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1 Geochemical characteristics of basalts related to incipient oceanisation: the example from the
2 Alpine-Tethys OCTs

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9
10 **Abstract**

11 Basalts exposed in the Platta and Tasna nappes (SE Switzerland) derive from the Alpine-Tethys Ocean-
12 Continent Transitions (OCT) and overlie SubContinental Lithospheric Mantle (SCLM). We show that the trace
13 element signatures of these basalts differ from Mid-Ocean Ridge Basalts (MORB). Two types of basalts occur
14 in the OCT: a type-1 showing a “garnet signature” that can be modeled by the partial melting of the SCLM in
15 the spinel stability field and a type-2 characterized by an enrichment in incompatible elements that can be
16 explained by the mixing between garnet-pyroxenite-derived melts and the melting of either a depleted MORB
17 mantle or a refertilized SCLM. Based on geological and geochemical observations we propose that the basalts
18 from the Alpine-Tethys OCTs result from a poly-phase magmatic system that carry an inherited SCLM signature.
19 These basalts should therefore be referred to as OCT-basalts rather than as MOR-basalts.

20 *Keywords: Ocean-Continent Transition, Alpine-Tethys, basalt, inheritance, subcontinental mantle, OCT-*
21 *basalt.*

22 1. Introduction

23 The geochemical composition of mid-ocean ridge basalts (MORB) is due to the partial melting of a
24 depleted mantle source (DMM). One remaining key question is whether basalts from magma-poor ocean-
25 continent transitions (OCTs) are formed by similar melting processes as MORB or if they reflect more complex
26 mantle processes associated with lithospheric thinning of a sub-continental lithospheric mantle (SCLM).
27 Basaltic samples derived from present-day magma-poor OCTs are rare and limited to few samples drilled at
28 the Iberia-Newfoundland (Chazot et al., 2005; Cornen et al., 1999; Robertson, 2007), the South China Sea (Sun
29 et al., 2016) or the Diamantine OCT (Beslier et al., 2004; Chatin et al., 1998). In contrast, Alpine-Apennine
30 ophiolites expose remnants of the OCTs of the ancient magma-poor Jurassic Alpine Tethys, which provide
31 information not only on the basalts but also on the underlying mantle (Picazo et al., 2016). Previous studies on
32 basalts derived from the fossil Alpine-Tethys OCT show that they are characterized by a wide range of trace
33 element compositions, from depleted basalts (Bill et al., 2000; Desmurs et al., 2002; Durand-Delga et al., 1997;
34 Kramer et al., 2003; Montanini et al., 2008; Piccardo, 2008; Saccani et al., 2008) to enriched basalts (Frisch et
35 al., 1994; Renna et al., 2018; Steinmann and Stille, 1999; Vannucci et al., 1993) leading to various
36 interpretations concerning the parental sources of these basalts. Over the last decade, a considerable data set
37 has been acquired on the mantle rocks from Alpine-Tethys OCTs (Müntener et al., 2010; Picazo et al., 2016;
38 Piccardo, 2016; Rampone et al., 2008). This new knowledge, combined with the geological observations,
39 enable to better discriminate between the various interpretations and to understand the processes involved
40 in magma-production at deeper parts of the lithosphere during mantle exhumation and lithospheric breakup.

41 In this paper, we present a compilation of Alpine-Tethys basalt chemistry including new data from the
42 Platta and Tasna nappes (SE Switzerland, Fig. 1a). The lack of a major Alpine metamorphic overprint, the
43 excellent exposure and the understanding of the mantle evolution prior and during breakup of these two
44 nappes allowed to propose a tectono-magmatic model that may explain the peculiar geochemistry of the
45 basalts of the Alpine-Tethys OCT.

46 **2. Geological setting**

47 **2.1. Remnants of OCTs in the Alps**

48 The Tasna and Platta nappes represent well preserved remnants of a fossil OCT of the ancient European
49 and Adriatic conjugate rifted margins (Manatschal et al., 2006; Manatschal and Nievergelt, 1997) that were
50 related to the opening of the Jurassic Liguro-Piemonte/Alpine Tethys ocean (Fig. 1b). This ocean probably was
51 a narrow basin (<600km-wide) that developed as a result of an ultra-slow spreading activity (Marroni and
52 Pandolfi, 2007; Li et al., 2013). The Tasna nappe is part of the northeastern European OCT preserving a wedge
53 of continental crust that is laterally replaced by exhumed subcontinental mantle and grades into proto-oceanic
54 crust with first magmatic additions (Manatschal et al., 2006). The Penninic Platta nappe is subdivided into two
55 Alpine tectonic units, the Upper and the Lower Platta units (Supplementary material S1). The Upper Platta
56 nappe is bounded at its top by the Lower Austroalpine Err nappe (part of the Adria distal continental margin)
57 and overlies Middle and Lower Penninic nappes (remnant of the European distal margin; Manatschal and
58 Nievergelt, 1997). Both the Platta and Tasna nappes, as well as the majority of Alpine-Apennine ophiolites,
59 consist of highly serpentinized peridotites intruded by large gabbroic bodies and overlain locally by massive
60 basaltic lava flows, continent-derived extensional allochthons and post-rift sediments (supplementary material
61 S2, Desmurs, 2002; Desmurs et al., 2001; Peters et al., 2008).

62 **2.2. Nature of the exhumed mantle**

63 Two main types of mantle have been distinguished all around the Alpine-Tethys and more particularly
64 in the Tasna and Platta nappes (Fig. 2a, Picazo et al., 2016). The first type identified in the Tasna and Upper
65 Platta units consists of spinel lherzolites containing abundant garnet-bearing pyroxenites (Desmurs, 2002;
66 Müntener et al., 2010; Peters et al., 2008). Major and trace element compositions in clinopyroxene
67 demonstrate the fertile character of these mantle rocks (Picazo et al., 2016). This mantle, referred to as the
68 inherited subcontinental mantle, is similar to the mantle exhumed in the proximal parts of the present-day
69 Iberia-Newfoundland OCTs (Müntener and Manatschal, 2006).

70 The second type of mantle composing the lower Platta unit consists of spinel lherzolites with minor
71 harzburgites and dunites and shows only few occurrences of pyroxenites (Desmurs, 2002; Müntener et al.,
72 2010, 2004; Peters et al., 2008). This peridotite is pervasively impregnated by ascending tholeiitic melts
73 (Müntener et al., 2010, 2004) that pre-date the Middle Jurassic mantle exhumation (Piccardo, 2016; Rampone
74 et al., 1995). This mantle type has also been described from the distal part of the Iberia margin (Müntener and
75 Manatschal, 2006) and is considered to be the refertilized part of exhumed subcontinental mantle (see type 2
76 mantle in Picazo et al. 2016).

77 **3. Basalts from the Platta and Tasna OCTs**

78 **3.1. Petrology of extrusive magmatic rocks and basaltic intrusions**

79 Extrusive basaltic rocks exposed in the Platta and Tasna nappes (supplementary material S3) occur as
80 hyaloclastites, pillow breccias, pillow lavas and massive lava flows that can reach up to 300 m thickness (Fig.
81 2a-b). Basalts lie on exhumed mantle and are covered by Upper Jurassic to Lower Cretaceous marine sediments
82 (Fig. 2c). Basalts exhibit intersertal to porphyric textures and contain plagioclase phenocrysts embedded in a
83 fine-grained matrix of plagioclase, chlorite, epidote, pumpellyite and calcite (Fig. 2d). Hyaloclastite breccias
84 consist of fragments of lava flows, which probably derive from the dislocation of former volcanic deposits. A
85 Jurassic age (167-161 Ma) is based on the dating of radiolarian cherts that are either interlayered or overlie
86 the basalts (Bill et al., 2001) and U-Pb on gabbros that are genetically linked to the basalts (Schaltegger et al.,
87 2002).

88 A younger magmatic event consisting of sills of 10 to 20 m thickness (Fig. 2e) has been found in the Platta
89 nappe. These sills intrude limestones and shales of Lower Cretaceous age (Peters et al., 2008), representing
90 the youngest pre-Alpine magmatic event identified so far in the Alpine OCT. The sills are characterized by fine-
91 grained textures at the rims and coarse-grained at the center. They are composed of plagioclase, chlorite,
92 pumpellyite and epidote (Fig. 2f).

93 3.2. Major and trace element composition of basalts

94 A comprehensive description of analytical methods for chemical analysis are given in supplementary
95 material S4. Major element content of our samples is similar to those of other basalts from the Platta nappe
96 (Desmurs et al., 2002; Frisch et al., 1994). Considering both their mineralogical and geochemical compositions,
97 they range between tholeiite and alkali basalt in nature (SiO_2 : 45-50 wt.%; Al_2O_3 : 15-19 wt.%; $\text{Na}_2\text{O}+\text{K}_2\text{O} < 5$
98 wt.%) and display relatively low Mg# (< 0.65) showing that they represent differentiated melts rather than
99 primitive liquids (Table 1).

