on the same mixing curve (Fig. 7). This hypothesis is also in line
693 with the model of Pilet et al. (2015), according to whom continental rift-related Na- to K-
694 alkaline magmatism could be decoupled from the metasomatic enrichment of the underlying
695 mantle source. Analogous speculations were made by Panter et al. (2018) to model the Cenozoic
696 (48-30 Ma) alkaline magmatism of the West Antarctic Rift System by melting of amphibole-rich
697 metasomes generated in the SCLM during the 550-150 Ma Ross subduction. In an analogous
698 way, the shoshonitic magmatism of the Southern Alps would have been generated by a Finero-

699 like mantle source, and would represent the exhaustion of the Variscan subduction-related
700 signature in the Southern Alps SCLM that predated the Late Triassic-Early Jurassic opening of
701 the Alpine Tethys (Fig. 11).

702

703 **10. Conclusive remarks**

704 A comprehensive study of the acme of the Middle Triassic shoshonitic magmatism in the
705 Dolomites enabled us to constrain the nature and evolution of the Southern Alps SCLM between
706 the Variscan subduction and the opening of the Alpine Tethys. Insights from major/trace
707 elements and Sr-Nd isotopes were brought together to frame the shoshonitic magmatism into a
708 comprehensive picture at both local and regional geodynamic scale. The main points are
709 summarized here:

- 710 - The shoshonitic magmatism in the Southern Alps at 239-237 Ma is mostly represented by
711 a huge effusive phase, during which SiO₂-saturated (SS) to SiO₂-unsaturated (SU) melts
712 partially crystallized and differentiated *en route* to the surface. The *T-P* paths recorded by
713 olivine and clinopyroxene phenocrysts in the least differentiated rocks encompass a
714 considerable interval, spanning from 1200-1210 °C at 830 MPa to 1010-1060 °C at 160-
715 230 MPa (i.e. polybaric crystallization during ascent). At 160-230 MPa, almost isobaric
716 crystallization conditions marked the onset of differentiation in discrete magma batches at
717 shallow crustal levels, the remnants of which are still visible in the Predazzo-Mt. Monzoni
718 plutons (Bonadiman et al., 1994; Casetta et al., 2018a).
- 719 - The primary transitional to alkali basaltic SS and SU melts in equilibrium with a fertile
720 SCLM were generated by 5-7% partial melting of a metasomatised amphibole+phlogopite-
721 bearing spinel lherzolite similar to the Finero peridotite. Such a source was probably located
722 at depth of 40-60 km, consistent with theoretical (Selverstone and Sharp, 2011) and
723 experimental data (e.g., Conceição and Green, 2004).

- 724 - The metasomatic modification of the Mid-Triassic Southern Alps SCLM by slab-related
725 components (Fig. 7) derived from the Variscan subduction could have been preserved into
726 the Southern Alps SCLM, being able to induce the shoshonitic magmatic phase during Mid-
727 Triassic (Fig. 11).
- 728 - The modelled Finero-like mantle sources are capable of retaining up to 600-800 ppm H₂O,
729 resulting in the production of primary transitional to alkali basaltic SS/SU melts with ~1.1
730 wt% H₂O. The shoshonitic magmatism of the Dolomites could fix at 237-239 Ma the
731 exhaustion of the Variscan subduction-related signature in the Southern Alps SCLM. This
732 event predated the Late Triassic onset of a major rifting phase related to the opening of the
733 Alpine Tethys, the expression of which are recorded in small volume alkaline magmas (e.g.,
734 Casetta et al., 2019).

735

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983

984 **Figure captions**

985 **Fig. 1. (colour online)**

986 (a) Map of the tectonic units of the eastern portion of the Alps (modified from Schmid et al.,
987 2020). LO: Ligurian Ophiolites; AM: deformed Adriatic margin; AD: Adriatic Microplate; SA:
988 Southern Alps; DI: Dinarides; SM: Southern margin of Meliata; HB: Eoalpine High-Pressure
989 Belt; TW: Tauern tectonic Window; EW: Engadine tectonic Window; OTW: Ossola-Tessin
990 tectonic Window; EA: Eastern Austroalpine; H: Helvetic domain; M: Molasse foredeep. The
991 Mid-Triassic magmatic occurrences in the Southern Alps domain are evidenced in black. (b)
992 Basaltic trachyandesitic dyke intruded in the volcanic sequences. (c) Ultramafic xenoliths
993 embedded in a trachybasaltic dyke. (d) Simplified geological map of the volcano-plutonic
994 sequences and dyke swarm cropping out near Predazzo (modified from Casetta et al., 2018a).
995 SS = Shoshonitic Silica Saturated; SU = Shoshonitic Silica Undersaturated; GU = Granitic Unit.
996 Stars indicate dykes where ultramafic xenoliths were found.

997

998 **Fig. 2. (colour online)**

999 Photomicrographs of representative thin sections of the studied lavas, dykes and ultramafic
1000 cumulates. (a) Triple junctions between clinopyroxene grains in a clinopyroxenitic xenolith. (b)
1001 Contact between a clinopyroxenitic xenoliths and the host SU trachybasaltic dyke. (c) Highly

1002 porphyritic SS trachybasaltic dyke. (d, e) Clinopyroxene, olivine and plagioclase phenocrysts
1003 in a SU trachybasaltic dyke; (f) Highly porphyritic SS trachybasaltic lava. Cpx: clinopyroxene;
1004 Plg: plagioclase; Sp: spinel. (a), (b), (d) and (e) are in transmitted cross-polarized light; (c) and
1005 (f) are in transmitted plane-polarized light. Scale bar is 1 mm.

1006

1007 **Fig. 3. (colour online)**

1008 (a) TiO₂ (wt%) vs. Mg# and (b) Wollastonite (Wo)-Enstatite (En)-Ferrosilite (Fs) (Morimoto,
1009 1988) diagrams for clinopyroxene in lavas, dykes and clinopyroxenites. (c, d) Chondrite-
1010 normalized (Sun and McDonough, 1989) incompatible element and REE patterns of
1011 clinopyroxene in clinopyroxenites. The composition of clinopyroxene near the contact with the
1012 host SU dyke is also discriminated. The major and incompatible element distribution of
1013 clinopyroxene phenocrysts in the Late Triassic lamprophyric dykes from the same area (Casetta
1014 et al., 2019) is also reported for comparison.

1015

1016 **Fig. 4. (colour online)**

1017 Major and trace element compositions of SS lavas and SS/SU dykes. (a) Total Alkali vs. Silica
1018 (TAS) (Le Maitre et al., 2002), (b) K₂O (wt%) vs. SiO₂ (wt%) (Peccerillo and Taylor, 1975),
1019 (c) CaO/Al₂O₃ vs. SiO₂ (wt%), and (d) Ni (ppm) vs. MgO (wt%) diagrams. The composition of
1020 the SS, SU and GU bodies of the Predazzo Intrusive Complex (PIC) and of the Late Triassic
1021 lamprophyric dykes from the same area (Casetta et al., 2018a; 2018b; 2019) are also reported
1022 for comparison. Asterisks indicate the composition of the primary SS transitional basalt and SU
1023 alkali basalt resulted from the backward fractionation models (see text).

1024

1025 **Fig. 5. (colour online)**

1026 (a, b) Primitive Mantle- (PM) and Chondrite-normalized (Sun and McDonough, 1989)
1027 incompatible element and REE diagrams showing the composition of the SS lavas, SS/SU

1028 dykes and clinopyroxenites. (c, d) Primitive Mantle- (PM) and Chondrite-normalized
1029 incompatible element and REE patterns of the primary SS transitional basalt and SU alkali
1030 basalt resulted from the backward fractionation models. The patterns of the Late Triassic
1031 lamprophyres from the same area (Casetta et al., 2019) are also reported for comparison in (c)
1032 and (d).

1033

1034 **Fig. 6. (colour online)**

1035 Trace element discrimination diagrams for the studied SS lavas and SS/SU dykes in comparison
1036 with the Late Triassic lamprophyres from the same area (Casetta et al., 2019). (a) Th_N vs. Nb_N
1037 (Saccani, 2015), (b) Nb/La vs. La/Yb diagrams (Smith et al., 1999) and (c) Th-Hf-Ta ternary
1038 diagram (Wood, 1980).

1039

1040 **Fig. 7. (colour online)**

1041 $^{87}Sr/^{86}Sr$ vs. $^{143}Nd/^{144}Nd$ diagram for the studied SS lavas and SS/SU dykes (corrected to 238
1042 Ma) compared to the Permian Campore dyke (280 Ma; Sinigoi et al., 2016), Ulten peridotite
1043 (330 Ma; Tumiati et al., 2003), Finero phlogopite peridotite (FPP, 300 Ma; Voshage et al.,
1044 1987), metasomatic amphiboles in the FPP (300 Ma; Giovanardi et al., 2020, and references
1045 therein), and the Southern Alps basement (gneisses, paragneisses, stromalites, 300-330 Ma;
1046 Voshage et al., 1990; Tumiati et al., 2003; Sinigoi et al., 2016). Fields indicate the composition
1047 of Balmuccia-Baldissero peridotites (Voshage et al., 1987; Mukasa and Shervais, 1999),
1048 Southern Alps Mid-Triassic magmatic rocks (Lustrino et al., 2019) and the Predazzo Intrusive
1049 Complex (PIC) SS/SU rocks (Casetta et al., 2018a). DMM = depleted MORB mantle
1050 (Workman and Hart, 2005; corrected to 238 Ma); EM I = enriched mantle (Zindler and Hart,
1051 1986); GLOSS and GLOSS-II = average global subducting sediments (Plank, 2014); Average
1052 crust = average composition of the Southern Alps Permo-Triassic crust (Voshage et al., 1990;
1053 Tumiati et al., 2003; Sinigoi et al., 2016). Curves represent the mixing trends between DMM

1054 and GLOSS, GLOSS-II and the Permo-Triassic average crust. Numbers indicate the mixing
1055 percentages.

1056

1057 **Fig. 8. (colour online)**

1058 Chondrite-normalized (Sun and McDonough, 1989) incompatible element patterns of the
1059 inferred Mid-Triassic mantle sources that generated SS/SU magmas. SS- and SU-like sources
1060 are constituted by 96% DMM + 4% GLOSS-II and 98% DMM + 2% GLOSS components,
1061 respectively. Whole-rock patterns of FPP (Giovanardi et al., 2020), Ulten peridotites (Ionov et
1062 al., 2017), DMM (Workman and Hart, 2005), PM (Sun and McDonough, 1989) and Etna-like
1063 enriched mantle (EM; Casetta et al., 2020) are also reported.

1064

1065 **Fig. 9. (colour online)**

1066 Chondrite-normalized (Sun and McDonough, 1989) incompatible element diagrams showing
1067 the composition of the modelled (a) SS and (b) SU mantle sources and primary melts. In each
1068 panel, the composition of the mantle source obtained by mass balance calculations and that
1069 resulted from the mixing models are indicated. The composition of the primary (a) SS
1070 transitional basalt and (b) SU alkali basalt and of the melt generated at 7% and 5% partial
1071 melting from the corresponding sources are also indicated.

1072

1073 **Fig. 10. (colour online)**

1074 (a) $K_{\text{Fe-Mg}}^{\text{Cpx-melt}}$ vs. Mg# diagram for clinopyroxene crystals in equilibrium (or brought to the
1075 equilibrium) with the SS and SU melts. The equilibrium range indicated by the black lines ($K_{\text{Fe-Mg}}^{\text{Cpx-}}$
1076 $\text{melt} = 0.24-0.30$) is from Putirka et al. (2003). (b) Chondrite-normalized (Sun and
1077 McDonough, 1989) incompatible element pattern of the hypothetical melt in equilibrium with
1078 clinopyroxene grains in clinopyroxenite (see Supplementary Material for the partition
1079 coefficients) compared to that of the natural SS trachybasaltic dyke. (c) Temperature (T , °C)

1080 vs. pressure (P , MPa) diagram showing the clinopyroxene crystallization conditions in the
1081 SS/SU magmas, and the corresponding melt H₂O content of the melts. The maximum T - P
1082 crystallization interval of the Late Triassic lamprophyres (Casetta et al., 2019; De Min et al.,
1083 2020) is also reported in (c).

1084

1085 **Fig. 11. (colour online)**

1086 Evolutionary sketch of the Southern Alps SCLM between Carboniferous and Late Triassic,
1087 adapted from Pilet (2015) and Panter et al. (2018). (a) Carboniferous (>320 Ma) subduction of
1088 the Paleotethys beneath the Adria plate resulting in the slab-derived metasomatism of the
1089 Southern Alps SCLM (Stampfli et al., 2013). (b) Permian extensional stage activated by
1090 brittle/ductile intra-lithospheric shear zones inducing thinning, partial melting of the SCLM.
1091 This resulted in the production of tholeiitic to calc-alkaline magmas which intruded the Lower
1092 Crust, and induced the hybrid anatectic to mafic-derived acidic magmatism throughout the
1093 Ivrea-Verbano Zone and the Athesina Volcanic District. (c) Middle Triassic prolonged
1094 extensional phase inducing the exhaustion of the slab-derived enrichment retained in the
1095 SCLM, and resulting in the production of calc-alkaline to shoshonitic magmatism. (d) Late
1096 Triassic production of alkaline magmas throughout the Southern Alps caused by the rift-related
1097 asthenospheric upwelling predating the opening of the Alpine Tethys.

1 **The Variscan subduction inheritance in the Southern Alps Sub-Continental**
2 **Lithospheric Mantle: clues from the Middle Triassic shoshonitic magmatism of the**
3 **Dolomites (NE Italy)**

4
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27 **The Variscan subduction inheritance in the Southern Alps Sub-Continental**
28 **Lithospheric Mantle: clues from the Middle Triassic shoshonitic magmatism of the**
29 **Dolomites (NE Italy)**

31 **Abstract**

32 Although often speculated, the link between the Middle Triassic shoshonitic magmatism at the
33 NE margin of the Adria plate and the subduction-related metasomatism of the Southern Alps
34 Sub-Continental Lithospheric Mantle (SCLM) has never been constrained. In this paper, a
35 detailed geochemical and petrological characterization of the lavas, dykes and ultramafic
36 cumulates belonging to the shoshonitic magmatic event that shaped the Dolomites (Southern
37 Alps) was used to model the composition and evolution of the underlying SCLM in the time
38 comprised between the Variscan subduction and the opening of the Alpine Tethys.
39 Geochemical models and numerical simulations enabled us to define that 5-7% partial melting
40 of an amphibole + phlogopite-bearing spinel lherzolite, similar to the Finero phlogopite
41 peridotite, can account for the composition of the primitive Mid-Triassic SiO₂-saturated to -
42 undersaturated melts with shoshonitic affinity (⁸⁷Sr/⁸⁶Sr_i = 0.7032-0.7058; ¹⁴³Nd/¹⁴⁴Nd_i =
43 0.51219-0.51235; Mg #~ 70; ~1.1 wt% H₂O). By taking into account the H₂O content
44 documented in mineral phases from the Finero phlogopite peridotite, it is suggested that the
45 Mid-Triassic SCLM source was able to preserve a significant enrichment and volatile content
46 (600-800 ppm H₂O) for more than 50 Ma, i.e. since the slab-related metasomatism connected
47 to the Variscan subduction. The partial melting of a Finero-like SCLM represents the
48 exhaustion of the subduction-related signature in the Southern Alps lithosphere that predated
49 the Late Triassic-Early Jurassic asthenospheric upwelling related to the opening of the Alpine
50 Tethys.

51
52 **Keywords**

53 Southern Alps Sub-Continental Lithospheric Mantle; Subduction-related metasomatism;
54 Middle Triassic shoshonitic magmatism; Finero phlogopite peridotite

55

56 **1. Introduction**

57 The genesis of shoshonitic magmas is controversial, as it can be variably ascribed to syn- to
58 post-collisional tectonic settings or to subduction-related geodynamic contexts (e.g., Conticelli
59 et al., 2009; Peccerillo and Frezzotti, 2015). The typical Large-Ion Lithophile Elements (LILE),
60 Light Rare Earth Elements (LREE) enrichments and High Field Strength Elements (HFSE)
61 depletion of shoshonitic rocks, coupled with the recorded radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ and unradiogenic
62 $^{143}\text{Nd}/^{144}\text{Nd}$ ratios, always reflects enrichment of the underlying Sub-Continental Lithospheric
63 Mantle (SCLM) domains (Conceição and Green, 2004; Zheng, 2019). Shoshonitic melts are
64 thought to be generated by SCLM portions modified by the interaction with slab-derived fluids
65 or melts, or crustal recycled lithophile components (Plank, 2014; Zheng, 2019), and can be
66 powerful tools to track the composition and evolution of the lithosphere during major
67 geodynamic cycles.

68 In this regard, of paramount interest is the Middle Triassic (242-237 Ma) occurrence of
69 shoshonitic magmatism in the Dolomites (NE Italy; Fig. 1) and throughout the Southern Alps
70 (see Lustrino et al., 2019). The Mid-Triassic magmatism of the Southern Alps is particularly
71 interesting because of its timing, that is in between two major geodynamic episodes, i.e. the
72 Variscan subduction (430-340/300 Ma) and the Late Triassic/Early Jurassic opening of the
73 Alpine Tethys (Stampfli et al., 2013; Kroner and Romer, 2014). This last rifting phase was
74 marked by a strong asthenospheric upwelling that induced the production of highly alkaline
75 anorogenic magmas at 225-215 Ma throughout the Southern Alps, i.e. 20 Ma after the
76 termination of the shoshonitic magmatic cycle (Stahle et al., 2001; Casetta et al., 2019).

77 Notwithstanding its concomitance with dominant strike-slip to extensional tectonics (e.g.,
78 Doglioni, 2007), the Mid-Triassic magmatic event that shaped the Dolomites was historically

79 defined as “shoshonitic” due to the potassic affinity/orogenic-type of the magmatic products
80 (Sloman, 1989; Casetta et al., 2018a; 2018b; Lustrino et al., 2019). Despite its geodynamic
81 implications, this term was adopted throughout the text to indicate the magmatic affinity, in line
82 with previous studies, although multiple and articulated geodynamic/tectonic models for the
83 Southern Alps magmatism have been proposed so far. They include: i) aborted continental
84 rifting (Bernoulli and Lemoine, 1980); ii) sinistral strike-slip tectonics (Doglioni, 2007); iii)
85 arc/back-arc systems located at the Paleotethys NW limb (Zanetti et al., 2013); iv) subduction
86 of a Palaeozoic oceanic basin located between Austroalpine and Southern Alps (Bianchini et
87 al., 2018); v) reactivation, during an anorogenic rifting phases, of old SCLM domains
88 metasomatized during the Variscan subduction (Sloman, 1989; Bonadiman et al., 1994; De Min
89 et al., 2020).

90 Although the impact of the Variscan subduction on the Southern Alps/Austroalpine SCLM
91 portions is widely documented, as in case of the Finero and Ulten peridotitic massifs (Zanetti
92 et al., 1999; Tumiatì et al., 2003; Ionov et al., 2017; Giovanardi et al., 2020), a direct link
93 between the metasomatic event and the subsequent genesis of magmas with orogenic-like
94 affinity is still missing, feeding relevant questions:

- 95 - What was the composition of the Southern Alps SCLM during Middle Triassic?
- 96 - How did this mantle source evolve from the Variscan subduction to the Alpine Tethys
97 opening?
- 98 - In which way is Mid-Triassic shoshonitic magmatism related to the geochemical features
99 of the Southern Alps SCLM?

