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1 Petrological and geochemical constraints on the origin of apatite  
2 ores from Mesozoic alkaline intrusive complexes, Central High-  
3 Atlas, Morocco

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18 **Abstract**

19 Five main apatite deposits from the Moroccan Central High-Atlas —including  
20 Anemzi, Tirrhist-Inouzane, Tassent, Ait-Daoud-Toumliline and Tasraft— have been  
21 investigated through integrated field, petrographic, mineralogical, and geochemical  
22 studies. The apatite ore occurs as veins, few mm to several cm thick, spatially  
23 associated with syenite–quartz monzonite (dominant) to monzodiorite–gabbrodiorite  
24 bodies emplaced during the Atlas Jurassic-Cretaceous alkaline magmatic event. Gem-  
25 quality apatite crystals, up to 15 cm in size, are associated with albite, clinopyroxene  
26 (hedenbergite–augite–diopside), amphibole (hornblende–edenite–actinolite), K-feldspar,  
27 quartz, magnetite, titanite, epidote, prehnite, and calcite. Two types of apatite have  
28 been distinguished according to their halogen contents: F-rich and Cl-rich apatites. Both  
29 types are enriched in REE (up to 2 wt.%  $\Sigma$ REE), and have major and trace element  
30 contents consistent with a magmatic origin, corroborated by the chemical similarity with  
31 accessory apatite in the host alkaline intrusions, as well as with other reference igneous  
32 apatites. However, hydrothermal fluids percolating through country sedimentary rocks  
33 were also involved, as indicated by the striking Na enrichment observed in the Cl-rich  
34 apatites, likely inherited from Triassic evaporites, showing close spatial relationships  
35 with the alkaline intrusions. These fluids would also account for pervasive Na-  
36 metasomatism of alkaline magmatic rocks, as observed in the wall rocks of the apatite  
37 veins.

38 **Keywords:** Moroccan Central High-Atlas, alkaline magmatic event, vein-type apatite  
39 ores, F-rich and Cl-rich apatites, magmatic origin/hydrothermal fluids.

## 41 1. Introduction

42 Apatite is the most ubiquitous accessory phosphorus(P)-bearing mineral found in  
43 sedimentary, igneous and metamorphic rocks. It is also sometimes a major rock-forming  
44 mineral in some peculiar rock types: e.g. sedimentary phosphorites and some igneous  
45 rocks of alkaline, carbonatitic and anorthositic affinities (e.g., [McConnell 1973](#); [Glenn et al., 1994](#);  
46 [Dymek and Owens 2001](#); [Martin and Rakovan, 2013](#); [Duchesne and Liégeois, 2015](#);  
47 [Pufahl and Groa et al., 2017](#)). Apatite is the only major potential resource of P  
48 used in the manufacturing of e.g., nitrogen-phosphorus-potassium (NPK) fertilizers,  
49 animal feed supplements, ceramics, pharmaceuticals, textiles and batteries. The  
50 sedimentary phosphorites make the largest contribution to the world production of  
51 phosphate, whereas igneous apatite rocks account for only ~ 10 % (e.g., [Jasinski, 2016](#);  
52 [Pufahl and Groa et al., 2017](#)). The igneous phosphate ores are nevertheless  
53 economically valuable since they provide high quality phosphate concentrates with low  
54 contents of unwanted contaminants such as Cd, As, Pb, Si and Al, and they are rich in  
55 Rare Earth Elements (REE) (e.g., [Webster and Piccoli, 2015](#); [Chakhmouradian et al 2017](#)).  
56 In terms of world resources, Morocco is the largest holder and producer of  
57 sedimentary phosphates, with reserves estimated to represent about ~ 76 % of world  
58 reserves (e.g., [Jasinski, 2016, 2019](#); [Pufahl and Groa et al., 2017](#); [www.ocpgroup.ma](#)).  
59 However, Morocco also bears igneous phosphate ores, especially in the High-Atlas  
60 mountain range. Over the last two decades, gem-quality apatite crystals well-  
61 appreciated by mineral collectors and known as "Moroccan High-Atlas Imilchil-Anemzi  
62 apatite" were made available on the international mineral market (e.g.,  
63 <https://www.minfind.com/mineral-569173.html>; [3](https://www.mchminerals.com/fine-</a></p></div><div data-bbox=)

64 [mineral-specimens/apatite-morocco](#); [Rakovan, 2015](#)). Despite their economic and  
65 scientific interests, only one recent study ([Dumańska-Słowik et al., 2018](#)) was devoted  
66 to the Anemzi apatite, chiefly dealing with organic inclusions, and a detailed  
67 characterization of the apatite ores remains to be done. This paper provides the first  
68 description of the main apatite deposits in the High-Atlas region. We carried out an  
69 exhaustive petrographic, mineralogical, and geochemical study (major, volatile, and  
70 trace elements) of the apatites and associated minerals, and of the host lithologies.  
71 These data are used to identify suitable sources for the phosphate mineralization.

