



HAL
open science

Petrology and geochemistry of the Texenna ophiolites, northeastern Algeria: Implications for the Maghrebian flysch suture zone

El Hachemi Boukaoud, Gaston Godard, Moulley Charaf Chabou, Youcef Bouftouha, Sidali Doukkari

► To cite this version:

El Hachemi Boukaoud, Gaston Godard, Moulley Charaf Chabou, Youcef Bouftouha, Sidali Doukkari. Petrology and geochemistry of the Texenna ophiolites, northeastern Algeria: Implications for the Maghrebian flysch suture zone. *Lithos*, 2021, 390, p.382-401. 10.1016/j.lithos.2021.106019. insu-03590043

HAL Id: insu-03590043

<https://insu.hal.science/insu-03590043>

Submitted on 24 Apr 2023

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

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



Distributed under a Creative Commons Attribution - NonCommercial 4.0 International License

1 **Petrology and geochemistry of the Texenna ophiolites, northeastern Algeria:**
2 **Implications for the Maghrebian flysch suture zone**

3 El Hachemi Boukaoud ^{a,b}, Gaston Godard ^{c,*}, Moulley Charaf Chabou ^d, Youcef Bouftouha ^a, SidAli
4 Doukkari ^{e,f}.

5

6 ^a Laboratoire de Génie Géologique, Université Mohamed Seddik Benyahia - Jijel, BP 98 Ouled Aissa,
7 18000 Jijel, Algeria.

8 ^b Laboratoire Géologie et Environnement, Université Frères Mentouri Constantine1, Algeria.

9 ^c Université de Paris, Institut de physique du globe de Paris, CNRS, F-75005 Paris, France.

10 ^d Emerging Materials Research Unit and Institute of Architecture and Earth Sciences, Department of
11 Earth Sciences, Ferhat Abbas University, Setif 1, Algeria.

12 ^e Laboratoire de Géodynamique, Géologie de l'Ingénieur et de Planétologie, FSTGAT-U.S.T.H.B., B.P.
13 32, El Alia, Dar el Beida, 16111-Alger, Algeria.

14 ^f Département S.N.V, Faculté des Sciences, Université Alger 1, 2 rue Didouche Mourad, Alger, Algeria.

15 *Corresponding author: godard@ipgp.fr

16 **Abstract**

17 This study presents the first geochemical, mineralogical and petrological data on the mafic and
18 ultramafic rocks of Texenna (Lesser Kabylia, north-eastern Algeria) with the aim of constraining their
19 tectonic setting in the context of the Maghrebide belt. The magmatic-sedimentary complex of
20 Texenna comprises serpentinite, metabasites (metagabbro, metadolerite and metabasalt with
21 pillow-lava structures), and oceanic metasediments (radiolarite and calcschists). Serpentinites consist
22 mainly of mesh textures and large bastites of lizardite, Cr-spinels altered to “ferritchromit” and
23 chlorite, with chrysotile and chlorite veins. Their features are very similar to those of serpentinites
24 collected from the seafloor. Pillow-lava metabasalts contain mineral assemblages of albite, epidote,
25 chlorite, actinolite and titanite, and, despite metamorphism, have preserved many of the pillow-lava
26 microstructures (former glassy, variolitic and spherulitic zones; former devitrification spherulites and
27 gas vesicles; pseudomorphs after phenocrysts and microliths). Metadolerites and metagabbros have
28 sometimes preserved their magmatic plagioclase, but are generally transformed into an assemblage
29 of albite, calcic amphiboles, chlorite, epidote and titanite, a mineral assemblage typical of the static
30 greenschist-facies hydrothermal metamorphism of an oceanic crust. Some deformation in the
31 conditions of phengite-bearing greenschist-facies metamorphism is linked to the tectonic
32 emplacement of the ophiolites in their current setting during the Maghrebide orogeny.

33 Geochemically, the whole-rock compositions of the serpentinites indicate a harzburgite protolith,
34 while the metabasites are tholeiitic and show the N-MORB signature for the pillow lavas and T-MORB
35 to E-MORB affinities for the intrusive rocks (metadolerites and metagabbros). These results
36 demonstrate for the first time that the magmatic-sedimentary complex of Texenna belongs to a true
37 ophiolitic slice and indicate that the sedimentary cover of the Maghrebian Flysch domain was
38 deposited over a true oceanic rather than a thinned continental crust. The Texenna ophiolites show
39 analogies with many other ophiolites in the Western Mediterranean region, indicating that they
40 represent remnants of the ancient oceanic lithosphere of the western Tethys (*i.e.*, Maghrebian
41 Tethys Ocean) that formed between Africa and Iberia during the Middle-Late Jurassic.

42 **Keywords:** Ophiolites, Maghrebian Flysch, Texenna, Maghrebide belt, Algeria, Western Tethys.

43 **Introduction**

44 Ophiolites are geological markers of prime importance for understanding the geodynamic evolution
45 of orogenic belts (*e.g.*, Furnes and Safonova, 2019). They often occur in suture zones that mark the
46 closure sites of the ancient oceanic domain. In the southern part of the circum-Mediterranean Alpine
47 belt, the Maghrebide chain would be such a suture zone, resulting from the Cenozoic collision of
48 fragments of the AlKaPeCa microcontinent (*i.e.*, Alboran, Kabyle, Peloritan and Calabrian blocks) with
49 Africa, respectively at the origin of the internal (*i.e.*, northern) and external (southern) zones of the
50 chain (Durand-Delga and Fontobé, 1980; Bouillin, 1986). The suture zone is underlined by
51 allochthonous flysch deposits (Frizon de Lamotte et al., 2011), with slices of mafic rocks with
52 ophiolite affinities (Fig. 1), well known in the Rif (Morocco) and Sicily (*e.g.*, Durand-Delga et al.,
53 2000). Similar mafic rocks have been reported at the base of the Maghrebian flysch, in various
54 localities of northern Algeria, both in Greater Kabylia (Chellata massif) and Lesser Kabylia (Texenna,
55 Moul Ed-Demamene massif, Fedj el Mekta') (Durand-Delga, 1948, 1949, 1950, 1954, 1955; Coutelle
56 and Gélard, 1968; Bouillin, 1974, 1986; Bouillin and Hernandez, 1975; Bouillin et al., 1977). In the
57 Texenna region, in particular, an association of mafic and ultramafic rocks occurs beneath the
58 Maghrebian flysch nappes at the southern front of the internal zones of Lesser Kabylia (Durand-
59 Delga, 1949, 1950, 1955; Bouillin et al., 1977). The occurrence near Texenna of serpentinites,
60 metagabbros, pillow lavas and radiolarites likely represents the only complete ophiolite suite of the
61 Maghrebian flysch suture zone (Bouillin, 1986). However, the last studies carried out on this
62 formation date back to the 1970s (*e.g.*, Durand-Delga, 1971; Bouillin et al., 1977), so that new
63 investigations are necessary to better characterize the only ophiolite suite known to date in the
64 Algerian part of the Maghrebide chain.

65 Major and trace element geochemistry of ophiolitic mafic rocks provides important clues for
66 unravelling the tectonic setting of ophiolites. Furthermore, immobile element geochemistry is an
67 important tool for investigating ophiolitic rocks, since ophiolites experience intense alteration and
68 metamorphism before, during and after being involved in orogenic processes (*e.g.*, Pearce, 2014).
69 Very little geochemical data are available for the mafic rocks associated with the Maghrebien flysch
70 zone. Durand-Delga et al. (2000) presented a few geochemical analyses of Jurassic basic rocks
71 associated with the Maghrebien flyschs of the Rif (Morocco) and Sicily (Italy), indicating their E-MORB
72 affinity. Roman'ko et al. (1998) gave geochemical composition of basalt from Moul Ed-Demamene
73 massif (Lesser Kabylia, Algeria). However, no geochemical analysis has been published for the mafic
74 and ultramafic rocks of the Texenna region. In addition, although petrographic descriptions of the
75 Texenna rocks have already been made using a polarizing optical microscope (Durand-Delga, 1955,
76 1971; Bouillin et al., 1977), a detailed petrological study remains necessary to decipher the formation
77 of these magmatic rocks and their metamorphic evolution.

78 The purpose of this paper is to present the first mineralogical, petrological and geochemical data on
79 the mafic and ultramafic rocks of Texenna and to discuss their geodynamic implications in the
80 context of the Maghrebide chain.

81 **Regional geology**

82 **Maghrebide chain**

83 The Maghrebide chain, which extends in northern Morocco and Algeria (Fig. 1), includes three main
84 parallel tectonic domains, from north to south (for a recent review, see Leprêtre et al., 2018): (i) the
85 internal zones, belonging to the ALKaPeCa microcontinent; (ii) the Maghrebien flysch nappes,
86 consisting of remnants of the Ligurian-Maghrebien Tethys oceanic basin; and (iii) the external zones,
87 interpreted as the North-African palaeomargin inverted during the Cenozoic collision. The ophiolitic
88 rocks studied here are in the vicinity of the Maghrebien flysch units, which, depending upon their
89 position within the initial basin, are traditionally divided into two main stratigraphic sequences
90 (Bouillin et al., 1970): the Mauretania series, originally close to the southern margin of the ALKaPeCa
91 microcontinent and the Massylian series, near the African palaeomargin.

92 ***Internal zones***

93 In the study area (Fig. 2a), the internal zones consist of two metamorphic groups (Bouillin et al.,
94 1977): a retro-metamorphosed lower group composed of gneiss and white-micaschist, and an upper

95 group with granulite-facies rocks, including garnet-kyanite “leptynites” (*i.e.*, leucocratic gneiss),
96 garnet pyroxenites, metagabbros and forsterite marbles.

97 In various places near Texenna, the Kabyle basement contains intercalations of pyroxenite,
98 amphibolite and layered metagabbro, from a few centimetres to a metre in thickness (Bouillin et al.,
99 1977; Djellit, 1987). Their mineralogical assemblages, essentially anhydrous, are dominated by
100 pyroxenes with oriented granoblastic texture, and were formed under granulite- and/or amphibolite-
101 facies metamorphic conditions. A layered metagabbro, dated by the U-Pb method on titanite,
102 provided ages at 315 and 239 Ma (unpublished data), and seems therefore linked to the Hercynian
103 orogeny. Peridotite meter-thick levels, with olivine + Cpx + Opx ± plagioclase ± spinel, were also
104 observed along the thrust that separates the Kabyle basement from the Texenna ophiolites (Fig. 2).
105 Subject to a petrological and geochemical study in progress, these peridotites also seem to be linked
106 to the Kabyle basement. These mafic and ultramafic rocks of the basement contrast with those of the
107 Texenna ophiolites by their high-grade anhydrous metamorphism and their probable Hercynian age,
108 and are therefore not considered in this study.

109 The Kabyle metamorphic basement is overlain by the so-called “Dorsale calcaire” (Raoult, 1974),
110 which consists of unmetamorphosed sediments, of either Palaeozoic (Baudelot et al., 1981) or Upper
111 Jurassic to Lutetian age, interpreted as deposited along the passive margin between the AlKaPeCa
112 continent and the Ligurian Tethys by several authors (*e.g.*, Bouillin, 1992). Two kilometres northeast
113 of the gorge of Oued Djendjene, Silurian graptolitic shales lie unconformably on the Kabyle basement
114 (Durand-Delga, 1955), which is also unconformably overlain by the much more recent clastic deposits
115 of the Oligo-Miocene Kabyle unit (Djebel el Kelaa; Fig. 2a) dated from Late Oligocene to Burdigalian
116 (Bouillin, 1979).

117 *Flysch domain*

118 Cretaceous turbidite sequences (“flyschs”) are widespread in the region of Texenna, where they form
119 the Massylian and Mauritanian units. According to Bouillin et al. (1977), the Mauritanian flysch of
120 Djebel Es-Sendouah is in continuity with the Texenna green rocks and the overlying radiolarites and
121 carbonate rocks. It is composed of interbedded thick layers of slightly metamorphosed lustrous
122 pelitic schists and quartzitic sandstones. Some unmetamorphosed Mauritanian flysch (“Flysch de
123 Guerrouch”) outcrops in the western and northern part of the study area (Fig. 2a). Further south,
124 some Albian-Aptian flysch in 10-cm thick layers, of Massylian affinity, are overlain by Albian-
125 Cenomanian shales with microbreccias and black phanites. A tectonic contact separates the
126 Mauritanian flysch of Djebel Es-Sendouah (Mauritanian flyschs) from the Massylian flysch (Bouillin et
127 al., 1977).

128 The occurrence in the region of Texenna of large undeformed and unmetamorphosed klippen of
129 Mauritanian flysch ("Flysch de Guerrouch") over the Kabyle basement and its Oligo-Miocene cover,
130 have led Andrieux and Djellit (1989) and Andrieux et al. (1989) to suggest an internal origin of the
131 "Flysch de Guerrouch", from basins located north of the Kabyle domain ("ultra" flysch). However, this
132 hypothesis has been challenged, spurring intense debates during the 1970s-1980s (*e.g.*, Coutelle and
133 Delteil, 1989; Bouillin, 1989).