100 To contribute in answering these questions, a geochemical and petrological investigation of
101 mafic to intermediate lava flows, dykes and entrained ultramafic cumulates cropping out in the
102 Dolomites was put forward. These products, located close to the coeval Predazzo Intrusive
103 Complex (Fig. 1; Casetta et al., 2018a; 2018b), belong to the acme of the Mid-Triassic
104 magmatism (from 239.04 ± 0.04 to 237.58 ± 0.04 ; Storck et al., 2020), which was typified by the

105 emission of effusive products recording a mantle-derived signature, with limited crustal
106 contamination (Storck et al., 2020). These volcanic sequences are ideal objects for investigating
107 the evolution of the Southern Alps SCLM, and modelling its possible links to the main
108 geodynamic processes recorded in the neighbouring orogenic peridotites (Zanetti et al., 1999;
109 Tumiati et al., 2003). Mineral phase, whole-rock major and trace element analyses and Sr-Nd
110 isotopes were used to constrain the *T-P* path of shoshonitic magmas during ascent and
111 differentiation, as well as to model the nature of the Southern Alps SCLM during Mid-Triassic.
112 Numerical models were performed to: i) reconstruct the composition of the primitive magmas
113 in equilibrium with potential mantle sources; ii) trace the evolution of the SCLM beneath Adria
114 during Mid-Triassic, in the large context of what recorded in the Finero, Balmuccia, Baldissero
115 and Ulten peridotitic massifs; iii) determine the geometry of the feeding system and the main
116 processes that led to ascent/differentiation of the shoshonitic magmas in the crust.

117

118 **2. Geological setting**

119 The Mid-Triassic magmatic event that took place throughout the Southern Alps between ~243
120 and 227 ± 6 Ma was characterized by the emission of significant amounts of igneous rocks with
121 calc-alkaline to shoshonitic affinity (Lustrino et al., 2019; Storck et al., 2020 and references
122 therein). In the central and eastern Southern Alps, an earlier volcanoclastic phase, with
123 discontinuous emission of acidic products was followed by a dominantly effusive stage,
124 characterized by the eruption of mafic lavas and emplacement of shallow intrusive bodies
125 (Bonadiman et al., 1994; Casetta et al., 2018a; Storck et al., 2020). The oldest magmatic
126 manifestations, i.e. tuff layers embedded in the sedimentary Buchenstein Fm. in the Brescian
127 Alps, record zircon U-Pb ages of 242.65 ± 0.04 Ma, whereas the youngest episodes, represented
128 by tuff layers inside the Wengen Fm. in the Dolomites, yield an age of 237.58 ± 0.04 Ma
129 (Wotzlaw et al., 2018; Storck et al., 2020).

130 The most productive phase of the Mid-Triassic magmatic event is best preserved in the thick
131 sequences of mafic to mildly evolved volcanics of the Dolomites (Fig. 1), emplaced after an
132 earlier (242.01 ± 0.05 Ma) phreatomagmatic episode recorded by lapilli tuffs and tuff deposits.
133 The effusive phase lasted <0.9 Ma (Storck et al., 2020), and was characterized by the production
134 of submarine pillow lavas, pillow breccias, lava breccias, hyaloclastites, and subaerial lava
135 flows, rarely intercalated with volcanoclastic sandstones. The production of volcanites was
136 associated to the emplacement of articulated dyke swarms and plutonic bodies, which are in
137 magmatic contact with the overlying lava flows and/or cut by dykes with the same composition
138 of the volcanites (Casetta et al., 2018b).

139

140 **3. Materials and methods**

141 The volcanic products (lava flows and lava breccias) and dykes crop out in the Dolomites, and
142 are spatially associated to the Predazzo Intrusive Complex (Fig. 1). Due to the widespread
143 hydrothermal alteration affecting the magmatic products in this area, an initial collection of
144 more than 200 samples was reduced to ~50 specimens for the purposes of this study, after
145 detailed screening based on thin section observations. The least altered samples were selected
146 for detailed whole-rock and in-situ chemical determinations. Whole-rock major and trace
147 element chemistry was determined by means of an ARL Advant-XP automated X-Ray
148 Fluorescence (XRF) spectrometer hosted at the Department of Physics and Earth Sciences of
149 the University of Ferrara. Full matrix correction procedure and intensities were completed
150 following Traill and Lachance (1966). Accuracy and precision were better than 2-5% for major
151 elements and 5-10% for trace elements. Detection limits were 0.01 wt% and 1-3 ppm for most
152 of the major and trace element concentrations, respectively. Rb, Sr, Y, Zr, Nb, Hf, Ta, Th, U,
153 and REE analyses were carried out at the Department of Physics and Earth Sciences of the
154 University of Ferrara by means of a Thermo Series X inductively coupled plasma-mass
155 spectrometer (ICP-MS). Precision and accuracy were better than 10% for all elements, well

156 above the detection limit. Mineral phase major element chemistry was determined by using a
157 CAMECA SX100 electron microprobe equipped with four WD and one ED spectrometers
158 hosted at the Department of Lithospheric Research of the University of Wien. The operating
159 conditions were as follows: 15 kV accelerating voltage, 20 nA beam current, and 20 s counting
160 time on peak position. For feldspars, a 5 μm defocused beam and 10 s counting time on peak
161 position for Na and K were used. Natural and synthetic standards were used for calibration, and
162 PAP corrections were applied to the intensity data (Pouchou and Pichoir, 1991). Trace element
163 concentrations of clinopyroxene were measured at the CNR - Istituto di Georisorse of Pavia by
164 laser ablation microprobe-inductively coupled plasma-mass spectrometry (LAM-ICP-MS)
165 following the protocols described by Tiepolo et al. (2003). NIST 610 and NIST 612 standard
166 glasses were used to calibrate relative element sensitivity. Precision and accuracy for trace
167 element analyses were assessed by standard sample BCR-2 (reference values from USGS
168 Geochemical Reference Materials Database). Each analysis was corrected with internal
169 standards using CaO. The detection limit, function of the ablation volume and counting time
170 were calculated for each analysis. Since ablation volume greatly depends on laser type and
171 sample (matrix), the detection limit decreased with decreasing spot size, beam power and cell
172 gas flow. A 40-100 μm beam diameter and 20 $\mu\text{m/s}$ scanning rate were used. The theoretical
173 detection limit ranges from 10 to 20 ppb for REE, Ba, Th, U, Zr and are about 2 ppm for Ti.
174 Whole-rock $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ analyses were made at the Scottish Universities
175 Environmental Research Centre (SUERC) by thermal ionization mass spectrometry (TIMS)
176 following procedures described by Casetta et al. (2018a). Eight measurements of SRM-987 and
177 twelve of JNdi-1 made during the course of this analytical programme yielded mean values of
178 0.710244 ± 0.000016 for $^{87}\text{Sr}/^{86}\text{Sr}$ and 0.512079 ± 0.000018 (2 SD) for $^{143}\text{Nd}/^{144}\text{Nd}$, consistent
179 with the consensus values of ~ 0.71025 and ~ 0.51210 .

180

181 **4. Petrography and mineral chemistry**

182 Lava flows with trachybasaltic to trachyandesitic composition crop out in thick sequences
183 emplaced over the Permo-Triassic volcano-sedimentary basement, and/or in direct contact with
184 the coeval plutonic counterparts. Multiple dykes, ranging in thickness from 10-20 cm up to 1-
185 2 m, cut both the basement and the volcano-plutonic bodies (Fig. 1). The crosscutting
186 relationships between the dykes and the plutonic bodies have been used to reconstruct the
187 temporal evolution of the multi-pulse Predazzo Intrusive Complex, which is composed of SiO₂-
188 saturated (SS) to SiO₂-undersaturated (SU) batches (Casetta et al., 2018b). Geochemical and
189 petrographic data (see below) support the discrimination between the two SS and SU suites for
190 the dykes studied in the present work, which mirror the geochemical features of the plutonic
191 rocks. They include hypersthene- to nepheline-normative trachybasalts, differentiating towards
192 quartz- and nepheline-normative trachytes, respectively. All lavas, on the other hand, are SiO₂-
193 saturated, highlighting how the occurrence of SU products was volumetrically limited.

194

195 ***4.1. Lava flows***

196 The texture of the SS trachybasaltic to trachyandesitic lavas is generally porphyritic to
197 glomeroporphyritic, with Porphyricity Index (P.I.) ranging from 5-10 up to 55-60%; these
198 values suggest near hypabyssal conditions of the last stages of their crystallization (Fig. 2). The
199 main phenocrysts are clinopyroxene (up to 3 mm), plagioclase (200 μm-2 mm) and spinel (50-
200 500 μm), together with some rare olivine (up to 1 mm), often altered in iddingsite or serpentine,
201 in the least differentiated samples (see Supplementary Material). Clinopyroxene ranges from
202 diopside to augite (Morimoto, 1988), with Mg# [calculated as Mg/(Mg+Fe_{TOT}) mol%] and TiO₂
203 contents of 70.0-74.1 and 0.5-1.3 wt%, respectively (Fig. 3). Its Al₂O₃ content varies between
204 2.5 and 4.7 wt%, whereas Cr₂O₃ is always < 0.1 wt%. Occasionally, glomerophyric aggregates
205 of clinopyroxene + spinel are observed. Spinel ranges in composition from magnetite to Ti-
206 magnetite (TiO₂ = 0.1-17.4 wt%). Plagioclase, ranging in composition from bytownitic to
207 labradoritic (An₅₈₋₈₇), is often sericitized in the less differentiated rocks, while it is quite fresh

208 in the trachyandesites. The groundmass of all lavas ranges from microcrystalline to
209 hypohyaline, constituted by the same mineral assemblage as the phenocrysts. Groundmass
210 plagioclase in the trachyandesitic lavas has andesinic composition (An_{37-44}).

211

212 **4.2. Dykes**

213 The P.I. of the trachybasaltic to trachyandesitic dykes varies from 5-10% up to 50-55%. SS
214 trachybasaltic to trachyandesites have paragenesis and texture comparable to those of the lavas,
215 except for the groundmass, which in some cases has coarser grain size and hosts some K-
216 feldspar crystals (Or_{79-90}). Plagioclase ranges in composition from bytownitic to labradoritic,
217 spinel is magnetite to Ti-magnetite, and clinopyroxene is similar in composition to the
218 phenocrysts in the SS lavas (Fig. 3) (see Supplementary Material).

219 Some trachybasaltic dykes, cropping out in proximity of the Latemar area (Fig. 1), have the
220 most primitive composition, and likely belong to the SU suite. They are constituted by
221 microcrystalline to hypohyaline groundmass, and clinopyroxene (400 μm -2 mm), plagioclase
222 (100-400 μm) and spinel (<200 μm) phenocrysts, together with fresh, zoned olivine phenocrysts
223 and xenocrysts (500 μm -2 mm; Fo_{85-92} cores to Fo_{78-81} rims, at NiO 0.07-0.15 wt%; Fig. 2). In
224 these dykes, clinopyroxene phenocrysts are Al-Cr-diopsides, with Mg# between 78.1 and 85.0,
225 TiO_2 contents up to 2.0 wt% and Cr_2O_3 concentrations ranging from 0.1 to 1.0 wt% (Fig. 3).
226 Some phenocrysts exhibit compositional zoning, with Cr-diopsidic cores (Mg# up to 88.4; TiO_2
227 = 0.2 wt%; Cr_2O_3 = 0.5 wt%) graded in diopsidic-augitic rims (Mg# down to 67.6; TiO_2 up to
228 1.0 wt%; Cr_2O_3 ~0.02 wt%). Throughout SU dykes, plagioclase phenocrysts range in
229 composition from bytownite to labradorite, whereas spinel encompass the magnetite to Ti-
230 magnetite compositional range. Spinel composition is different, however, in the least
231 differentiated SU trachybasaltic dyke, and ranges from Ti-magnetite to Cr-spinel (TiO_2 = 0.8-
232 16.9 wt%; Cr_2O_3 = 4.0-28.3 wt%). Groundmass K-feldspar ranges in composition from Or_{79} to
233 Or_{90} .

234 The trachytic dykes, distinguishable in the field by their bright pink to grey (SS quartz-
235 trachytes) or light pink to greenish (SU nepheline-trachytes) colours, have low P.I. (usually
236 <25%). Quartz-trachytes are mainly constituted by K-feldspar (300 μm -4 mm), plagioclase (up
237 to 7 mm) and spinel (<200 μm) phenocrysts in a micro- to cryptocrystalline groundmass made
238 by apatite, feldspars, biotite, quartz and Fe-Ti oxides. Nepheline-trachytes are usually highly
239 altered, and composed of K-feldspar (1-5 mm), plagioclase (up to 1 mm), and rare nepheline
240 phenocrysts in a cryptocrystalline to hypohyaline groundmass. Plagioclase phenocrysts and
241 groundmass crystals in SS-SU trachytes have large compositional variations, ranging from
242 labradorite to albite. K-feldspar is quite homogeneous, ranging from Or₉₂ to Or₉₉.
243 Overall, the composition of clinopyroxene phenocrysts is the best proxy for discriminating
244 between the SS and SU suites: clinopyroxene in SS rocks is generally more depleted in TiO₂
245 with respect to those in SU samples, at comparable Mg# (Fig. 3). Such a feature further
246 evidences the more pronounced “alkaline” character of SU samples with respect to the SS ones,
247 fostering a discrimination between the two suites.

248

249 ***4.3. Ultramafic cumulates***

250 In the Latemar area, some SU trachybasaltic dykes contain ultramafic xenoliths (Figg. 1-2). The
251 least altered clinopyroxenites have cumulitic texture, with small (200-300 μm) to medium-sized
252 (2-3 mm) fresh clinopyroxene crystals in contact by means of triple junctions, and sometimes
253 enclosing small olivine grains (Fig 2). Clinopyroxene is diopside (Mg# = 85.5-87.6; Wo = 47-
254 48), with TiO₂ and Cr₂O₃ contents of ~0.2 wt% and 0.1-0.3 wt%, respectively (Supplementary
255 Material). At the edges of the xenolith, the contact with the host SU trachybasalt induced
256 marked TiO₂ (up to 2.3 wt%) and Cr₂O₃ (up to 1.2 wt%) enrichments, coupled with a general
257 decrease in the Mg# (from 84.0 down to 62.5). Beside these contact areas, clinopyroxene show
258 compositional analogies with the phenocrysts of the SS suites (Fig. 3). In chondrite-normalized
259 (Sun and McDonough, 1989) incompatible element patterns, these clinopyroxene have Rb-Ba

260 and Nb-Ta depletions, as well as marked Zr-Hf and Ti negative anomalies. Their REE patterns
261 have the convex-upward shape typical of alkaline magmas, with $(La/Yb)_N$ values of 1.91-2.00
262 (Fig. 3). Although preserving the significant HFSE paucity, clinopyroxene grains in proximity
263 of the contact with the host basalt are significantly REE-enriched, and have $(La/Yb)_N$ values of
264 ~1.6 (Fig. 3).

265

266 **5. Whole-rock geochemistry**

267 ***5.1. Major and trace element chemistry***

268 Lavas and dykes belong to both the alkaline and transitional series with K-affinity (Fig. 4). On
269 the Total Alkali vs. Silica (TAS) diagram (Fig. 4), lavas plot between the basaltic and the
270 trachyandesitic fields ($SiO_2 = 49.2-59.9$ wt%; $Na_2O+K_2O = 3.1-8.5$ wt%), whereas dykes show
271 wider compositional variations, from basaltic to trachytic and phonolitic ($SiO_2 = 48.1-63.7$
272 wt%; $Na_2O+K_2O = 3.4-12.2$ wt%) (Fig. 4; Table 1). The Mg# [calculated as $MgO/(MgO+FeO)$
273 mol%, considering a Fe_2O_3/FeO ratio of 0.15, in agreement with fO_2 around +1 ΔFMQ (Kress
274 and Carmichael, 1991; Casetta et al., 2018a)] of the lavas ranges from 64.3 in trachybasaltic
275 samples down to 41.5 in trachyandesites, whereas dykes have Mg# of 65.4 (trachybasalts) to
276 23.7 (trachytes). The TiO_2 and FeO_{TOT} contents, as well as the CaO/Al_2O_3 ratio of the samples
277 are inversely correlated with the silica content, and decrease from 1.5 to 0.2 wt%, from 10.5 to
278 2.0 wt% and from 0.78 to 0.03, respectively (Fig. 4). Similarly, Ni decreases from 89 to 1 ppm,
279 while Zr, Nb and Rb concentrations increase from 77 to 779 ppm, from 5 to 51 ppm and from
280 12 to 344 ppm, respectively, with increasing SiO_2 (Fig. 4). As for the corresponding intrusive
281 suites (Casetta et al., 2018a; 2018b), SU and SS rocks can be easily discriminated by their FeO
282 contents, CaO/Al_2O_3 and Na_2O/K_2O ratios, as well as by their Rb, Zr and Nb concentrations at
283 comparable silica content (Fig. 4).

284 In the Primitive Mantle- (PM; Sun and McDonough, 1989) normalized incompatible element
285 diagram, all lavas and dykes are enriched in LILE, Th-U and LREE, and display the Nb-Ta and

286 Ti negative anomalies typical of magmatic suites generated by subduction-modified mantle
287 source (Fig. 5). Moving from trachybasaltic to trachytes, the incompatible element patterns are
288 increasingly enriched in LILE and depleted in Ti. All samples have flat Middle (M-) to Heavy
289 (H-) REE profiles in chondrite-normalized (Sun and McDonough, 1989) diagrams, and
290 moderate enrichment in LREE, with $(La/Yb)_N$ values comprised between 8.5 and 9.5 for the
291 least differentiated rocks of both suites (Fig. 5). Despite the similar incompatible element
292 concentrations with respect to the most primitive SS lavas and dykes, SU trachybasalts have
293 slightly higher Zr-Hf abundances, and less pronounced Nb-Ta-Ti anomalies. Significant
294 differences can be envisaged by comparing the most differentiated trachytic products of the two
295 suites: SU trachytes have extremely higher Th-U, Zr and LREE concentrations with respect to
296 SS ones, this latter parameter resulting in a marked difference in their $(La/Yb)_N$ ratios (Fig. 5;
297 Table 1). On the Th_N vs. Nb_N tectonic discrimination diagram (Fig. 6a), the SS/SU volcanic
298 series fall within the calc-alkaline, (i.e. orogenic) field. Their HFSE distribution is well
299 consistent with magmatic suites from active continental margins, as also testified by the
300 distribution of the least differentiated SS-SU rocks on the Th-Hf-Ta diagram (Fig. 6c). In both
301 diagrams, the marked difference between Mid-Triassic SS/SU rocks and Late Triassic
302 lamprophyres from the same area (Casetta et al., 2019) supports the hypothesis that different
303 mantle sources were activated within a period of ~20 Ma to generate the distinct magmatic
304 events. This feature is also highlighted by the Nb/La vs. La/Yb ratios of the least differentiated
305 SS/SU rocks (Fig. 6b), which suggest that the contribution of lithospheric mantle in their
306 genesis was more pronounced than for lamprophyres, which bear witness of a marked
307 asthenospheric involvement.

308

309 **5.2. $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$**

310 Whole-rock $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ isotopic ratios were measured on representative samples
311 of lavas and dykes from both the SS and SU suites (Table 1). Initial isotopic ratios, respectively

312 named $^{87}\text{Sr}/^{86}\text{Sr}_i$ and $^{143}\text{Nd}/^{144}\text{Nd}_i$, were corrected to an age of 238 Ma, in accordance with the
313 recently published U-Pb datings (from 238.19 ± 0.05 to 237.58 ± 0.04 Ma; Storck et al., 2020).
314 Lavas and dykes from the SS suite have $^{87}\text{Sr}/^{86}\text{Sr}_i$ values ranging from 0.7043 to 0.7057, and
315 $^{143}\text{Nd}/^{144}\text{Nd}_i$ from 0.51219 to 0.51235 (Fig. 7). Dykes belonging to the SU suite are
316 characterized by slightly higher $^{87}\text{Sr}/^{86}\text{Sr}_i$ (0.7055-0.7058) with respect to SS rocks, at
317 $^{143}\text{Nd}/^{144}\text{Nd}_i$ of 0.51226-0.51232. On the whole, Sr-Nd isotopic data for the SS and SU lavas
318 and dykes are similar to those of the corresponding intrusive rocks, lying in proximity of the
319 enriched mantle (EM I) end-member, in line with most of the Southern Alps Mid-Triassic
320 magmatic rocks (Fig. 7; Lustrino et al., 2019). They are also well distinct from the Late Triassic
321 lamprophyres, which plot close to the depleted MORB mantle (DMM) end-member (Fig. 7;
322 Workman and Hart, 2005). The only available analysis for cumulitic clinopyroxenite xenoliths
323 yields an $^{87}\text{Sr}/^{86}\text{Sr}_i$ value of 0.7044, i.e. similar to those of the SS rocks, but slightly lower with
324 respect to the initial Sr ratios typical of SU dykes.