100 On the basis of trace element compositions, two different types of Jurassic basalts can be distinguished
101 (Fig. 3). Type-1 is depleted in incompatible elements, shows a positive Zr anomaly, a low $(\text{La}/\text{Sm})_N$ and is
102 comparable to “normal” MORB. Type-1 shows a large fractionation of HREE over MREE, $(\text{Sm}/\text{Yb})_N=0.82-1.76$,
103 and a widespread Nb/La ratio (0.55 to 3.25). Samples characterized by medium to high HREE over MREE and a
104 low Nb/La ratio are volumetrically predominant in the Platta-Tasna nappes as well as in the entire Alpine
105 Tethys margins (Fig. 5). The type-2 is enriched in incompatible elements (>10 times PUM concentrations) and
106 high field strength elements (HFSE, Nb, Ta, Zr, Hf; Fig. 3). Type-2 basalts have high $(\text{La}/\text{Sm})_N > 1.3$ and high
107 $(\text{Sm}/\text{Yb})_N > 1.5$ ratios, and are similar to present-day enriched MORB (E-MORB; Gale et al., 2013). Although of a
108 different apparent age, trace element patterns of Cretaceous sills (type-3) are similar to those of Jurassic E-
109 MORB. This geochemical diversity is not restricted to Platta and Tasna basalts but have been described in
110 several ophiolites of the Alpine-Tethys domain (e.g. Bill et al., 2000 and Supplementary material S6).

111 4. Origin of OCT-basalts

112 Basalts from this study are representative of basalts from the Alpine-Tethys (Fig. 4). The Cr vs. Y diagram
113 shows that all these basalts result from 5% to 15% partial melting followed by various degrees of fractional
114 crystallization (Fig. 4). However, their variable trace element patterns (Fig. 3), illustrate that the Platta-Tasna
115 basalts (types-1, 2 and 3) as well as other basalts from the Alpine Tethys may not derived from the melting of
116 a depleted mantle source but rather may reflect the melting of another mantle source.

117 4.1. The depleted “N-MORB”-like basalts (type-1)

118 Type-1 basalts depletion in LREE makes them to a first order comparable to N-MORB (**Erreur ! Source**
119 **du renvoi introuvable.a**). However, most of these basalts as well as most of the Alpine-Tethys basalts
120 substantially differ from present-day MORB by lower Nb/La ratios and higher $(Sm/Yb)_N$ ratios (**Erreur ! Source**
121 **du renvoi introuvable**. and supplementary material S5). High $(Sm/Yb)_N$ ratios are classically used as a proxy to
122 identify the presence of a garnet-bearing component in the MORB source (Hellebrand et al., 2002; Montanini
123 et al., 2008; Saccani, 2015; Sobolev et al., 2007; Stracke et al., 1999). The melting of a DMM-like source in the
124 spinel stability field hardly produces basalts with $(Sm/Yb)_N > 1.2$, even at low degrees of melting, while most of
125 the Alpine basalts display $(Sm/Yb)_N > 1.2$ (see supplementary material S5). Such high $(Sm/Yb)_N$ ratios are easily
126 simulated by melting a depleted mantle source in the garnet stability. However, both modeling approaches
127 fail to reproduce the low Nb/La ratio that characterize most of the Alpine basalts (**Erreur ! Source du renvoi**
128 **introuvable.b**). Since Nb is more incompatible than La ($D_{Nb} < D_{La}$), Nb/La ratios tend to decrease in both the
129 residual mantle and the extracted melts with increasing melt degree. Thus, basalts with $Nb/La < 0.7$ must be
130 derived from the melting of a mantle source more refractory than the asthenospheric DMM.

131 The inherited SCLM can hardly be considered as a potential source for type-1 basalts since it is too fertile
132 (mantle type 1a from Picazo et al., 2016) or too depleted (mantle type 1b from Picazo et al., 2016). Similarly,
133 the oceanic mantle (type 3 from Picazo et al., 2016) is too depleted and occurs too oceanward from the location
134 where basalts erupted. However, the refertilized mantle section (mantle type 2; Picazo et al., 2016) may be a
135 plausible source. The geochemical composition of this refertilized mantle is calculated following the modelling
136 approach of Müntener et al. (2010). The starting mantle composition is the PUM, which has undergone 4%
137 partial melting in the garnet (4%) stability field, followed by 4% in the spinel stability field. This residual mantle
138 is then refertilized by ascending melts (up to 12% in volume) deriving from the melting of an asthenospheric
139 depleted source (DMM) in the spinel stability field. Fig. 5a, b shows that the melting of such a refertilized
140 source is able to produce melts similar to type-1 basalts with low Nb/La and high $(Sm/Yb)_N$. The calculated
141 melting degree of the refertilized mantle is rather low (<15%), which is consistent with previous estimates

142 made on Alpine basalts (Desmurs et al., 2002). This model is very similar to the one proposed previously by
143 Meyzen et al., (2003) to explain the peculiar REE signature of some MORB from the northern magma-poor
144 section of the ultraslow-spreading Southwest Indian Ridge. More importantly, this model is consistent with
145 geological observations made by previous studies (Müntener et al., 2010; Picazo et al. 2016).

146 **4.2. The formation of “E-MORB”-like basalts (types-2 and 3)**

147 Aside from type-1 basalts, a few percent of extrusive basalts can be ascribed as enriched-basalts (E-
148 MORB, types-2 and 3). Additional occurrences of E-MORBs have been reported previously in other fossil OCT
149 in the Alps (External Ligurides and Corsica; Montanini et al., 2008; Renna et al., 2018; Sacconi et al., 2008). The
150 enriched nature of these basalts was attributed to the simultaneous melting of the Jurassic ascending
151 asthenosphere (i.e. the DMM) and garnet-pyroxenites. This hypothesis was tested by applying a similar
152 modeling approach to that of Montanini et al., (2008): variable amounts of garnet-pyroxenite derived melts
153 were mixed with (i) aggregated melts of a depleted spinel peridotite source (red mixing lines in **Erreur ! Source**
154 **du renvoi introuvable.**c, d) and (ii) type-1 basalts extracted from the refertilized SCLM (blue mixing lines in
155 **Erreur ! Source du renvoi introuvable.**c, d). Three different partial melting degrees are considered for the
156 formation of depleted melts ($F=0.05$, 0.1 and 0.15), while a high partial melting degree of 0.25 is considered
157 for the pyroxenite melt (Montanini et al., 2008). Other modeling features (partition coefficients, pyroxenite
158 source composition, etc.) are given in **Erreur ! Source du renvoi introuvable.** caption. Results highlight that E-
159 MORB melts may derive from the mixing between 80-90% of depleted MORB and 10-20% partial melts of
160 garnet-pyroxenite, these results being in good agreement with previous calculations of Sobolev et al., (2007)
161 and Montanini et al., (2008).

162 **4.3. Geodynamic constraints and conclusions**

163 Ophiolites from the Alpine-Tethys domain have provided strong evidence for processes of interaction
164 between inherited SCLM and the ascending asthenosphere (Bodinier et al., 1991; Montanini et al., 2008;
165 Müntener et al., 2004; Müntener and Piccardo, 2003; Piccardo et al., 2004b, 2007; Tribuzio, 2004), which is

166 different from basalts formed at steady state, mature MORs. Almost all extrusive basalts from the Alpine-
167 Tethys OCT overlie an exhumed, refertilized SCLM (Picazo et al., 2016). Only a few overlie the inherited SCLM
168 in Central Alps (Upper Platta, Malenco, Totalp, Picazo et al., 2016). We therefore suggest here that the specific
169 geochemistry of basalts from the Alpine-Tethys OCT is related to their emplacement over the exhumed,
170 refertilized SCLM. The upward melt infiltration coming from the partial melting of the DMM in the garnet
171 stability field creates the widespread refertilized SCLM (type 2 mantle of Picazo et al. 2016). This melting
172 episode is enhanced by a near-adiabatic decompression regime (McKenzie and Bickle, 1988) during the very
173 first stages of lithospheric thinning and asthenospheric uplift (Piccardo et al., 2004a, Fig. 5a). Melt percolation
174 occurred during lithospheric/crustal thinning and was controlled by tectonic processes and the thermal
175 structure of the extending lithosphere, both of which are not well constrained. Kaczmarek and Müntener
176 (2010) proposed, using the example of the Lanzo shear zone, that percolation occurred in the footwall of a
177 major extensional shear zone that acted as a permeability barrier. The thermal advection related to both
178 asthenospheric upwelling and melt percolation is likely able to increase the thermal gradient in order to reach
179 the solidus of the refertilized SCLM at shallower depths in the spinel stability field (Piccardo et al., 2004a,
180 2004b). Integrating this inherited history of the SCLM may explain the peculiar geochemical signature
181 characterizing most of basalts from the Alpine-Tethys OCTs that were erupted on exhumed SCLM in the distal
182 part of the hyperextended margin (i.e. type-1 basalts, Fig. 5b). In contrast, the enriched basalts may correspond
183 to low degree melts of the DMM with small amounts of garnet-pyroxenite, both before and after the onset of
184 seafloor spreading (Fig. 5b-c). This study supports the idea that basalts erupting in the Alpine-Tethys OCT have
185 long-lived, poly-phase and variable mantle sources. The peculiar trace element geochemistry of basalts and
186 the well-established geodynamic setting of OCT in the Alps with complex mantle inheritance enable to explain
187 the differences between OCT and MOR basalts. Therefore, basalts formed along OCTs should be referred to as
188 OCT-basalts rather than MOR-basalts.