## 72 **2. Geological setting**

73 The High-Atlas (up to 4000 m elevation) constitutes a typical intra-continental  
74 Alpine orogen, forming a thrust chain resulting from the Africa–Eurasia convergence. Its  
75 growth started during the early Mesozoic with the reactivation of Pangea paleo-faults  
76 which delimited subsidence sedimentary basins, leading to the Atlantic Ocean opening  
77 at the NW Africa margin (e.g., [Stampfli and Borel, 2002](#); [Piqué et al., 2002](#); [Teixell et al.,](#)  
78 [2003](#); [Laville et al., 2004](#); [Frizon de Lamotte et al., 2008, 2009](#); [Domènech et al., 2015](#)).  
79 This chain includes several segments, such as the Western, Central and Eastern High-  
80 Atlas (Fig. [1a](#)). In contrast with the two other segments, the Central High-Atlas  
81 corresponds to the deepest part of the orogen (Fig. [1b](#); e.g., [Teixell et al., 2003](#); [Frizon](#)  
82 [de Lamotte et al., 2008, 2009](#); [Gouiza et al., 2010](#); [Michard et al., 2011](#)). Several works  
83 have described the lithostratigraphical sequence of this belt (Fig. [1b–c](#); [Charrière et al.,](#)  
84 [2005, 2009](#); [Haddoumi et al., 2002, 2010](#); [Ettaki et al., 2007](#); [Frizon de Lamotte et al.,](#)  
85 [2008, 2009](#); [Cavin et al., 2010](#); [Saura et al., 2014](#); [Charrière and Haddoumi, 2016,](#)  
86 [2017](#); [Cavallina et al., 2018](#)). The first Mesozoic sediments are Triassic red argillites and

87 evaporites (mostly halite and gypsum) forming more than 1000 m-thick series which  
88 deposited within subsident basins and which cover unconformably a leveled  
89 Precambrian-Paleozoic basement (e.g., [Du Dresnay, 1987](#); [Piqué et al., 2002](#); [Eddif et](#)  
90 [al., 2007](#); [Ouabid et al., 2017, 2020](#); [Boukerrou et al., 2018](#); [Karaoui et al., 2021](#)). These  
91 deposits are generally interbedded with tholeiitic lava flows linking to the Central Atlantic  
92 Magmatic Province (e.g., [Marzoli et al., 2019](#) and references therein). Subsequently, a  
93 regional marine transgression led to the deposit of a series of Early Jurassic carbonates  
94 (mostly limestones and marls). During the Middle Jurassic and Lower Cretaceous,  
95 conformable red beds of continental sandstone settled, indicating the filling of basins  
96 and a marine regression. Overall, these post-rift basins were controlled by sinistral  
97 strike-slip faults, and separated by Triassic diapiric anticlinal ridges with dominant NE  
98 trending (Fig. 1b–c; e.g., [Laville and Pique, 1992](#); [Saura et al., 2014](#)). In the Middle  
99 Cretaceous, a new marine carbonated platform sets up (e.g., [Froitzheim et al. 1988](#);  
100 [Cavallina et al., 2018](#) and references therein), and at the end of the Cretaceous, the  
101 sediments consist of terrigenous red beds, expressing the early manifestation of the  
102 Alpine orogeny (e.g., [Laville et al. 1977](#); [Froitzheim et al. 1988](#); [Ettachfini and Andreu,](#)  
103 [2004](#); [Ettachfini et al., 2005](#); [Michard et al., 2011](#); [Cavallina et al., 2018](#)). The Central  
104 High-Atlas Mesozoic ridges are often unconformably overlain by Paleocene non-marine  
105 red bed formation which draws the synclinal geometries attributed to Late  
106 Cretaceous–Cenozoic shortening and halokinesis (Fig. 1c; e.g., [Charriere et al., 2009](#);  
107 [Michard et al., 2011](#)). The famous Cenozoic alkaline-carbonatite magmatic episode (~  
108 45–35 Ma) that occurred at NW Central High-Atlas Tamazight area (Fig. 1b) coincides

109 with the African–European collisional event (e.g., [Bouabdli et al., 1988](#); [Bernard-Griffiths](#)  
110 [et al., 1991](#); [Bouabdellah et al., 2010](#)).

111 In addition, the core of the Central High-Atlas anticlinal ridges is formed by an  
112 important Jurassic–Cretaceous alkaline to transitional magmatism between ~ 165 Ma  
113 and ~ 125 Ma (Fig. [1b–c](#); e.g., [Hailwood and Mitchell 1971](#); [Westphal et al., 1979](#);  
114 [Beraâouz and Bonin, 1993](#); [Rahimi et al. 1997](#); [Armando, 1999](#); [Haddoumi et al., 2010](#);  
115 [Michard et al., 2011, 2013](#); [Bensalah et al., 2013](#)). This magmatism occurs as  
116 numerous dykes, sills and laccoliths, either intrusive along faults and narrow anticlinal  
117 ridges or intercalated in the thick Mesozoic sedimentary series (Fig. [1b–c](#); e.g.,  
118 [Armando, 1999](#); [Teixell et al., 2003](#)). Basaltic lava flows within wide synclines are also  
119 reported (e.g., [Guezal et al., 2011](#); [Bensalah et al., 2013](#); [Michard et al., 2013](#)). The  
120 magma emplacement was mainly controlled by faults coeval with halokinetic tectonic  
121 and moderate syn-schistous deformation, might link to the widespread exhumation  
122 resulting from the orogen during the Upper Jurassic–Early Cretaceous (e.g., [Laville and](#)  
123 [Pique, 1992](#); [Frizon de Lamotte et al., 2008, 2009](#); [Michard et al., 2011](#); [Saura et al.,](#)  
124 [2014](#); [Torres-Lopez et al., 2016](#); [Essaifi and Zayane, 2018](#)). Metamorphic aureole  
125 around the intrusions is almost absent at the contact with the country sedimentary  
126 rocks, which are mostly faulted (e.g., Fig. [1b–c](#)). The tectonic boundaries are mostly  
127 vertical/sub-vertical faults marked by brecciated formation composed of fragments of  
128 igneous intrusive and carbonate rocks. However, some rare normal contacts between  
129 the intrusions and the sedimentary sequence are observed with chilled marginal  
130 magmatic rocks with carbonates recrystallization that occurred at about 300–500 °C  
131 ([Lhachmi, 1992](#), [Armando, 1999](#); [Lhachmi et al., 2001](#); [Essaifi and Zayane, 2018](#)). The