325

326 **6. The primary shoshonitic magmas and their SCLM source**

327 ***6.1. Reconstruction of primary shoshonitic melts***

328 The variable extent of differentiation that affected the Mid-Triassic shoshonitic rocks makes it
329 difficult to understand the compositional features of the primary melts, and from there to
330 understand the nature of the underlying SCLM. The least evolved lavas and dykes have Mg#
331 always <65, and are often characterized by moderate to high crystal cargoes (P.I. up to 50-55),
332 which result in “doped” bulk rock analyses, where the high Mg# and the low REE abundances
333 are function of the hypabyssal nature of the rocks rather than of their primitivity. No melt
334 inclusions were found in the early-forming phases (olivine or clinopyroxene), being the
335 composition of the most Mg-rich crystals inside lavas, dykes and cumulitic xenoliths (Fig. 3)
336 the only petrological evidence for the existence of primitive melts. For these reasons, the most
337 Mg-rich and least porphyritic dykes from the two magmatic suites (SS-SU) were chosen and

338 traced back to the ideal condition of equilibrium with a fertile mantle assemblage, in line with
339 what proposed by Casetta et al. (2020) for Mt. Etna magmas. Olivine and clinopyroxene plus
340 small amounts of Cr-bearing spinel were re-added to the melt in a backward fractionation
341 process, being aware of respecting the Fe-Mg equilibrium conditions between the progressively
342 reconstructed melt and each added mineral phases (i.e. $^{Ol-Melt}Kd_{Fe-Mg} = 0.27-0.33$ and $^{Cpx-}$
343 $^{Melt}Kd_{Fe-Mg} = 0.24-0.30$; Roeder and Emslie, 1970; Putirka, 2008). The occurrence of high-Fo
344 olivine xenocrysts and olivine inclusions in cumulitic clinopyroxene suggests that olivine
345 dominated the earlier crystallization stages, probably in a restricted *T-P* interval, followed in a
346 rapid sequence by clinopyroxene in the possible liquid line of descent of these magmas.
347 Therefore, olivine + clinopyroxene (diopside-augite) assemblages, together with small amounts
348 of Cr-bearing spinel, were progressively re-added to the chosen starting SS and SU rocks until
349 their major element composition resulted in equilibrium with a fertile mantle source (i.e. Mg#
350 68-70, in equilibrium with a Fo₈₈₋₉₀ olivine; see McDonough, 1990). The trace element
351 composition of the ideal primitive SS and SU melts was modelled by means of the Rayleigh
352 fractionation equation (Shaw, 1979), following Casetta et al. (2020). The adopted mineral/melt
353 trace element partition coefficients are listed in the Supplementary Materials. The primary
354 transitional to alkali basaltic SS and SU melts were obtained after 10.5-11.4% backward
355 fractionation of olivine (6.6-6.9 vol.%), clinopyroxene (2.4-3.0 vol%) and spinel (1.5 vol.%)
356 (Table 2). The modelled primitive melts have Mg# ~70 and preserve the above-mentioned
357 differences between the SS and SU suites, in terms of both major and trace element features
358 (Fig. 4). Their PM-normalized incompatible element patterns are characterized by LILE-
359 enrichment as well as Nb-Ta-Ti negative anomalies, these latter less marked in the primitive
360 SU melt, which preserve the slightly higher Zr-Hf abundance with respect to the SS one (Fig.
361 5). When compared to the Late Triassic lamprophyres, SS and SU rocks can be easily
362 discriminated on the basis of their HFSE abundances (Fig. 5).

363

364 **6.2. The composition of the mantle source**

365 The shoshonitic affinity and incompatible element distribution of the studied magmas indicate
366 their derivation from an enriched SCLM source modified by subduction-related components.
367 In fact, Sr-Nd, Pb and Hf isotopic signatures point towards an enriched mantle-derived
368 fingerprint, and rule out the effect of large (>10%) crustal assimilation during the peak of
369 effusive magmatism at 238 Ma (Casetta et al., 2018a; De Min et al., 2020; Storck et al., 2020).
370 Several authors hypothesized a decoupling between the timing of the shoshonitic magmatism
371 and the metasomatism of the lithospheric mantle source, relating this latter process to the
372 Variscan subduction (Bonadiman et al., 1994; Lustrino et al., 2019). Direct evidence of the
373 impact of the Variscan subduction on the Southern Alps SCLM is visible in the Ulten and Finero
374 peridotitic massifs, where modal (i.e. presence of amphibole, phlogopite, Cl-rich apatite) and/or
375 cryptic subduction-related metasomatism is documented (e.g., Zanetti et al., 1999; 2016;
376 Tumiatì et al., 2003; Ionov et al., 2017; Tommasi et al., 2017).
377 To simulate the slab-derived modification of the Southern Alps SCLM, Sr-Nd isotopic mixing
378 models (Faure, 1986; see also Melchiorre et al., 2020) between DMM and selected subduction-
379 related components were performed. The average composition of the global subducting
380 sediments (GLOSS/GLOSS-II; Plank, 2014) and the average composition of the Southern Alps
381 crust (Voshage et al., 1990) were chosen to represent possible end-members of subducting
382 crustal components, for modelling the Sr-Nd isotopic variability caused by slab-derived
383 influxes (see Kessel et al., 2005; Plank, 2014). The SS rocks lie on a mixing trend between
384 DMM and GLOSS-II, whereas the relative Sr-Nd isotopic enrichment of SU magmas
385 approaches the mixing curve between DMM and GLOSS, confirming that two different mantle
386 sources are associated to the genesis of SS and SU magmas. By contrast, the Finero
387 amphibole+phlogopite-bearing peridotite samples plot towards the GLOSS-II end-member,
388 resulting slightly more enriched in radiogenic Sr with respect to SS magmas (Fig. 7). The Ulten
389 peridotites span a large Sr-Nd isotopic range, as result of complex interactions between mantle

390 and metasomatising fluids with crustal signature (Fig. 7; see Tumiati et al., 2003). Metasomatic
391 amphibole and phlogopite in the Finero phlogopite peridotite, recording post-Variscan ages as
392 old as 310 Ma (Zanetti et al., 2016; Malitch et al., 2017, and references therein), mostly plot
393 along the mixing trend between DMM and the average crust (Fig. 7), enhancing the role played
394 by subduction-related melts/fluids in metasomatising the Southern Alps SCLM prior to 290 Ma
395 (Hartmann and Wedepohl, 1993; Zanetti et al., 1999; 2016; Selverstone and Sharp, 2011;
396 Giovanardi et al., 2020). Following the Sr-Nd isotopic mixing models, the ideal trace element
397 composition of the enriched SCLM domains that generated the SS/SU primary melts has been
398 firstly approached by mixing DMM with 4% GLOSS-II (SS magmas) and 2% GLOSS (SU
399 magmas) components. The inferred mantle source of SS magmas has LILE-enrichment, Nb-
400 Ta-Ti negative troughs, slight LREE-enrichment and PM-like MREE-HREE contents (Fig. 8).
401 The calculated mantle source of SU magmas is relatively depleted in all trace elements with
402 respect to the SS one, being typified by PM-like LREE concentrations and DMM-like MREE-
403 HREE profiles (Fig. 8).

404 Subsequently, the mineral phases of highly fertile composition from the Southern Alps
405 peridotitic massifs (Mazzucchelli et al., 2010; Giovanardi et al., 2020) were used to model the
406 trace element content and modal composition of the ideal “SS” and “SU” sources *via* mass
407 balance calculations. Results showed that the trace element concentrations of the ideal “SS”
408 and “SU” sources are matched by considering two amphibole+phlogopite-bearing spinel-
409 lherzolites with the following modal compositions: 55.4% olivine + 23.0% orthopyroxene +
410 17.7% clinopyroxene + 1.1% spinel + 2.0% amphibole + 0.8% phlogopite (SS source); 57.5%
411 olivine + 24.3% orthopyroxene + 15.1% clinopyroxene + 1.1% spinel + 1.4% amphibole +
412 0.6% phlogopite (SU source) (Table 3). It should be highlighted that all the considered phases
413 belong to the Finero phlogopite peridotite, except clinopyroxene, which is from the Baldissero
414 peridotite. Clinopyroxene in the Finero phlogopite peridotite in fact is highly depleted,
415 recording up to ~20% melt extraction (Coltorti and Siena, 1984; Hartmann and Wedepohl,

416 1993), being thus unsuitable for simulating an enriched mantle source. By considering the
417 maximum H₂O contents measured in the mineral phases of the Finero peridotite (3.2 ppm
418 olivine; 111 ppm orthopyroxene; 267 ppm clinopyroxene; 1.93 wt% amphibole; 4.3 wt%
419 phlogopite; Tommasi et al., 2017), the amounts of H₂O retained in the modelled SS and SU
420 mantle sources would result of 784 and 598 ppm, respectively (Table 3). These values are well
421 within the range proposed for other worldwide SCLM domains, such as the Subei Basin in
422 China (328-1440 ppm H₂O; Hao et al., 2019) or the Norther Victoria Land in Antarctica (910-
423 1486 ppm H₂O; Giacomoni et al., 2020). In line with the trace element and Sr-Nd isotopes
424 mixing results, the compositional discrimination between the modelled SS and SU sources is
425 mirrored by different incompatible elements and H₂O concentrations, in turn related to the
426 lower modal amounts of clinopyroxene, amphibole and phlogopite (Fig. 8).

427

428 **7. The genesis of shoshonitic magmatism**

429 ***7.1. Mantle source melting model***

430 The partial melting conditions of the mantle sources required to produce the primary SS and
431 SU transitional to alkali basaltic melts were simulated by using the mineral phase modal
432 proportions and the eutectic composition of the peridotitic systems as function of the degree of
433 partial melting (see Casetta et al., 2020). A “melting+recrystallization” mass balance model
434 enabled us to simulate the contribution of the fusible components (e.g., TiO₂, Al₂O₃ and alkalis)
435 from each mineral phase of the lherzolitic source, and then reproduce the major element
436 composition of the produced magmas. In this process, the contribution during melting of each
437 phase with fertile composition is counterbalanced by recrystallization of a restitic residuum
438 (e.g., melting of Al-rich clinopyroxene + recrystallization of Al-poor clinopyroxene), in
439 accordance with the conditions of melting of a amphibole+phlogopite-bearing lherzolite in the
440 spinel stability field (Gaetani and Grove, 1998; Conceição and Green, 2004; Liu et al., 2006;

441 Condamine and Médard, 2014). The relative contribution of each mineral phase at the eutectic
442 was evaluated, for major elements (oxide, wt%) by least square method, following the equation:

443
$$S = \sum_{i=1}^n r_i^2 .$$

444 Here, the optimal parameter S is fitted by minimizing the sum of squared residuals r^2 between
445 the predicted values for each oxide “i” (e.g., SiO₂ wt% of the primitive melt) and the values
446 calculated from the melting model (e.g., SiO₂ wt% resulting from the
447 “melting+recrystallization” process). For our purposes, the model was considered accurate
448 when $S < 1$, i. e. when the difference between modelled and calculated parameters, expressed as
449 oxides wt%, is well within the analytical errors (i.e. XRF and/or EPMA determinations). The
450 partial melting degree (F) associated to the production of primitive melts was then calculated
451 as the difference between the melted fertile assemblage and the recrystallized restitic residuum,
452 checking the consistency with the experimental data on genesis of shoshonitic melts (e.g.,
453 Conceição and Green, 2004). Once fixed the mineral phase proportions required to produce the
454 primary SS/SU magmas at the eutectic, the non-modal batch melting equation of Shaw (1979)
455 was applied to model their trace element distribution. In this equation, expressed as:

456
$$C_L = \frac{C_0}{[D+F \times (1-P)]} ,$$

457 C_L represents the concentration of each trace element in the primary melt, C_0 its concentration
458 in the modelled source, D its partition coefficients weighted for the modal composition of the
459 source and P the partition coefficients weighted for the eutectic mineral proportions (see Table
460 2 and Supplementary Material for the partition coefficients and mineral melting proportions).

461 The application of this equation to elements with highly incompatible behaviour, i.e. with $D \rightarrow 0$
462 and $P \rightarrow 0$, approximated by the ratio:

463
$$F = \frac{C_0}{C_L} ,$$

464

465 was also used throughout the calculations to check the consistency of the partial melting
466 degrees.

467

468 ***7.2. Partial melting conditions and role of H₂O***

469 The partial melting models suggest that 7% partial melting of a fertile spinel-lherzolite bearing
470 small proportions of amphibole (1.98%) and phlogopite (0.76%) can produce the major/trace
471 element content of primary SS transitional basalt. Hydrous phases were totally consumed
472 during this melting stage, significantly contributing to the eutectic of the system (Table 4). A
473 negligible contribution of spinel accompanied by positive melting proportions of
474 orthopyroxene, clinopyroxene and by dominant recrystallization of olivine, is also required to
475 account for the major/trace element distribution of primary SS melts. On the other hand, the
476 production of the SU primary alkali basalt can be explained by 5% partial melting of a spinel-
477 lherzolite with slightly lower abundances of hydrous phases (1.36% amphibole + 0.62%
478 phlogopite). Again, hydrous phases were completely consumed, together with significant
479 proportions of orthopyroxene and clinopyroxene. Using experimental estimations of mineral-
480 melt water partitioning during peridotite melting, the estimated H₂O amounts of the produced
481 primary melts are in the range of 1.06-1.12 wt% (Table 4; Supplementary Material). In terms
482 of trace elements, the predicted and calculated SS/SU melts are very close, especially the LILE
483 and REE (Fig. 9). In both models, the main discrepancies are related to U, Zr-Hf and Sr, whose
484 concentrations in the sources resulted too low to match those of the primary melts (see
485 discussion below).

486

487 ***7.3. Peridotite melting and mantle source depth***

488 Conceição and Green (2004) showed that decompression melting of a metasomatised spinel-
489 peridotite bearing small amounts amphibole and phlogopite can produce shoshonitic magmas
490 at ~5% partial melting, $T = 1075^{\circ}\text{C}$ and $P = 1.0 \text{ GPa}$, by exhausting hydrous phases very close

491 to the solidus. Consistent with their findings, our model shows that 5-7% melting of an
492 amphibole+phlogopite-bearing spinel lherzolite can reproduce the composition of the primary
493 SS/SU melts, requiring the complete disappearance of both amphibole and phlogopite.

494 During the two melting stages, the contribution of the pyroxenes was opposite. Indeed, while
495 the SS melting stage was characterized by an orthopyroxene-dominated eutectic, in the SU
496 melting stage the maximum contribution at the eutectic was given by clinopyroxene (Table 4).
497 This behaviour can be explained by the role played by volatiles (H₂O, CO₂) and alkalis during
498 partial melting. Experimental studies showed that the amount of clinopyroxene and
499 orthopyroxene consumed during melting of spinel lherzolites at 1.1-1.2 GPa are negatively and
500 positively correlated to the amount of H₂O in the system, respectively (Gaetani and Grove,
501 1998; Condamine and Médard, 2014). Moreover, Liu et al. (2006) found that the effect of H₂O
502 is somewhat similar to that of K₂O, so that increasing the concentration of these elements in the
503 peridotitic source results in the genesis of SiO₂-saturated (to oversaturated) melts. Conversely,
504 decreasing H₂O and/or increasing Na₂O in the source system leads to the production of
505 nepheline-normative liquids (Liu et al., 2006). Accordingly, the slightly different H₂O contents
506 of the SS and SU sources are associated to the genesis of hypersthene- and nepheline-normative
507 melts, respectively.

508 According to Selverstone and Sharp (2011) results, the Finero amphibole-phlogopite peridotite
509 equilibrated at *P* of 1.2-1.6 GPa. These values are consistent with the experimental works on
510 the genesis of shoshonitic melts from peridotitic sources (e.g., Conceição and Green, 2004) and
511 with the experimental studies on amphibole/phlogopite stability in the SCLM (Condamine and
512 Médard, 2014; Mandler and Grove, 2016). In this framework, it is likely that the transitional to
513 alkali basaltic SS/SU melts were generated was located at 40-60 km of depth, i.e. well above
514 the spinel-garnet transition in SCLM domains (2-3 GPa, i.e. 60-90 km; Falloon and Green,
515 1988). This melting depth is consistent with the equilibration depth calculated by Carraro and
516 Visonà (2003) for the spinel-lherzolitic xenoliths sampled along the same SCLM column (i.e.

517 ~45 km), and quite shallower than the one from which alkaline lamprophyres were generated
518 during the subsequent Late Triassic cycle (70-80 km; Casetta et al., 2019).

519

520 **8. The ascent and emplacement of the Middle Triassic melts**

521 ***8.1. Equilibrium conditions and adopted thermobarometric/hygrometric models***

522 The *T-P-H₂O* paths followed by SS and SU melts in the interval comprised between segregation
523 from the mantle source and their emplacement/differentiation in shallow crust were tracked by
524 means of thermobarometric and hygrometric equations based on olivine- and clinopyroxene-
525 melt equilibria. To reconstruct the *T* conditions of crystallization of the least differentiated SU
526 magmas, the olivine- and clinopyroxene-melt thermometers of Putirka (2008; Equations 22 and
527 33) were applied to the phenocrysts yielding equilibrium with the whole-rock composition of
528 the host dykes (i.e. $^{Ol-Melt}Kd_{Fe-Mg} = 0.27-0.33$ and $^{Cpx-Melt}Kd_{Fe-Mg} = 0.24-0.30$; Fig. 10). Due to
529 the generally low crystal cargo of the SU trachybasaltic dyke (P.I. down to 5-10%; Table 1), its
530 whole-rock compositions was considered a good approximation of a melt composition. The
531 pressures of clinopyroxene crystallization were calculated by means of the Putirka (2008,
532 Equation 32b) H₂O-dependent barometer, in turn derived from the Putirka et al. (1996, Equation
533 T1) thermometer. Because of the sensitivity of the chosen thermobarometers to the H₂O content
534 of the melt, multiple calculations were performed using as input variable amounts of H₂O (from
535 1 to 5 wt%). These H₂O-dependent equations were then iterated with the *T-P*-dependant
536 clinopyroxene-based hygrometer of Perinelli et al. (2016) until matching results were obtained,
537 in a process that enable to skip circular assumptions of the *T-P-H₂O* conditions of the
538 crystallizing melts (see Casetta et al., 2018a; Lanzafame et al., 2020).

539 Due to the absence of unaltered olivine phenocrysts, the primary *T-P-H₂O* conditions of
540 crystallization of SS melts were modelled by applying thermobarometric and hygrometric
541 equations to clinopyroxene in early-derived clinopyroxenitic cumulates. Accordingly,
542 clinopyroxene grains inside clinopyroxenitic xenolith were traced back to the equilibrium

543 conditions with the least differentiated SS trachybasaltic dyke in terms of both major (Fig. 10a)
544 and trace elements (Fig. 10b; see Supplementary Material for the adopted mineral-melt partition
545 coefficients for trace elements). The most Mg-poor clinopyroxenes in trachybasaltic lavas
546 resulted in clear disequilibrium with their host rock (Fig. 10), likely because the high P.I. of the
547 studied lavas, crystallized in almost hypabyssal conditions, make them far from resembling a
548 melt in terms of whole-rock composition. Therefore, these crystals were traced back to
549 equilibrium conditions (i.e. $K_{\text{Fe-Mg}}^{\text{Cpx-Melt}} = 0.24\text{-}0.30$; Putirka, 2008) with less porphyritic
550 trachybasaltic and trachyandesitic rocks belonging to the same magmatic suite (see also Casetta
551 et al., 2018a). Afterwards, the chosen clinopyroxene-melt thermobarometers and hygrometers
552 (Putirka, 2008; Perinelli et al., 2016) were iteratively applied to determine the T - P - H_2O space
553 of clinopyroxene crystallization. Unfortunately, no reliable P - H_2O results were obtained by the
554 iterative procedures for the clinopyroxene crystals inside clinopyroxenitic xenoliths. Therefore,
555 to obtain a rough estimate of their crystallization P , the Putirka et al. (2003) H_2O -independent
556 barometer was used. Further details, together with a complete list of the obtained results and of
557 the errors associated to each adopted equation are listed in Table 5.