134 *External Zones*

135 The Tellian units belong to the external zones and are exposed to the south of the Flysch domain; in
136 the studied area, they correspond to the mid-Tellian units of Wildi (1983). The sedimentary
137 sequence, which ranges in age from Triassic to Lutetian (Paleogene) (Bouillin, 1979), includes Triassic
138 sandstones, Jurassic carbonate rocks, conglomerates and microbreccias of the Upper Cretaceous,
139 and finally black marls with yellowish "balls" of the Upper Cretaceous, Palaeocene and Eocene
140 (Bouillin, 1979). Finally, the region is characterized by large exposures of the Triassic complex
141 composed of red clays, evaporites and mafic vein rocks ("ophites": Durand-Delga, 1955).

142 **Ophiolitic complex of Texenna**

143 Ehrmann (1946) was the first to report on the Tamesguida geological map at 1/50 000 the
144 occurrences of mafic and ultramafic rocks in the region of Texenna. Durand-Delga (1949, 1950, 1955)
145 included these rocks in the Precambrian basement of Lesser Kabylia. Later, by re-examining in detail
146 the outcrops of Texenna, he recognized the possible ophiolitic nature of the mafic-ultramafic
147 complex, affected by a low-grade metamorphism, which he referred to as "Sendouah-Tabellout unit"
148 and placed at the base of the Cretaceous Massylian flysch (Durand-Delga, 1971). Bouillin et al. (1977)
149 reinvestigated the magmatic complex of Texenna and the overlying series, and concluded that the
150 magmatic complex of Texenna could be considered as the upper part of an ophiolitic series.

151 The magmatic-sedimentary complex of Texenna crops out discontinuously over an area of
152 approximately 21 km² (14 km × 1.5 km). It extends from the Es-Sendouah massif in the northwest to
153 Dra-El-Fertasa in the southeast, through the village of Texenna, following the tectonic contact line
154 between the Kabyle basement and the Flysch domain (Fig. 2a, b). At the top of the Es-Sendouah
155 massif, brecciated serpentinites are crosscut by metadolerite veins (Fig. 3a), which in turn form sill-
156 like bodies, a few cm to several m thick, interspersed between calcschist and radiolarite layers near
157 the village of Texenna. Further SSE, Bouillin et al. (1977) and Djellit (1987) subdivided the volcano-
158 sedimentary complex of Texenna into two units: (1) The lower unit is composed of layered igneous
159 rocks showing variable textures (aphyric, porphyritic and coarse-grained); this unit begins with

160 pillow-lavas associated with recrystallized radiolarites, followed by igneous beds of metabasalts,
161 metadolerites and metagabbros with calcschist intercalations. (2) The upper unit consists of quartzo-
162 pelitic layers, pillow-lavas and dolerites; these rocks are overlain by an alternation of limestone and
163 lustrous calcschist layers, 1 cm to 10 cm thick. These two units, totalling 500 m in thickness, are well
164 exposed in the Tabellout massif, at Djebel Dra-El-Fertasa and in the gorge of Oued Djendjene (Fig. 3e-
165 h).

166 The magmatic complex of Texenna is overlain by carbonate rocks and radiolarites of presumed Upper
167 Jurassic-Berriasian age, in turn overlapped by Cretaceous pelites and quartzites of the Mauretanian
168 flysch (Bouillin et al., 1977; Bouillin, 1979). All these formations, which are overthrust by the high-
169 grade basement of the internal zones of Lesser Kabylia, are overturned. They are all schistosed and
170 metamorphosed, in particular the magmatic rocks of the Texenna complex which have undergone a
171 greenschist-facies metamorphism (Bouillin et al., 1977).

172 **Material and methods**

173 About 100 samples were collected along a NW-SE profile from Djebel Es-Sendouah to the gorge of
174 the Djendjene valley (Fig. 2a). Many outcrops located in the Djendjene valley to the southwest of the
175 Tabellout dam are now under water, but were surveyed before the impoundment; on the other
176 hand, the dam worksite has revealed new outcrops. Eighteen samples were selected for detailed
177 mineralogical and geochemical analyses (see details on the techniques in Sections 2 and 3 of the
178 Supplementary materials [SM]). The investigated rocks comprise ultramafic rocks (serpentinite),
179 metabasites (metagabbro, metadolerite and metabasalt with some pillow-lava structures), and
180 metasediments (radiolarite and calcschists). Thin sections were investigated under the SEM and at
181 the electron microprobe. Representative microprobe analyses of the various minerals are reported in
182 Tables A–F in SM. Serpentinite samples were also characterized by X-Ray powder diffraction (XRPD:
183 Fig. A in SM). Fifteen samples were selected for whole-rock analyses of major, trace and rare earth
184 elements (REE), using the ICP-AES and ICP-MS methods (see Section 3 and Table G in SM). In order to
185 unravel the *P-T* conditions during the tectonic emplacement, the thermodynamic modelling of a *P-T*
186 pseudosection was carried out for a foliated metabasite, using the Thermocalc software with an
187 upgraded version of the thermodynamic dataset of Holland and Powell (2011) (Section 5 in SM).

188 **Results**

189 **Field occurrence, petrology and mineral chemistry**

190 *Serpentinites*

191 Small outcrops of serpentinite are scattered in the Sendouah-Tabellout unit. The main outcrop is
192 located in the northeast, at Es-Sendouah crest, and is associated with metadolerites. The
193 serpentinites extend over 700 m in length for a maximum width of 50 m; they are generally compact,
194 sometimes brecciated, dark green to green-black in colour, contain whitish veins of chrysotile and
195 are crosscut by metadolerite dykes, 10 to 20 cm thick (Fig. 3a-b). Several serpentinite lenses are also
196 visible on the southern slope of the Tabellout Massif, about 3 km to the SSE of Texenna (Fig. 2a; *e.g.*,
197 36°38'27.9"N–5°48'6.3"E; 36°38'17.9"N–5°48'15.9"E). Durand-Delga (1955, pp. 80-83) reported a few
198 other occurrences that are no more visible, mostly along the SW tectonic contact with the
199 Cretaceous flysch: at ca. 36°38'52"N–5°47'34"E; 36°37'31.6"N–5°48'46.3"E (a 200-m long level on the
200 southern bank of Oued Djendjene); 36°37'11.3"N–5°49'21.8"E (a meter-thick lens along the contact
201 with the flysch); and in the vicinity of 36°35'29"N–5°53'26"E.

202 Under the optical microscope and the scanning electron microscope (SEM), these rocks generally
203 show mesh textures resulting from the serpentinization of olivine grains, large “bastites” formed at
204 the expense of pyroxene, pseudomorphs after Cr-rich spinel and small veins of chrysotile scattered in
205 the rock (Fig. 4).

206 The mesh texture consists of cells about 0.5 mm in diameter, with a discontinuous edge rich in tiny
207 crystals of magnetite enveloping one or more concentric layers of serpentine towards the nucleus,
208 where no relict olivine has been observed (Fig. 4b). The serpentine is here almost Al- and Ca-free and
209 of fairly homogeneous composition (Table A in SM), despite changes in the intensity of the
210 backscattered electrons (BSE) (Fig. 4b), which seem to be linked to slight variations in the H₂O and Fe
211 contents (rim vs. core in Table A). When calculated on the basis of 3 cations, in order to be compared
212 with Si₁(Mg, Fe)₂O₄ of the former olivine, the serpentine composition is on average Si_{1.337±0.069}
213 Al_{0.047±0.018} Cr_{0.001±0.000} Mg_{1.484±0.083} Fe_{0.118±0.007} Mn_{0.002±0.000} Ca_{0.007±0.002} Na_{0.002±0.002} K_{0.002±0.001} (sample
214 Serp2); the higher values of Si and Mg ($X_{Mg} = Mg/[Mg+Fe+Mn] = 0.926±0.005$) compared to a
215 common olivine can be explained by the concomitant crystallisation of magnetite, which contains
216 only traces of Si and Mg (Si_{0.008} Ti_{0.000} Cr_{0.001} Fe³⁺_{1.986} Fe²⁺_{0.992} Mn_{0.002} Mg_{0.008} O₄). XRPD analyses of
217 serpentinite shows that serpentine is predominantly lizardite (Fig. A in SM).

218 Bastites are generally 1 to 4 mm in size, but can reach 1 cm according to Durand-Delga (1955). They
219 are mainly made of serpentine, but also contain chlorite and a Ca-rich and Al-poor clin amphibole
220 (*i.e.*, tremolite: [Si_{7.54} Al_{0.46}]_T [Al_{0.28} Ti_{0.05} Cr_{0.01} Mg_{3.15} Fe²⁺_{1.51}]_C [Mn_{0.04} Ca_{1.90} Na_{0.06}]_B [Na_{0.06} K_{0.02}]_A O₂₂
221 [OH_{1.99} F_{0.01}]), arranged in parallel grooves (dashed line *in* Fig. 4c). The overall composition of bastite,

222 obtained by scanning during analysis (see 2 in SM) and calculated on the basis of 4 cations ($\text{Si}_{1.714}$
223 $\text{Al}_{0.188} \text{Cr}_{0.019} \text{Ti}_{0.002} \text{Mg}_{1.757} \text{Fe}_{0.137} \text{Mn}_{0.001} \text{Ca}_{0.177} \text{Na}_{0.004} \text{K}_{0.002}$), can be compared with some Al-rich Ca-
224 bearing orthopyroxene (ideally, $\text{Si}_{2-x} [\text{Al}, \text{Cr}]_{2x} [\text{Mg}, \text{Fe}]_{2-x-y} \text{Ca}_y \text{O}_6$), stable at elevated temperatures.
225 The Ca-rich tremolite-bearing grooves are probably derived from clinopyroxene lamellae exsolved
226 during the cooling of such a high-*T* orthopyroxene protocystal.

227 Former spinel crystals have been altered into the so-called “ferritchromit” and are surrounded by a
228 chlorite corona (Fig. 4d). Tiny areas in these “ferritchromit” pseudomorphs are rich in Cr-magnetite
229 (shown in white *in* Fig. 4d), which, being mixed with its matrix, only gave imprecise microprobe
230 analyses (“Cr-magnetite” in Fig. 5e). Other inclusions are made of Al-rich and Si-poor chlorite
231 (corundophyllite in Fig. 5f). Such a “ferritchromit” would in fact consist of a complex cryptocrystalline
232 association of Cr-magnetite, chlorite and lizardite (Mellini et al., 2005). The overall composition of
233 one of these “ferritchromit” grains (Fig. 5b; “bulk” in Table B of SM), obtained by scanning at the
234 microprobe, corresponds quite exactly to a combination of Cr-magnetite (70 mol% of theoretical
235 $\text{Cr}_{0.81}\text{Fe}^{3+}_{1.19}\text{Fe}^{2+}_{1.00}\text{O}_4$), chlorite (22 mol%) and serpentine (8 mol%), the relative amounts of which can
236 vary, thus explaining the concentric chemical variations of the pseudomorph (Fig. 4d). The Mg-rich
237 chlorite coronae around the former spinel grains present, from the matrix inwards, a progressive
238 zoning with a not surprising decrease in Si (from 6.87 to 6.00 atoms per formula unit [apfu]) and
239 increases in Al (from 2.69 to 4.15 apfu) and Cr (0.08 to 0.15 apfu), ranging from penninite to
240 clinochlore end-members (triangles *in* Fig. 5f; Table D in SM), according to Hey’s (1954)
241 nomenclature; some chlorite included in the former spinel even reaches the composition of a
242 corundophyllite, particularly poor in Si and rich in Al (Fig. 5f). This microstructure can be explained by
243 chemical gradients and exchanges between the former Cr-rich spinel and its serpentine-rich matrix,
244 Al being released from the spinel to form chlorite, while some Si and H₂O from the matrix
245 contributed to the spinel replacement.

246 Finally, numerous veinlets filled with serpentine (likely chrysotile), tremolite and/or chlorite (close to
247 the former spinel), crosscut the former microstructures (Fig. 4a, d).

248 The normative mineralogy of the peridotite protolith could be estimated by the least-square method,
249 using the bulk-rock chemical analyses of the serpentinite samples (Table G in SM). The result yields
250 an Opx-rich peridotite, with (in mol%) 53.2 ± 8.8 [1 σ] olivine, 40.1 ± 3.9 Opx, 2.4 ± 2.9 Cpx and $4.3 \pm$
251 2.2 spinel, which projects into the harzburgite field of Le Bas and Streckeisen (1991) (Fig. 6). Although
252 this normative mineralogy only gives an approximate picture of the protolith, the composition of
253 which could have been modified during serpentinization, it clearly indicates that the original
254 peridotite was a spinel harzburgite, which is corroborated by the low Ca content of the rocks and the

255 abundance of bastite pseudomorphs after Opx. Most of the serpentinite samples described by
256 Durand-Delga (1955) are rich in bastite and also derive from harzburgite.

257 XRPD spectra of serpentinite samples Serp7, Serp17 and Serp18 (Fig. A in SM) show a predominance
258 of lizardite against antigorite, reflecting an hydrothermal serpentinization at low temperature (<350
259 °C: Reynard, 2013). This serpentinization frequently preserved microstructures inherited from the
260 harzburgite protolith, like the mesh texture after olivine grains, bastite after Opx enclosing former
261 Cpx exsolution lamellae and pseudomorphs after Cr-rich spinel grains. Commonly, these
262 microstructures are almost intact, being affected only by the formation of veinlets linked to some
263 brittle deformation of the serpentinite (Fig. 4d).