132 intrusive complexes include almost all terms of alkaline magma differentiation series  
133 (e.g., [Armando, 1999](#); [Lhachmi et al., 2001](#); [Zayane et al., 2002](#); [Essaifi and Zayane,](#)  
134 [2018](#)). The intrusions are made up to 60–70% of troctolites, gabbros, and  
135 monzogabbros, while the remaining volume is represented by differentiated rocks  
136 (mostly monzonites and syenites). Depside the fragmentation of the intrusive bodies  
137 into several blocks which hampers to infer more information about the primary  
138 relationships, the differentiated unit seems to be intrusive into the mafic unit (e.g.,  
139 [Armando, 1999](#); [Lhachmi et al., 2001](#); [Essaifi and Zayane, 2018](#)). The evolution of the  
140 magma sequence from mafic to felsic members would be mainly linked to fractional  
141 crystallization, whereas crustal contamination during the magma ascent cannot be  
142 excluded (e.g., [Armando, 1999](#); [Lhachmi et al., 2001](#), [Zayane et al., 2002](#); [Essaifi and](#)  
143 [Zayane, 2018](#)). The magmatism is interpreted to be formed in a transpressional tectonic  
144 setting where the Central High-Atlas experienced a significant uplift and orogenesis  
145 during the Jurassic (e.g., [Laville and Pique, 1992](#)), or in an asthenosphere mantle  
146 upwelling context beneath the orogen with partial melting of enriched upper mantle  
147 components without significant crustal extension (e.g., [Frizon de Lamotte et al., 2009](#);  
148 [Michard et al., 2013](#); [Essaifi and Zayane, 2018](#)). From our field observations, the major  
149 apatite deposits of the Moroccan Central High-Atlas —enclosing Anemzi, Tirrhist-  
150 Inougzane, Tassent-Tarstaft, Tarstaft, and Ait Daoud-Toumliline deposits, Figs. [1b, c](#)—  
151 are always spatially associated to these Jurassic-Cretaceous alkaline intrusions. The  
152 apatite ores form mm- to several cm-thick veins without any preferential trending and  
153 the current mining is done only in an artisanal way and the only economic opportunities  
154 are for mineral collectors. Unfortunately, the limited extension of the exposed apatite

155 veins hides more information about the lateral extensions, but the scraping of the  
156 apatite ores generated by the mineral collectors could vary between 2 and 40 m long.  
157 These apatite-producing localities are isolated and less abundant. The contacts  
158 between the apatite veins and their host magmatic rocks are mostly sharp without any  
159 clear zonation and the mineralogical compositions of the vein-type apatite ores and its  
160 host rocks from different localities are slightly variable, whereas the main components  
161 are always alkali feldspars, pyroxene, amphibole, quartz, prehnite, calcite, magnetite,  
162 epidote, and titanite (detail below).

### 163 **3. Sampling and analytical methods**

164 The five Moroccan Central High-Atlas apatite deposits cited above (Fig. 1) were  
165 investigated. For each deposit, we sampled apatite mineralized veins and their direct  
166 country magmatic rocks (see Table 1). Apatite crystals have been separated by hand,  
167 included in epoxy mounts and polished. These apatites, the gangue-forming minerals,  
168 and the country rocks were also studied in polished thin-sections. For petrographical  
169 observations, we used standard microscope Leica DM2700P in the Geo-AnalyticalLab  
170 of Geology and Sustainable Mining Department, Mohammed VI Polytechnic University  
171 (UM6P, Benguerir, Morocco).

172 Seven representative samples of the magmatic rocks hosting the apatite  
173 mineralization veins were selected for whole-rock geochemical analyses of major  
174 elements and some trace elements performed at the Instituto Andaluz de Ciencias de la  
175 Tierra (IACT-CSIC, Granada, Spain). The samples include Anemzi AP45 and AP54,  
176 Tirrhist-Inouzane AP10, Ait Daoud-Toumliline AP08 and AP09, Tassent AP27 and

177 Tasraft AP61 (Table 1). Rock powders were obtained using a jaw crusher and an agate  
178 ring mill. Loss on ignition (L.O.I) was determined by drying the samples at 900°C, and  
179 ranges from 0.07 to 5.86 wt.%. (Cl, Cr, Co, Ni, Cu, Zn, Ga, Zr, Sr, Y, Nb) The analyses  
180 were performed using a sequential spectrometer Bruker AXS S4 Pioneer equipped with  
181 three analyzers (LiF200, OVO-55, PET). For X-ray fluorescence (XRF) major element,  
182 Cr, Ni, Zr, Sr and Y analyses, rock powders (c. 1 g) were weighed with di-lithium  
183 tetraborate flux, and then the mixture was fused at 1000 °C for 15 min. Within run  
184 precision (% RSD), measured by repeated analyses of USGS BHVO-2 and AGV-2  
185 reference materials as external standards was better than 1.5% for all the elements  
186 except P (2.7%) (Varas-Reus et al., 2018). The other minor elements and SO<sub>3</sub> were  
187 analysed in pressed powder pellets in an independent calibration. The analyses of  
188 certified reference materials (BIR-1) have been used as external standard and the  
189 accuracies are <5% for major elements and <11% for minor elements.