558

559 **8.2. T - P - H_2O paths**

560 Thermometric models showed that the phenocrysts of olivine (Fo_{85}) and clinopyroxene (Mg\#_{Cpx}
561 85) in equilibrium with the SU trachybasaltic dyke (Mg\#_{Liq} 64) record the most congruent
562 crystallization T (from 1195 to 1204-1207 °C), by considering a H_2O content in the melt of 1.5-
563 1.9 wt%. The corresponding crystallization P obtained from the clinopyroxene-melt barometer
564 is ~830 MPa (Fig. 10). The deepest conditions of crystallization of SS magmas ($P = 570\text{-}710$
565 MPa) are from clinopyroxene grains inside cumulitic xenoliths (Mg\#_{Cpx} 86-88) in equilibrium
566 with the SS trachybasaltic dyke (Mg\#_{Liq} 65), at T between 1100 and 1128 °C (Fig. 10). The
567 most Mg-poor clinopyroxene phenocrysts (Mg\#_{Cpx} 70-83) analysed in SS trachybasaltic to
568 trachyandesitic lavas and dykes (equilibrium melts = Mg\#_{Liq} 41-53) record shallower

569 conditions. Their crystallization occurred along a polybaric path at P comprised between 470
570 and 230-170 MPa, until the onset of a nearly isobaric cooling regime took place at P of 50-170
571 MPa (Fig. 10). Although plausible and in line with previous estimates for the same magmatic
572 suite (e.g., Casetta et al., 2018a; De Min et al., 2020), these T - P results have to be considered
573 together with the associated errors (i.e. ± 29 - 42°C ; ± 170 - 260 MPa; Table 5).

574

575 **9. Discussion**

576 ***9.1. The Southern Alps SCLM and the genesis of shoshonitic melts***

577 A possible relationship between the metasomatized orogenic peridotites and the Mid-Triassic
578 orogenic-like magmatism in the Southern Alps has been often hypothesized (Zanetti et al.,
579 1999; 2013; Mazzucchelli et al., 2010; De Min et al., 2020). As shown in Fig. 7, the shoshonitic
580 magmas of the Dolomites share a common enriched (EM I-like) Sr-Nd isotopic signature
581 (Bonadiman et al., 1994; Casetta et al., 2018a; Lustrino et al., 2019), mirroring the fingerprint
582 of the underlying mantle source. The existence of K-H₂O-rich mantle-derived melts in the
583 Southern Alps SCLM during Mid-Triassic is also supported by the intrusion of gabbroic dykes
584 with similar Sr-Nd isotopic signature to the SS/SU magmas inside the Finero peridotite
585 (Giovanardi et al., 2020). According to many authors, the slab-derived enrichment of the
586 Southern Alps SCLM has to be ascribed to the Variscan subduction (Bonadiman et al., 1994;
587 Lustrino et al., 2019), the effect of which are also preserved in the metasomatised Finero and
588 Ulten peridotites (Zanetti et al., 1999; Ionov et al., 2017; Giovanardi et al., 2020). In the Finero
589 peridotite, several generations of hydrous phases (amphibole and phlogopite) formed after
590 pervasive and prolonged metasomatism of depleted harzburgites (Voshage et al., 1987;
591 Hartmann and Wedepohl, 1993; Zanetti et al., 1999; 2013; 2016; Tommasi et al., 2017;
592 Giovanardi et al., 2020). The infiltration of slab-related metasomatic agents in the Finero main
593 peridotitic body is also suggested by the Nb-Ta-Ti negative anomalies and enrichment in
594 radiogenic Sr of amphibole grains, which lie on the mixing trend between DMM and the

595 average crustal (and/or GLOSS/GLOSS-II) components (Fig. 7; see also Giovanardi et al.,
596 2020). Although partially overprinted by the complex *T-P-t* trajectories followed from
597 depletion to the subsequent Palaeozoic metasomatism and the Mesozoic (~180 Ma) exhumation
598 at shallower levels (Hartmann and Wedepohl, 1993; Zanetti et al., 2016; Langone et al., 2017),
599 the Finero peridotite is the best approximation of the Southern Alps SCLM during
600 MiddleTriassic, i.e. at the time of the shoshonitic magmatism. Accordingly, our models enabled
601 us to set out that an assemblage composed of phases with most fertile composition from the
602 Finero (and Baldissero) peridotite is able to approach, in terms of major/trace elements, a bulk-
603 rock source composition similar to that resulting from a mixing between DMM (98-96%) and
604 GLOSS/GLOSS-II (2-4%) components (Fig. 9; Table 3). Besides confirming the role played
605 by subductive crustal components in the modification of the Southern Alps SCLM, the
606 modelled fertile mantle source perfectly encompasses the Sr-Nd isotopic variations of the Mid-
607 Triassic shoshonitic magmas (Fig. 7; Table 1). The modelled sources are consistently able to
608 account for the major/trace element composition of the primary SS/SU magmas at 5-7% partial
609 melting. These melting degrees are higher than those hypothesized by Lustrino et al. (2019),
610 who suggested that a supra-subductive peridotite interacting with slab-derived components is
611 able to generate the Mid-Triassic shoshonitic magmatism at 2% slab + 1.1% peridotite melting.
612 According to these authors, although reliable in terms of geodynamic interpretation, such a
613 model is limited by the use of a non-primary target melts (Mg# 61), resulting in excessively
614 low degrees of partial melting.

615 The incompatible element patterns of the modelled SS/SU melts mirror those of the primary
616 magmas, except for some discrepancies in the Th-U, Sr and Zr-Hf distributions (Fig. 9), which
617 likely reflect the low concentrations of these elements in the modelled sources, i.e. in the
618 mineral phases in the Finero peridotite (see Giovanardi et al., 2020). The “ideal” mantles
619 resulting from the mixing between DMM and GLOSS/GLOSS-II components are enriched in
620 Th-U and Zr-Hf with respect to the modelled Finero-like sources (Fig. 9). The positive Sr spike

621 in the primary melts, a peculiar feature of magmas generated from subduction-modified sources
622 (Zheng, 2019), cannot be reproduced neither by the modelled sources nor by the “mixed”
623 DMM+GLOSS mantles. This discrepancy could be explained by the variable concentration of
624 Sr in worldwide sediments (see Plank, 2014) or, alternatively, by the extremely variable and T -
625 P -dependent mobility of Sr in metasomatic melts/fluids and/or supercritical fluids in
626 subduction-related settings (Kessel et al., 2005).

627

628 ***9.2. Ascent and emplacement of the Middle Triassic magmas***

629 The maximum P of crystallization of SS/SU magmas yielded by clinopyroxene-melt
630 barometers are comprised between 710 and 830 MPa (i.e. 24.5-28.6 km assuming a $\Delta P/\Delta z$ of
631 ~ 29 MPa/km; Fig. 10). During ascent, the crystallization processes continued in a polybaric
632 regime until 160-230 MPa, where the nearly isobaric cooling from T of 1060 °C down to T
633 < 1000 °C reflects the formation, in a shallow crustal environment, of discrete magma batches,
634 dominated by the onset of fractional crystallization processes (Fig. 10). Evidence of the
635 progressive alkali- and H₂O-enrichment during the shallow evolution of the melts are widely
636 documented in the Predazzo and Mt. Monzoni plutons, which record emplacement P of 40-165
637 MPa (Bonadiman et al., 1994; Casetta et al., 2018a). The hygrometric results show that
638 clinopyroxene crystallization in SS and SU melts took place at progressively increasing H₂O
639 amounts, from 1.7 wt% in trachybasalts at 710-830 MPa up to 3.9 wt% in trachyandesites at
640 380 MPa (Fig. 10; Table 5). Such H₂O concentrations are not far from what expected if a
641 completely incompatible behaviour of H₂O is assumed during progressive differentiation of
642 mantle-derived transitional to alkali basaltic melts with ~ 1.1 wt% H₂O. At shallower P , volatiles
643 exsolution promoted crystallization of clinopyroxene, feldspar and Fe-Ti oxides, followed by
644 amphibole and phlogopite in the case of magmas differentiating and cooling in
645 hypabyssal/plutonic conditions (Fig. 10).

646

647 ***9.3. Geodynamic implications: a mantle perspective***

648 Our study provides a robust link between the shoshonitic magmatism of the Dolomites and the
649 metasomatised Finero peridotite, enabling speculation about the evolution of the Southern Alps
650 SCLM. In fact, 7% partial melting of a Finero-like mantle containing ~800 ppm H₂O is able to
651 generate primary transitional basaltic SS melts resembling those emplaced during the most
652 productive phase of the Mid-Triassic (239-237 Ma) magmatism in the Dolomites. On the other
653 hand, 5% partial melting of a slightly different SCLM domain (H₂O ~600 ppm) produced the
654 SU primary alkali basaltic melts, which emplaced mostly in form of dykes and small intrusive
655 bodies, for which no age data have been reported so far. The progressive depletion of the SCLM
656 beneath the Southern Alps continued towards Late Triassic, when the asthenospheric influx
657 related to the Alpine Tethys opening phase led to the production of alkaline lamprophyres with
658 anorogenic signature at 219.22±0.73 Ma (Casetta et al., 2019). The extreme volatile-rich nature
659 of lamprophyres can be ascribed to small degree partial melts (1-0-2.5%) of their amphibole-
660 garnet-lherzolitic source (Casetta et al., 2019). Overall, the shifting from shoshonitic to alkaline
661 magmatism recorded in the Dolomites is a very effective way to track progressive depletion, in
662 terms of both major/trace elements and volatiles content, of the Southern Alps SCLM prior to
663 the rifting phase connected to the Alpine Tethys opening.

664 On the other hand, the retention of the modal/compositional enrichment in the lithosphere for a
665 time interval as long as 50 Ma, i.e. from the Variscan subduction to the Mid-Triassic
666 magmatism, can be matter of debate. In particular, it is complicated by the Permian (~294 to
667 274 Ma) onset of a large magmatic phase, during which mantle-derived anhydrous basaltic
668 melts with tholeiitic to subduction-related affinity were produced by the SCLM and/or the
669 asthenosphere (Sinigoi et al., 2016; Boscaini et al., 2020). These magmas underplated the thick
670 crustal piles and induced extensive anatexis, leading to the emplacement of mafic/ultramafic to
671 granitic/rhyolitic rocks throughout the Ivrea-Verbano Zone, particularly in Val Sesia (Voshage
672 et al., 1990; Sinigoi et al., 2016; Karakas et al., 2019) and the Athesina Volcanic District (Barth

673 et al., 1993; Marocchi et al., 2008). Furthermore, the gabbroic bodies of the Finero Mafic
674 Complex record the intrusion of multiple discrete tholeiitic/transitional magmatic batches
675 between Permian and Triassic, suggesting the existence of a multi-stage thermal evolution for
676 the Southern Alps lithosphere (Lu et al., 1997; Zanetti et al., 2013; Langone et al., 2017). In
677 this framework, the timing between the enrichment of Southern Alps SCLM and the Mid-
678 Triassic shoshonitic magmatism can belong to two different scenarios:

679 a) The existence of a Triassic arc/back-arc system at the NW Paleotethys limb (Zanetti et al.,
680 2013). In this view, the infiltration of slab-derived components into a depleted SCLM would
681 promote the formation of a Finero-like suprasubductive mantle domain during Mid-Triassic,
682 making the shoshonitic magmatism a simple expression of an active subduction-related setting.
683 This model implies that the enriched signature related to the Variscan subduction in the
684 Southern Alps SCLM was exhausted by the voluminous Permian magmatic event (see Storck
685 et al., 2020).

686 b) The preservation, in the Southern Alps lithospheric mantle, of an enrichment induced by the
687 Variscan subduction (Bonadiman et al., 1994; Zanetti et al., 1999; Lustrino et al., 2019). In this
688 view, the SCLM would have been capable of generating a transitional to alkaline magmatic
689 phase delayed with respect to the metasomatic events. The Sr-Nd isotopes support the
690 plausibility of such a scenario, showing how the hypothetical slab-derived components (i.e.
691 GLOSS/GLOSS-II), the Finero peridotite, the metasomatic amphiboles and the Mid-Triassic
692 magmatic rocks would lie on the same mixing curve (Fig. 7). This hypothesis is also in line
693 with the model of Pilet et al. (2015), according to whom continental rift-related Na- to K-
694 alkaline magmatism could be decoupled from the metasomatic enrichment of the underlying
695 mantle source. Analogous speculations were made by Panter et al. (2018) to model the Cenozoic
696 (48-30 Ma) alkaline magmatism of the West Antarctic Rift System by melting of amphibole-rich
697 metasomes generated in the SCLM during the 550-150 Ma Ross subduction. In an analogous
698 way, the shoshonitic magmatism of the Southern Alps would have been generated by a Finero-

699 like mantle source, and would represent the exhaustion of the Variscan subduction-related
700 signature in the Southern Alps SCLM that predated the Late Triassic-Early Jurassic opening of
701 the Alpine Tethys (Fig. 11).

702

703 **10. Conclusive remarks**

704 A comprehensive study of the acme of the Middle Triassic shoshonitic magmatism in the
705 Dolomites enabled us to constrain the nature and evolution of the Southern Alps SCLM between
706 the Variscan subduction and the opening of the Alpine Tethys. Insights from major/trace
707 elements and Sr-Nd isotopes were brought together to frame the shoshonitic magmatism into a
708 comprehensive picture at both local and regional geodynamic scale. The main points are
709 summarized here:

- 710 - The shoshonitic magmatism in the Southern Alps at 239-237 Ma is mostly represented by
711 a huge effusive phase, during which SiO₂-saturated (SS) to SiO₂-unsaturated (SU) melts
712 partially crystallized and differentiated *en route* to the surface. The *T-P* paths recorded by
713 olivine and clinopyroxene phenocrysts in the least differentiated rocks encompass a
714 considerable interval, spanning from 1200-1210 °C at 830 MPa to 1010-1060 °C at 160-
715 230 MPa (i.e. polybaric crystallization during ascent). At 160-230 MPa, almost isobaric
716 crystallization conditions marked the onset of differentiation in discrete magma batches at
717 shallow crustal levels, the remnants of which are still visible in the Predazzo-Mt. Monzoni
718 plutons (Bonadiman et al., 1994; Casetta et al., 2018a).
- 719 - The primary transitional to alkali basaltic SS and SU melts in equilibrium with a fertile
720 SCLM were generated by 5-7% partial melting of a metasomatised amphibole+phlogopite-
721 bearing spinel lherzolite similar to the Finero peridotite. Such a source was probably located
722 at depth of 40-60 km, consistent with theoretical (Selverstone and Sharp, 2011) and
723 experimental data (e.g., Conceição and Green, 2004).

- 724 - The metasomatic modification of the Mid-Triassic Southern Alps SCLM by slab-related
725 components (Fig. 7) derived from the Variscan subduction could have been preserved into
726 the Southern Alps SCLM, being able to induce the shoshonitic magmatic phase during Mid-
727 Triassic (Fig. 11).
- 728 - The modelled Finero-like mantle sources are capable of retaining up to 600-800 ppm H₂O,
729 resulting in the production of primary transitional to alkali basaltic SS/SU melts with ~1.1
730 wt% H₂O. The shoshonitic magmatism of the Dolomites could fix at 237-239 Ma the
731 exhaustion of the Variscan subduction-related signature in the Southern Alps SCLM. This
732 event predated the Late Triassic onset of a major rifting phase related to the opening of the
733 Alpine Tethys, the expression of which are recorded in small volume alkaline magmas (e.g.,
734 Casetta et al., 2019).

735

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983

984 **Figure captions**

985 **Fig. 1. (colour online)**

986 (a) Map of the tectonic units of the eastern portion of the Alps (modified from Schmid et al.,
987 2020). LO: Ligurian Ophiolites; AM: deformed Adriatic margin; AD: Adriatic Microplate; SA:
988 Southern Alps; DI: Dinarides; SM: Southern margin of Meliata; HB: Eoalpine High-Pressure
989 Belt; TW: Tauern tectonic Window; EW: Engadine tectonic Window; OTW: Ossola-Tessin
990 tectonic Window; EA: Eastern Austroalpine; H: Helvetic domain; M: Molasse foredeep. The
991 Mid-Triassic magmatic occurrences in the Southern Alps domain are evidenced in black. (b)
992 Basaltic trachyandesitic dyke intruded in the volcanic sequences. (c) Ultramafic xenoliths
993 embedded in a trachybasaltic dyke. (d) Simplified geological map of the volcano-plutonic
994 sequences and dyke swarm cropping out near Predazzo (modified from Casetta et al., 2018a).
995 SS = Shoshonitic Silica Saturated; SU = Shoshonitic Silica Undersaturated; GU = Granitic Unit.
996 Stars indicate dykes where ultramafic xenoliths were found.

997

998 **Fig. 2. (colour online)**

999 Photomicrographs of representative thin sections of the studied lavas, dykes and ultramafic
1000 cumulates. (a) Triple junctions between clinopyroxene grains in a clinopyroxenitic xenolith. (b)
1001 Contact between a clinopyroxenitic xenoliths and the host SU trachybasaltic dyke. (c) Highly

1002 porphyritic SS trachybasaltic dyke. (d, e) Clinopyroxene, olivine and plagioclase phenocrysts
1003 in a SU trachybasaltic dyke; (f) Highly porphyritic SS trachybasaltic lava. Cpx: clinopyroxene;
1004 Plg: plagioclase; Sp: spinel. (a), (b), (d) and (e) are in transmitted cross-polarized light; (c) and
1005 (f) are in transmitted plane-polarized light. Scale bar is 1 mm.

1006

1007 **Fig. 3. (colour online)**

1008 (a) TiO₂ (wt%) vs. Mg# and (b) Wollastonite (Wo)-Enstatite (En)-Ferrosilite (Fs) (Morimoto,
1009 1988) diagrams for clinopyroxene in lavas, dykes and clinopyroxenites. (c, d) Chondrite-
1010 normalized (Sun and McDonough, 1989) incompatible element and REE patterns of
1011 clinopyroxene in clinopyroxenites. The composition of clinopyroxene near the contact with the
1012 host SU dyke is also discriminated. The major and incompatible element distribution of
1013 clinopyroxene phenocrysts in the Late Triassic lamprophyric dykes from the same area (Casetta
1014 et al., 2019) is also reported for comparison.

1015

1016 **Fig. 4. (colour online)**

1017 Major and trace element compositions of SS lavas and SS/SU dykes. (a) Total Alkali vs. Silica
1018 (TAS) (Le Maitre et al., 2002), (b) K₂O (wt%) vs. SiO₂ (wt%) (Peccerillo and Taylor, 1975),
1019 (c) CaO/Al₂O₃ vs. SiO₂ (wt%), and (d) Ni (ppm) vs. MgO (wt%) diagrams. The composition of
1020 the SS, SU and GU bodies of the Predazzo Intrusive Complex (PIC) and of the Late Triassic
1021 lamprophyric dykes from the same area (Casetta et al., 2018a; 2018b; 2019) are also reported
1022 for comparison. Asterisks indicate the composition of the primary SS transitional basalt and SU
1023 alkali basalt resulted from the backward fractionation models (see text).

1024

1025 **Fig. 5. (colour online)**

1026 (a, b) Primitive Mantle- (PM) and Chondrite-normalized (Sun and McDonough, 1989)
1027 incompatible element and REE diagrams showing the composition of the SS lavas, SS/SU

1028 dykes and clinopyroxenites. (c, d) Primitive Mantle- (PM) and Chondrite-normalized
1029 incompatible element and REE patterns of the primary SS transitional basalt and SU alkali
1030 basalt resulted from the backward fractionation models. The patterns of the Late Triassic
1031 lamprophyres from the same area (Casetta et al., 2019) are also reported for comparison in (c)
1032 and (d).