264 *Metadolerites and metagabbros*

265 Metadolerites occur in the form of thin dykes that crosscut the Es-Sendouah serpentinites (Fig. 3a)
266 and sills or dykes intrusive into the calcschists (Fig. 3c). They range from a few meters to several tens
267 of meters in length with a thickness up to a few meters. The metadolerites are green to green-yellow
268 and are penetrated by whitish and greenish veins. They frequently retained the doleritic texture of
269 the protolith, showing mm-sized tabular crystals of plagioclase arranged in a matrix mainly consisting
270 of clin amphiboles (Fig. 7d-e). Other metabasites have also partially retained their magmatic
271 doleritic texture (*e.g.*, Mb11: Fig. 7a-c), but instead of occurring in the form of sills or dykes, they
272 belong to larger masses of metabasites. These rocks, referred to as metagabbros, are common in
273 Oued Missa and Tabellout massif areas (Fig. 2a).

274 The metadolerites may have preserved their primary plagioclase crystals (*e.g.*, sample Md4: Fig. 7d-
275 e), the magmatic origin of which is attested by their subhedral tabular forms, their zoning with An-
276 rich cores (Md4: An₆₈ to An₄₀, from core to rims; Fig. 5d), as well as their double Carlsbad-albite twins
277 (the Carlsbad-twin planes are visible as longitudinal grooves in Fig. 7d-e). Overgrowths of andesine-
278 oligoclase (An₃₈₋₁₇: blue areas in Fig. 7e) may have formed at least partially during metamorphism. In
279 most cases, however, the former magmatic plagioclase was completely transformed by
280 "saussuritization" into albite (An_{0.3-2.5}) associated with some epidote (sample Mb11; Fig. 7a-c).
281 Epidote, common in the most altered samples, shows a ferric molar fraction ($X_{Fe^{3+}} = Fe^{3+} / [Fe^{3+} + Al +$
282 $Cr]$), that ranges from 0.15 to 0.32 with an average of $X_{Fe^{3+}} = 0.28$, $X_{Fe^{3+}}$ of 1/3 corresponding to the
283 epidote s.s. end-member.

284 The interstices between the former plagioclase laths are mostly filled with amphibole in the less
285 metamorphosed samples (*e.g.*, Md4; Fig. 7e), or with amphibole, chlorite and titanite in the more
286 altered ones (*e.g.*, Mb11; Fig. 7b-c). In the most preserved metadolerites, calcic amphiboles, which
287 replaced the former clinopyroxene of the protolith, show a trend of composition varying between

288 actinolite, magnesiohornblende, up to ferro-pargasite and Ti-rich pargasite (Fig. 5a-c), preferentially
289 located in the vicinity of plagioclase, which would have facilitated the pargasitic substitution
290 ($\text{Na}_{+1}\text{Al}^{\text{IV}}_{+1}\text{Si}_{-1}$) by providing Al and Na. On the other hand, actinolite dominates in the most altered
291 samples, where it replaces magnesio-hornblende on the crystal edges (Fig. 7c). Chlorite is
292 predominantly pycnochlorite (Fig. 5f). The accessory minerals are chlorapatite, fluorapatite,
293 ilmenorutile and ilmenite frequently surrounded by titanite coronae. Veinlets are filled with albite,
294 epidote (Fig. 7a), rarely with potassium feldspar (Or_{96-97}), calcite and dolomite.

295 Although the normative mineralogy of the original gabbros and dolerites may be distorted by
296 changes in rock composition during metamorphism, the calculation of the C.I.P.W. norm, applied to
297 bulk-rock compositions (Table G in SM), reveals that the gabbroic protoliths were close or slightly
298 above the "critical plane of silica undersaturation", with a magmatic paragenesis of $\text{Pl} + \text{Cpx} + \text{Ol} \pm$
299 Opx . The average normative composition is in fact typical of an olivine-tholeiite composition (in
300 wt.%): 58.3 Pl ($\text{An}_{51}\text{Ab}_{44}\text{Or}_5$) + 16.3 Cpx + 14.8 Ol + 2.3 Opx + 2.9 Ilm + 1.5 Mag + 0.3 Ap + 0.1 Chr + 3.5
301 water.

302 *Pillow lavas*

303 Stacks of pillow lavas are observed in the gorge of the Djendjene valley, near the Tabellout water
304 dam (*e.g.*, 36°36'49.1"N–5°51'00.7"E), where they extend horizontally over $1.2 \times 1 \text{ km}^2$ (Fig. 2a), with
305 a thickness from a few to ~25 meters. They are interlayered in places with cm- to m-thick radiolarite
306 and calcschist beds. Because of deformation, the pillows are ellipsoidal (Fig. 3g) and their size varies
307 from a few decimetres to 1.5 metre. They are well recognizable by their concentric structure and the
308 occasional presence of a variolitic zone (Fig. 3h; *e.g.*, 36°36'58.1"N–5°51'11.4"E). Their core is
309 massive, grey and may contain partly-saussuritised plagioclase phenocrysts, whereas the rims are
310 commonly schistose, aphyric and greenish (Fig. 3e).

311 These metabasalts exhibit a hydrous greenschist-facies metamorphic paragenesis, made of albite,
312 epidote, chlorite, actinolite, titanite and apatite (Fig. 7f-k). They ordinarily do not show relicts of the
313 magmatic assemblage, but instead have commonly preserved microstructures that make it possible
314 to recognize the typical original features of an unmetamorphosed pillow basalt, like the one dredged
315 at the Southwest Indian Ridge and shown in Figure B (in SM):

316 (i) *Textural zoning*: The detailed study of a section through a pillow shows several concentric zones
317 (Fig. 7f): (a) A thin outer zone with microcrystalline interstitial texture, consisting predominantly of
318 chlorite and minor amounts of epidote and titanite (Z1 in Fig. 7f); (b) variolitic zones, composed of
319 chlorite, actinolite and epidote (Z2 in Fig. 7f), grading into epidote, albite and actinolite (Z3 in Fig. 7f);
320 (c) an inner zone, where epidote with inclusions of albite is dominating (Z4 in Fig. 7f). These zones

321 probably correspond to the classical concentric zoning that frequently characterizes unaltered pillow
322 lavas, with glassy, variolitic and spherulitic zones (Fig. B in SM).

323 (ii) *Magmatic flow*: Trails of micro-crystals, especially of titanite, visible in the Z₁ outer zone, close to
324 the surface of the pillow and subparallel to it, could be the vestiges of magmatic flow (1 in Fig. 7g).

325 (iii) *Phenocrysts*: Polycrystalline clusters of epidote show regular external shapes (2 in Fig. 7g-h),
326 which strongly suggest that they are pseudomorphs after mm-sized phenocrysts of An-rich
327 plagioclase, transformed into epidote during metamorphism. Some porphyritic pillow lavas contain
328 still-preserved zoned plagioclase phenocrysts, ranging in size from a few to ~10 mm. In sample Pil13,
329 these incompletely-saussuritised phenocrysts still show a clear chemical zoning between the
330 apparent centre (An₂₃) and the rims (An₄).

331 (iv) *Devitrification spherulites*: The greenish varioles visible to the naked eye appear as globular
332 epidote clusters with many albite micro-inclusions under the microscope (3 in Fig. 7g-k). We interpret
333 them as deriving from devitrification spherulites of the basaltic protolith (see, *e.g.*, Arndt and Fowler,
334 2004), for the following reasons: (a) These globules increase in size and density from the pillow edge
335 (Z₁), where they are isolated within chlorite (3 in Fig. 7g), towards the centre of the pillow, where
336 they become contiguous (3 in Fig. 7i) and finally merge (3 in Fig. 7j); this reflects a common feature of
337 the devitrification spherulites, which are progressively less abundant towards the glassy quenched
338 crust (Fig. B). (b) Albite micro-inclusions are often aligned and sheave-shaped inside each epidote
339 globule, similarly to the common fibroradiate microstructure of former spherulites (3 in Fig. 7i). (c)
340 Irregular thin coronae of the same Ep+Ab association almost completely surround some
341 pseudomorphs after plagioclase phenocrysts (3 around 2 in Fig. 7h), which finds a simple explanation
342 by considering that devitrification commonly initiates on the surface of pre-existing plagioclase
343 crystals (Fig. B). (d) Devitrification spherulites in basaltic rocks are essentially made of
344 cryptocrystalline plagioclase, which indeed should evolve into Ep+Ab during low-grade
345 metamorphism.

346 (v) *Microliths*: In the inner zone of the pillow (Z₄), tabular 0.1-mm sized albite microcrystals could
347 derive from former plagioclase microliths (4 in Fig. 7j).

348 (vi) *Gas vesicles*: Some 0.1-mm spherules behaved like microgeodes, filled with concentric micro-
349 layers of epidote (Pil15: $X_{Fe^{3+}} = Fe^{3+} / [Fe^{3+} + Al + Cr] = 0.326 \pm 0.011$), clinozoisite/epidote ($X_{Fe^{3+}} =$
350 0.23–0.13) and chlorite (5 in Fig. 7k). In some samples (*e.g.*, Pil13), eye-shaped vacuoles of several
351 mm are now filled, at the margins, with concentric euhedral growth zones evolving from epidote
352 ($X_{Fe^{3+}} = 0.32$) to clinozoisite ($X_{Fe^{3+}} = 0.11$) and, towards the centre, with chlorite that occupies most of
353 the cavity. All are certainly derived from gas vesicles in the basalt protolith.

354 In most metabasalts, plagioclase is almost pure albite ($Ab_{98.7\pm 1.1}$), but its composition also extends in
355 the oligoclase range, up to An_{28} when plagioclase saussuritization was incomplete (*e.g.*, Pil13; red
356 squares in Fig. 5d). Epidote is close to the epidote s.s. end-member (typically, $X_{Fe^{3+}} \approx 0.32$); it is
357 homogeneous in all sites except in the former gas vesicles (see above). Amphibole is most commonly
358 actinolite or magnesio-hornblende (Fig. 5b), and chlorite ranges in composition essentially from
359 ripidolite to pynochlorite (Fig. 5f). Titanite is common and often occurs as coronae around relict
360 ilmenite. Fluorapatite, subhedral magnetite and chalcopyrite have been observed. A few scarce
361 euhedral crystals of spinel, whose chemical composition varies between Al-chromite and picotite
362 (sample Pil13: Fig. 5e; Table B in SM), are corroded and surrounded by coronae of Al-rich Cr-bearing
363 chlorite (Pil13 in Table D; ripidolite in Fig. 5f), similarly to what is observed in serpentinites.

364 The major-element composition of the pillow lavas (Table G in SM) is highly variable, and the
365 calculation of their C.I.P.W. norm gives irregular results. The normative composition of feldspars in
366 particular is erratic and more often Ab-rich than An-rich ($An_{37\pm 16} Ab_{57\pm 18} Or_{6\pm 3}$), which is manifested in
367 the rocks by the changing behaviour of the original plagioclase phenocrysts, sometimes preserved
368 and quite sodic (Pil13) or sometimes replaced by epidote pseudomorphs and thus initially Ca-rich
369 (*e.g.*, Pil16). Such characteristics, in particular an enrichment in Na, can be attributed to some
370 submarine weathering of the pillow lavas, which would therefore deserve the name “spilites” rather
371 than “metabasalts”.

372 *Foliated metabasites*

373 In some foliated and banded metabasites (Fig. 7l), the deformation has destroyed all traces of
374 the magmatic microstructures, making it difficult to diagnose what the original rock was,
375 except when a transition with an undeformed metabasite is observed, which is sometimes the
376 case in the vicinity of pillow lavas (Fig. 3e). These metabasites generally consist of a fine-
377 grained assemblage of clinoamphibole (actinolite to magnesio-hornblende), chlorite
378 (pynochlorite), Fe^{3+} -rich epidote, albite (typically, $Ab_{99.8-96.6}$) sometimes coexisting with
379 oligoclase, titanite, calcite, with some accessory minerals like apatite and magnetite
380 surrounded by hematite. In one sample (Pil14), small crystals of phengite
381 ($Ms_{56}Ce_{132}FeCe_{15}Pg_3Mrg_5$), with a significant phengitic $Si_{+1}Al^{IV}_{-1}Al^{VI}_{-1}(Mg, Fe)_{+1}$ substitution
382 ($Si = 6.525 \pm 0.101$ [1 σ] apfu), were observed.

383 Variations in the relative abundance of these minerals and grain size determine the banding.
384 Fairly homogeneous ribbons, rich in calcite, in epidote (Fig. 7l) or more rarely in dolomite or
385 quartz, seem to result from the transposition parallel to the foliation of early veinlets. In the

386 transposed veins of epidote, the latter is truncated and its gaps filled with albite and/or calcite
387 (Fig. 7l). On the other hand, late veinlets of microcline intersect the foliation.

388 ***Radiolarite and calcschist***

389 Metasediments, made up of metaradiolarites and calcschists, overlie stratigraphically the pillow-
390 lavas. The gradual transition between these two formations is best seen on both sides of the gorge of
391 Djendjene river where alternating layers of metaradiolarites, lavas and calcschists are observed (Fig.
392 3f).