190 In situ analyzes of major and minor elements for minerals were carried out using  
191 the Cameca SX100 Electron Microprobe at the Geosciences Montpellier (Montpellier,  
192 France). Analyses were obtained using a beam size of 30–40 µm for apatite and 5 µm  
193 for the minerals of gangue and country rocks, an accelerating voltage of 15 kV and a  
194 probe current of 20 nA. The counting time was fixed at 20 s for major elements and 40 s  
195 for F, Cl, and S. Combination of natural—including fluorapatite (F, P), chlorapatite (Cl),  
196 orthoclase (K), albite (Al, Na), wollastonite (Si, Ca), forsterite (Mg, Fe), baryte (S),  
197 rhodochrosite (Mn)— and synthetic (TiO<sub>2</sub> and Cr<sub>2</sub>O<sub>3</sub>) standards are used for the  
198 calibrations. The apatite and the associated pyroxene and amphibole were analyzed for  
199 trace elements using the Laser Ablation Coupled Plasma-Mass Spectrometry (LA-ICP-

200 MS) at Geosciences Montpellier (AETE platform, OSU-OREME). Signals were acquired  
201 in Time Resolved Acquisition. Each analysis consisted of a pre-ablation pulse of 5 s and  
202 2 min for the blank and 1 min for measurement. The laser was fired employing an  
203 energy density of 15 J/cm<sup>2</sup> at a frequency of 5 Hz and using a spot size of 50 µm.  
204 Reference material NIST612 glass (Pearce et al., 1997) was used as external standard  
205 and CaO and SiO<sub>2</sub> contents determined by electron probe were used as internal  
206 standards. Data were processed using the GLITTER software (Griffin et al., 2008). LA-  
207 ICP-MS data accuracy and precision were assessed using the obtained results from the  
208 BIR-1G glass and show good agreement with published values for this international  
209 reference material (Gao et al., 2002; Jochum et al., 2005, 2016).

## 210 4. Petrography

### 211 4.1. Anemzi

212 In the Anemzi veins, the apatites (~ 10–30 vol.% of the vein) occur as hexagonal  
213 prisms with variable sizes that can reach more than 6 cm length and more than 3 cm  
214 width (Fig. 2). They are present in two different types of gangue (Fig. 2a–b). The first  
215 one is white in color and it mainly consists of calcite, plagioclase (albite), K-feldspar,  
216 magnetite, and quartz (Figs. 2c–d). Calcite occurs as medium- to fine subhedral to  
217 anhedral crystals (~ 0.1–0.8 mm in size) and is the dominant gangue-forming mineral  
218 (up to 40 vol.%), filling the spaces between the other minerals. Albite and K-feldspar are  
219 abundant (up to 25 vol.%) euhedral crystals (~ 0.2–1 cm in size) with sericitized cores  
220 and fresh rims. Magnetite is omnipresent and occurs as red-brownish euhedral  
221 prismatic to anhedral crystals up to 0.5 cm in size. Quartz is less abundant, euhedral to

222 subhedral, and small in size (< 0.5 mm). The second gangue type is dark, and it is  
223 made almost exclusively of needle-shaped amphibole (mostly hornblende) with  
224 subordinate chlorite as a product of amphibole transformation (Figs. 2b, e). The most  
225 beautiful Moroccan apatite gems are extracted from veins within white gangue: they are  
226 characterized by their perfect euhedral habit, transparency, green-yellow color and  
227 glassy luster (Fig. 2a). In contrast, the apatites associated with dark gangue are cloudy  
228 crystals, pale green to grey in color, with little or no luster and enclosed many euhedral  
229 needles of amphibole similar to those present in the matrix veins (Figs. 2b, e).

230 The Anemzi apatite veins are mainly hosted in medium- to coarse-grained  
231 mesocratic to leucocratic magmatic rock which consists of 40 to 60 vol.% of albitic  
232 plagioclase ( $An < 10\%$ ), 20–25 vol.% of K-feldspar and 2–8 vol.% of quartz (Fig. 2c).  
233 Chlorite and Fe-oxides are omnipresent: they result from the complete transformation of  
234 ferromagnesian minerals (i.e., pyroxene, amphibole) which are no longer present in the  
235 rock. Accessory minerals include rutile, apatite, and zircon. Calcite is also observed as  
236 secondary mineral phase.

#### 237 4.2. Tirrhist-Inouzane

238 The Tirrhist-Inouzane veins provide gem-quality apatites (up to 15 vol.%)  
239 transparency and luster with similar color than apatites from the Anemzi region (Figs.  
240 3a, b). The size of apatite crystals is variable but can reach 3 cm length. The gangue  
241 consists of clinopyroxene, albite, epidote, quartz, K-feldspar, and subordinate chlorite  
242 (Fig. 3c–d). Clinopyroxene occurs as euhedral to subhedral crystals (up to 1 cm in size),  
243 forming the main gangue component (up to 35 vol.%). Zonation of some clinopyroxene

244 crystals is highlighted by the alternance of dark and light green domains. Epidote is  
245 abundant (up to 20 vol.%) and it appears as subhedral to euhedral crystals (~ 0.2–1 cm  
246 in size). Albite (dominant), quartz and K-feldspar (rare) represent around 30 vol.% and  
247 occur as medium- to fine subhedral to anhedral crystals (~ 0.1–1 mm in size), filling the  
248 interstices among clinopyroxene, epidote, and apatite. These apatite veins crosscut  
249 medium-grained to pegmatitic leucocratic magmatic rock rich in albitic plagioclase (up to  
250 85 vol.%) (Figs. 3a, e, f) and also contain clinopyroxene (5–10 vol.%), and subordinate  
251 (< 5 vol.%) quartz and K-feldspar. Magnetite (sometimes with ilmenite exsolutions),  
252 stilpnomelane, apatite and zircon are also observed. Chlorite, epidote and sericite occur  
253 as secondary mineral assemblage.