1033

1034 **Fig. 6. (colour online)**

1035 Trace element discrimination diagrams for the studied SS lavas and SS/SU dykes in comparison
1036 with the Late Triassic lamprophyres from the same area (Casetta et al., 2019). (a) Th_N vs. Nb_N
1037 (Saccani, 2015), (b) Nb/La vs. La/Yb diagrams (Smith et al., 1999) and (c) Th-Hf-Ta ternary
1038 diagram (Wood, 1980).

1039

1040 **Fig. 7. (colour online)**

1041 $^{87}Sr/^{86}Sr$ vs. $^{143}Nd/^{144}Nd$ diagram for the studied SS lavas and SS/SU dykes (corrected to 238
1042 Ma) compared to the Permian Campore dyke (280 Ma; Sinigoi et al., 2016), Ulten peridotite
1043 (330 Ma; Tumiati et al., 2003), Finero phlogopite peridotite (FPP, 300 Ma; Voshage et al.,
1044 1987), metasomatic amphiboles in the FPP (300 Ma; Giovanardi et al., 2020, and references
1045 therein), and the Southern Alps basement (gneisses, paragneisses, stromalites, 300-330 Ma;
1046 Voshage et al., 1990; Tumiati et al., 2003; Sinigoi et al., 2016). Fields indicate the composition
1047 of Balmuccia-Baldissero peridotites (Voshage et al., 1987; Mukasa and Shervais, 1999),
1048 Southern Alps Mid-Triassic magmatic rocks (Lustrino et al., 2019) and the Predazzo Intrusive
1049 Complex (PIC) SS/SU rocks (Casetta et al., 2018a). DMM = depleted MORB mantle
1050 (Workman and Hart, 2005; corrected to 238 Ma); EM I = enriched mantle (Zindler and Hart,
1051 1986); GLOSS and GLOSS-II = average global subducting sediments (Plank, 2014); Average
1052 crust = average composition of the Southern Alps Permo-Triassic crust (Voshage et al., 1990;
1053 Tumiati et al., 2003; Sinigoi et al., 2016). Curves represent the mixing trends between DMM

1054 and GLOSS, GLOSS-II and the Permo-Triassic average crust. Numbers indicate the mixing
1055 percentages.

1056

1057 **Fig. 8. (colour online)**

1058 Chondrite-normalized (Sun and McDonough, 1989) incompatible element patterns of the
1059 inferred Mid-Triassic mantle sources that generated SS/SU magmas. SS- and SU-like sources
1060 are constituted by 96% DMM + 4% GLOSS-II and 98% DMM + 2% GLOSS components,
1061 respectively. Whole-rock patterns of FPP (Giovanardi et al., 2020), Ulten peridotites (Ionov et
1062 al., 2017), DMM (Workman and Hart, 2005), PM (Sun and McDonough, 1989) and Etna-like
1063 enriched mantle (EM; Casetta et al., 2020) are also reported.

1064

1065 **Fig. 9. (colour online)**

1066 Chondrite-normalized (Sun and McDonough, 1989) incompatible element diagrams showing
1067 the composition of the modelled (a) SS and (b) SU mantle sources and primary melts. In each
1068 panel, the composition of the mantle source obtained by mass balance calculations and that
1069 resulted from the mixing models are indicated. The composition of the primary (a) SS
1070 transitional basalt and (b) SU alkali basalt and of the melt generated at 7% and 5% partial
1071 melting from the corresponding sources are also indicated.

1072

1073 **Fig. 10. (colour online)**

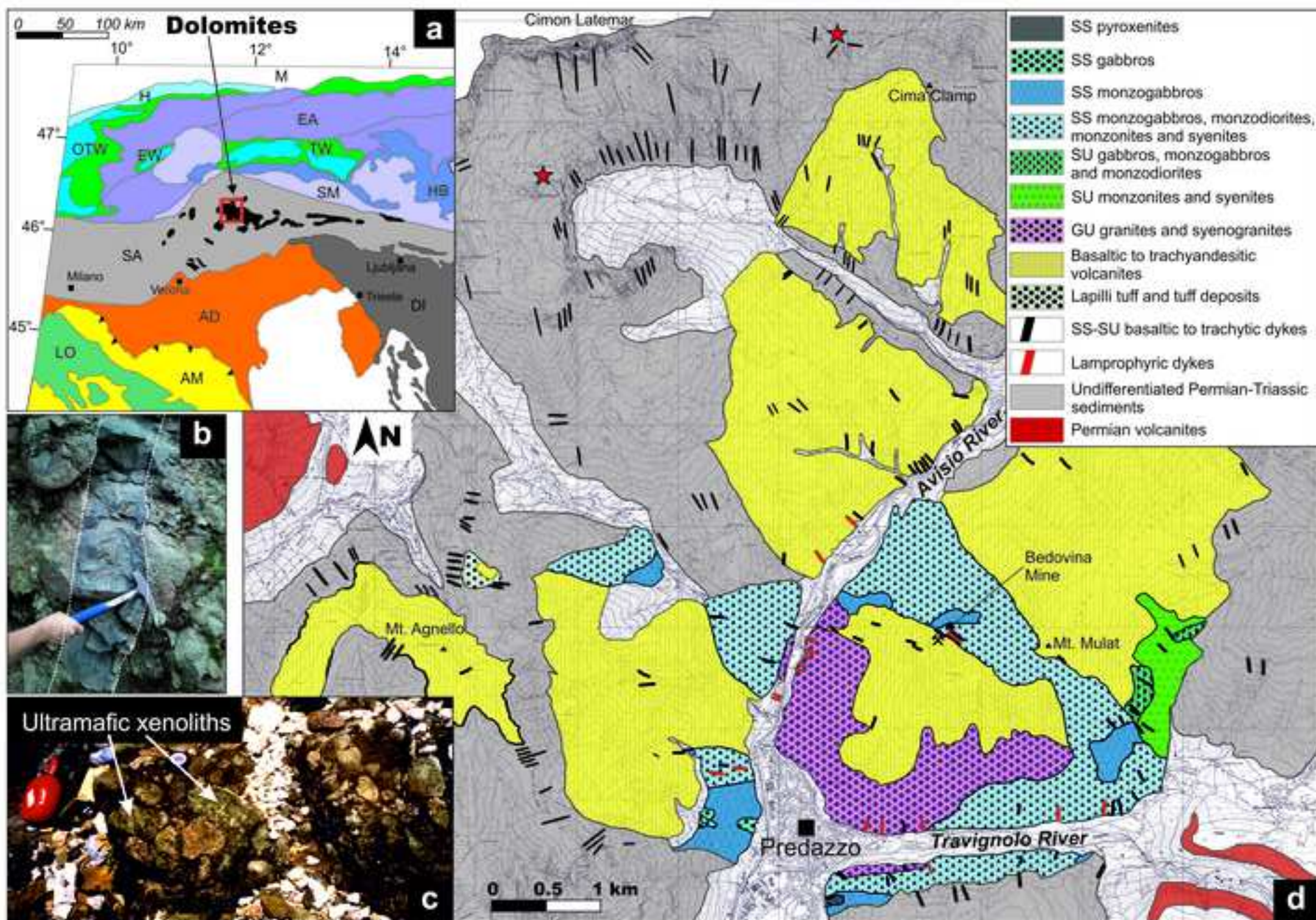
1074 (a) $K_{\text{Fe-Mg}}^{\text{Cpx-melt}}$ vs. Mg# diagram for clinopyroxene crystals in equilibrium (or brought to the
1075 equilibrium) with the SS and SU melts. The equilibrium range indicated by the black lines ($K_{\text{Fe-Mg}}^{\text{Cpx-}}$
1076 $^{\text{melt}} = 0.24-0.30$) is from Putirka et al. (2003). (b) Chondrite-normalized (Sun and
1077 McDonough, 1989) incompatible element pattern of the hypothetical melt in equilibrium with
1078 clinopyroxene grains in clinopyroxenite (see Supplementary Material for the partition
1079 coefficients) compared to that of the natural SS trachybasaltic dyke. (c) Temperature (T , °C)

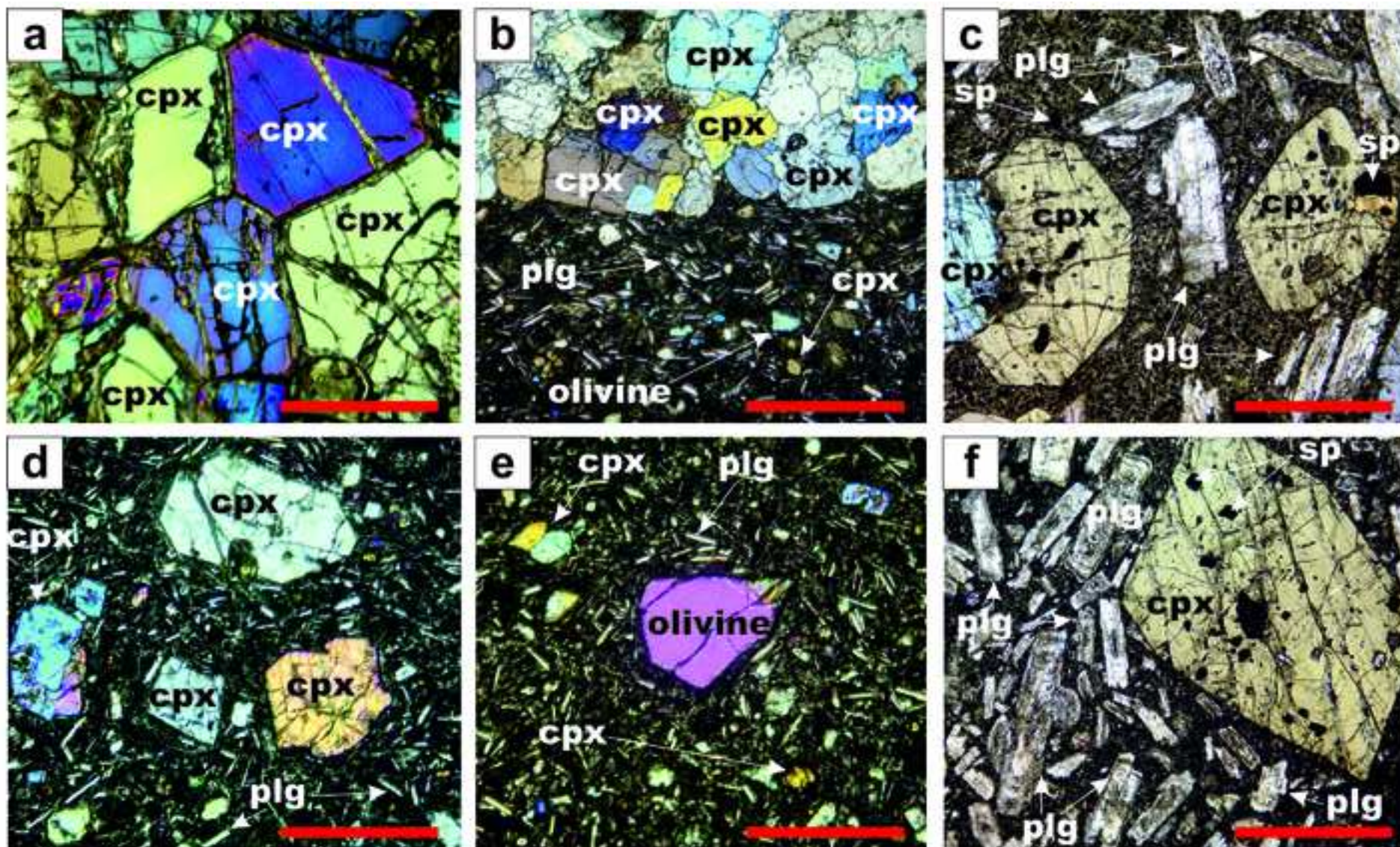
1080 vs. pressure (P , MPa) diagram showing the clinopyroxene crystallization conditions in the
1081 SS/SU magmas, and the corresponding melt H₂O content of the melts. The maximum T - P
1082 crystallization interval of the Late Triassic lamprophyres (Casetta et al., 2019; De Min et al.,
1083 2020) is also reported in (c).

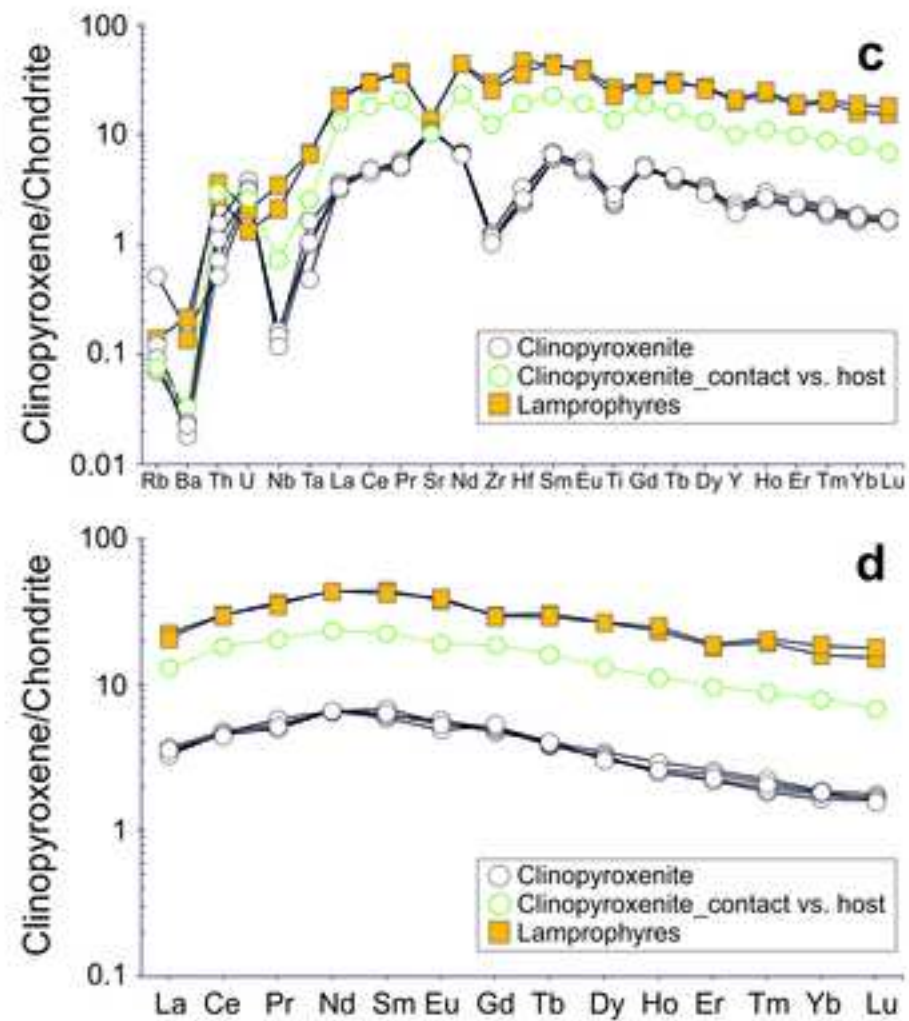
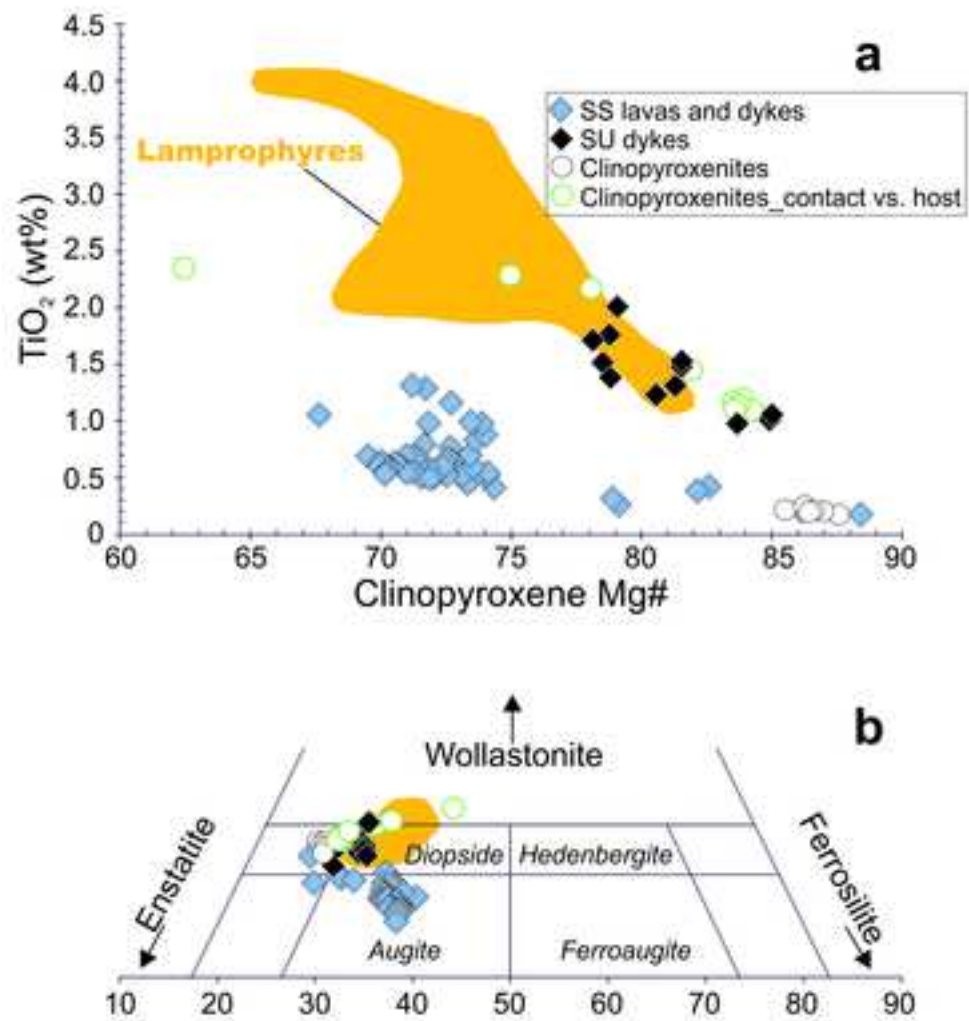
1084