393 Metaradiolarites, called “jaspéroïdes” by Durand-Delga (1955), deposited as 1- to 10-cm thick
394 varicoloured beds. Reddish to greenish beds consists of quartz, ankerite ($\text{Ca}_{52}\text{Mg}_{32}\text{Sd}_{15}\text{Rds}_1$)
395 commonly concentrated in veinlets with nuclei of Fe-bearing dolomite ($\text{Ca}_{53}\text{Mg}_{43}\text{Sd}_4\text{Rds}_0$), oxidized
396 botryoidal pyrite, minute inclusions of clay, apatite and TiO_2 . Whitish beds are composed of
397 microcrystalline quartz with chalcedony cement. No radiolarian fossils were found in our samples
398 due to intense recrystallisation. However, further west, in the Col de Chellata massif (Fig. 1), Gélard
399 (1969) reported on the discovery in a red radiolarite layer of 21 radiolarian species of Late Jurassic
400 (Kimmeridgian) age. This layer is located at the base of the Mauretanian flysch, in the same
401 stratigraphic position as the Texenna metaradiolarites.

402 The calcschist beds range in thickness from a few decimetres to a few metres. Sample Cal consists of
403 quartz, calcite ($\text{Ca}_{98}\text{Mg}_2\text{Sd}_0$), dolomite ($\text{Ca}_{51}\text{Mg}_{46}\text{Sd}_3$), chlorite, fluorapatite, hematite and white
404 mica ($\text{Ms}_{63}\text{Cel}_{30}\text{FeCel}_2\text{Pg}_4\text{Mrg}_1$) with a significant phengitic substitution ($\text{Si} = 6.654 \pm 0.052 [1 \sigma]$ apfu).

405 **Geochemistry**

406 Most of the studied rocks have undergone a greenschist-facies metamorphism implying selective
407 element mobility, especially for K, Na and the large ion lithophile elements (LILE: Cs, Rb, Ba, Sr....).
408 For this reason, we use high field strength elements (HFSE: Ti, P, Zr, Y, Nb, Th...), some transition
409 metals (*e.g.*, Ni, Co, Cr, V) and rare-earth elements (REE) in our geochemical and petrogenetic
410 interpretations (Table G in SM), since these elements are identified as almost immobile during rock
411 weathering and metamorphism in general, especially in the case of ophiolites (Pearce, 2014).

412 Loss-on-ignition (LOI) values of the serpentinite samples range from 11.8 to 13.6 wt.%. These
413 elevated values are indeed due to hydration during serpentinization. The analysed serpentinites are
414 highly magnesian (32.16–36.48 wt.% MgO) and have moderate Fe_{tot} concentrations ranging from
415 7.68 to 8.45 wt.% of Fe_2O_3 . Their Mg number [$\text{Mg\#} = 100 \times \text{Mg}^{2+} / (\text{Mg}^{2+} + \text{Fe}^{2+})$] is elevated and varies
416 from 90 to 92, which is typical of upper mantle peridotites. The Al_2O_3 and CaO concentrations are
417 very low, ranging from 1 to 3.22 wt.% and from 0.04 to 1.54 wt.% respectively, in accordance with

418 the absence of plagioclase and the low amount of clinopyroxene in the harzburgite protolith (Fig. 6;
419 see above).

420 Metabasites show a wide range of LOI value (from 1.18 to 7.81 wt.%), which is lowest in
421 metadolerites and highest in metabasites with pillow-lava structures. This shows that the latter have
422 undergone an intensive alteration, as also indicated by the normative calculation (see above). The
423 metabasites have low SiO₂ and low to moderate TiO₂ contents, ranging from 44.6 to 49.2 wt.% and
424 from 1.05 to 2.17 wt.%, respectively. The MgO concentrations range from 3.47 to 9.29 wt.%, with the
425 lowest concentrations observed in the pillow lavas (3.47 to 3.81 wt.%). These low values do not
426 result from magmatic differentiation, but are rather due to the stronger alteration of the pillow-lava
427 samples. The metabasites display moderate Fe (7.41–11.75 wt.% of Fe₂O₃) and CaO (8.89–12.93
428 wt.%) contents and high Al₂O₃ (13.95–17.65 wt.%).

429 On the Zr/Ti vs. Nb/Y diagram (Pearce, 1996), the metabasites have basaltic subalkaline compositions
430 (Fig. 8a). Their tholeiitic nature is suggested by the Zr–P₂O₅ diagram (Floyd and Winchester, 1975)
431 used for distinguishing between alkaline and tholeiitic affinities (Fig. 8b), and by Zr–Y, La–Yb and Th–
432 Yb diagrams (Ross and Bédard, 2009) used for discriminating calc-alkaline and tholeiitic rocks (Fig. 8c–
433 e). In these latter three diagrams, some samples are plotted in the transitional field, a word used
434 here for rocks that plot at the tholeiitic/calc-alkaline boundary on the AFM diagram (Ross and
435 Bédard, 2009).

436 Chondrite-normalized REE patterns of the metabasites vary between two extremes (Fig. 9a, c, e). The
437 patterns of the metagabbro Mb9 and metadolerite Md5 show erratic variations probably due to
438 important secondary alteration and will not be considered hereafter. Pillow-lavas have nearly flat to
439 weakly depleted REE patterns with La_N/Yb_N ranging from 0.66 to 0.97 (Fig. 9a), whereas metagabbros
440 are enriched in Light REEs (LREE) with La_N/Yb_N ranging from 2.18 to 2.24, and show flat High REE
441 (HREE) profiles (Fig. 9e). The REE patterns of metadolerites are comparable to those of pillow lavas,
442 but with a slight enrichment in LREEs (La_N/Yb_N between 1.07 and 1.66: Fig. 9c). Thus, the REE patterns
443 of pillow-lavas, metadolerites and metagabbros are similar to those observed respectively in normal
444 middle-ocean ridge basalts (N-MORBs), transitional T-MORBs and enriched E-MORBs (*e.g.*, Gale et al.,
445 2013).

446 The same characteristics are shown in the N-MORB-normalized trace-element patterns (Fig. 9b, d, f),
447 in which pillow-lava samples are very close to 1 (with a slight Nb-Ta depletion) indicating their N-
448 MORB affinity whereas the metagabbros show E-MORB patterns.

449 Discussion

450 **Petrogenetic evolution of the Texenna ophiolites**

451 *Protoliths and geodynamic context*

452 As pointed out above, the REE and multielement patterns of the Texenna metabasites correspond
453 mainly to N-MORB for the pillow lavas, T-MORB for the metadolerites and E-MORB for the
454 metagabbros. We use hereafter some discrimination diagrams to further refine the type of oceanic
455 environment in which the rocks studied were formed.

456 In the V vs. Ti discrimination diagram (Shervais, 1982), most Texenna metabasic rocks plot in the
457 MORB/back-arc basin basalt (BABB) fields ($Ti/V = 18-48$; Fig. 10a). On the Nb–Zr–Y diagram
458 (Meschede, 1986), the pillow lava samples plot in the field of N-type MORB, metadolerites in within-
459 plate tholeiites (WPT) and volcanic-arc basalt (VAB) fields, whereas metagabbros plot in E-MORB field
460 (Fig. 10b). All the samples plot in the oceanic basalt domain of the $La/10-Y/15-Nb/8$ discrimination
461 diagram of Cabanis and Lecolle (1989) (Fig. 10c), which defines also the pillow lava samples as N-
462 MORB, the metagabbros as E-MORB and the metadolerite rocks as E-MORB falling between the
463 pillow lavas and the metagabbros (Fig. 10c). The same results are seen in the Th–Ta–Hf/3 diagram of
464 Wood (1980), where pillow-lava and Md6 metadolerite samples plot in N-MORB field, and other
465 metadolerites and metagabbros in E-MORB field (Fig. 10d).

466 In recent years, powerful diagrams to fingerprint tectonic settings of ophiolites have become
467 available (Pearce, 2008, 2014). These include the Th/Yb vs. Nb/Yb diagram where rocks from mid-
468 ocean ridge and ocean island basalts (OIB) are plotted diagonally along a MORB-OIB array, and the
469 Nb/Yb vs. TiO_2/Yb diagram, which allows discriminating between different types of MORBs. In the
470 Th/Yb vs. Nb/Yb diagram, the Texenna mafic rocks plot within the MORB-OIB array (Fig. 10e),
471 whereas in the Nb/Yb vs. TiO_2/Yb diagram pillow lavas are clearly N-MORB in character and
472 metagabbros are E-MORB (Fig. 10f).

473 Discrimination diagrams corroborate an oceanic crust affinity of the studied metabasites, with N-
474 MORB signature for the pillow lava samples, T-MORB for metadolerites and E-MORB for
475 metagabbros. Furthermore, the association in the Texenna area of serpentinite, metagabbros,
476 metadolerites, pillow lavas, radiolarites and calcschists is indeed typical of ophiolites, formed in an
477 oceanic environment. Following the classification scheme of Dilek and Furnes (2011, 2014), these
478 may be classified as subduction-unrelated ophiolites that evolved during rift drift and seafloor
479 spreading.

480 *Greenschist-facies hydrothermal oceanic metamorphism*

481 Despite complete serpentinization, Texenna serpentinites very commonly show microstructures
482 inherited from the original peridotites, such as serpentine pseudomorphs after olivine (*i.e.*, mesh
483 texture) and pseudomorphs after orthopyroxene (*i.e.*, bastite), in which even traces of former
484 exsolution lamellae are still visible. Their mineralogical composition is characterized by lizardite in
485 both mesh textures and bastites, magnetite, “ferritchromit” with chlorite coronae replacing former
486 Cr-spinel, and small veins of chrysotile and chlorite. All these textural and mineralogical features, in
487 particular the alteration of Cr-spinel (*e.g.*, Burkhard, 1993; Mellini et al., 2005; Kapsiotis et al., 2007),
488 are well documented in ophiolite serpentinites. They are also typical of the ocean floor serpentinites,
489 the hydrothermal serpentinization of which occurs in the range of 300–500°C (Mével, 2003).

490 Metadolerites, metagabbros and pillow lavas are weakly deformed since they have preserved their
491 magmatic microstructures, only intersected by fractures filled with albite, epidote and/or calcite.
492 Their mineral assemblages, with albite, chlorite, epidote, titanite and actinolite, are typical of
493 greenschist-facies metamorphic alteration of basic rocks. Thompson (1991), in particular, described
494 such alteration of pillowed basalts in the dredged metabasalts from the ocean floor. Furthermore,
495 most of the chlorites from the Texenna metabasites plot in the field corresponding to greenschist-
496 facies seafloor hydrothermal alteration in the Si vs. Fe/(Fe+Mg) diagram (Fig. 5f; see Nimis et al.,
497 2004, and references therein).

498 In summary, many rocks of Texenna, either serpentinites, metabasites or pillow lavas, have
499 preserved the original texture of their protolith and are therefore weakly deformed. The magmatic
500 minerals of the protoliths were very generally replaced by low-grade greenschist-facies hydrous
501 metamorphic assemblages, accompanied by secondary mineral precipitation in veins and fractures.
502 These features are typical of hydrothermal oceanic metamorphism, during which strain is reduced to
503 fracturing, under temperatures in the range of 300–500°C (Hekinian, 1982; Mével, 2003). The
504 abundance of carbonates, both in metasediments (calcschists and ankerite-bearing radiolarites) and
505 as veinlets in metabasites, indicates that this hydrothermalism occurred above the carbonate
506 compensation depth.

507 ***Metamorphism and deformation related to the Maghrebide orogen***

508 Some of the Texenna metabasites are deformed, showing prominent banding and foliation (see
509 above). The foliation of these rocks is sub-parallel to the foliation in the neighbouring geological
510 formations, such as the Kabyle basement, the associated calcschists and the Massylian flysch, so the
511 deformation appears to be related to the tectonic emplacement of these units, well after the oceanic
512 hydrothermal metamorphism. The mineralogy of these metabasites is not fundamentally different
513 from that of undeformed metabasites, metamorphosed under greenschist-facies hydrothermal

514 oceanic conditions. One notable difference, however, appears to be the presence of phengite (Pil14),
515 also observed in some calcschists (sample Cal); this relatively high-*P* mineral cannot have formed
516 under sub-surface conditions during the hydrothermal oceanic metamorphism.

517 Specifying the *P-T* conditions of this metamorphism is not as simple as it seems. While modelling a *P-*
518 *T* pseudosection requires ideally a homogeneous rock that has reached equilibrium, most of these
519 rocks are banded and may have preserved minerals from the oceanic hydrothermal metamorphic
520 stage and even from the magmatic protolith, like ilmenite in the core of titanite crystals; they are
521 also crosscut by late calcite and feldspar veinlets, which have modified the bulk composition. We
522 have chosen to model a *P-T* pseudosection for the Pil14 metabasite, derived from pillow lavas but
523 strongly deformed and foliated during this metamorphism. The chemical composition used for the
524 modelling, including the Fe³⁺ content, was obtained from the molar modal quantities of the minerals
525 in the rock, calculated by the least-square method from the bulk-rock and mineral compositions, and
526 from which apatite and calcite, not taken into account in the model, were subtracted. Low albite (Ab)
527 was considered in addition to plagioclase (Pl), since the “peristerite” miscibility gap can occur in such
528 a low-grade metamorphism. Details on the thermodynamic database and solid-solution models are
529 given in Section 5 of the SM.