189

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196

197 **6. References**

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391

392 **7. List of figures and tables**

393

394 Fig. 1: (a) Location of major ophiolite massifs in the Alps-Apennine-Corsica (redrawn after Rampone and
395 Hofmann, (2012); Schaltegger et al., (2002). [1]: Tertiary basins; [2]: European Units; [3]: Penninic Units; [4]:
396 Ophiolitic Units; [5]: Adriatic Units. (b) Schematic paleogeographic reconstruction of the Alpine-Tethys during
397 the Late Jurassic (redrawn after Picazo et al., 2016). The reconstruction is based on structural (Manatschal and
398 Müntener, 2009) as well as mantle geochemical (Picazo et al., 2016; Rampone and Hofmann, 2012)
399 considerations. Ophiolite with (*) highlight ambiguous paleo-location. CE: Cecina Valley, CH: Chenaillet, EL:
400 External Ligurides, ET: Erro-Tobbio, GE: Gets Nappe, IL: Internal Ligurides, LA: Lanzo, LP: Lower Platta, MA:
401 Malenco, MM: Monte Maggiore, MV: Mont-Viso, UP: Upper Platta, TA: Tasna, TO: Totalp, ZS: Zermatt-Saas.
402 Br: Briançonnais, Sa: Sardinia.

403

404 Fig. 2: Macroscopic field and microscopic observations of basalts. (a) Panoramic view of the Err-Platta Ocean-
405 Continent Transition zone illustrating the position of the inherited and refertilized (infiltrated) subcontinental
406 mantle. UP: Upper Platta; LP: Lower Platta; SCLM: subcontinental lithospheric mantle. (b) Platta outcrop with
407 gabbro body and refertilized mantle, overlain by basalt flows and radiolarian chert. Note that the topographic
408 surface is controlled by both Alpine thrusts and detachment faults. (c) Radiolarian chert sealing a basalt flow
409 (inverse serie). (d) Photomicrograph of Jurassic extrusive basalt showing plagioclase intersertal texture with
410 occurrence of little epidote. (e) Early Cretaceous basaltic sill intruding post-rift sediments. (f) Photomicrograph
411 of Cretaceous basaltic sill showing subhedral granular albitite (doleritic texture). Ep: epidote, Pl: plagioclase.

412

413 Table 1: Major and trace element compositions of basaltic rocks from the Tasna and Platta nappes. Mg# =
414 Mg/(Mg+Fe) (calculated in molar proportions).

415

416 Fig. 3: Whole-rock incompatible trace element (normalized to the primitive mantle, PM, McDonough and
417 Sun, 1995) and rare-earth element compositions (normalized to C1 chondrite, Anders and Grevesse, 1989) of
418 basalts from the Platta-Tasna nappes. (a) Type-1 depleted Jurassic extrusive basalts, (b) Type-2 enriched
419 Jurassic extrusive basalts, (c) Type-3 Cretaceous basaltic sills. Dashed black lines N-MORB and E-MORB are
420 from Gale et al., (2013) and other basalts from Platta are from Desmurs et al., (2002).

421

422 Fig. 4: Cr (ppm) versus Y (ppm) of whole-rock Platta-Tasna basalts. The partial melting trend starts from the
423 mean DMM (Workman and Hart, 2005). The arrows indicate the calculated trends of the liquids for fractional
424 crystallization of clinopyroxene + plagioclase (cpx + plg) assemblages starting from the relatively primitive
425 basaltic sample 13-04 (residual liquid fraction F up to 0.5). Mineral mass fractions are 0.2cpx+0.8plg. Partition
426 coefficients sources for Y and Cr are from Ionov et al., (2002) and from Hart and Dunn, (1993) respectively.
427 Note that the starting mantle composition does not dramatically change the partial melting models nor the
428 crystal fractionation trends. 1 : Gale et al., 2013, 2 : Other Platta basalts are from Desmurs et al., (2002) and
429 Frisch et al., (1994). Alpine-Apennine basalts (little blue circles) are the overall Alpine-Tethys domain. Internal
430 Ligurides (Ferrara et al., 1976; Ottonello et al., 1984; Rampone et al., 1998; Venturelli et al., 1981), External
431 Ligurides (Marroni et al., 1998; Montanini et al., 2008; Vannucci et al., 1993; Venturelli et al., 1981), Corsica
432 (Beccaluva et al., 1977; Durand-Delga et al., 1997; Padoa et al., 2001; Renna et al., 2018; Saccani et al., 2008;
433 Venturelli et al., 1981, 1979), Western Alps (Bodinier et al., 1986; Chalot-Prat et al., 2003; Lewis and Smewing,
434 1980; Pognante et al., 1985), Central Alps (Desmurs et al., 2002; Frisch et al., 1994), Tuscany (Ferrara et al.,
435 1976; Mazzeo et al., 2016), Calabre (Liberi et al., 2006) and inferred Valais domain (Bill et al., 2000; Dürr et al.,
436 1993; Kramer et al., 2003; Mugnier et al., 2008; Steinmann and Stille, 1999).

437

438 Fig. 5: (a) Nb/La vs. La showing the compositions of Alpine basalts. The modelling consists of the calculation of
439 Nb/La ratio of aggregated non-modal fractional liquids that derive from the partial melting of the refertilized
440 SCLM. The chemistry of the refertilized mantle source is calculated following the study of Müntener et al.
441 (2010) : a fertile peridotite (primitive upper mantle, PUM; McDonough and Sun, 1995) experienced 4% partial
442 melting in the garnet stability field (A) followed by 4% of additional melting in the spinel stability field (B). The
443 residue is refertilized by adding a 4% fractional melt derived from a spinel DMM source (1% increment, C). A
444 volume of 1%, 3%, 6%, 9% and a maximum of 12% of melts stored in the peridotite is considered (Müntener
445 et al., 2010). The red curves correspond to melts extracted from the calculated refertilized mantle in the spinel
446 stability field at F up to 20%. Partition coefficients are from Johnson et al., (1998), starting mineralogy and
447 modes are from Johnson et al., (1990). (b) $(\text{Sm}/\text{Yb})_N$ vs. Yb_N of alpine basalts. Red curves are calculated in the
448 same way as in Fig. 5a. (c) Nb/La vs. La of E-MORB alpine basalts, assuming that they derive from the mixing of
449 aggregated non-modal fractional melts (F=5, 10, 15%) from a spinel DMM source (in red) or a refertilized SCLM
450 (8% refertilization, in blue) with garnet pyroxenite derived melts (F=25%; Montanini et al., 2008). The garnet
451 pyroxenite source derives from high degrees of batch melting (F=40%; Montanini et al., 2012) of a recycled
452 oceanic crust (i.e. an eclogite; Sobolev et al., 2007). The eclogite composition (C_{eclogite}) is calculated assuming
453 that eclogite corresponds to basaltic, N-MORB type crust ($C_{\text{N-MORB}}$) that experienced element mobility during
454 subduction prior being recycled into the mantle: $C_{\text{eclogite}} = C_{\text{N-MORB}} - \text{retention factors}$ where retention factors =
455 $1 - \text{mobility coefficients}$ (Kogiso et al., 1997). The calculated eclogite has a high Nb/La ratio (~ 2 , not shown
456 here), consistent with values calculated by Rudnick et al., (2000). Gt-pyroxenite mineralogy: 60% cpx, 30% opx,
457 10% gt (Hirschmann and Stolper, 1996); melting modes: cpx=gt=50%. Eclogite mineralogy: 50% gt, 50% cpx.
458 Numbers in *italic* represent the proportion of mantle derived melts in the mixing. (d) $(\text{Sm}/\text{Yb})_N$ vs. Yb_N of alpine
459 basalts. Red and blue curves are calculated in the same way as in Fig. 5c. N-star: N-MORB average $\pm 2\sigma$ (grey
460 area); Present-day MORBs and N-MORB average are from Gale et al. (2013).

461

462 Fig. 5 : Schematic representation of the geodynamic model leading to the OCT-basalts. (a) Melt percolation.
463 Near adiabatic upwelling of the asthenosphere (DMM) during this early rifting stage. Onset of asthenosphere

464 partial melting in the garnet stability field (pink circles), and initial percolation of melts towards the
465 subcontinental mantle (light brown area) during the stretching of the lithosphere. Porous-flow melt
466 percolation induces (i) the refertilization of Permian (?) depleted SCLM, (ii) the formation of plagioclase-
467 bearing peridotite at shallower levels, and (iii) the formation of a weakened mantle wedge for the lithospheric
468 breakup (Piccardo, 2016). The distal subcontinental mantle is melt-impregnated (up to 12%) and refertilized.
469 Major crustal/lithospheric thinning is followed by mantle exhumation and onset of sea-floor exposure of the
470 melt-modified peridotites. (b) Creation of OCT-basalts. Thermal advection (Piccardo et al., 2014) enhanced by
471 porous-flow melt percolation moves the solidus upwards allowing the melting of the refertilized
472 subcontinental mantle in the spinel stability field (light blue circles) forming the type-1 basalts (light blue pipes
473 and volcanoes). Mixing of melts from the partial melting of the DMM (dark blue circles) and garnet-pyroxenite
474 (black dashes, orange circles) creates type-2 basalts (red pipes and volcanoes). (c) Formation of type-3 off-axis
475 basalts (yellow volcanoes) by the same process as type-2. The mixing between the two sources is located at
476 the two edges of the upwelling asthenosphere and is probably related to thermal readjustments. Active
477 spreading ridge followed by MOR-basalt formation caused by the partial melting of the DMM under spinel
478 (shallow) stability field.