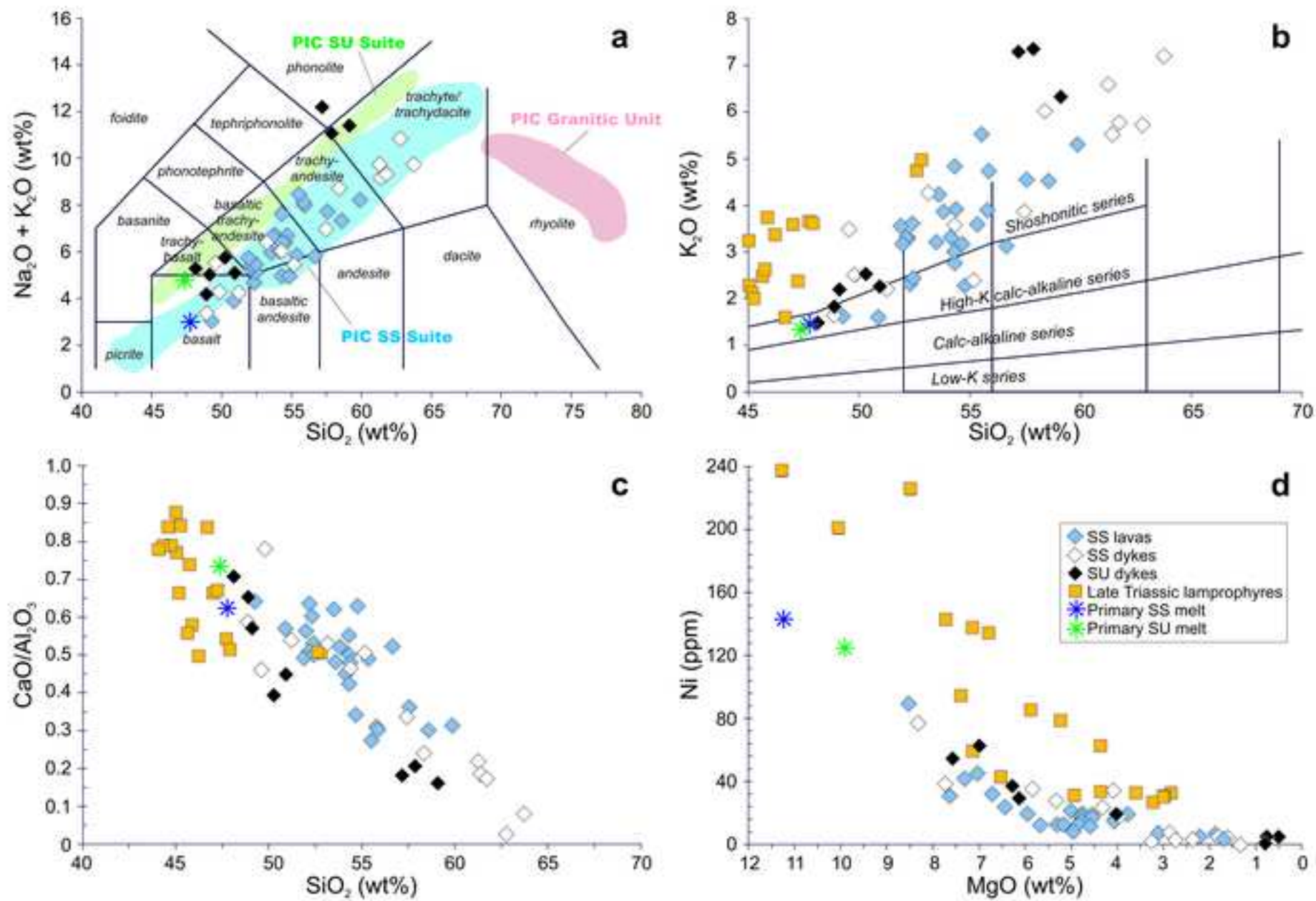
1085 **Fig. 11. (colour online)**

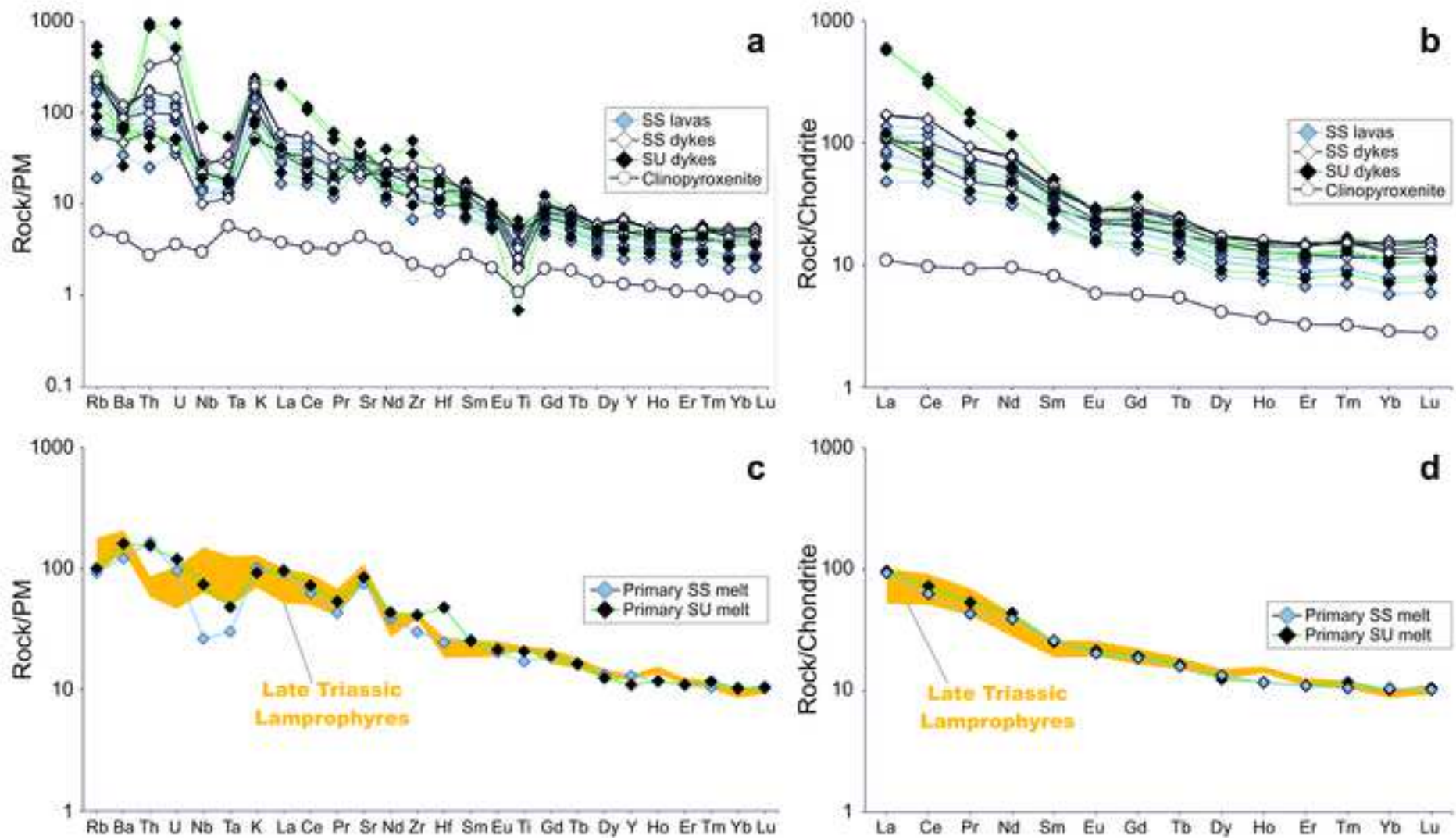
1086 Evolutionary sketch of the Southern Alps SCLM between Carboniferous and Late Triassic,
1087 adapted from Pilet (2015) and Panter et al. (2018). (a) Carboniferous (>320 Ma) subduction of
1088 the Paleotethys beneath the Adria plate resulting in the slab-derived metasomatism of the
1089 Southern Alps SCLM (Stampfli et al., 2013). (b) Permian extensional stage activated by
1090 brittle/ductile intra-lithospheric shear zones inducing thinning, partial melting of the SCLM.
1091 This resulted in the production of tholeiitic to calc-alkaline magmas which intruded the Lower
1092 Crust, and induced the hybrid anatectic to mafic-derived acidic magmatism throughout the
1093 Ivrea-Verbano Zone and the Athesina Volcanic District. (c) Middle Triassic prolonged
1094 extensional phase inducing the exhaustion of the slab-derived enrichment retained in the
1095 SCLM, and resulting in the production of calc-alkaline to shoshonitic magmatism. (d) Late
1096 Triassic production of alkaline magmas throughout the Southern Alps caused by the rift-related
1097 asthenospheric upwelling predating the opening of the Alpine Tethys.