530 The *P-T* conditions of the metamorphism related to deformation should correspond to one of the
531 fields where the minerals ubiquitous in the rock are stable (Ab, Ep, Chl, Act/Hbl, Tnt, Ms/Ph,
532 ±hematite), *i.e.*, in the vicinity of the low-*P*–low-*T* / high-*P*–high-*T* diagonal of the *P-T* pseudosection
533 (Fig. 11). Glaucofane-bearing high-*P*–low-*T* fields should be excluded, since this mineral was never
534 observed, as well as the low-*P*–high-*T* fields in which epidote is absent and plagioclase too An-rich
535 (An_{21–38}) relative to the real values. Because of the banding and heterogeneity of the rock, the use of
536 isopleths is somewhat hazardous. However, the Si-in-phengite isopleths (Si = 6.525 ± 0.101 [$\pm 1\sigma$])
537 make it possible to better specify the *P-T* domain for this metamorphism (Fig. 11). The stability of
538 some Kfs predicted by the model is in agreement with the presence of late veinlets of K-feldspar. It
539 therefore appears that the metamorphism linked to the tectonic emplacement of the Texenna
540 ophiolites in the Maghrebide chain occurred under low-grade greenschist-facies conditions, following
541 a medium geothermal gradient that is not linked to a subduction zone.

542 **Comparison with other ophiolites of the western Mediterranean region**

543 Numerous ophiolitic outcrops occur in the Alpine orogen of the Western Mediterranean region, *e.g.*,
544 in the Alps, in North-eastern Corsica, in the Northern and Southern Apennines, in Sicily, in the Rif
545 (Morocco) and in the Betic Cordillera (southern Spain) (Principi et al., 2004; Bortolotti and Principi,

546 2005; Guerrero et al., 2005). These ophiolites are considered to represent the records of the Jurassic
547 Western Tethys Ocean, and are characterized by the following features (Principi et al., 2004): (i)
548 Middle-Late Jurassic age; (ii) MORB signature of the mafic rocks; (iii) reduced and thin ophiolitic
549 successions; (iv) exposure of serpentinites at the base of sedimentary-volcanic covers; and (v)
550 widespread presence of radiolarites above the magmatic section. It is prominent to note that all of
551 these features can be seen in the Texenna volcano-sedimentary complex.

552 Recently, ophiolite slivers with gabbros, altered basalts and oceanic sediments have been discovered
553 in the core of the external zones of the Central Rif belt in Morocco, far from the Maghrebian Flysch
554 suture zone (Benzaggagh et al., 2014; Michard et al., 2014). These rocks define two major lineaments
555 in the Mesorif (Michard et al., 2018; Fig. 1): (i) The Mesorif suture zone (MSZ) shows slivers of
556 serpentinites, metagabbros, metabasalts, associated with oceanic metasediments. A gabbro of the
557 MSZ gave a radiometric age of 190 Ma (Michard et al., 2018). (ii) The Mesorif Basalts–Breccias
558 lineament (MBB), composed of Kimmeridgian–Berriasian basalt flows and carbonates breccias, is
559 more external and parallel to the MSZ.

560 The metabasic rocks of Alpine ophiolites of the Western Mediterranean region show compositions
561 from N-MORB to E-MORB:

562 (a) In the Chenaillet ophiolitic complex (Ligurian-Piemonte zone in the Western Alps), the mafic rocks
563 are predominantly N-MORB (Chalot-Prat, 2005) and this complex, as well as other Western Alpine
564 ophiolites exposed in NW Italy, is considered as the testimony of a segment of an axial volcanic ridge
565 in a slow-spreading ridge within a restricted ocean basin (Balestro et al., 2019; Chalot-Prat, 2005).

566 (b) Two groups of ophiolites are distinguished in Corsica (Saccani et al., 2008): (i) a most prevalent
567 group with N-MORB affinity and (ii) a group with T-MORB composition restricted to Santo di Tenda,
568 Nebbio and upper part of Balagne units. The N-MORB group is considered to have been formed in
569 the central part of the ocean after the onset of the oceanic spreading, whereas the T-MORB rocks
570 would have originated in the ocean-continent transition zone during the first stage of oceanic
571 spreading.

572 (c) The metabasites of the ophiolitic sequences of the Calabrian Arc (Southern Apennines) have N-
573 MORB (Sansone et al., 2011) to T-MORB tholeiitic affinities (Liberi et al., 2006).

574 (d) In Sicily and Moroccan Rif, the metabasites show E-MORB characteristics (Durand-Delga et al.,
575 2000) and are thought to be formed in a rift environment during the initial stage of oceanisation
576 (Durand-Delga et al., 2000).

577 (e) The metabasites of the Betic ophiolites show mainly E-MORB or T-MORB affinities (Bodinier et al.,
578 1987; Puga et al., 2011). They are interpreted as formed in ultraslow-spreading ridge environment,
579 during the initial stages of oceanisation, and are classified as “continental margin ophiolites” (Puga et
580 al., 2017). The Betic oceanic sector was located at the westernmost end of the Western Tethys
581 Ocean. However, a few authors proposed a continental setting for the metabasic igneous rocks of the
582 Betics (e.g., Gomez-Pugnaire, 2019, and references therein).

583 In Th/Yb vs. Nb/Yb diagram, the Texenna pillow lavas, with N-MORB affinity, lie mainly within the
584 Chenaillet and most of the Corsica mafic rocks array (Fig. 12), whereas metagabbros and
585 metadolerites show affinities with mafic ophiolitic rocks of the Rif, Sicily, the Betics and with the
586 recently studied metabasites of Eastern Elba Island (Northern Apennines, Italy; Bianco et al., 2019)
587 (Fig. 12). Finally, the Kimmeridgian–Berriasian basalts of the MBB and the gabbros of the MSZ of the
588 Central Rif belt in Morocco show E-MORB affinities and differ from the Texenna metabasites, in the
589 Th/Yb–Nb/Yb discrimination diagram (Fig. 12). The gabbros of the MSZ plot between E-MORB and
590 OIB, while basalts of the MBB plot above the MORB-OIB array, with high Th/Yb ratios probably due to
591 interaction of these basalts with crustal rocks of the African palaeomargin (Benzaggagh et al., 2014).

592 The Texenna serpentinites share the same features as many Alpine ophiolitic serpentinites, like those
593 at Chenaillet (Thiéblemont et al., 2019): high serpentinization degree; lizardite as serpentine
594 polymorph; alteration of Cr-spinel to “ferritchromit” surrounded by chlorite coronae; harzburgite
595 protolith. On the other hand, the studied serpentinites differ notably from the Collo peridotites and
596 serpentinites, which crop out within the metamorphic basement of Lesser Kabylia not far from the
597 Texenna area. Collo ultramafic rocks are much less serpentinized than those of Texenna, with
598 antigorite instead of lizardite as the dominant serpentine polymorph; they display anhydrous mineral
599 assemblages, unaltered Cr-spinels and their protolith is lherzolitic (Laouar et al., 2017). Collo
600 ultramafic rocks belong to Alpine-type peridotites that were emplaced within an orogenic setting
601 (Bouillin and Kornprobst, 1974; Laouar et al., 2017) and are similar to other peridotite massifs of the
602 Western Mediterranean region: Lherz (Pyrenees); Lanzo massif (Italian Alps); Ronda (Betic
603 Cordillera); and Beni Bousera (the Moroccan Rif).

604 **Geodynamic implications**

605 This study demonstrates for the first time that the Texenna rocks belong to a true ophiolitic slice.
606 Contrary to what has been suggested by some authors, the Maghrebian Flysch suture zone is marked
607 by outcrops of ophiolites, as any other suture zone of the Circum-Mediterranean Alpine Belts, and
608 the sedimentary cover of the Flysch domain was here deposited over a true oceanic rather than a

609 thinned continental crust. The Texenna ophiolites show analogies with other ophiolites from the
610 Western Mediterranean region, indicating that they are remnants of the ancient oceanic domain
611 formed in the southwestern part of the Western Tethys (Fig. 13a), more precisely in the Maghrebian
612 Tethys ocean that developed between Africa and Iberia during the Middle-Late Jurassic (*e.g.*,
613 Fernandez, 2019).

614 The Texenna metabasites show T-MORB to E-MORB affinities for the metadolerites and metagabbros
615 and N-MORB signature for the pillow lavas. This can be interpreted in two ways:

616 (a) The pillow lavas and the intrusive rocks are coeval and were emplaced in the same area, which is
617 supported by the association of N-MORB to E-MORB basalts in modern ultraslow-spreading ridge,
618 such as the Southwest Indian Ridge (Standish et al., 2008). This however seems contradictory with
619 the harzburgite composition of the serpentinite protoliths, which would rather indicate a formation
620 at a fast-spreading ridge.

621 (b) The intrusive rocks are not coeval with the extrusive igneous rocks (*i.e.*, the pillow lavas). The
622 gabbros and dolerites may have emplaced at an early stage of oceanic accretion, which could explain
623 their enriched geochemical signature. After the onset of oceanic spreading, basaltic magmas with an
624 N-MORB affinity would have produced the pillow lavas observed.

625 The latter scenario, proposed for the Betic ophiolites (Puga et al., 2011), would apply well to the
626 Texenna ophiolites; it is similar to the tectonic models of the Alpine Tethys evolution developed by
627 Balestro et al. (2019) and of the Mesozoic margin evolution of the Maghrebian Tethys invoked by
628 Michard et al. (2020) for the Rif belt (Morocco). According to this model, the Alpine Tethys was
629 formed in four stages between the Middle Jurassic to Early Cretaceous involving seafloor spreading
630 of a narrow ocean basin: (1) emplacement of gabbro bodies and dolerite dykes in the upper mantle
631 peridotites; (2) extensional deformation, upper mantle exhumation-denudation and doming; (3)
632 eruption of basaltic lavas into the exhumed peridotites and gabbroic plutons and (4) deposit of
633 pelagic sediments (radiolarites, marble and calcschist). The emplacement of the intrusive rocks
634 (gabbros, dolerites) in the Texenna complex would correspond to the phase 1 of this model, in
635 accordance with their E-MORB affinity and association with serpentinite in the field. The pillow lavas
636 emplacement is likely related to phase 3, which would explain their N-MORB composition. Finally,
637 the overlying radiolarites are linked to phase 4.

638 Bouillin (1992) proposed that the Texenna rocks were close to a NW-SE Jurassic transform fault that
639 brought the Lesser Kabylia basement, to the NE, into direct contact with the Maghrebian oceanic
640 basin, to the SW (Fig. 13b). This hypothesis elegantly accounts for (a) the apparent dextral shift
641 between the continental basements of Lesser and Greater Kabylia, (b) the absence near Texenna of

642 the “Dorsale calcaire”, considered elsewhere as the remnant of the former southern passive margin
643 of the Kabyle microcontinent, and (c) the thrusting of the Lesser Kabylia basement directly upon the
644 ophiolites (Bouillin, 1992). This NW-SE transform fault would be one of the many later proposed by
645 Schettino and Turco (2011) in their palaeogeographic reconstruction of the western Tethys (Fig. 13a).

646 The Maghrebide orogeny partly deformed, metamorphosed and dismembered the ophiolites of
647 Texenna, the original sequence of which is hardly recognizable, with serpentinites, on the one hand,
648 and pillow lavas and metasediments, on the other hand, unclearly concentrated on the SW and NE
649 edges, respectively, of the tectonic slice.

650 Ophiolitic outcrops are much less common in the Maghrebide chain of North Africa than in other
651 regions. A possible reason for the extreme scarcity of ophiolite remnants in the Maghrebian Flysch
652 suture zone is the fact that the basement of the Maghrebian Flysch nappes is rooted deeply beneath
653 the internal zone. Furthermore, the geological formations at the front of the internal zones are highly
654 dilacerated and dismembered, which led to the dispersion in the suture zone of the different units of
655 the ophiolitic sequence into slivers, separated by long distances.

656 These ophiolite slivers were often considered in ancient literature as part of the metamorphic
657 basement, like the Texenna rocks, long regarded as belonging to the Kabyle basement. Since the
658 suture(s) zone(s) of the Maghrebide chain is(are) usually outlined by Triassic outcrops, they were also
659 frequently mistaken for Triassic "ophites" (*e.g.*, Ben Yaich et al., 1989). The mafic and ultramafic
660 rocks that crop out in the core of the external zones of the Central Rif belt, supposed until recently to
661 be basement units or Triassic ophites, have been recognized as part of Middle-Upper Jurassic
662 ophiolite slivers, defining a second suture zone within the Maghrebide chain, between the Intrarif
663 and Mesorif (External Zones, Morocco) (Benzaggagh et al., 2014). A major challenge for the future is
664 therefore to investigate magmatic-sedimentary associations, very little studied so far, which could
665 reveal new sutures in the Maghrebide chain.

666 **Conclusions**

- 667 (i) An association of mafic and ultramafic rocks occurs beneath the Maghrebian flysch nappes at
668 the southern front of the internal zones of Lesser Kabylia, in the Texenna region (NE Algeria).
669 This association includes serpentinite, metagabbro, metadolerite, metabasalt with pillow-lava
670 structures, and oceanic metasediments (radiolarite and calcschists).
- 671 (ii) Serpentinites are mainly composed of lizardite pseudomorphs after olivine and
672 orthopyroxene, with small veins of chrysotile and chlorite scattered throughout the rocks.