#### 254 4.3. Tassent

255 The apatites (~ 10–20 vol.%) associated with the Tassent veins are elongated  
256 crystals up to 15 cm length and 4 cm wide (Figs. 4a, b). They are pale yellow to pale  
257 green with an opalescent luster. The gangue mineral assemblage consists mainly of  
258 prehnite and quartz (~ 60 vol.%), plus clinopyroxene and epidote (Figs. 4c, d). Prehnite  
259 and quartz occur as euhedral to anhedral crystals (up to 3 mm in size). Clinopyroxene  
260 appears as euhedral to subhedral destabilized crystals (up to 1 cm in size). Epidote is  
261 occasionally observed as yellow-greenish euhedral crystals (< 0.5 mm in size).

262 The apatite veins are hosted by medium-grained mesocratic to leucocratic  
263 magmatic rocks (Figs. 4a, e), which consists of albitic plagioclase (up to 60–70 vol.%),  
264 amphibole (15–20 vol.%), quartz (5–10 vol.%) and subordinate (< 5 vol.%) K-feldspar

265 and clinopyroxene. Titanite, apatite and zircon are the main accessory minerals, and  
266 epidote and chlorite occur as alteration products.

#### 267 4.4. Ait Daoud-Toumliline

268 The apatites (10–20 vol.%) from the Ait Daoud Toumliline veins are gem-quality  
269 dark green, with moderate vitreous luster and characterized by an abundance of  
270 amphibole inclusions (Figs. 5a to 5c). Their size can reach 3 cm. They are found in a  
271 gangue composed mainly of amphibole, prehnite, quartz, epidote, magnetite, albite and  
272 K-feldspar (Fig. 5d–e). Amphibole, prehnite and quartz represent ~ 60–70 vol.% of the  
273 vein and they occur as euhedral to subhedral crystals (up to ~ 4 mm in size). Epidote is  
274 abundant and crystallized as subhedral to euhedral crystals (< 1 mm in size). Albite, K-  
275 feldspar, and magnetite are also observed as medium to fine grains.

276 The apatite ores crosscut medium- to coarse-grained country mesocratic to  
277 melanocratic pale magmatic rocks with a primary paragenesis associating albitic  
278 plagioclase (60–65 vol.%) and amphibole (20–35 vol.%) (Fig. 5f). Clinopyroxene is also  
279 present and is frequently associated with Fe-Ti oxides (magnetite and ilmenite). Quartz  
280 and K-feldspar are rare. Titanite, rutile, and apatite are the main observed accessory  
281 minerals, and epidote, chlorite, and prehnite occur as secondary phases.

#### 282 4.5. Tasraft

283 Two types of apatite occur at Tasraft: (i) the first one is similar to the apatites  
284 found in Ait Daoud-Toumliline (up to 2 cm in size, dark green translucent crystals with  
285 moderate vitreous luster and often containing tiny needle-shaped amphibole inclusions,

286 Fig. 6a); (ii) the second one corresponds to fine crystals (less than 0.5 cm in size) of  
287 gem-quality apatite with a color, transparency and luster similar to those found in  
288 Tirrhist-Inouzane and Anemzi. The gangue of these fine apatite gems consists mainly of  
289 clinopyroxene, albite, quartz, K-feldspar, and titanite (Figs. 6b, c, d, e). As observed for  
290 Tirrhist-Inouzane, clinopyroxene (up to 45 vol.% of the vein) is often zoned and occurs  
291 as euhedral to subhedral crystals (up to 1 cm in size). Albite (dominant), K-feldspar, and  
292 quartz represent around 40 vol.% and occur as medium euhedral to anhedral crystals (~  
293 0.1–1 mm in size), interstitial between clinopyroxene and apatite. Titanite is present and  
294 occurs as green-yellowish to clear prismatic crystals (< 5 vol.%, up to 1 cm in size).

295 The host rocks of these apatite veins are coarse- to medium-grained leucocratic  
296 magmatic rocks containing up to 70–80 vol.% of albitic plagioclase, up to 10 vol.% of  
297 clinopyroxene and subordinate K-feldspar and quartz. Amphibole, magnetite, titanite,  
298 rutile, apatite and zircon are accessory minerals (Figs. 6b–c). Epidote, chlorite, and  
299 sericite also occur as secondary phases.

## 300 5. Geochemistry of magmatic host rocks

301 Whole-rock geochemical analyses for the magmatic facies directly hosting the  
302 Moroccan Central High-Atlas vein-type apatite ores are given in Table 2. In the TAS  
303 classification (Fig. 7a; Middlemost, 1994), these rocks have intermediate to felsic  
304 (dominant) signature with SiO<sub>2</sub> contents ranging from ~ 53 to 65 wt.% and plot in the  
305 fields for gabbrodiorite–monzodiorite and syenite–quartz monzonite (dominant). They  
306 belong mostly to the alkaline/sub-alkaline series with alkali (K<sub>2</sub>O + Na<sub>2</sub>O) content from ~  
307 4.6 to ~ 10 wt.%. All these facies have relatively large ranges in Fe<sub>2</sub>O<sub>3</sub><sup>tot</sup> (~ 2–10.3