479

480 **8. Supplementary material**

481

482 Supplementary material S1: (a) Location of the Platta and the Tasna nappes in the Alpine realm. (b) Tectonic
483 units and paleogeographic domains close to Platta and Tasna Nappes. (c) Geological map of Platta modified
484 after Schaltegger et al., (2002). with sample locations. Cretaceous magmatic sills (red) have been magnified
485 for clearance.

486 Supplementary material S2: Schematic section representing the lithologies and their stratigraphic position in
487 the Platta nappe. Note that the shear-zone in mantle and the sulfide deposit are truncated by the
488 detachment fault. Redrawn after Peters et al., (2008) and own fieldwork observations.

489 Supplementary material S3: Locations and mineralogical assemblage of studied magmatic samples.
490 Coordinates grid is WGS84 UTM 32N, decimal degrees. Fds (Ab): feldspar (albite), Cal: calcite, Chl: chlorite,
491 Cpx: clinopyroxene, Ep: epidote. Crosses indicate relative mineral proportions. Note that samples 13-03 and
492 13-04 are from the Tasna nappe.

493 Supplementary material S4: Analytical method

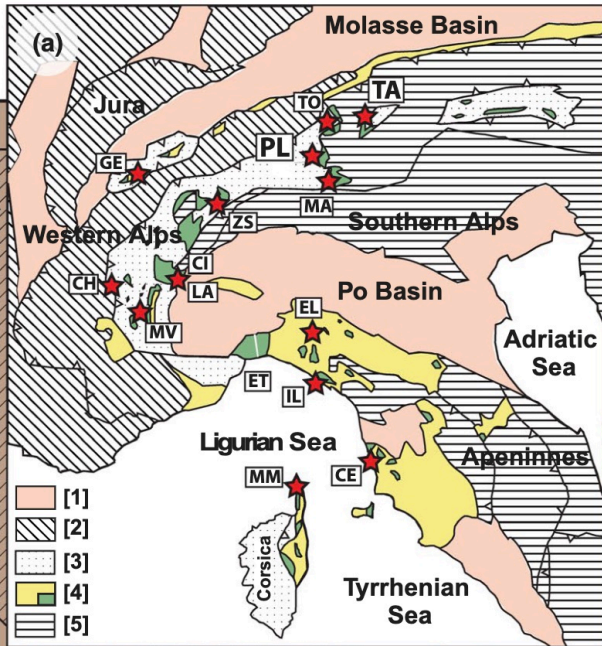
494 Major and trace elements were measured using the analytical method of Chauvel et al., (2011) at
495 Laboratoire d'Hydrologie et de Géochimie de Strasbourg (LHyGeS, University of Strasbourg, France). For major
496 elements, ~50 mg of sample powder was precisely weighted and dissolved in closed Savilex® beakers using a
497 mixture of 0.3 mL HF 24N and 0.85 mL HNO₃ 14N in a hot plate at 90°C for 5 days. 20 ml of boric acid (H₃BO₃
498 at 20 g.l⁻¹) was added to the solutions after cooling to neutralize excess HF and were further diluted with 250
499 ml of Milli-Q® water. The solutions were then stored in a refrigerator for at least 2 days to ensure complete
500 neutralization of HF by H₃BO₃ and analyzed within a week using an ICP-AES (Thermo Scientific ICAP 6500). For
501 trace elements, 100 mg of sample powder were dissolved in a HF-HClO₄ mixture (5:1) in closed teflon beaker
502 on a hot plate at 140°C for 7 days. The mixture was then completely evaporated at 150°C, and the residue was
503 taken up in concentrated HNO₃. After another complete evaporation, the residue is finally diluted in 40 mL of
504 HNO₃ 7N. The international standard PM-S (gabbro) was analysed repeatedly as unknown to ensure the validity
505 of our measurements. Results show that PM-S concentrations are consistent with the preferred reference
506 values from GeoREM (Jochum et al., 2008). Differences are less than 0.6% for SiO₂ and less than 2% for all
507 other major oxides except for Na₂O, K₂O and P₂O₅ which are present in very low concentrations. In addition,
508 sample 24-04 has been analyzed twice as a duplicate and results show no relevant differences in trace
509 elements between the two aliquots. It is noticed that the anomalous composition of one sample (20-03), which
510 is characterized by very low SiO₂ (36.92 wt.%), high CaO (15.33 wt.%) and high LOI (9.33 wt.%), is explained by
511 the amount of calcite veins and is not used in the discussion.

512

513 Supplementary material S5: (a) $(La/Sm)_N$ vs. $(Sm/Yb)_N$ of Alpine basalts compared to present-day MORBs (Gale
514 et al., 2013). Sample position reflects their shape of REE patterns, which depends on melting processes and/or
515 the source composition. Most of present-day MORBs (N-MORBs) are explained by moderate to high degrees
516 of partial melting (Gale et al., 2013) of a depleted mantle source in the spinel stability field (Workman and
517 Hart, 2005) as represented by the small graph in the lower left corner. In contrast, most Alpine basalts
518 (including basalts from the Platta and Tasna nappes) are characterized by high $(Sm/Yb)_N$, suggesting the
519 involvement of garnet in the mantle source. Samples that plot in the upper right part of the graph (i.e. E-MORB)
520 can be related to low degrees of melting of a garnet depleted mantle source although they more likely derive
521 from the melting of an heterogeneous, enriched source. N-star: N-MORB average $\pm 2\sigma$ (grey area); E-star: E-
522 MORB average (Gale et al., 2013). PM: partial melting, sp: spinel, gt: garnet. (b) Nb/La vs. $(Sm/Yb)_N$ of Alpine
523 basalts compared to present-day MORBs. The calculated aggregated non-modal fractional melts derived from
524 a spinel lherzolite (in red) and a garnet lherzolite (in green) are shown. The mantle source chemistry is the
525 DMM from Workman and Hart, 2005, and results using the enriched (E-) and depleted (D-)DMM are also shown
526 to not have a narrow vision of the heterogeneous geochemistry of the DMM. Numbers in italic indicate the
527 degree of partial melting. Partition coefficients are from Johnson et al., (1998), starting mineralogy and modes
528 are from Johnson et al., (1990).

529

530



(b)