673 Metadolerites and metagabbros have sometimes preserved their magmatic zoned plagioclase
674 (labrador to andesine), but are generally transformed into an assemblage of albite,
675 actinolite/magnesian hornblende, epidote, chlorite and titanite, in the same way as the
676 metabasalts, which furthermore commonly show inherited microstructures typical of pillow
677 lavas (concentric former glassy, variolitic and spherulitic zones; recrystallized devitrification
678 spherulites; epidote pseudomorphs after plagioclase phenocrysts; gas vesicles filled with
679 epidote and chlorite). The greenschist-facies static metamorphism that affected the Texenna
680 metabasites and serpentinites is typical of hydrothermal alteration of ocean floor rocks at
681 slow-spreading ridges. After the oceanic hydrothermal metamorphism, the Texenna
682 ophiolites underwent some deformation accompanied by greenschist-facies phengite-bearing
683 metamorphism, linked to the tectonic emplacement of the ophiolites in their current setting
684 during the Maghrebide orogen.

685 (iii) Whole-rock compositions of the serpentinites indicate a harzburgite protolith, whereas the
686 metabasic rocks are tholeiitic and show N-MORB signature for the pillow-lavas and from T-
687 MORB to E-MORB affinities for the intrusive rocks (metadolerites and metagabbros). This
688 duality is common among the ophiolites of the western Mediterranean region.

689 (iv) This study demonstrates for the first time that the Texenna rocks belong to a true ophiolitic
690 slice, and indicates that the sedimentary cover of the Maghrebian Flysch domain was
691 deposited over a true oceanic rather than a thinned continental crust, as several researchers
692 often suggest. The Texenna ophiolites represent a fragment of the ancient oceanic
693 lithosphere (Maghrebian Tethys Ocean) that formed between Africa and Iberia-AlKaPeCa
694 during the Middle-Upper Jurassic.

695 **Acknowledgments**

696 We would like to warmly thank Jean-Pierre Bouillin for his invaluable advices, especially concerning
697 the metabasites and pillow lavas of the Tabellout region, as well as the head office of the Tabellout
698 dam for allowing us to access the outcrops. Grateful thanks are also due to the *Direction Générale de*
699 *la Recherche Scientifique et du Développement Technologique* for funding this work. Moulley Charaf
700 Chabou is supported by the *Ministère de l'Enseignement Supérieur et de la Recherche Scientifique*,
701 PRFU project n°E04N01UN190120180001. André Michard and an anonymous reviewer are sincerely
702 thanked for their constructive reviews, as well as Marco Scambelluri for the editorial handling.

703 **Bibliography**

- 704 Andrieux, J., Djellit, H., 1989. Structure de la Petite Kabylie occidentale (Algérie): flyschs «ultra» et
705 flyschs externes. *Comptes rendus de l'Académie des sciences, Série II*, 309, 1191–1196.
- 706 Andrieux, J., Frizon de Lamotte, D., Braud, J., 1989. A structural scheme for the western
707 Mediterranean area in Jurassic and Early Cretaceous times. *Geodinamica Acta* 3, 5–15.
- 708 Arndt, N., Fowler, A., 2004. Textures in komatiites and variolitic basalts, in: Eriksson, P.G., Altermann,
709 W., Nelson, D.R., Mueller, W.U., Catuneanu, O. (Eds.), *The Precambrian Earth: tempos and*
710 *events*. Elsevier, *Developments in Precambrian Geology* 12, pp. 298–311.
- 711 Balestro, G., Festa, A., Dilek, Y., 2019. Structural architecture of the Western Alpine Ophiolites, and
712 the Jurassic seafloor spreading tectonics of the Alpine Tethys. *Journal of the Geological*
713 *Society* 176, 913–930.
- 714 Baudelot, S., Bouillin, J.-P., Coiffait, P., 1981. Découverte d'Ordovicien inférieur daté par Acritarches
715 dans l'Ouest de la Petite Kabylie: conséquences structurales. *Comptes Rendus de*
716 *l'Académie des Sciences* 293, 611–614.
- 717 Ben Yaich, A., Hervouët, Y., Duée, G., El Hatimi, N., Aubouin, J., 1989. Age jurassique et
718 réinterprétation des roches basiques au Nord de Ouezzane (Rif externe, Maroc);
719 signification géodynamique. *Comptes rendus de l'Académie des sciences. Série 2*, 309,
720 1197–1202.
- 721 Benzaggagh, M., Mokhtari, A., Rossi, P., Michard, A., El Maz, A., Chalouan, A., Saddiqi, O., Rjimati, E.-
722 C., 2014. Oceanic units in the core of the External Rif (Morocco): Intramargin hiatus or
723 South-Tethyan remnants? *Journal of Geodynamics* 77, 4–21.
- 724 Bianco, C., Godard, G., Halton, A., Brogi, A., Liotta, D., Caggianelli, A., 2019. The lawsonite-
725 glaucophane blueschists of Elba Island (Italy). *Lithos* 348-349, 105–198.
- 726 Bodinier, J.-L., Morten, L., Puga, E., de Federico, A.D., 1987. Geochemistry of metabasites from the
727 Nevado-Filabride Complex, Betic Cordilleras, Spain: Relics of a dismembered ophiolitic
728 sequence. *Lithos* 20, 235–245.
- 729 Bortolotti, V., Principi, G., 2005. Tethyan ophiolites and Pangea break-up. *Island Arc* 14, 442–470.
- 730 Bouillin, J., 1974. Présence de flyschs Maurétaniens épimétamorphiques dans le massif du Moul Ed
731 Demamène (Constantinois, Algérie). *Comptes Rendus de l'Académie des Sciences, Série D*,
732 279, 1059–1062.

- 733 Bouillin, J.-P., 1979. La transversale de Collo et d'El Milia (Petite Kabylie): une région-clef pour
734 l'interprétation de la tectonique alpine de la chaîne littorale d'Algérie. Mémoires de la
735 Société Géologique de France (nouv. s.) 135, 83 p.
- 736 Bouillin, J.-P., 1986. Le "bassin maghrébin"; une ancienne limite entre l'Europe et l'Afrique à l'ouest
737 des Alpes. Bulletin de la Société géologique de France (8^{ème} s.) 2, 547–558.
- 738 Bouillin, J.-P., 1989. La suture alpine en Méditerranée occidentale. Remarques sur une synthèse et
739 rappel d'une autre conception. II: réponses. Bulletin de la Société géologique de France
740 (8^{ème} s.) 5, 859–567.
- 741 Bouillin, J.-P., 1992. La répartition des affleurements de la Dorsale kabyle: héritage d'une
742 segmentation mésozoïque de la marge nord-téthysienne? Comptes rendus de l'Académie
743 des sciences, Série 2, 315, 1127–1132.
- 744 Bouillin, J.-P., Hernandez, J., 1975. Découverte de roches volcaniques associées aux phtanites
745 cénomaniennes du flysch massylien dans le Nord du Constantinois (Algérie). Compte rendu
746 sommaire des séances de la Société géologique de France 17, 14–16.
- 747 Bouillin, J.-P., Kornprobst, J., 1974. Associations ultrabasiques de Petite Kabylie: péridotites de type
748 alpin et complexe stratifié; comparaison avec les zones internes bético-rifaines. Bulletin de
749 la Société géologique de France (7^{ème} s.) 16, 183–194.
- 750 Bouillin, J.-P., Durand-Delga, M., Gélard, J.P., Leikine, M., Raoult, J.F., Raymond, D., Tefiani, M., Vila,
751 J.M., 1970. Définition d'un flysch massylien et d'un flysch maurétanien au sein des flyschs
752 allochtones de l'Algérie. Comptes Rendus de l'Académie des Sciences 270, 2249–2252.
- 753 Bouillin, J.-P., Kornprobst, J., Raoult, J.F., 1977. Données préliminaires sur le complexe volcano-
754 sédimentaire de Rekkada Metletine (ex-*Texenna*), en Petite Kabylie (Algérie). Bulletin de la
755 Société géologique de France (7^{ème} s.) 19, 805–813.
- 756 Burkhard, D.J.M., 1993. Accessory chromium spinels: Their coexistence and alteration in
757 serpentinites. *Geochimica et Cosmochimica Acta* 57, 1297–1306.
- 758 Cabanis, B., Lecolle, M., 1989. Le diagramme La/10–Y/15–Nb/8: un outil pour la discrimination des
759 séries volcaniques et la mise en évidence des processus de mélange et/ou de
760 contamination crustale. Comptes Rendus de l'Académie des Sciences, Série 2, 309, 2023–
761 2029.
- 762 Chalot-Prat, F., 2005. An undeformed ophiolite in the Alps: Field and geochemical evidence for a link
763 between volcanism and shallow plate tectonic processes, in: Fouger, G.R., Natland, J.H.,

- 764 Presnall, D.C., Anderson, D.L. (Eds.), Plates, Plumes, and Paradigms. Geological Society of
765 America special Paper 388, pp. 751–780.
- 766 Chalouan, A., Michard, A., El Kadiri, Kh., Negro, F., Frizon de Lamotte, D., Soto, J.I., Saddiqi, O., 2008.
767 The Rif Belt, in: Michard, A., Saddiqi, O., Chalouan, A., Frizon de Lamotte, D. (Eds.),
768 Continental Evolution: The Geology of Morocco. Springer, Lecture Notes in Earth Sciences
769 116, pp. 203–302.
- 770 Coutelle, A., Gélard, J.P., 1968. Existence de radiolarites à la base du "flysch schisto-gréseux
771 tithonique-néocomien" du col de Chellata (Grande Kabylie). Compte rendu sommaire des
772 séances de la Société géologique de France 3, 79–81.
- 773 Coutelle, A., Delteil, J., 1989. La suture alpine en Méditerranée occidentale. Remarques sur une
774 synthèse et rappel d'une autre conception. Bulletin de la Société géologique de France (8^{ème}
775 s.) 5, 859–867.
- 776 Dilek, Y., Furnes, H., 2011. Ophiolite genesis and global tectonic fingerprinting of ancient oceanic
777 lithosphere. The Geological Society of America Bulletin 123, 387–411.
- 778 Dilek, Y., Furnes, H., 2014. Ophiolites and their origins. Elements 10, 93–100.
- 779 Djellit, H., 1987. Évolution tectono-métamorphique du socle kabyle et polarité de mise en place des
780 nappes de flysch en Petite Kabylie occidentale (Algérie). Thèse de Doctorat, Université Paris
781 XI, 206 p.
- 782 Durand-Delga, M., 1948. Sur la structure du Moul ed Demamène (Chaîne Numidique, Algérie).
783 Comptes Rendus de l'Académie des Sciences 226, 1826–1827.
- 784 Durand-Delga, M., 1949. Notes sur la structure géologique des environs de Texenna (Petite Kabylie,
785 Algérie). Bulletin de la Société géologique de France (5^{ème} s.) 19, 271–278.
- 786 Durand-Delga, M., 1950. Le chevauchement bordier de Petite Kabylie (Algérie). Comptes Rendus de
787 l'Académie des Sciences 231, 1522–1524.
- 788 Durand-Delga, M., 1954. Les rapports des venues doléritiques du Moul ed Demamène (chaîne
789 Numidique, Algérie) et de la mise en place de la nappe bordière de Petite Kabylie. XIX^{ème}
790 Congrès géologique international, Alger, 1952, fasc. XVII, C. R. section XV, 37–41.
- 791 Durand-Delga, M., 1955. Etude géologique de l'Ouest de la chaîne numidique. Bulletin du Service de
792 la Carte Géologique de l'Algérie, 2^{ème} série, Stratigraphie-Descriptions régionales 24, 533 p.

- 793 Durand-Delga, M., 1971. Les unités à Mésozoïque métamorphique d'El Milia à Texenna (Algérie) et
794 leur cadre structural. Bulletin de la Société géologique de France (7^{ème} s.) 13, 328–337.
- 795 Durand-Delga, M., Fontobé, J.M., 1980. Le cadre structural de la Méditerranée occidentale. 26^{ème}
796 Congrès Géol. Int. Paris, Coll. 5. Mémoires du BRGM 115, 67–87.
- 797 Durand-Delga, M., Rossi, P., Olivier, P., Puglisi, D., 2000. Situation structurale et nature ophiolitique
798 de roches basiques jurassiques associées aux flyschs maghrébins du Rif (Maroc) et de Sicile
799 (Italie). Comptes Rendus de l'Académie des Sciences, Série IIA 331, 29–38.
- 800 Ehrmann, E.F., 1946. Feuille au 50000^e « Tamesguida », avec notice explicative. Service de la Carte
801 géologique de l'Algérie, Alger.
- 802 Fernandez, O., 2019. The Jurassic evolution of the Africa-Iberia conjugate margin and its implications
803 on the evolution of the Atlantic-Tethys triple junction. Tectonophysics 750, 379–393.
- 804 Floyd, P.A., Winchester, J.A., 1975. Magma type and tectonic setting discrimination using immobile
805 elements. Earth and Planetary Science Letters 27, 211–218
- 806 Frizon de Lamotte, D., Raulin, C., Mouchot, N., Wrobel-Daveau, J.-C., Blanpied, C., Ringenbach, J.-C.,
807 2011. The southernmost margin of the Tethys realm during the Mesozoic and Cenozoic:
808 Initial geometry and timing of the inversion processes. Tectonics 30, TC3002, doi:
809 10.1029/2010TC002691.
- 810 Furnes, H., Safonova, I., 2019. Ophiolites of the Central Asian Orogenic Belt: Geochemical and
811 petrological characterization and tectonic settings. Geoscience Frontiers 10, 1255–1284.
- 812 Gale, A., Dalton, C.A., Langmuir, C.H., Su, Y., Schilling, J.G., 2013. The mean composition of ocean
813 ridge basalts. Geochemistry, Geophysics, Geosystems 14, 489–518.
- 814 Gélard, J.P., 1969. Le flysch à base schisto-gréseuse de la bordure méridionale et orientale du massif
815 de Chellata; le flysch maurétanien (Grande-Kabylie, Algérie). Bulletin de la Société
816 Géologique de France (7^{ème} s.) 11, 676–686.
- 817 Gomez-Pugnaire, M.T., Sanchez-Vizcaino, V.L., Fernandez-Soler, J.M., Acosta-Vigil, A., 2019. Mesozoic
818 and Cenozoic magmatism in the Betics, in: Quesada, C., Oliveira, J.T. (Eds.), The Geology of
819 Iberia: a geodynamic approach, Vol. 3: The alpine cycle. Springer, Regional Geology
820 Reviews, pp. 545–566.