308 wt.%), MgO (Mg# ~ 0.04–0.8), CaO (~ 1.7–10.6 wt.%), Al<sub>2</sub>O<sub>3</sub> (14.2–16.5 wt.%), TiO<sub>2</sub> (~  
309 0.4–2.3 wt.%) and P<sub>2</sub>O<sub>5</sub> (~ 0.03–0.9 wt.%). Almost all these rocks are peralkaline [mol.  
310 (Al<sub>2</sub>O<sub>3</sub>/CaO+Na<sub>2</sub>O + K<sub>2</sub>O) = ACNK = 0.51-0.94] with low K contents [K# = mol.  
311 (K<sub>2</sub>O/Na<sub>2</sub>O+K<sub>2</sub>O) = 0.01–0.1], except for Anemzi samples that belong to the high-K  
312 serie (K#: 0.26– 0.36) (Fig. 7b, Table 2, e.g., Rickwood, 1989; Maniar and Piccoli,  
313 1989). Compared to the available data from the literature for the same Central High-  
314 Atlas magmatic intrusions further away from the apatite veins (Figs. 7a–c; data from  
315 Armando, 1999; Lhachmi et al., 2001; Zayane et al., 2002; Essaifi and Zayane, 2018),  
316 the studied rocks have geochemical compositions similar to the more differentiated  
317 members and differ significantly from the more primitive members (mostly gabbros).  
318 Evaluation of the alteration effects using the discriminated diagram of Hughes (1973)  
319 shows that almost all these differentiated rocks plot toward the field of sodic series  
320 (Na<sub>2</sub>O up to ~ 9 wt.%), suggesting that they might have been affected by general Na-  
321 metasomatism conditions (Fig. 7c). In terms of volatile element compositions, Cl and  
322 SO<sub>3</sub> content reaches, respectively, a maximum of 0.25 wt.% and 0.04 wt.% in the Ait  
323 Daoud-Toumliline samples and are lowest in the other samples (0.05–0.13 wt.% Cl and  
324 0.01 wt.% SO<sub>3</sub>) (Table 2). Besides, all investigated rock samples have moderate  
325 contents of MnO (0.03–0.12 wt.%), Co (53–155 ppm), Ni (29–133 ppm), Cu (17–55  
326 ppm), Zn (14–50 ppm), Ga (18–47 ppm), Sr (213–407 ppm), and Y (15–103 ppm).  
327 However, the more felsic rocks have slightly high Zr (up to 748 ppm) and Nb (up to 120  
328 ppm) concentrations, and low values of Cr (less than 32 ppm) (Table 2).

## 329 **6. Mineralogy**

### 330 *6.1. Apatite*

#### 331 6.1.1. Major element composition

332 Table 3 shows representative major element analyses of the studied apatites.  
333 The two major constituents —i.e., CaO and P<sub>2</sub>O<sub>5</sub>— show rather little variations: CaO  
334 and P<sub>2</sub>O<sub>5</sub> vary between 53 and 57 wt.% and between 39 and 44 wt.%, respectively. The  
335 F content of the Anemzi apatite associated with calcite, feldspars, and magnetite matrix  
336 (white gangue; see above) is quite homogenous with about 2 wt.% and low Cl content  
337 (0.67–0.90 wt.%). In contrast, the Anemzi apatites associated with dark gangue  
338 (amphibole matrix), are rather heterogeneous: their F contents decrease from core to  
339 rim (from ~ 1.9 to ~ 0.8 wt.%) while Cl increases (from ~ 0.6 to ~ 2.5 wt.%) (Table 3). All  
340 analyzed Tirrhist-Inouzane apatites have relatively homogeneous F and Cl contents (~  
341 1.9–2.4 and ~ 0.34–0.47 wt.%, respectively). Similarly, the apatites from Tassent and  
342 Tasraft (fine gems) have F concentrations between 1.7 and 2.8 wt.% and low to  
343 moderate Cl contents (0.5–1.47 wt.%). On the other hand, distinct apatite compositions  
344 have been yielded by amphibole-bearing apatites from both Tasraft and Ait Daoud-  
345 Toumliline: the first ones show compositions intermediate between those of Cl-rich  
346 apatites and those of Cl-F apatites (F contents vary between 0.47 and 1.47 wt.% and Cl  
347 between 3.44 and 1.5 wt.%); the second ones are Cl-rich apatites with high Cl (up to 3.3  
348 wt.%) and low F (~ 0.6–1 wt.%) contents. If we considered all apatites analyses, they  
349 show a good linear trend and progressive compositional variations from the F-rich to Cl-  
350 rich apatites as shown by Fig 8a.