Eastern Alps

Western and Central Alps sections

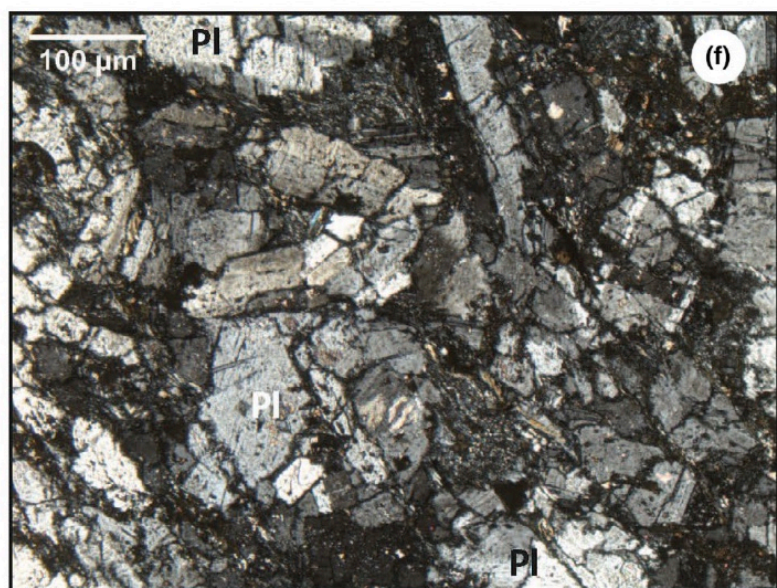
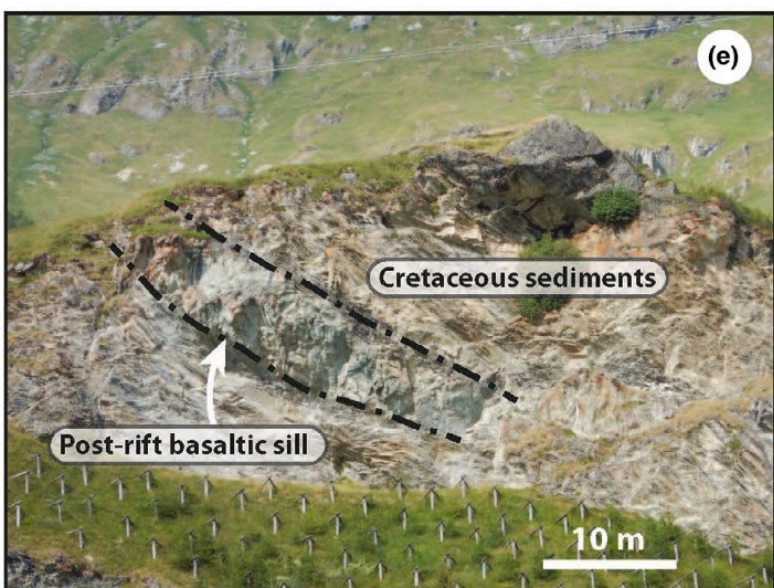
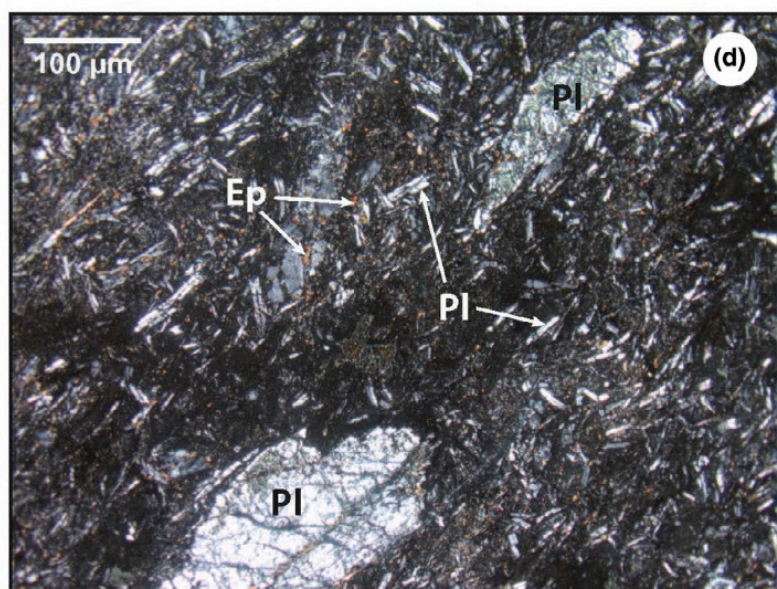
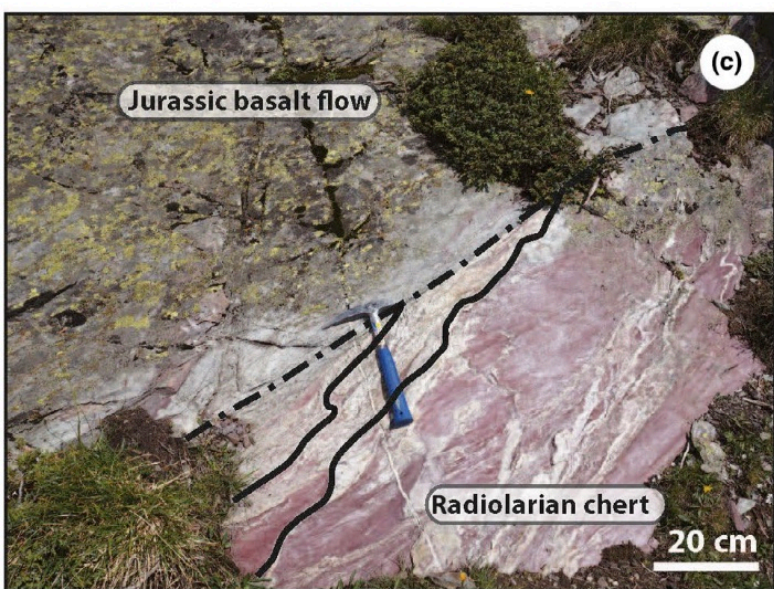
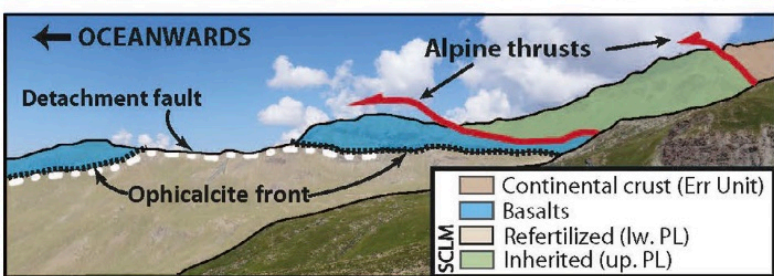
Europe

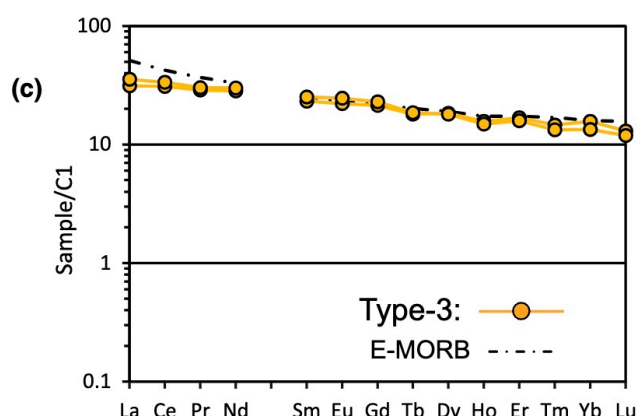
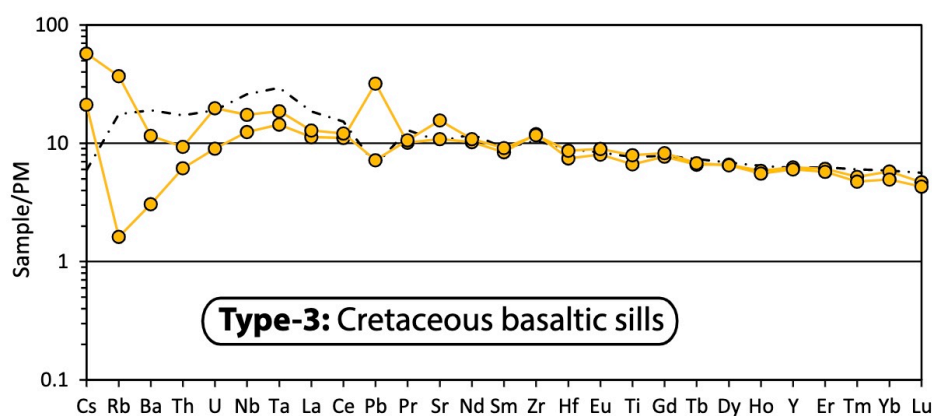
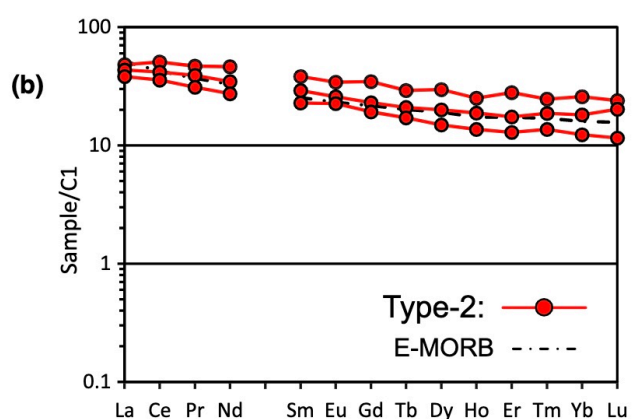
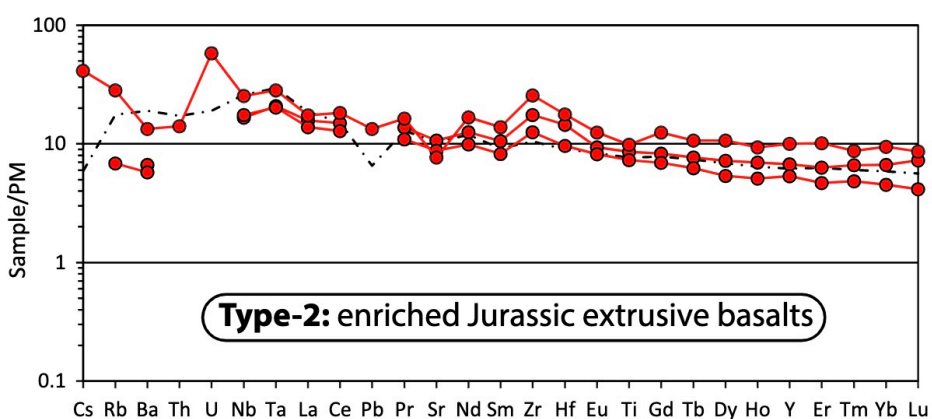
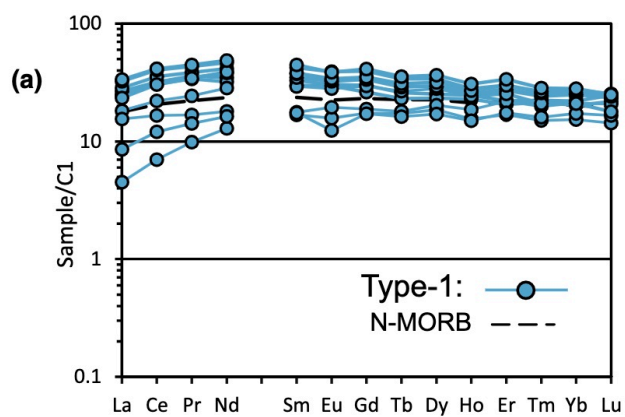
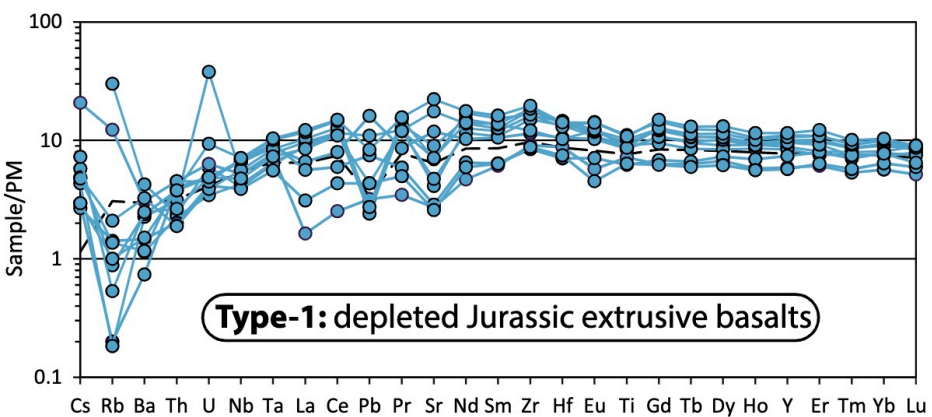
Pyrenees