- 821 Guerrera, F., Martín-Martín, M., Perrone, V., Tramontana, M., 2005. Tectono-sedimentary evolution
822 of the southern branch of the Western Tethys (Maghrebian Flysch Basin and Lucanian
823 Ocean): consequences for Western Mediterranean geodynamics. *Terra Nova* 17, 358–367.
- 824 Hawthorne, F.C., Oberti, R., Harlow, G.E., Maresch, W.V., Martin, R.F., Schumacher, J.C., Welch, M.D.,
825 2012. Nomenclature of the amphibole supergroup. *American Mineralogist* 97, 2031–2048.
- 826 Hekinian, R., 1982. *Petrology of the Ocean Floor*. Elsevier Oceanography Series 33, Elsevier, Academic
827 Press, 393 p.
- 828 Hey, M.H., 1954. A new review of the chlorites. *Mineralogical Magazine and Journal of the*
829 *Mineralogical Society* 30, 277–292.
- 830 Holland, T.J.B., Powell, R., 2011. An improved and extended internally consistent thermodynamic
831 dataset for phases of petrological interest, involving a new equation of state for solids.
832 *Journal of Metamorphic Geology* 29, 333–383.
- 833 Kapsiotis, A., Tsikouras, B., Grammatikopoulos, T., Karipi, S., Hatzipanagiotou, H., 2007. On the
834 metamorphic modification of Cr-spinel compositions from the ultrabasic rocks of the Pindos
835 ophiolite complex (NW Greece). *Bulletin of the Geological Society of Greece* 40, 781–793.
- 836 Laouar, R., Satouh, A., Salmi-Laouar, S., Abdallah, N., Cottin, J.-Y., Bruguier, O., Bosch, D., Ouabadi, A.,
837 Boyce, A.J., Fallick, A.E., 2017. Petrological, geochemical and isotopic characteristics of the
838 Collo ultramafic rocks (NE Algeria). *Journal of African Earth Sciences* 125, 59–72.
- 839 Le Bas, M.J., Streckeisen, A.L., 1991. The IUGS systematics of igneous rocks. *Journal of the Geological*
840 *Society* 148, 825–833.
- 841 Leprêtre, R., Frizon de Lamotte, D., Combier, V., Gimeno-Vives, O., Mohn, G., Eschard, R., 2018. The
842 Tell-Rif orogenic system (Morocco, Algeria, Tunisia) and the structural heritage of the
843 southern Tethys margin. *BSGF-Earth Sciences Bulletin* 189, 10,
844 doi.org/10.1051/bsgf/2018009.
- 845 Liberi, F., Morten, L., Piluso, E., 2006. Geodynamic significance of ophiolites within the Calabrian Arc.
846 *Island Arc* 15, 26–43.
- 847 Mellini, M., Rumori, C., Viti, C., 2005a. Hydrothermally reset magmatic spinels in retrograde
848 serpentinites: formation of “ferritchromit” rims and chlorite aureoles. *Contributions to*
849 *Mineralogy and Petrology* 149, 266–275.

850 Meschede, M., 1986. A method of discriminating between different types of mid-ocean ridge basalts
851 and continental tholeiites with the Nb–Zr–Y diagram. *Chemical Geology* 56, 207–218.

852 Mével, C., 2003. Serpentinization of abyssal peridotites at mid-ocean ridges. *Comptes rendus*
853 *Geoscience* 335, 825–852.

854 Michard, A., Mokhtari, A., Chalouan, A., Saddiqi, O., Rossi, P., Rjimati, E.-C., 2014. New ophiolite
855 slivers in the External Rif belt, and tentative restoration of a dual Tethyan suture in the
856 western Maghrebides. *Bulletin de la Société géologique de France* 185, 313–328.

857 Michard, A., Mokhtari, A., Lach, P., Rossi, P., Chalouan, A., Saddiqi, O., Rjimati, E.-C., 2018. Liassic age
858 of an oceanic gabbro of the External Rif (Morocco): Implications for the Jurassic continent–
859 ocean boundary of Northwest Africa. *Comptes Rendus Geoscience* 350, 299–309.

860 Michard, A., Saddiqi, O., Chalouan, A., Chabou, M.C., Lach, P., Rossi, P., Bertrand, H., Youbi, N., 2020.
861 Comment on “The Mesozoic Margin of the Maghrebian Tethys in the Rif Belt (Morocco):
862 Evidence for Polyphase Rifting and Related Magmatic Activity” by Gimeno-Vives et al.
863 *Tectonics* 39, e2019TC006004.

864 Nimis, P., Tesalina, S.G., Omenetto, P., Tartarotti, P., Lerouge, C., 2004. Phyllosilicate minerals in the
865 hydrothermal mafic–ultramafic-hosted massive-sulfide deposit of Ivanovka (southern
866 Urals): comparison with modern ocean seafloor analogues. *Contributions to Mineralogy*
867 *and Petrology* 147, 363–383.

868 Pearce, J.A., 1996. A user’s guide to basalt discrimination diagrams. Geological Association of Canada
869 Special Publication 12, 79–113.

870 Pearce, J.A., 2008. Geochemical fingerprinting of oceanic basalts with applications to ophiolite
871 classification and the search for Archean oceanic crust. *Lithos* 100, 14–48.

872 Pearce, J.A., 2014. Immobile element fingerprinting of ophiolites. *Elements* 10, 101–108.

873 Principi, G., Bortolotti, V., Chiari, M., Cortesogno, L., Gaggero, L., Marcucci, M., Sacconi, E., Treves, B.,
874 2004. The pre-orogenic volcano-sedimentary covers of the western Tethys oceanic basin: a
875 review. *Ofioliti* 29, 177–211.

876 Puga, E., Fanning, M., De Federico, A.D., Nieto, J.M., Beccaluva, L., Bianchini, G., Díaz Puga, M.A.,
877 2011. Petrology, geochemistry and U–Pb geochronology of the Betic Ophiolites: Inferences
878 for Pangaea break-up and birth of the westernmost Tethys Ocean. *Lithos* 124, 255–272.

- 879 Puga, E., De Federico, A.D., Fanning, M., Nieto, J.M., Martínez-Conde, J.A.R., Díaz Puga, M.A., Lozano,
880 J.A., Bianchini, G., Natali, C., Beccaluva, L., 2017. The Betic Ophiolites and the Mesozoic
881 Evolution of the Western Tethys. *Geosciences* 7, 31, doi: 10.3390/geosciences7020031.
- 882 Raoult, J.-F., 1974. Géologie du centre de la chaîne numidique: Nord du Constantinois, Algérie.
883 Mémoires de la Société géologique de France (nouv. s.) 121, 1-162 + dépliants.
- 884 Reynard, B., 2013. Serpentine in active subduction zones. *Lithos* 178, 171–185.
- 885 Roman'ko, E.F., Roman'ko, A.E., Meskhi, A.M., 1998. The geology and geochemistry of Meso-
886 Cenozoic magmatic formations in Northeastern Algeria. *Doklady Earth Sciences* 362, 925–
887 927.
- 888 Ross, P.-S., Bédard, J.H., 2009. Magmatic affinity of modern and ancient subalkaline volcanic rocks
889 determined from trace-element discriminant diagrams. *Canadian Journal of Earth Sciences*
890 46, 823–839.
- 891 Saccani, E., Principi, G., Garfagnoli, F., Menna, F., 2008. Corsica ophiolites: geochemistry and
892 petrogenesis of basaltic and metabasaltic rocks. *Ofioliti* 33, 187–207.
- 893 Sansone, M.T.C., Rizzo, G., Mongelli, G., 2011. Petrochemical characterization of mafic rocks from the
894 Ligurian ophiolites, southern Apennines. *International Geology Review* 53, 130–156.
- 895 Schettino, A., Turco, E., 2011. Tectonic history of the western Tethys since the Late Triassic. *The*
896 *Geological Society of America Bulletin* 123, 89–105.
- 897 Shervais, J.W., 1982. Ti-V plots and the petrogenesis of modern and ophiolitic lavas. *Earth and*
898 *Planetary Science Letters* 59, 101–118.
- 899 Standish, J.J., Dick, H.J., Michael, P.J., Melson, W.G., O'Hearn, T., 2008. MORB generation beneath
900 the ultraslow spreading Southwest Indian Ridge (9–25°E): Major element chemistry and the
901 importance of process versus source. *Geochemistry Geophysics Geosystems* 9, Q05004,
902 doi.org/10.1029/2008GC001959.
- 903 Sun, S.S., McDonough, W.F., 1989. Chemical and isotopic systematics of oceanic basalts: Implications
904 for mantle composition and processes. *Geological Society of London, Special Publication* 42,
905 313–345.
- 906 Thiéblemont, D., Duron, J., Plunder, A., 2019. Atlas pétrographique obstiné de l'ophiolite du
907 Chenaillet. *Géologie de la France* 2019 (4), 1–78; [http://geolfrance.brgm.fr/mise-en-ligne-](http://geolfrance.brgm.fr/mise-en-ligne-dun-atlas-petrographique-roches-massif-chenaillet-brianconnais-hautes-alpes)
908 [dun-atlas-petrographique-roches-massif-chenaillet-brianconnais-hautes-alpes.](http://geolfrance.brgm.fr/mise-en-ligne-dun-atlas-petrographique-roches-massif-chenaillet-brianconnais-hautes-alpes)

909 Thompson, G., 1991. Metamorphic and hydrothermal processes: basalt-seawater interactions, in:
910 Floyd, P.A. (Ed.), *Oceanic basalts*. Springer, New York, pp. 148–173.

911 Wildi, W., 1983. La chaîne tello-rifaine (Algérie, Maroc, Tunisie): structure, stratigraphie et évolution
912 du Trias au Miocène. *Revue de Géographie physique et de Géologie dynamique* 24, 201–
913 297.

914 Wood, D.A., 1980. The application of a Th Hf Ta diagram to problems of tectonomagmatic
915 classification and to establishing the nature of crustal contamination of basaltic lavas of the
916 British Tertiary Volcanic Province. *Earth and Planetary Science Letters* 50, 11–30.

917 **Figure Captions**

918 **Fig. 1.** Betic-Maghrebian Chain (modified from Chalouan et al., 2008) with the location of mafic rocks
919 with ophiolite affinities associated to the flysch zone.

920 **Fig. 2.** Geological overview of the Texenna region (Lesser Kabylia). **(a)** Geological map compiled from
921 Durand-Delga (1955), Andrieux and Djellit (1989) and original geological survey; S_0 : stratification in
922 flysch and sediments; S_1 : foliation or schistosity in metamorphic rocks. **(b)** Panoramic view from
923 $36^{\circ}39'33''\text{N}-5^{\circ}46'25''\text{E}$, before the filling of the Tamellout dam, showing the contacts between the
924 Kabyle basement (KB), the ophiolite complex (Sendouah-Tabellout Unit: STU) and the flysch domain
925 (FD).

926 **Fig. 3.** Outcrop photographs of the Texenna ophiolite complex. **(a)** Metadolerite vein crosscutting the
927 serpentinite. **(b)** Chrysotile veinlets in serpentinite. **(c)** Metadolerite dyke inside calcschists. **(d)**
928 Metagabbroic rock mainly composed of chlorite (Chl), epidote (Ep) and actinolite (Act), intersected
929 by whitish veins of calcite (Cal) and quartz (Qz). **(e)** Pillow lava showing a massive nucleus of
930 porphyritic metabasalt, surrounded by greenish, aphyritic and schistose rims. **(f)** Alternating layers of
931 radiolarite and metabasalt on top of the pillow lavas. **(g)** Pillows with deformed ellipsoidal shapes. **(h)**
932 Pillow lava with external chlorite-rich and variolitic zones. Outcrops are at Djebel Es-Sendouah (a-b),
933 Tabellout (c-d) and Oued Djendjene gorge, near the Tabellout water dam (e-h).

934 **Fig. 4.** Petrology and microstructures of serpentinite. **(a)** Serpentine (Srp) mesh texture, bastite and
935 altered chromite (Chr). **(b)** Detail of the mesh microstructure showing cells derived from olivine
936 grains, with an edge rich in tiny crystals of magnetite (Mag) enveloping concentric layers of
937 serpentine. **(c)** Detail of the bastite microstructure showing tremolite (Tr) grooves in serpentine
938 (dashed arrow), interpreted as former Cpx lamellae exsolved in the Opx protocystal. **(d)** Altered
939 chromite grains surrounded by a chlorite (Chl) corona, and intersected by late veinlets of tremolite

940 and/or chlorite. BSE images (a-c) and (d) composite EDS elemental map of Fe (red), Cr (green) and Al
941 (blue).