351 In all studied apatites, the element substitutions are always very low, and they  
352 vary between the different apatite deposits and even between the crystals of the same  
353 deposit (see Table 3 and Figs. 8b, c, d). The major substituent element is Si ( $0.1 < \text{SiO}_2$   
354 wt.%  $< 0.8$ ).  $\text{Na}_2\text{O}$  contents can reach 0.55 wt.%, whereas  $\text{FeO}^{\text{tot}}$  and  $\text{MgO}$  never  
355 exceed 0.11 and 0.08 wt.%, respectively. The S contents are particularly variable:  $\text{SO}_2$   
356 varies between 0 and 0.57 wt.%. In detail, the amphibole-bearing apatites from Ait  
357 Daoud-Toumiline and Tasraft show quite similar compositions with relatively high  $\text{Na}_2\text{O}$   
358 and  $\text{MgO}$  contents (up to 0.57 wt.% and 0.9 wt.%, respectively),  $\text{FeO}^{\text{tot}}$  contents  
359 between 0.04 and 0.11 wt.% and low concentrations of  $\text{SiO}_2$  ( $< 0.4$  wt.%) and  $\text{SO}_2$  ( $\sim 0$ –  
360 0.3 wt.%). Similar  $\text{FeO}^{\text{tot}}$  and  $\text{SiO}_2$  abundances (0.04–0.10 wt.% and 0.18–0.40 wt.%,  
361 respectively) have been measured in Tassent apatite but with generally low  $\text{Na}_2\text{O}$   
362 (0.06–0.29 wt.%) and  $\text{SO}_2$  ( $< 0.04$  wt.%) contents and no  $\text{MgO}$ . The apatites from  
363 Tirrhist-Inouzane and Tasraft (fine gems) have  $\text{MgO}$  below detection limit and they  
364 generally contain similar low  $\text{Na}_2\text{O}$  ( $\leq 0.08$  wt.%) and  $\text{FeO}^{\text{tot}}$  ( $\leq 0.04$  wt.%)  
365 concentrations, whereas  $\text{SiO}_2$  (up to 0.76 wt.%) and  $\text{SO}_2$  (0.30–0.57 wt.%) abundances  
366 are relatively high. The Anemzi apatites from both white and dark gangues have  
367 comparable  $\text{Na}_2\text{O}$  (0–0.15 wt.%),  $\text{MgO}$  ( $\leq 0.01$  wt.%) and  $\text{FeO}^{\text{tot}}$  (0.02–0.09 wt.%)  
368 concentrations. However, these two types of apatite are distinguished by their  $\text{SiO}_2$  and  
369  $\text{SO}_2$  contents: the apatites from the dark gangue have  $\text{SiO}_2$  (0.12–0.32 wt.%) and  $\text{SO}_2$   
370 (0–0.04 wt.%) significantly lower than those of the white gangue apatites (0.62–0.70  
371 and 0.01–0.42 wt.%, respectively).

372 In addition, the rim–core chemical variations (e.g.,  $\text{SiO}_2$ ,  $\text{FeO}^{\text{tot}}$ ,  $\text{Na}_2\text{O}$ , Cl, and F)  
373 in some apatite crystals of veins are shown in Figs. 8e–h and Table 3. Despite their

374 large size, almost all fluorapatites —e.g., Tirrhist-Inouzane and Anemzi (white  
375 gangue)— have no chemical zoning. In contrast, the Anemzi amphibole-bearing  
376 apatites (dark gangue) display F-rich cores and Cl-rich rims and the SO<sub>2</sub> contents  
377 decrease from core to rim while Na<sub>2</sub>O contents increase (Table 3).

378 To further constrain the relationships between the apatites and their magmatic  
379 host rocks, the apatites in Tirrhist-Inouzane syenites have also been analyzed (see  
380 Table 3). These igneous apatite grains have similar compositions compared to the  
381 Tirrhist-Inouzane, Anemzi (white gangue), Tassent and Tasraft (fine gems) apatite  
382 crystals in veins in terms of F (2–2.3 wt.%), Cl (0.5–0.8 wt.%), Na<sub>2</sub>O (0.1–0.19 wt.%),  
383 MgO (< 0.01) and FeO<sup>tot</sup> (0.04–0.09 wt.%), and they are clearly distinguished from both  
384 Ait Daoud-Toumliline and Tasraft amphibole-bearing apatite gems in veins as shown by  
385 Figs. 8a–c and Table 3. On the other hand, SiO<sub>2</sub> and SO<sub>2</sub> abundances in the syenite  
386 apatites are similar to those found in the gemmy apatites in veins from Tassent and  
387 Anemzi (dark gangue) (SO<sub>2</sub> < 0.04 wt.%, SiO<sub>2</sub> ~ 0.35 wt.%) (Fig. 8d).

#### 388 6.1.2. Trace element composition

389 The trace element abundances of studied apatites given in Table 3 show that  
390 these minerals can incorporate a wide range of trace elements and often in large  
391 quantities. The most abundant trace elements in the Central High-Atlas gem-quality  
392 apatites in veins are REE (mainly La, Ce and Nd), Sr, Y, V, Mn, Th and U, whereas Sc,  
393 Co, Rb, Zr, Nb, Ba and Pb contents are very low (< 6 ppm) (Fig. 9 and Table 3). The  
394 highest REE concentrations are observed in the Ait Daoud-Toumliline apatites (ΣREE:  
395 18190–22587 ppm), with variable abundances in the Anemzi (white gangue) and

396 Tasraft amphibole-bearing apatites ( $\Sigma\text{REE}$ : 3846–17432 ppm), moderate contents in  
397 Tassent apatites ( $\Sigma\text{REE}$ : 13140–13554 ppm) and low to moderate values ( $\Sigma\text{REE}$ :  
398 4823–8495 ppm) in Tirrhist-Inouzane, Anemzi (dark gangue) and Tasraft (fine gems)  
399 apatites. Figs. 10 and 11 show, respectively, their REE patterns and multi-element  
400 spider diagrams normalized to chondrite abundances from McDonough and Sun (1995).  
401 In all apatites, a pronounced negative Eu-anomaly can be observed ( $0.11 < \text{Eu}/\text{Eu}^* <$   
402  $0.63$ ; Fig. 10). Almost all analyses show consistent enrichment in light REE relative to  
403 heavy REE with  $(\text{Ce}/\text{Yb})_n$  ratios ranging from 12 to 78.