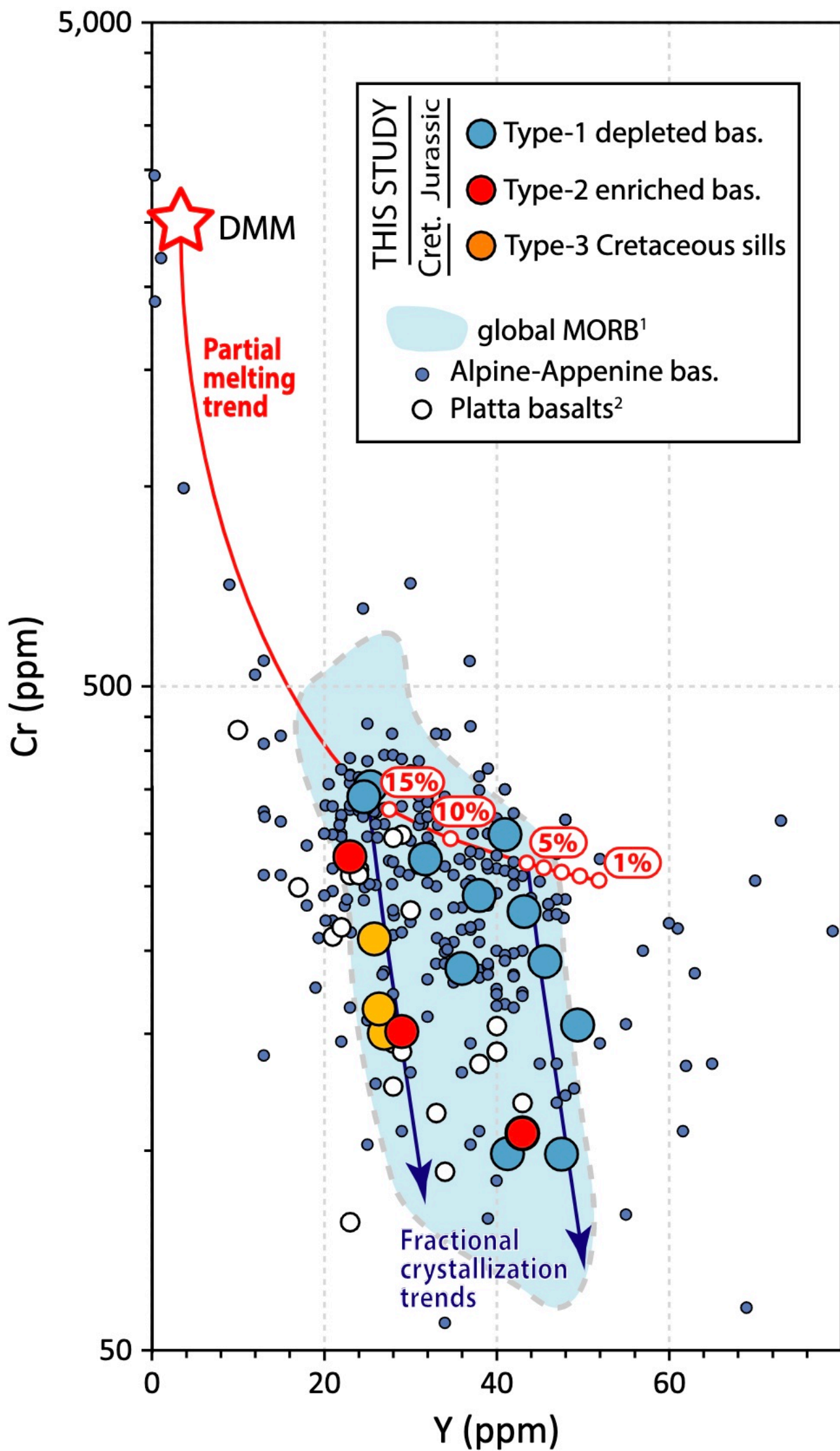
Iberia

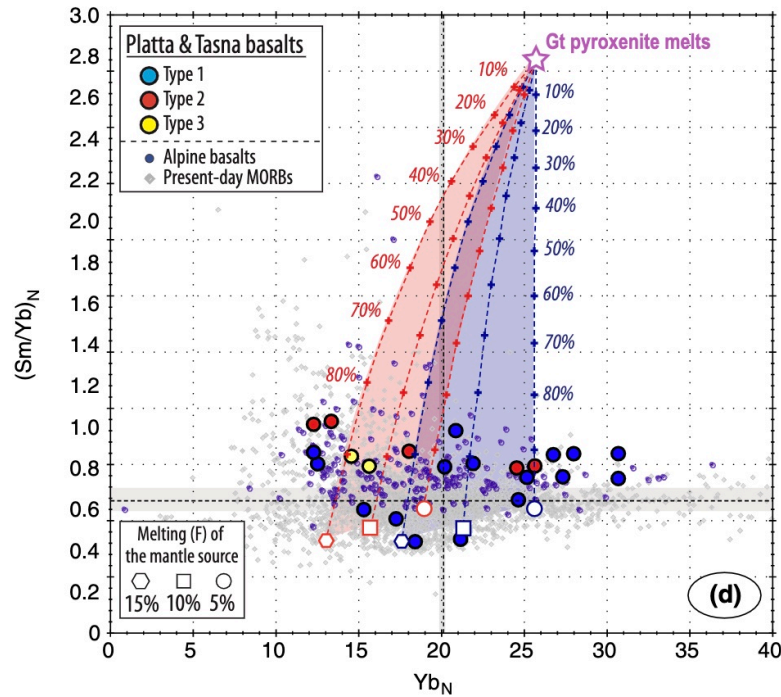
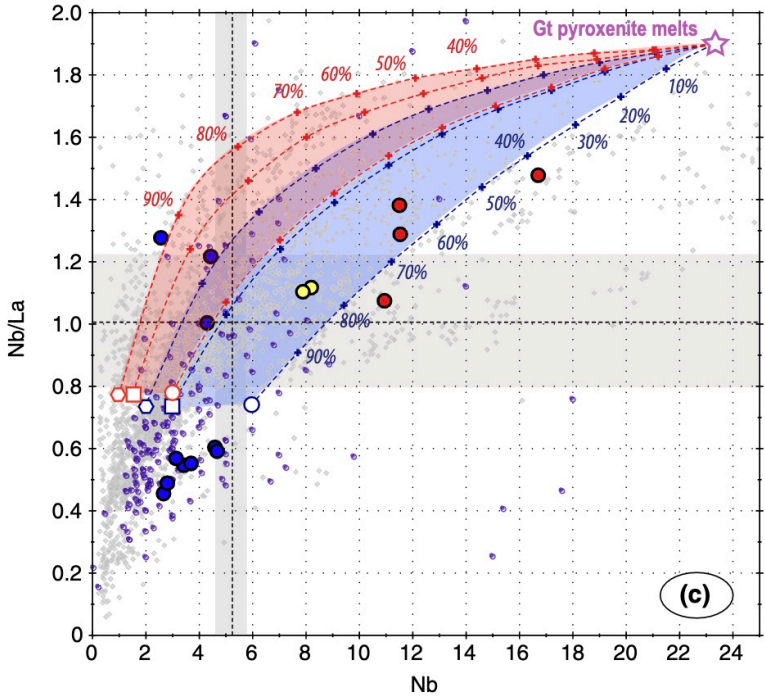
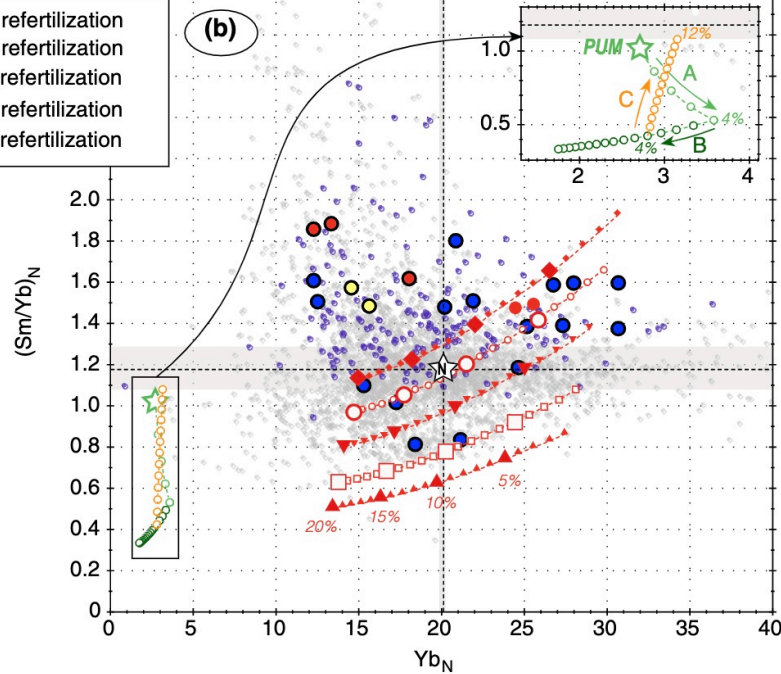
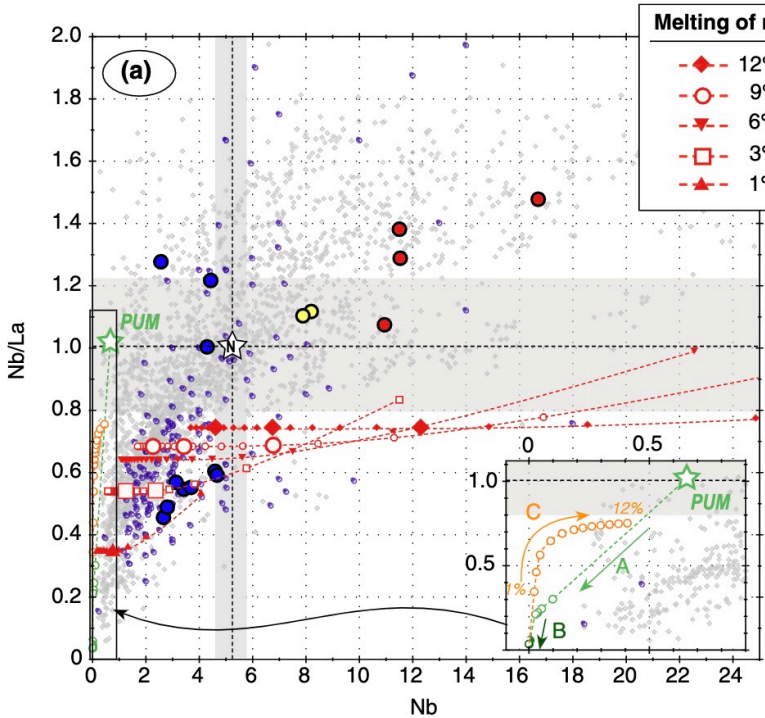
Liguria-Piemonte Basin