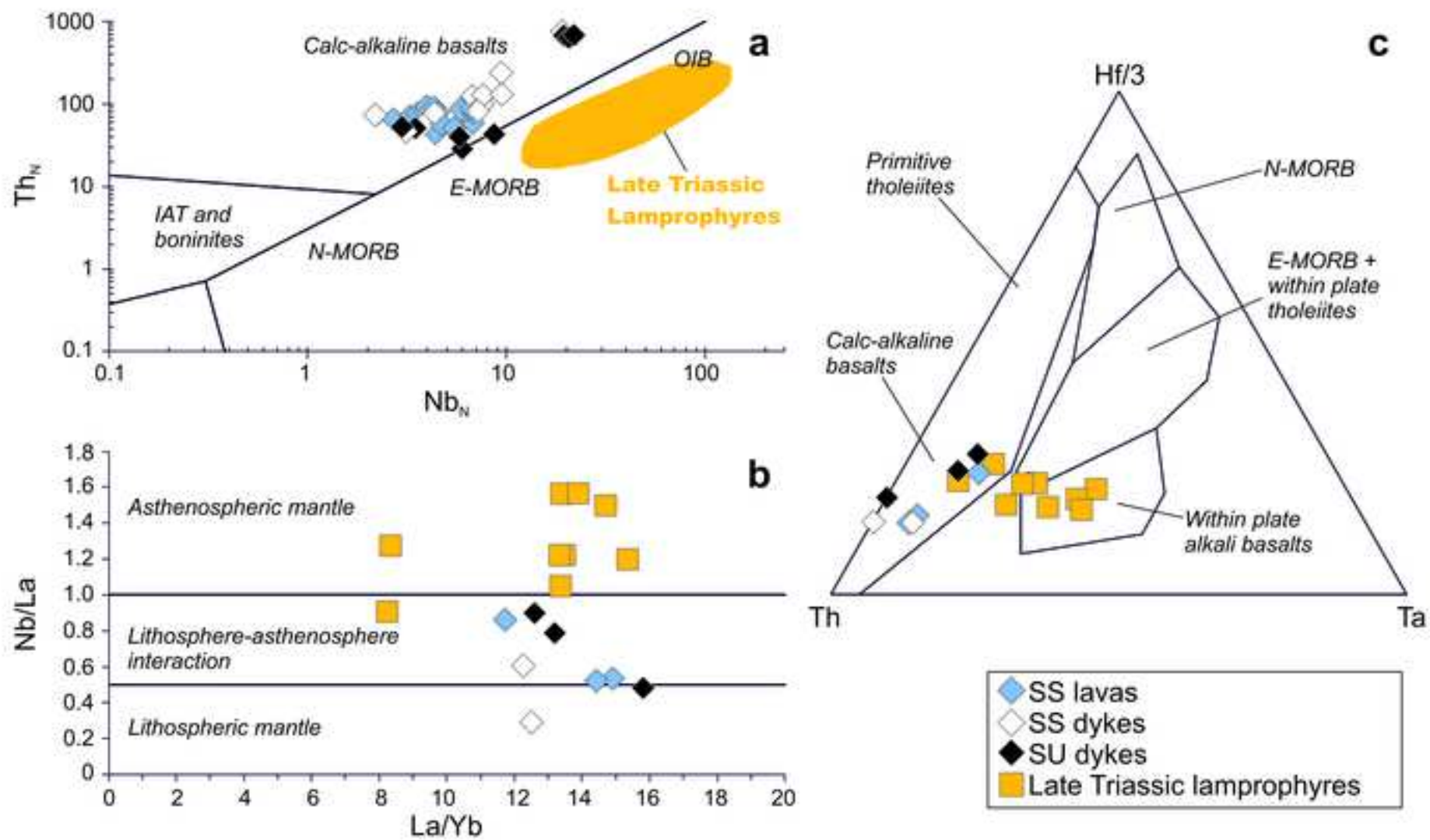


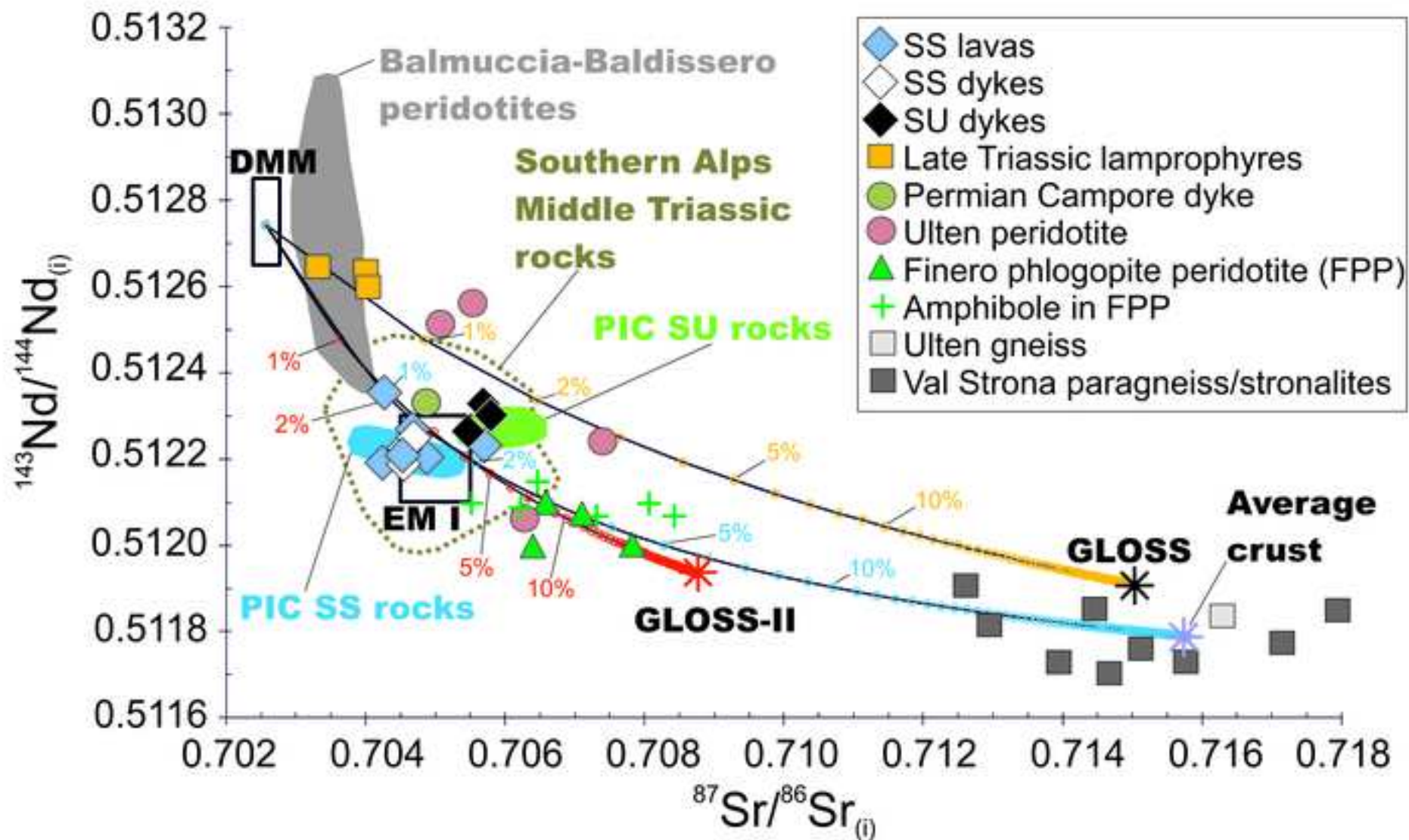


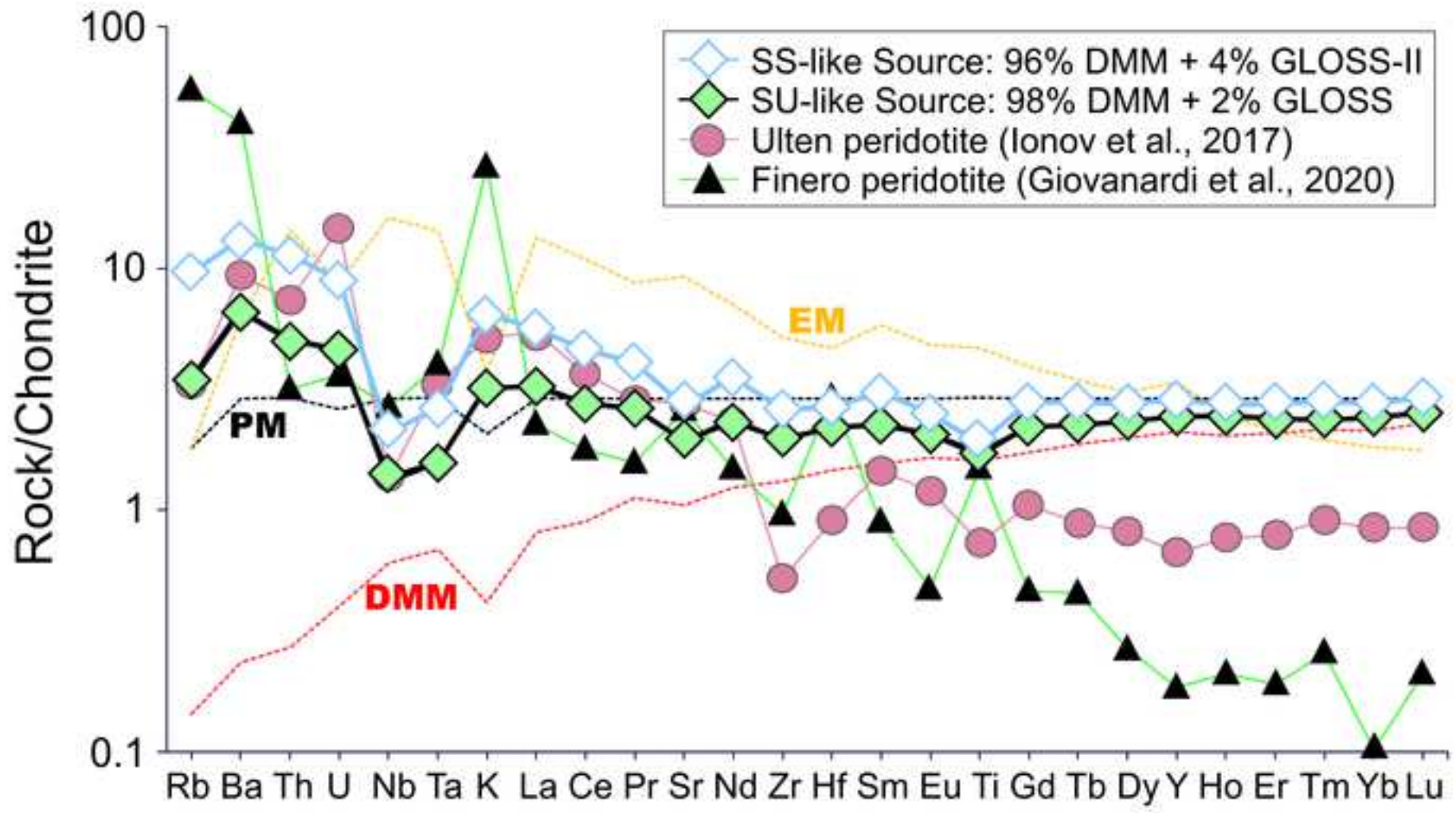


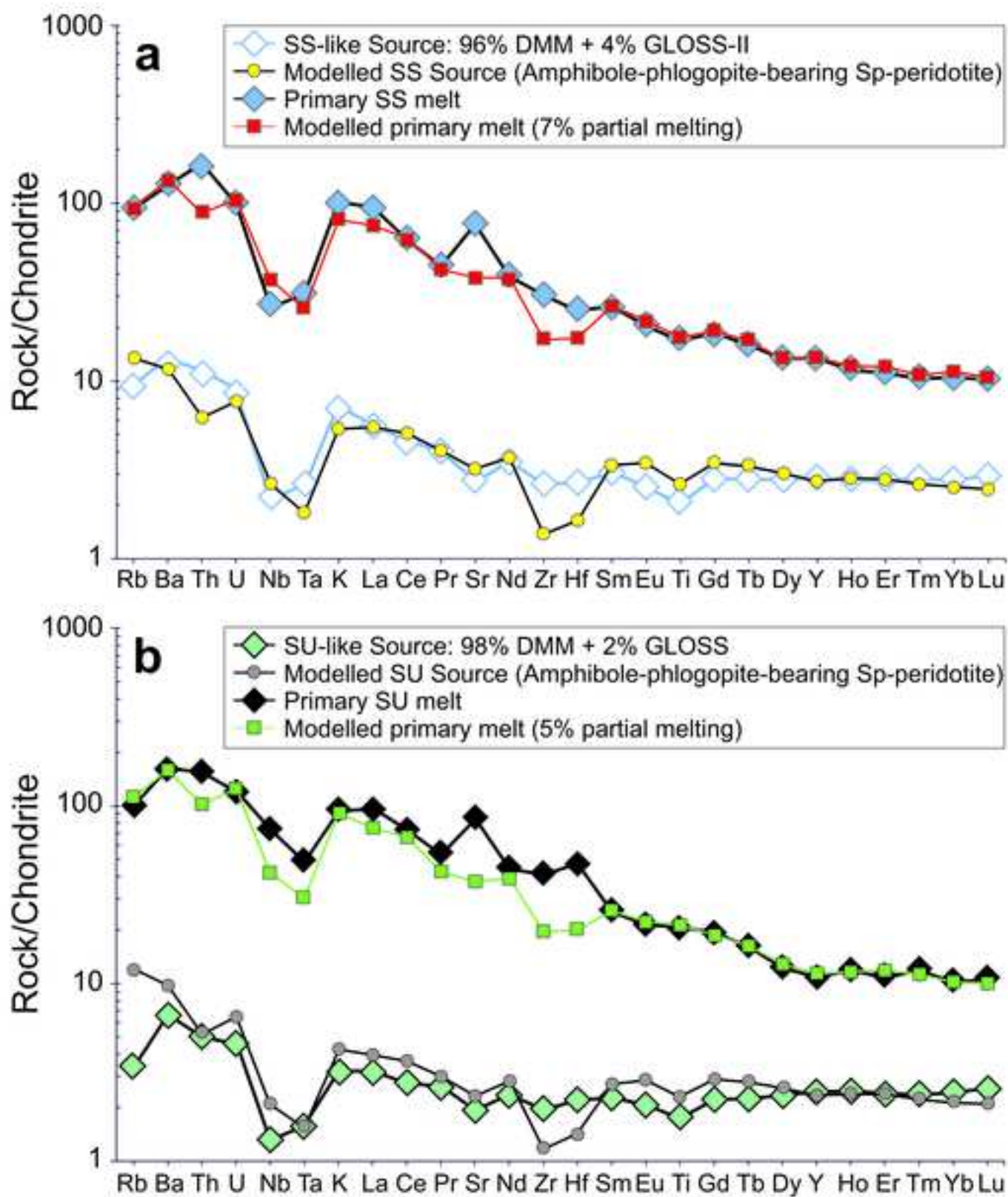


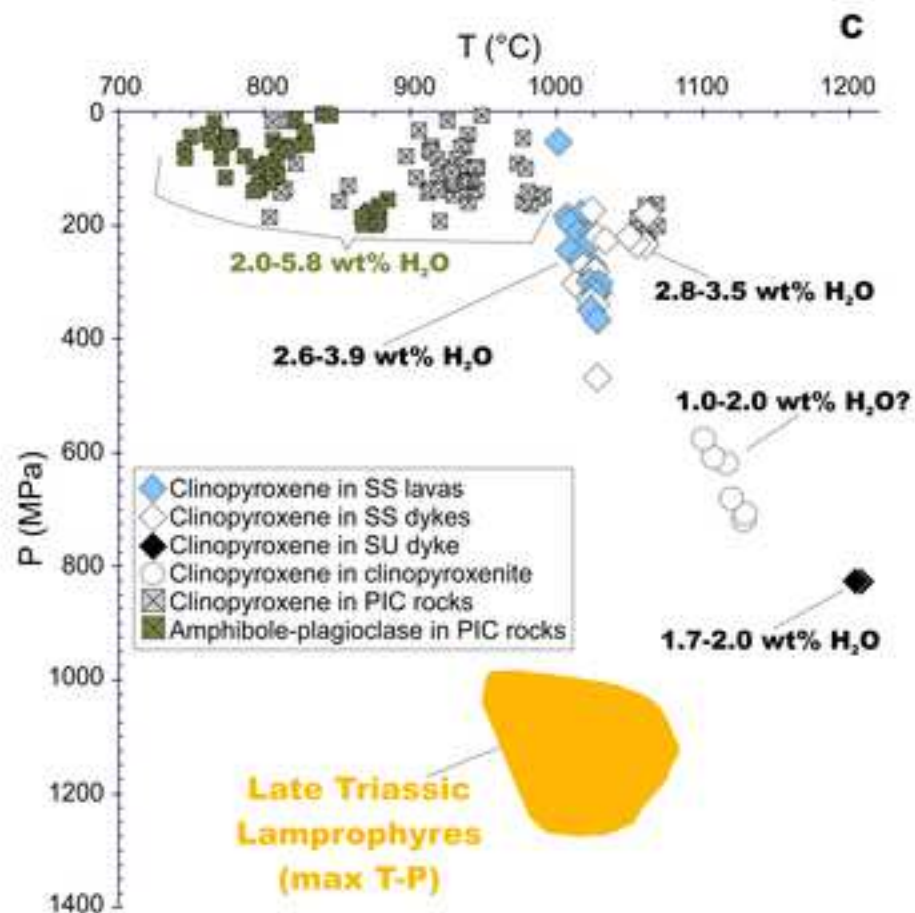
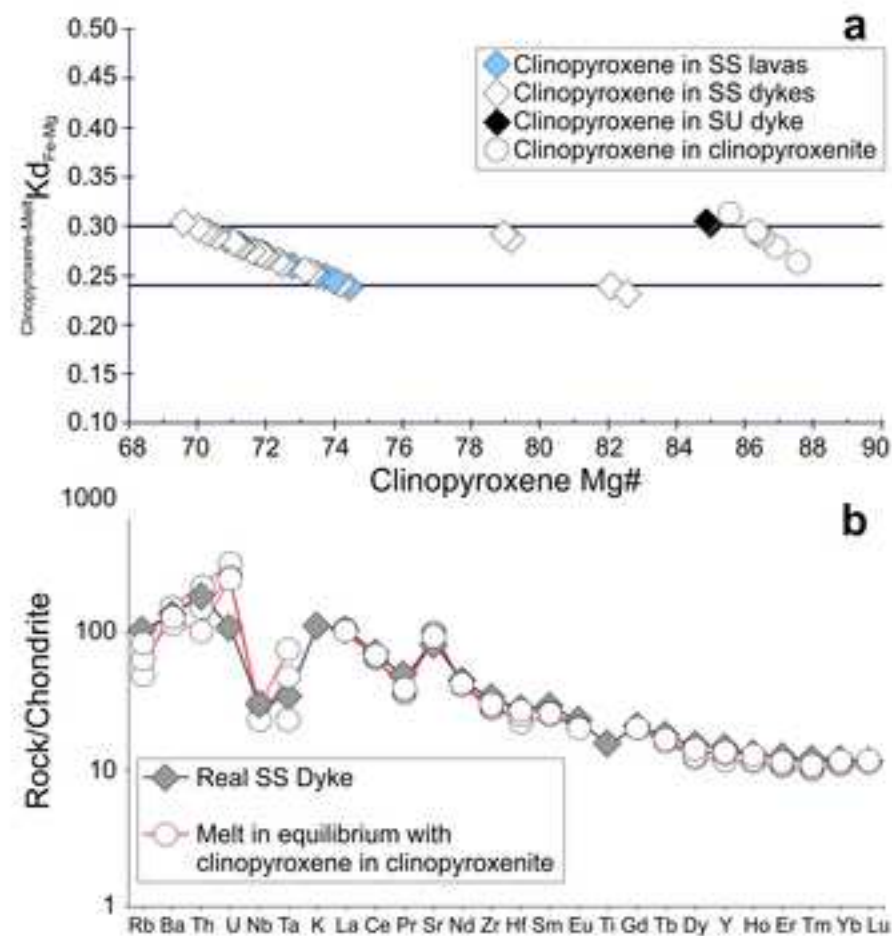












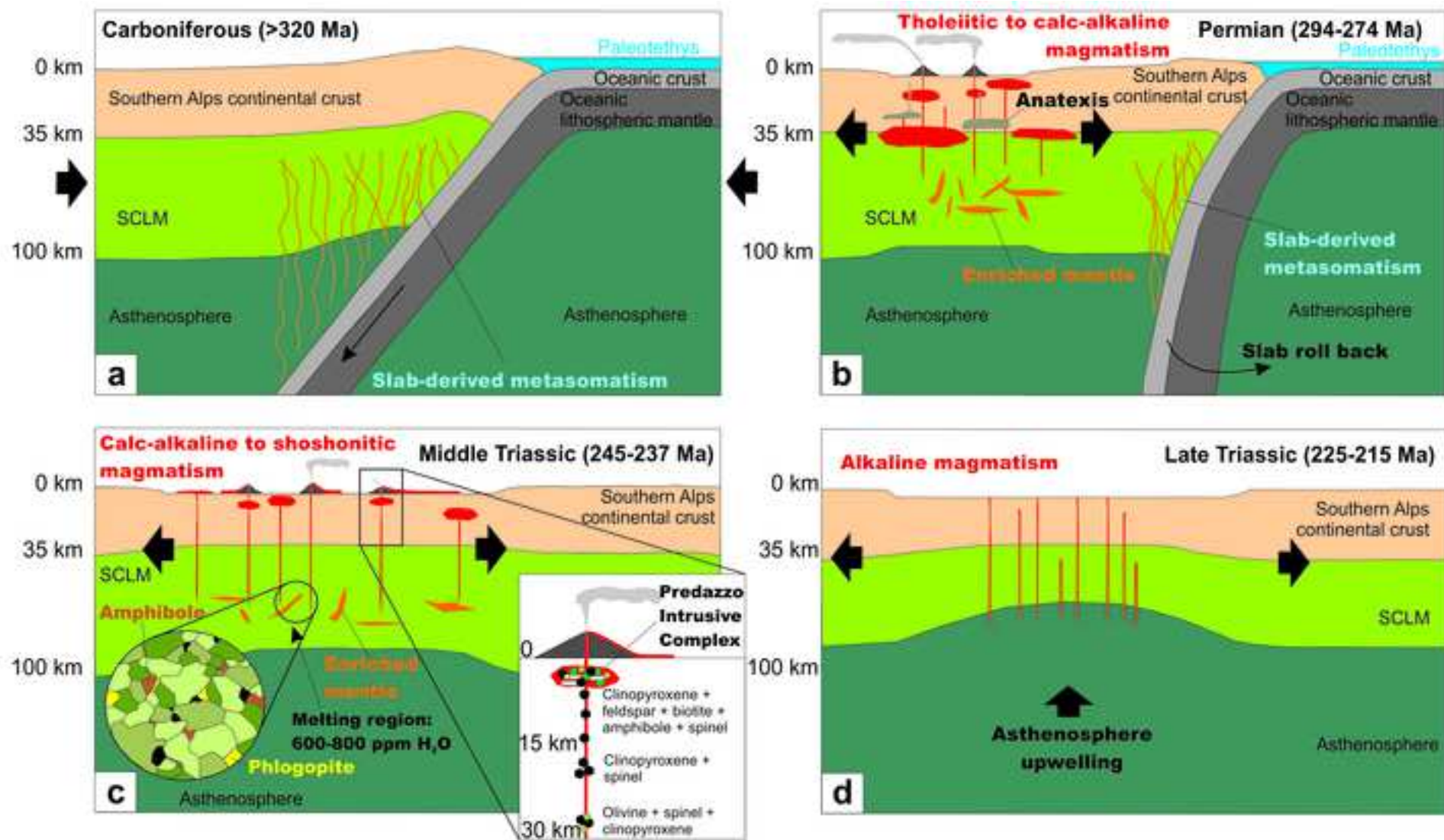


Table 1 Whole rock major, trace element composition and $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ isotopes of representative lavas, dykes and ultramafic cumulates from the Predazzo area (Dolomites, Southern Alps). Fe_2O_3 and FeO were calculated by considering a $\text{Fe}_2\text{O}_3/\text{FeO}$ ratio of 0.15, in agreement with a $f\text{O}_2$ around FMQ buffer (Kress and Carmichael, 1991). $\text{Mg}\# = \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ mol%; n.d. = not detected; LOI = Loss On Ignition. All trace element (ppm) were analysed by ICP-MS except Pb, Zn, Ni, Co, Cr, V and Ba (XRF). The trace element composition of samples labelled with (*) was entirely determined by XRF. Sr-Nd isotopic ratios were corrected for 238 Ma of radiogenic ingrowth using the trace element abundances determined by ICP-MS, the decay rates of Rotenberg et al. (2012) and Lugmair and Marti (1978). Initial ratios (i) and uncertainties (2σ) were propagated according to Ickert (2013). P.I. = Porphyricity Index; SS = Silica-saturated; SU = Silica-undersaturated; Tr-bas = Trachybasalt; Bas tr-and = Basaltic trachyandesite; Tr-and = Trachyandesite; Tra = Trachyte; Cpxite = Clinopyroxenite.

Lithology	Tr-bas	Tr-bas	Tr-bas	Bas tr-and	Bas tr-and	Bas tr-and	Bas tr-and	Bas tr-and	Bas tr-and	Bas tr-and	Tr-and	Tr-and	Tr-and	Tr-bas	Tr-bas	Tr-bas
Type	Lava	Lava	Lava	Lava	Lava	Lava	Lava	Lava	Lava	Lava	Lava	Lava	Lava	Dyke	Dyke	Dyke
Suite	SS	SS	SS	SS	SS	SS	SS	SS	SS	SS	SS	SS	SS	SS	SS	SS
P.L. (%)	45-50	25-30	40-45	5-10	55-60	30-35	30-35	35-40	35-40	25-30	20-25	20-25	25-30	40-45	25-30	
Sample	FC39	A43 (*)	A54 (*)	A11 (*)	FC7	A12	FC9	FC10 (*)	FC43 (*)	FC14	FC44	A22 (*)	A46	A75 (*)	EM93	
SiO₂ (wt%)	50.25	51.42	48.56	54.12	51.41	53.07	53.06	52.69	52.97	59.30	57.15	57.28	48.13	48.98	47.33	
TiO₂	1.27	0.86	1.17	0.86	0.85	0.79	0.79	1.18	0.82	0.49	0.57	0.51	1.11	0.83	1.08	
Al₂O₃	16.59	16.36	16.55	16.53	15.15	17.19	14.40	16.50	15.60	18.36	18.23	19.35	17.75	14.38	18.57	
Fe₂O₃	1.22	1.26	1.24	1.23	1.15	1.12	1.19	1.12	1.09	0.62	0.83	0.69	1.16	1.36	1.10	
FeO	8.10	8.39	8.33	8.24	7.68	7.48	7.94	7.49	7.29	4.14	5.51	4.58	7.71	9.08	7.36	
MnO	0.15	0.22	0.17	0.18	0.17	0.21	0.17	0.17	0.23	0.11	0.09	0.19	0.14	0.19	0.12	
MgO	7.53	5.87	8.40	5.11	6.62	3.99	6.38	7.06	5.22	1.70	2.19	1.82	8.18	7.60	5.58	
CaO	9.45	8.30	10.63	5.65	9.63	7.75	8.94	8.75	8.13	5.75	6.61	5.85	10.41	11.22	8.58	
Na₂O	2.26	2.12	1.41	3.53	2.22	2.68	2.76	2.73	2.82	2.90	3.18	2.76	1.72	1.76	1.97	
K₂O	1.57	3.26	1.61	3.13	3.23	3.24	3.19	2.45	3.80	5.26	4.52	4.42	1.62	2.45	3.32	
P₂O₅	0.32	0.52	0.54	0.40	0.37	0.46	0.33	0.34	0.43	0.43	0.44	0.33	0.51	0.49	0.46	
LOI	1.81	1.18	1.67	1.25	0.61	1.26	0.55	0.65	1.53	0.79	0.87	0.79	2.04	0.70	3.82	
Total	100.52	99.76	100.28	100.23	99.09	99.24	99.70	101.13	99.93	99.85	100.19	98.57	100.48	99.04	99.29	
Mg#	62.3	55.5	64.2	52.5	60.6	48.7	58.9	62.7	56.1	42.2	41.4	41.5	65.4	59.8	57.4	
Pb (ppm)	15.8	24.7	6.40	15.5	14.1	36.6	16.80	12.6	14.8	30.0	19.3	31.7	7.40	7.9	5.2	
Zn	90.5	132.5	86.7	117	70.1	213	70.5	73.9	99.1	64	52.6	170.3	73.6	96.1	51.2	
Ni	31.4	19.7	89.4	12.6	32.5	15.8	23.9	45.3	12.9	3.30	5.30	6.50	77.3	38.5	35.3	
Co	31.6	29.7	38.1	35.1	35.0	28.8	31.3	30.6	24.6	33.7	7.8	10.6	30.3	53.5	25.6	
Cr	45.8	82.4	201	17.2	101.7	44.5	166	94.8	117	0.80	4.70	8.40	128	238	49.8	
V	255	192	262	207	258	202	231	212	193	62	92	82.4	225	292	221	
Rb	12.4	115	38.2	121	41.7	106	43.3	70.3	122	136	119	143	36.5	65.2	146	
Ba	243	556	395	615	455	518	505	393	556	676	621	679	331	531	620	
Sr	541	722	597	608	513	682	550	712	689	793	813	1012	613	611	855	
Nb	10.0	7.9	6.4	7.8	10.0	13.6	10.9	12.9	13.9	16.6	15.4	14.4	7.30	5.20	16.0	
Zr	76.7	148	158	145	132	139	143	173	150	267	195	206	130	116	184	
Hf	2.49				3.54	4.12	3.58			5.83	5.23		2.96		4.26	
U	0.75				1.72	2.43	1.80			1.95	2.65		0.87		1.99	
Th	2.18	6.70	7.60	8.20	5.70	10.5	6.60	6.00	10.30	10.20	11.90	9.70	5.40	8.60	8.53	
Y	11.4	26.8	24.5	24.0	14.7	23.2	16.1	27.4	31.7	19.7	22.8	26.7	23.0	23.7	30.7	
Ta	0.48				0.54	0.75	0.57			0.80	0.76		0.48		n.d.	
La	11.6	27.9	26.5	18.2	18.9	28.4	20.3	28.7	47.7	27.0	32.7	58.1	24.9	6.00	26.5	
Ce	29.6	62.5	51.3	61.2	39.3	55.6	41.3	49.0	71.1	71.4	81.3	70.1	43.2	50.9	51.5	
Pr	3.31				4.83	6.89	5.15			6.18	7.11		4.59		n.d.	
Nd	14.63				20.47	28.28	21.64			24.94	28.39		20.4		30.2	
Sm	3.06				4.13	5.49	4.32			4.66	5.23		4.47		6.23	
Eu	0.89				1.07	1.26	1.05			1.22	1.25		1.32		1.71	
Gd	2.76				3.68	4.86	3.82			4.23	4.74		4.29		5.62	
Tb	0.43				0.56	0.75	0.59			0.63	0.70		0.67		0.91	
Dy	2.09				2.69	3.58	2.76			3.06	3.32		3.80		4.46	
Ho	0.43				0.55	0.73	0.56			0.64	0.69		0.74		0.92	
Er	1.11				1.44	1.94	1.48			1.76	1.86		2.03		2.42	
Tm	0.18				0.24	0.32	0.24			0.30	0.31		0.30		0.40	
Yb	0.98				1.31	1.79	1.36			1.71	1.79		1.99		2.16	
Lu	0.15				0.21	0.27	0.21			0.27	0.28		0.29		0.32	
$^{87}\text{Sr}/^{86}\text{Sr}$	0.704490				0.705460	0.707224	0.705615			0.705941	0.705990					0.706349
$^{87}\text{Rb}/^{86}\text{Sr}$	0.0662				0.2352	0.4502	0.2277			0.4971	0.4246					0.4951
$^{87}\text{Sr}/^{86}\text{Sr}(i)$	0.704270				0.704677	0.705725	0.704857			0.704286	0.704576					0.704700
2σ	0.000020				0.000017	0.000027	0.000019			0.000020	0.000027					0.000025
$^{143}\text{Nd}/^{144}\text{Nd}$	0.512548				0.512465	0.512424	0.512392			0.512363	0.512392					0.512447
$^{147}\text{Sm}/^{144}\text{Nd}$	0.1264				0.1221	0.1173	0.1207			0.1128	0.1114					0.1245
$^{143}\text{Nd}/^{144}\text{Nd}(i)$	0.512351				0.512275	0.512241	0.512204			0.512187	0.512218					0.512253
2σ	0.000006				0.000051	0.000006	0.000007			0.000019	0.000007					0.000006

Table 1 (continued)