942 **Fig. 5.** Mineral compositions in serpentinites and metabasites. **(a-c)** Calcic clin amphiboles in the
943 metabasites, following the nomenclature of Hawthorne et al. (2012). **(d)** Feldspars in the
944 metabasites. The inset rectangle shows the composition profile A–B across a magmatic plagioclase in
945 metadolerite Md4; the red line indicates the position of the Carlsbad twin plane on this profile,
946 whose position is shown in Figure 7d-e; chemical analyses are given in Table C of SM (2: rim, core,
947 rim). **(e)** Composition of Cr-spinels and their pseudomorphs, in serpentinites and metabasalts. **(f)**
948 Compositions of chlorites from serpentinites and metabasites in the classification diagram of Hey
949 (1954). “Sea floor hydrothermally altered rocks” from Nimis et al. (2004).

950 **Fig. 6.** Ol–Cpx–Opx normative molar composition of the studied serpentinites, following the
951 nomenclature of Le Bas and Streckeisen (1991).

952 **Fig. 7.** Petrology and microstructures of the metabasites. **(a-c)** Greenschist-facies metabasite, with
953 albite (Ab), epidote (Ep), actinolite (Act), hornblende (Hbl), chlorite (Chl) and titanite (Tnt), showing
954 the preserved doleritic texture of the gabbroic protolith crosscut by albite and epidote veins. **(d-e)**
955 Metadolerite, showing the still-preserved magmatic laths of plagioclase in a matrix of actinolite (Act)
956 and pargasite (Prg); A–B: composition profile across plagioclase (see Fig. 5d). **(f-k)** Microstructures of
957 the pillow lavas (see text for the meaning of the Z1–Z4 and 1–5 labels; to be compared with Fig. B in
958 SM); (f) pillow zoning (Z1, chlorite-rich outer zone after glassy crust; Z2 and Z3, variolitic zones, rich in
959 former devitrification spherulites; Z4, internal massive zone); (g) preserved magmatic flow in the
960 outer zone; (h) epidote pseudomorph after plagioclase phenocrysts; (j) epidote+albite association
961 after devitrification spherulites, around former plagioclase microliths; (k) microgeode after gas
962 bubble, filled with epidote, clinozoisite and chlorite. **(l)** Deformed metabasite, with early transposed
963 veins of calcite (Cal) and truncated epidote (Ep+Ab), and late veins of K-feldspar (Kfs). SEM images (a,
964 c, d, g-j, l) and composite BSE and EDS elemental maps of Al, Ca, Fe, and Ti (b, e-f, k).

965 **Fig. 8.** Geochemical discrimination diagrams of the metabasites. Red squares: pillow lava samples
966 (Pil12-13-14-15); green diamonds: metadolerite samples (Md4-5-6); blue circles: metagabbro
967 samples (Mb9-10-11); ppm means $\mu\text{g/g}$. **(a)** Zr/Ti vs. Nb/Y diagram (Pearce, 1996). **(b)** P_2O_5 vs. Zr
968 diagram (Floyd and Winchester, 1975). **(c)** Zr vs. Y; **(d)** La vs. Yb and **(e)** Th vs. Yb diagrams (Ross and
969 Bédard, 2009).

970 **Fig. 9.** REE and trace-element patterns for the metabasites. **(a, c, e)** Chondrite-normalized REE
971 patterns and **(b, d, f)** N-MORB-normalized incompatible-element patterns. (a, b) Pillow lavas; (c, d)
972 metadolerites; (e, f) metagabbros. The normalizing values for chondrites and N-MORB are from Sun

973 and McDonough (1989) and Gale et al. (2013), respectively. N-MORB and E-MORB compositions are
974 from Gale et al. (2013).

975 **Fig. 10.** Tectonic setting discrimination diagrams. Red squares: pillow lava samples (Pil12-13-14-15);
976 green diamonds: metadolerite samples (Md4-5-6); blue circles: metagabbro samples (Mb9-10-11).
977 **(a)** Ti vs. V diagram of Shervais (1982). **(b)** $Zr/4-2Nb-Y$ discrimination diagram of Meschede (1986);
978 WPA: within-plate alkali basalts; WPT: within-plate tholeiites; VAB: volcanic-arc basalts. **(c)** $Y/15-$
979 $La/10-Nb/8$ discrimination diagram of Cabanis and Lecolle (1989). **(d)** $Th-Hf/3-Ta$ discrimination
980 diagram (Wood, 1980). **(e)** Th/Yb vs. Nb/Yb and **(f)** TiO_2/Yb vs. Nb/Yb diagrams (Pearce, 2008).

981 **Fig. 11.** $P-T$ pseudosection modelled for the composition of the Pil14 metabasite. The minerals are
982 listed in decreasing order of abundance, those in brackets being in negligible quantity (< 2 vol.%).
983 Since the sample was subjected to deformation and metamorphism during the emplacement of the
984 ophiolites in the Maghrebide belt, the model applies to this episode.

985 **Fig. 12.** Th/Yb vs. Nb/Yb diagram showing the compositional fields of some ophiolites of the Western
986 Mediterranean region. (1) Kimmeridgian–Berriasian basalt flows of the Mesorif Basalts–Breccias
987 lineament (MBB) and (2) Bou Adel gabbros of the Mesorif suture zone (MSZ) (Morocco) (Benzaggagh
988 et al., 2014); (3) Metabasites of the Betic ophiolites (Puga et al., 2011); (4) Metabasites of Eastern
989 Elba Island (Northern Apennines, Italy) (Bianco et al., 2019); (5) Metabasic rocks of the Chenaillet
990 ophiolitic complex (Western Alps) (Chalot-Prat, 2005); (6) Metabasic rocks of Alpine Corsica
991 ophiolites (Sacconi et al., 2008).

992 **Fig. 13.** The western Tethys at the Upper Jurassic. **(a)** Palaeogeographic reconstruction after
993 Schettino and Turco (2011). NPF: North Pyrenean Fault; GiF: Gibraltar Fault; red lines: middle-ocean
994 ridges; orange lines: rifts. **(b)** Schematic reconstruction of the north-tethysian margin of the Kabylia
995 after Bouillin (1992).

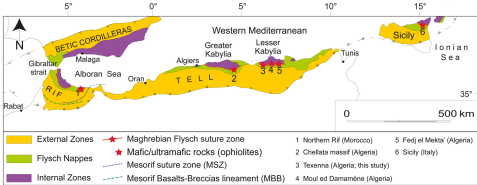


Figure 1

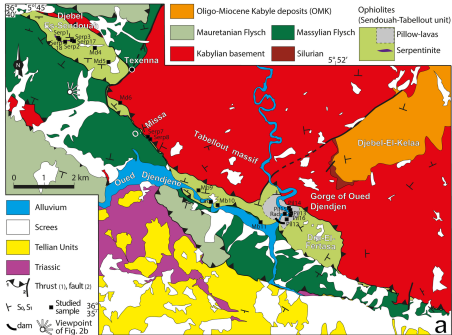


Figure 2

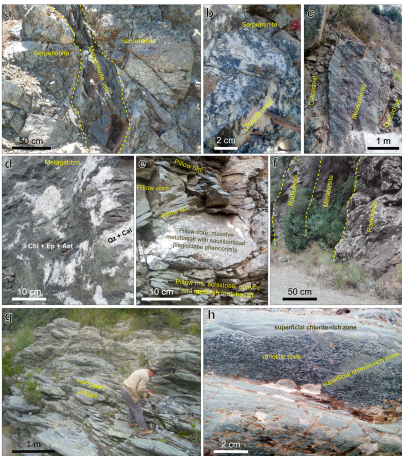


Figure 3

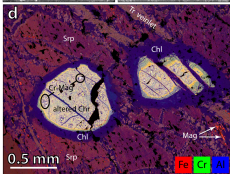
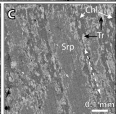
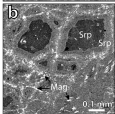
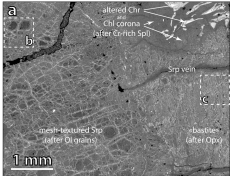


Figure 4

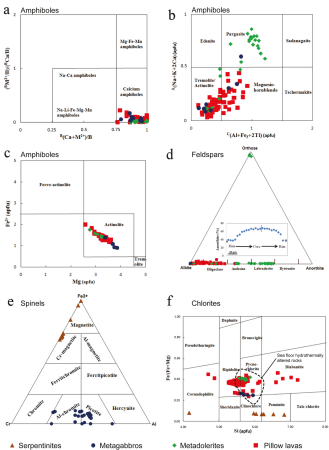


Figure 5

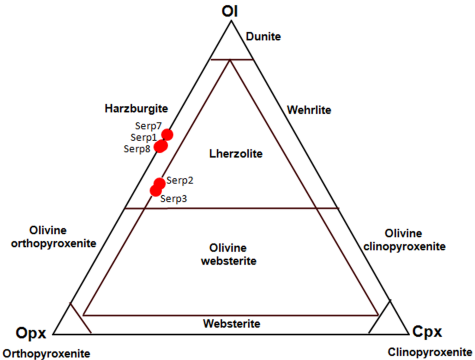


Figure 6

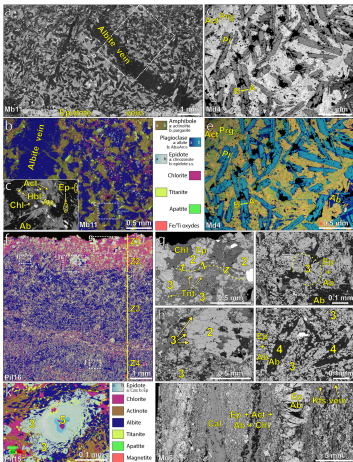


Figure 7

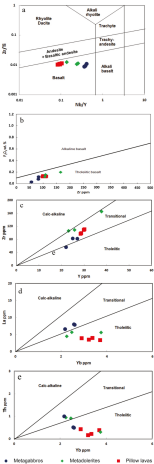


Figure 8

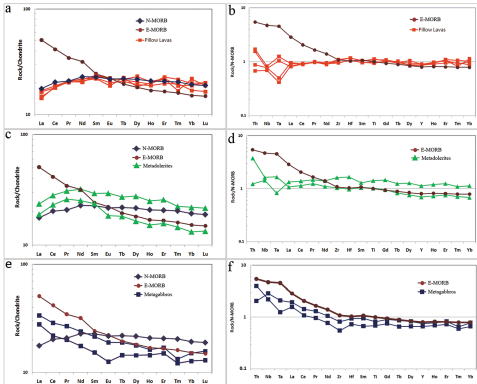


Figure 9

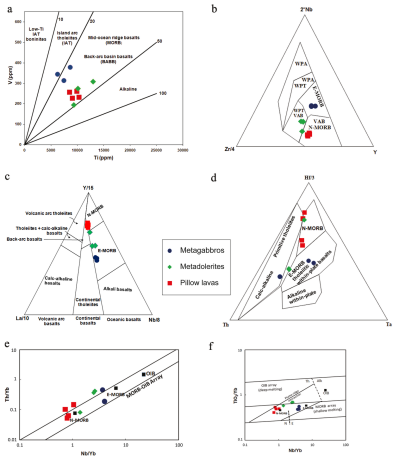


Figure 10

SiO₂ Al₂O₃ CaO MgO FeO K₂O Na₂O TiO₂ O +H₂O
 57.38 10.30 8.24 6.52 8.11 0.60 5.66 1.41 1.78 mol%

Ab: low albite; Act: actinolite; Chl: chlorite; Ep: epidote; Gln: glaucophane; Hem: hematite;
 Hbl: hornblende; Kfs: K-feldspar; Ms: muscovite; Ph: phengite; Pl: plagioclase; Tnt: titanite

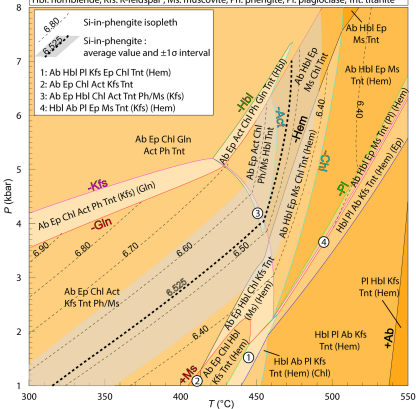
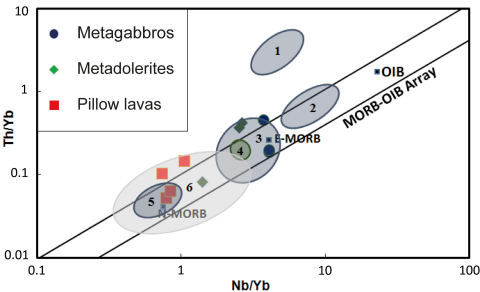


Figure 11



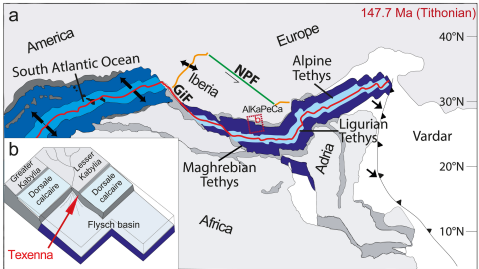


Figure 13