- Thinned/Undeformed continental crust
- Inherited and/or depleted SCLM
- Refertilized SCLM
- Oceanic lithospheric mantle (DMM)

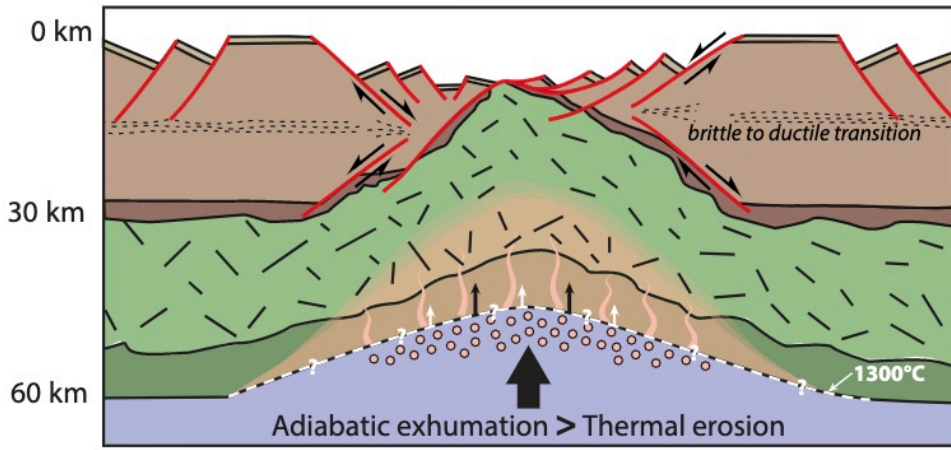




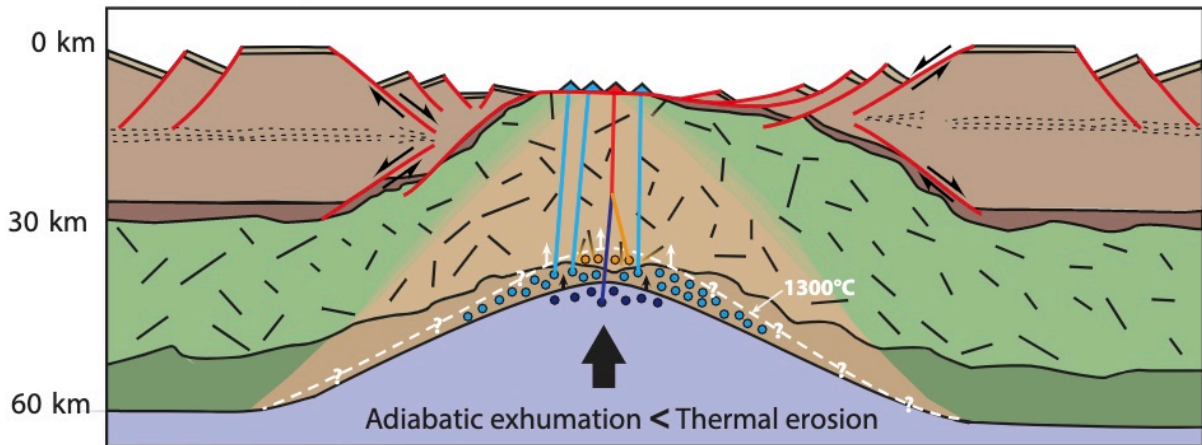




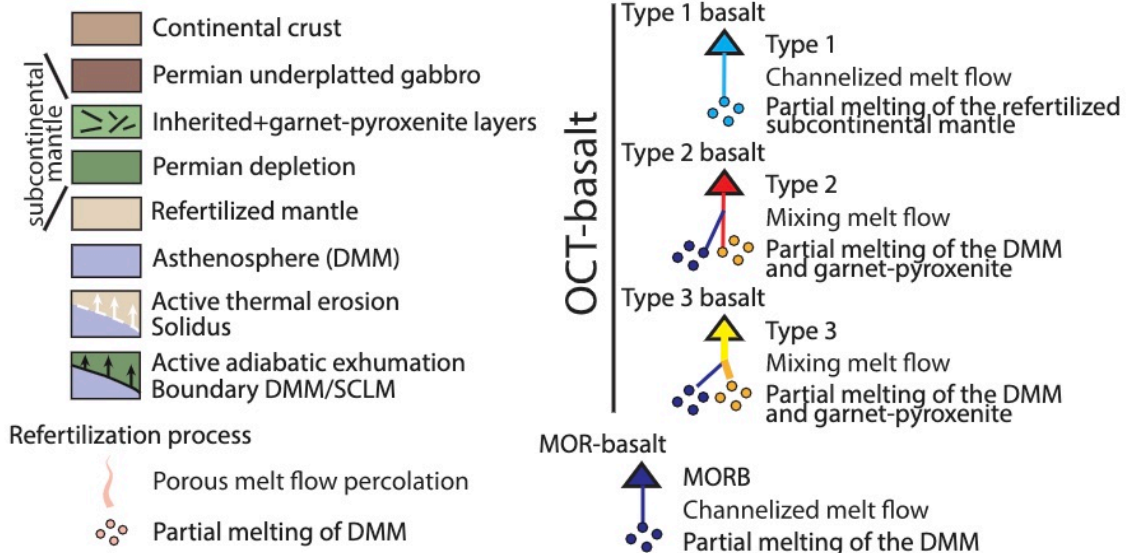
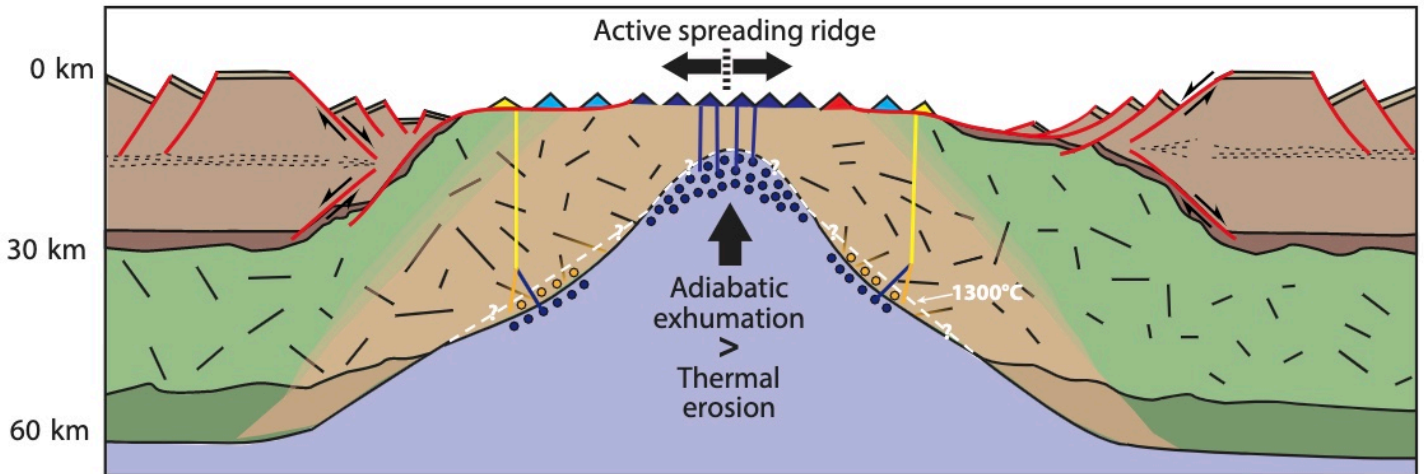
(a) Melt percolation