Lithology	Bas tr-and	Bas tr-and	Tr-and	Tr-and	Tra	Tra	Tra	Tr-bas	Tr-bas	Tr-bas	Tr-bas	Tra	Tra	Cpxite
Type	Dyke	Dyke	Dyke	Dyke	Dyke	Dyke	Dyke	Dyke	Dyke	Dyke	Dyke	Dyke	Dyke	Xenolith
Suite	SS	SS	SS	SS	SS	SS	SS	SU	SU	SU	SU	SU	SU	SS
P.L. (%)	25-30	5-10	50-55	5-10	10-15	35-40	15-20	5-10	45-50	25-30	45-50	5-10	20-25	-
Sample	EM80A	P96B (*)	EM35 (*)	FC41A	A23 (*)	A96	FC54	CT55	EM51 (*)	A61	A47 (*)	EM104	EM11	CT8
SiO ₂ (wt%)	53.78	51.75	56.61	57.46	62.55	58.82	60.67	46.75	47.76	48.37	49.54	52.84	56.80	51.67
TiO ₂	0.88	0.85	1.08	0.71	0.60	0.54	0.43	1.44	1.28	1.16	1.19	0.14	0.44	0.24
Al ₂ O ₃	16.38	16.27	17.81	16.39	16.71	16.64	18.14	16.68	16.52	18.96	20.12	21.54	20.58	2.51
Fe ₂ O ₃	1.29	1.18	0.89	0.97	0.69	0.67	0.50	1.04	1.18	1.07	1.01	0.24	0.43	0.70
FeO	8.58	7.86	5.88	6.50	4.57	4.48	3.35	6.93	7.86	7.14	6.78	1.60	2.84	4.70
MnO	0.18	0.20	0.12	0.16	0.11	0.08	0.24	0.13	0.19	0.15	0.13	0.12	0.13	0.12
MgO	3.96	3.97	2.82	3.19	1.79	1.55	2.70	6.79	7.39	5.90	3.92	0.70	0.50	18.52
CaO	7.59	8.64	6.00	3.93	1.35	3.63	3.36	11.80	10.81	7.49	9.03	3.96	3.35	19.51
Na ₂ O	2.46	2.08	3.09	2.68	2.54	3.04	3.63	3.72	2.32	3.12	2.79	4.52	4.89	0.20
K ₂ O	3.54	4.15	3.80	5.92	7.05	6.32	5.45	1.43	1.79	2.42	2.20	6.73	6.08	0.14
P ₂ O ₅	0.39	0.39	0.47	0.51	0.17	0.22	0.28	0.46	0.59	0.40	0.54	0.00	0.07	0.14
LOI	1.73	1.12	1.00	2.38	1.33	3.07	2.59	3.67	1.57	3.78	2.99	7.06	2.98	1.62
Total	100.76	98.46	99.57	100.80	99.46	99.06	101.34	100.84	99.26	99.96	100.24	99.45	99.09	100.07
Mg#	45.1	47.4	46.0	46.6	41.1	38.2	58.9	63.6	62.6	59.6	50.8	43.7	23.7	87.5
Pb (ppm)	22	17.89	15.80	31.30	19.00	10.90	37.10	5.00	13.70	8.80	10.00	183.00	36.50	6
Zn	130	122	81.2	96.6	77.8	66.5	163	64.0	91.9	130	71.2	551	84.4	25.0
Ni	18.9	34.3	7.60	2.10	4.90	3.80	2.70	62.9	54.7	28.9	19.2	5.00	5.20	104
Co	36.3	33.6	16.50	16.90	6.60	10.3	6.50	36.7	30.7	27.3	21.5	1.00	3.00	41.0
Cr	53.6	59.99	14.00	2.30	4.00	3.00	8.20	266	122	32.0	25.4	2.00	3.50	897
V	205	192	144	97.9	37.3	38.4	39.0	175	274	213	221	12.0	43.8	95.0
Rb	119	116	127	162	196	155	158	39.0	68.6	77.7	58.1	286	344	3.24
Ba	577	678	684	861	849	763	566	434	531	498	473	184	444	30.0
Sr	689	819	685	635	476	404	458	684	631	989	746	421	656	93.78
Nb	10.0	n.d.	17.9	15.9	18.0	18.0	22.0	20.2	7.10	13.7	13.9	49.0	50.6	2.16
Zr	145	169	225	209	278	259	297	177	145	110	218	561	407	24.95
Hf				5.57		n.d.	7.31	5.64		3.37	4.89	7.24	3.67	0.57
U				3.13		n.d.	8.27	1.07		1.07	1.11	20.3	10.8	0.08
Th	9.60	10.2	11.9	14.5	15.0	15.0	28.2	5.02	6.50	4.77	3.59	75.0	81.9	0.24
Y	27.8	25.5	31.2	29.3	33.7	32.1	22.2	19.2	19.1	25.8	14.1	20.0	29.9	6.12
Ta				0.76		1.43	1.13	0.75		n.d.	0.69	n.d.	2.26	0.24
La	41.3	50.4	32.8	40.9	43.2	39.8	24.5	25.6	14.2	28.5	15.4	143	137	2.62
Ce	60.8	75.6	77.3	96.3	103.5	96.0	61.7	49.4	47.2	49.2	34.2	189	209	5.98
Pr				8.96		8.87	7.19	5.64		n.d.	3.88	14.1	17.0	0.90
Nd				36.8		35.1	28.9	22.7		29.6	16.5	38.0	54.8	4.49
Sm				6.84		6.66	5.67	4.33		5.88	3.29	4.43	7.79	1.26
Eu				1.67		1.60	1.37	1.38		1.71	0.94	1.03	1.59	0.34
Gd				6.14		5.97	4.91	4.41		5.23	3.09	5.49	7.50	1.19
Tb				0.91		0.83	0.79	0.68		0.83	0.48	0.58	0.94	0.20
Dy				4.34		4.50	3.94	3.52		3.97	2.33	3.05	4.21	1.07
Ho				0.89		0.87	0.83	0.74		0.81	0.49	0.62	0.88	0.21
Er				2.42		2.53	2.34	2.02		2.09	1.30	1.94	2.50	0.54
Tm				0.40		0.38	0.41	0.33		0.33	0.22	0.35	0.43	0.08
Yb				2.29		2.65	2.47	1.94		1.80	1.22	2.69	2.54	0.49
Lu				0.36		0.40	0.39	0.30		0.27	0.19	0.41	0.37	0.07
⁸⁷ Sr/ ⁸⁶ Sr				0.707044		0.708271	0.708238			0.706576	0.706242	0.712254		0.704715
⁸⁷ Rb/ ⁸⁶ Sr				0.7396		1.1118	0.9970			0.2272	0.2252	1.9677		0.0999
⁸⁷ Sr/ ⁸⁶ Sr(i)				0.704581		0.704568	0.704918			0.705820	0.705492	0.705702		0.704382
2σ				0.000033		0.000053	0.000036			0.000016	0.000022	0.000067		0.000020
¹⁴³ Nd/ ¹⁴⁴ Nd				0.512405		0.512373	-			0.512484	0.512445	0.512430		-
¹⁴⁷ Sm/ ¹⁴⁴ Nd				0.1124		0.1147	-			0.1200	0.1207	0.0705		-
¹⁴³ Nd/ ¹⁴⁴ Nd(i)				0.512230		0.512194	-			0.512297	0.512257	0.512320		-
2σ				0.000006		0.000008	-			0.000007	0.000006	0.000006		-

Table 2 Major and trace element composition of the calculated SS and SU melts in equilibrium with a fertile mantle assemblage. For each magmatic suite, the starting sample, the amount of backward fractionation (F%) required to reach the primitive composition, the percentage of each fractionated mineral phase and its compositional interval are also indicated. Fo = forsterite; Mg# = Mg/[Mg+Fe²⁺] mol%; Cr# = Cr/[Cr+Al] mol%.

Calculated primitive melts					
Suite	SS	SU	Suite	SS	SU
Composition	Transitional basalt	Alkali basalt	Starting sample	A46	CT55
SiO₂ (wt%)	47.71	47.25	Backward F%	11.4	10.5
TiO₂	1.26	1.53	Olivine	6.9	6.6
Al₂O₃	16.14	15.64	Clinopyroxene	3.0	2.4
Fe₂O₃	1.28	1.12	Spinel	1.5	1.5
FeO	8.56	7.46			
MnO	0.16	0.15			
MgO	11.23	9.89			
CaO	10.07	11.50			
Na₂O	1.55	3.45			
K₂O	1.46	1.33			
P₂O₅	0.46	0.43			
Cr₂O₃	0.09	0.23			
NiO	0.02	0.02			
Total	100.00	100.00			
Mg#	70.0	70.3			
Rb (ppm)	32.4	34.9			
Ba	293.6	388			
Th	4.8	4.49			
U	0.8	0.96			
Nb	6.5	18.1			
Ta	0.4	0.67			
La	22.1	22.9			
Ce	38.4	44.3			
Pr	4.1	5.06			
Sr	545.2	614			
Nd	18.2	20.4			
Zr	115.9	159			
Hf	2.6	5.07			
Sm	4.0	3.88			
Eu	1.2	1.24			
Gd	3.8	3.97			
Tb	0.6	0.61			
Dy	3.4	3.17			
Y	20.5	17.2			
Ho	0.7	0.67			
Er	1.8	1.82			
Tm	0.3	0.30			
Yb	1.8	1.74			
Lu	0.3	0.27			

Mineral phase compositional interval	
Olivine	Fo = 85-92
Clinopyroxene	Mg# = 86-89
Spinel	Mg# = 7; Cr# = 49

Table 3 Major/trace element and modal composition of the mantle sources modelled by using real mineral phases from the Finero and Baldissero peridotites compared to those of the ideal sources inferred from the Sr-Nd isotopes mixing models (DMM + GLOSS/GLOSS-II; Workman and Hart, 2005; Plank, 2014). The major/trace element composition of the fertile mineral phases from Finero (olivine, orthopyroxene, spinel, amphibole and phlogopite) and Baldissero (clinopyroxene) peridotites used to reconstruct the composition of the SS/SU sources are also reported. The H₂O concentration in the modelled sources, obtained by weighting the H₂O amounts in each mineral phase (according to the estimates made on the Finero peridotite by Tommasi et al., 2017) are also indicated. Olivine, orthopyroxene, spinel and phlogopite compositions are from Giovanardi et al. (2020); clinopyroxene composition is from Mazzucchelli et al. (2010); amphibole composition is from Morishita et al. (2008). * = concentration of trace elements speculated from adjacent elements; ** = U-Th concentrations speculated by considering the maximum amounts reported for clinopyroxene from the Southern Alps orogenic massifs (i.e. Ulten peridotite; Ionov et al., 2017).

	Mantle sources				Mineral phase major/trace element compositions and modal amounts							
	Modelled using real mineral phases		Resulted from mixing models based on Sr-Nd isotopes		Modal amount (SS source)	Modal amount (SU source)	Olivine	Orthopyroxene	Clinopyroxene	Spinel	Amphibole	Phlogopite
	SS source	SU source	SS source (0.96 DMM + 0.04 GLOSS-II)	SU source (0.98 DMM + 0.02 GLOSS)								
SiO₂ (wt%)	46.05	45.94	-	-	40.80	57.51	55.40	23.01	17.71	1.14	1.98	0.76
TiO₂	0.20	0.17	-	-	0.00	0.00	0.00	0.12	0.82	0.15	0.42	1.74
Al₂O₃	2.25	1.99	-	-	0.00	0.00	0.00	1.49	6.81	30.14	11.06	15.86
FeO_{TOT}	7.53	7.72	-	-	9.27	9.27	9.27	6.67	2.66	25.06	3.60	2.98
MnO	0.14	0.14	-	-	0.16	0.16	0.16	0.15	0.09	0.19	0.00	0.00
MgO	38.35	39.30	-	-	49.20	49.20	49.20	34.05	14.11	11.30	19.00	23.52
CaO	4.23	3.58	-	-	0.01	0.01	0.01	0.35	21.98	0.00	12.31	0.00
Na₂O	0.30	0.25	-	-	0.00	0.00	0.00	0.00	1.39	0.00	2.36	1.28
K₂O	0.08	0.07	-	-	0.00	0.00	0.00	0.00	0.00	0.00	0.74	8.29
Cr₂O₃	0.64	0.61	-	-	0.01	0.01	0.01	0.29	0.82	33.41	1.93	0.57
NiO	0.24	0.23	-	-	0.38	0.38	0.38	0.07	0.05	0.00	0.00	0.00
Total	100	100	-	-	99.83	99.83	99.83	99.97	100.05	100.25	96.70	93.65
Rb (ppm)	4.66	4.19	3.40	1.19	-	-	-	7.46	0.04*	-	6.20	370
Ba	28.04	23.80	31.98	16.07	-	-	-	29.7	0.05	-	276	2071
Th	0.18	0.15	0.33	0.15	-	-	-	0.09	0.59**	-	2.90	0.00
U	0.06	0.05	0.07	0.04	-	-	-	0.01	0.29**	-	0.42	0.00
Nb	0.65	0.52	0.52	0.32	-	-	-	0.13	2.02	-	11.0	5.46
Ta	0.03	0.02	0.04	0.02	-	-	-	0.00	0.10*	-	0.18	0.48
La	1.31	0.93	1.35	0.76	-	-	-	0.03	1.00	-	57.0	0.03
Ce	3.11	2.24	2.83	1.69	-	-	-	0.07	3.41	-	126	0.05
Pr	0.38	0.28	0.39	0.25	-	-	-	0.01	0.70*	-	13.0	0.01
Sr	23.53	16.89	19.44	14.05	-	-	-	1.39	17.04	-	977	112
Nd	1.72	1.31	1.66	1.11	-	-	-	0.05	4.37	-	47.0	0.00
Zr	5.34	4.50	10.04	7.58	-	-	-	1.05	25.22	-	32.0	0.33
Hf	0.18	0.15	0.29	0.24	-	-	-	0.12	0.75*	-	0.85	0.05
Sm	0.51	0.42	0.47	0.35	-	-	-	0.04	2.04	-	7.20	0.00
Eu	0.20	0.17	0.15	0.12	-	-	-	0.01	0.91	-	1.90	0.04
Gd	0.72	0.60	0.58	0.46	-	-	-	0.02	3.41	-	5.50	0.03
Tb	0.13	0.10	0.10	0.08	-	-	-	0.01	0.60*	-	0.92	0.00
Dy	0.77	0.65	0.70	0.59	-	-	-	0.09	3.77	-	4.30	0.02
Y	4.34	3.64	4.53	3.86	-	-	-	0.31	21.34	-	25.0	0.01
Ho	0.16	0.13	0.15	0.14	-	-	-	0.01	0.80*	-	0.80	0.00
Er	0.46	0.39	0.46	0.40	-	-	-	0.07	2.27	-	2.30	0.00
Tm	0.07	0.06	0.07	0.06	-	-	-	0.02	0.32*	-	0.35	0.00
Yb	0.43	0.36	0.47	0.41	-	-	-	0.06	2.12	-	2.20	0.03
Lu	0.06	0.05	0.07	0.07	-	-	-	0.01	0.31*	-	0.32	0.00
H₂O (ppm)	784	598	-	-	3.00	3.00	3.00	111	267.00	-	19300	43000

Table 4 Major, trace element composition and H₂O amount of the SS/SU magmas obtained from partial melting models and mineral phases eutectic melting proportions. The minimization of the sum of squared residuals between the predicted values for each oxide and the values calculated from the melting models is expressed by the parameter S. Partial melting degrees are expressed as F (%).

Suite	Melting models results		F (%)	SS	SU
	SS	SU			
Composition	Transitional basalt	Alkali basalt		7	5
SiO₂ (wt%)	47.98	47.48	Eutectic melting proportions		
TiO₂	1.56	1.60	Olivine	-0.26	-0.11
Al₂O₃	16.58	16.05	Orthopyroxene	0.54	0.22
FeO_{TOT}	9.06	8.41	Clinopyroxene	0.32	0.52
MnO	0.04	0.08	Spinel	0.01	-0.02
MgO	11.50	10.24	Amphibole	0.28	0.27
CaO	10.01	11.64	Phlogopite	0.11	0.12
Na₂O	1.66	3.00		1.00	1.00
K₂O	1.17	1.31			
Cr₂O₃	0.39	0.16			
NiO	0.05	0.03			
Total	100.00	100.00			
S (should be <1)	0.98	0.44			
Rb (ppm)	33.0	38.9			
Ba	329.3	382.8			
Th	2.56	2.96			
U	0.85	1.00			
Nb	9.13	10.32			
Ta	0.36	0.42			
La	17.8	17.7			
Ce	38.8	40.9			
Pr	4.03	4.04			
Sr	278.4	272.7			
Nd	17.8	18.3			
Zr	66.2	76.0			
Hf	1.86	2.13			
Sm	4.03	3.91			
Eu	1.27	1.28			
Gd	3.93	3.84			
Tb	0.63	0.61			
Dy	3.42	3.26			
Y	21.4	17.9			
Ho	0.69	0.66			
Er	2.00	1.96			
Tm	0.28	0.29			
Yb	1.92	1.74			
Lu	0.27	0.25			
H₂O (wt%)	1.06	1.12			

Table 5 *P*, *T* and H₂O conditions of crystallization of SS and SU magmas recorded by mineral-melt thermobarometric/hygrometric equations. For each adopted method, the chosen equation, corresponding reference and associated error on estimates are reported together with the results interval. The temperature of olivine crystallization was calculated by applying the Putirka et al. (2007) equation. *T* and *P* of clinopyroxene crystallization were obtained by means of the mineral-melt equations of Putirka et al. (2003) and Putirka (2008). The water content of the melt during clinopyroxene crystallization was calculated using the Perinelli et al. (2016) hygrometer. This latter equation was iterated with the Putirka (2008) methods to unveil the T-P-H₂O conditions of crystallization of SS/SU magmas.

Sample	Suite	Method	Phenocryst composition	Reference	T (°C)	Error (°C)	P (MPa)	Error (MPa)	H ₂ O (wt%)	Error (wt%)
CT55 (trachybasalt)	SU	Olivine-Liq.	Fo 85	Putirka et al. (2007), Eq. 4	1195	±29	-	-	1.5 (input)	-
CT55 (trachybasalt)	SU	Clinopyroxene-Liq.	Mg# 85	Putirka (2008), Eq. 32b/33; Perinelli et al. (2016)	1204-1207	±42	827-829	±260	1.7-2.0	±0.45
CT8 (clinopyroxenite)	SS	Clinopyroxene-Liq.	Mg# 86-88	Putirka et al. (2003); Putirka (2008), Eq. 33	1099-1128	±42	572-714	±170	1.3 (input)	-
A75 (trachybasalt)	SS	Clinopyroxene-Liq.	Mg# 79-83	Putirka (2008), Eq. 32b/33; Perinelli et al. (2016)	1051-1061	±42	181-235	±260	2.8-3.5	±0.45
A75 (trachybasalt)	SS	Clinopyroxene-Liq.	Mg# 70-74	Putirka (2008), Eq. 32b/33; Perinelli et al. (2016)	1012-1035	±42	169-467	±260	2.9-3.8	±0.45
FC39 (trachybasalt)	SS	Clinopyroxene-Liq.	Mg# 71-74	Putirka (2008), Eq. 32b/33; Perinelli et al. (2016)	1016-1028	±42	181-325	±260	2.6-3.2	±0.45
FC7 (basaltic trachyandesite)	SS	Clinopyroxene-Liq.	Mg# 72-74	Putirka (2008), Eq. 32b/33; Perinelli et al. (2016)	1009-1025	±42	246-330	±260	3.4-3.8	±0.45
FC9 (basaltic trachyandesite)	SS	Clinopyroxene-Liq.	Mg# 70-74	Putirka (2008), Eq. 32b/33; Perinelli et al. (2016)	1010-1028	±42	200-368	±260	3.2-3.7	±0.45
FC41 (trachyandesite)	SS	Clinopyroxene-Liq.	Mg# 74	Putirka (2008), Eq. 32b/33; Perinelli et al. (2016)	1001-1008	±42	49-184	±260	3.2-3.8	±0.45



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**Supplementary material/Appendix (Files for online
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Supplementary Material.xlsx



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