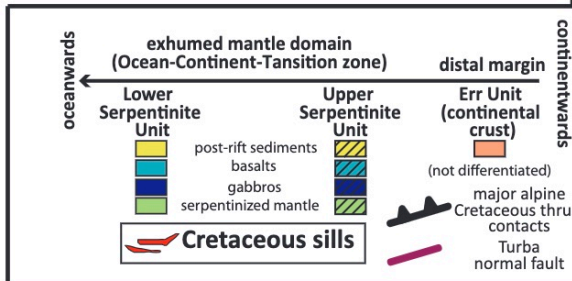
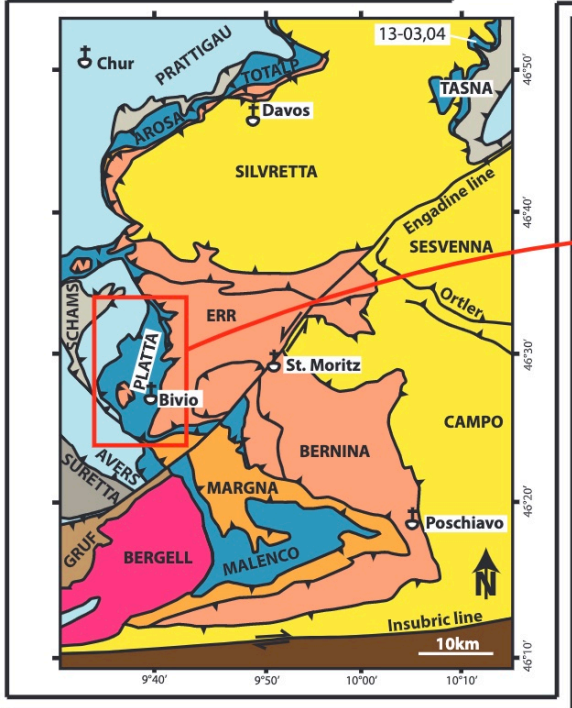
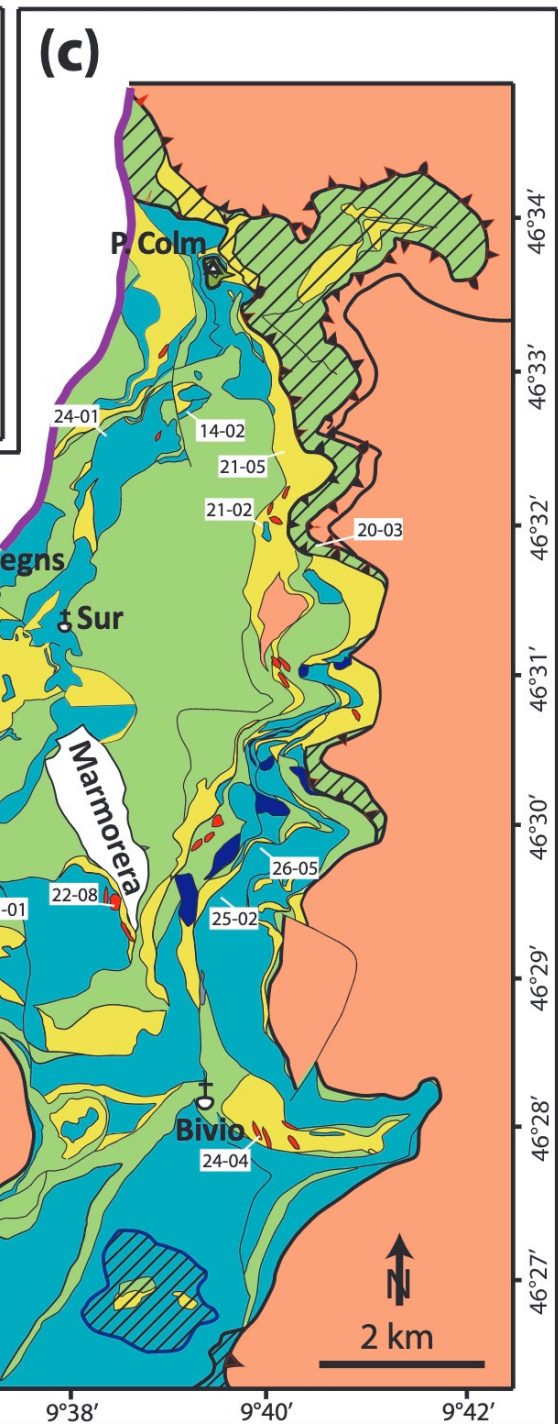
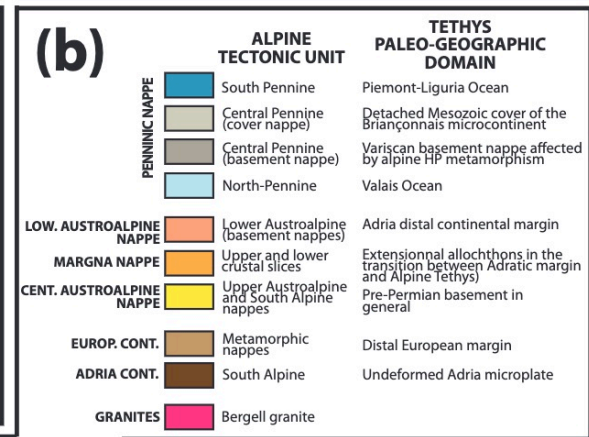
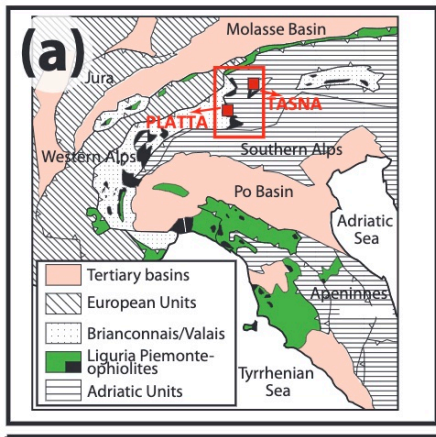


(b) OCT-basalts

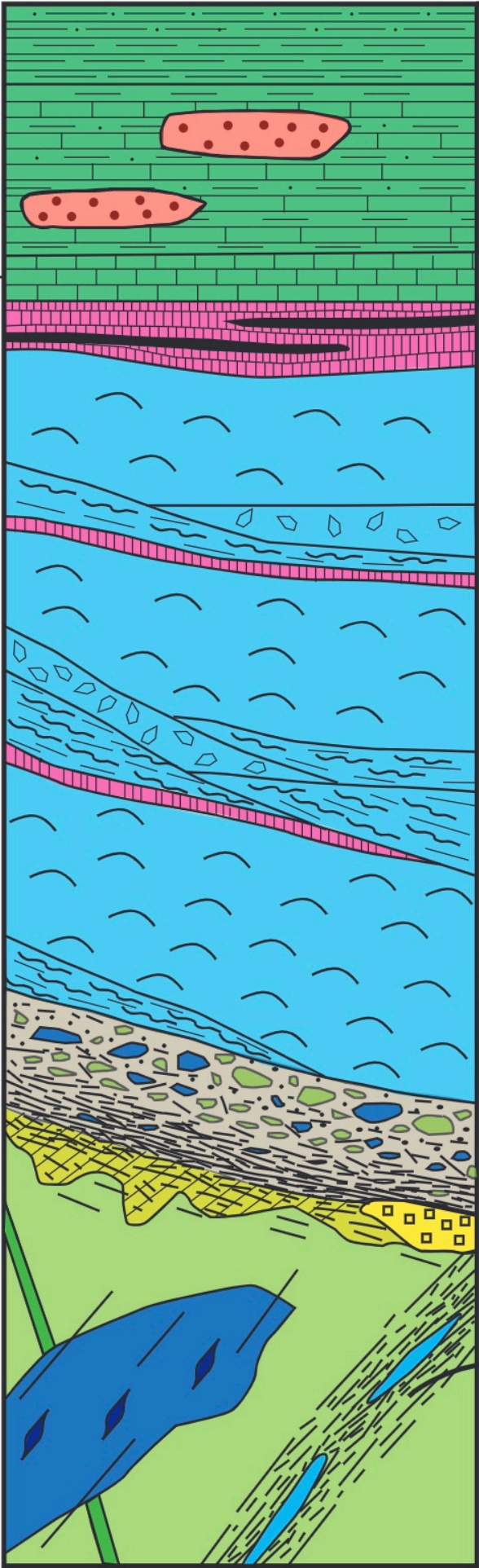


(c) MOR-basalts






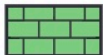



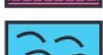







JURASSIC
CRETACEOUS



Tonschiefer and Palombini

Aptychus limestone

-  phyllite, schist
-  alkali-gabbro
-  calc-schist and calcarenite
-  limestone
-  Mn-deposit
-  hyaloclastite flow/pillow breccia
-  radiolarian cherts
-  basalt flow and pillow basalt
-  sulfide deposit
-  opihicalcite
-  rodingite dike
-  gabbro with sigmoidal magmatic pocket
-  mantle-derived rocks and garnet-pyroxenite layers

Detachment fault
(gouge, breccia with gabbro and serp. clasts)

Shear-zone
(rodingitized basaltic dikes)