404 The V contents of apatites vary greatly between the different deposits. They are  
405 very low (between 3 and 4 ppm) in the Tassent apatites and slightly higher (between 18  
406 and 34 ppm) in the Tirrhist-Inouzane and high in Ait Daoud-Toumliline (55–77 ppm)  
407 (Fig. 9a). We also find these high values but with variability in both Tasraft apatite types:  
408 3–128 ppm in the amphibole-bearing apatites and 96–146 ppm in the fine apatite gems.  
409 In Anemzi, the V contents of the apatites from the dark gangue are low but relatively  
410 constant (9–14 ppm), while these from the white gangue are surprisingly variable and  
411 they can even reach high values (0.17–128 ppm) as in Tasraft. The Mn contents are  
412 very variable in the Anemzi apatites from the white gangue (56–397 ppm), while they  
413 are relatively stable in those from the dark gangue (274–312 ppm) (Fig. 9b). These  
414 latter values are relatively close to that found in the Tassent apatites (222–241 ppm  
415 Mn). The two types of apatite from Tasraft have Mn contents significantly different: 63–  
416 96 ppm for the fine gem-crystals and 138–250 ppm for the Amp-bearing crystals. The  
417 Mn contents of the apatites from Tirrhist-Inouzane (140–148 ppm) and Ait Daoud-  
418 Toumliline (116–131 ppm) are homogeneous and quite similar. Strontium

419 concentrations are moderate to high in all studied apatites (Fig. 9c): 407–686 ppm in  
420 Anemzi white gangue, 673–792 ppm in Anemzi dark gangue, 763–781 ppm in Tirrhist-  
421 Inouzane, 720–730 ppm in Tassent, 382–764 ppm in Tasraft and 386–483 ppm in Ait  
422 Daoud-Toumliline. Yttrium concentration ranges from 230 ppm to 4118 ppm (Fig. 9e):  
423 1567–1672 ppm in Tassent, 1454–1692 ppm in Ait Daoud-Toumliline, 230–4118 ppm in  
424 Anemzi (white gangue), 425–947 ppm in Anemzi (dark gangue), 358–696 ppm in  
425 Tirrhist-Inouzane and 240–979 ppm in Tasraft. Thorium concentrations are moderate to  
426 high and usually ranges from 76 ppm to 530 ppm (Fig. 9f): 90–227 ppm in Anemzi white  
427 gangue, 79–112 ppm in Anemzi dark gangue, 140–182 ppm in Tirrhist-Inouzane, 76–83  
428 ppm in Tassent, 63–253 ppm in Tasraft and 183–253 ppm in Ait Daoud-Toumliline.  
429 Uranium concentrations vary generally from 1.5 to 26 ppm (Fig. 9g). Rather similar high  
430 U values (16–26 ppm) are found in the apatites from the Anemzi (white gangue),  
431 Tirrhist-Inouzane and Tasraft (fine gems). The lowest U contents (1.5–7 ppm) are found  
432 in Tassent and mostly in amphibole-bearing apatites from Anemzi (dark gangue) and in  
433 these from Ait Daoud-Toumliline and Tasraft.

434 In addition, if we considered all apatites analyses from all studied veins, they  
435 show a rough positive correlation between total REE and Na<sub>2</sub>O from the less to more  
436 Cl-rich apatites as shown by Fig 12. Although some compositional variations can be  
437 observed among the different apatites in veins (Figs. 9, 10, 11, 12, Table 3), their REE  
438 and multi-trace element patterns overlap and are globally parallel with those of the  
439 magmatic host rock apatites investigated here (Figs. 10–11).

## 440 6.2. The gangue minerals

### 441 6.2.1. Clinopyroxene

442 Clinopyroxene is exclusively observed in the F-rich apatites veins (e.g., Figs. 3, 4  
443 and 6). The crystals are euhedral, up to 3 cm in size and dark to light green in color,  
444 sometimes with well-marked zoning corresponding to an alternation of dark and light  
445 green domains. Dark and light green grains also occur in almost all magmatic rocks  
446 hosting the apatite veins: they appear as euhedral to subhedral grains (0.5 and 2 mm,  
447 Figs. 3–4, 6). Major and trace element abundances of clinopyroxenes are given in  
448 Tables 4–5. The two major constituents —i.e., SiO<sub>2</sub> and CaO— show little variations;  
449 the representative values are ~ 49–54 wt.% SiO<sub>2</sub> and 20–26 wt.% CaO. In detail, all  
450 analyzed dark green clinopyroxene zones and crystals from both apatite-bearing veins  
451 and country alkaline rocks are similar and have hedenbergite to Fe-rich augite  
452 compositions (Wo<sub>44–49</sub>, En<sub>12–26</sub>, Fs<sub>27–43</sub>, see Fig. 13a and Table 4; e.g., Morimoto et al.,  
453 1988) with low Mg# [Mg/(Mg + Fe<sup>2+</sup>) ~ 0.32–0.66], high total FeO content (~ 16–23 wt.  
454 %), and moderate to high Na<sub>2</sub>O values (up to ~ 3 wt. %) and MnO (up to 1.4 wt. %).  
455 However, the light green zones and crystals, are mainly diopside in compositions (Wo<sub>46–</sub>  
456 <sub>51</sub>, En<sub>27–41</sub>, Fs<sub>8–25</sub>) with high Mg# ~ (up to 0.95), and low total FeO content (~ 5–14 wt.  
457 %), MnO (~ 0.1–0.4 wt. %) and Na<sub>2</sub>O values (mostly < 2 wt. %). All these clinopyroxene  
458 types have similar variable compositional range, with, Al<sub>2</sub>O<sub>3</sub> (~ 0.1–2.6 wt. %), TiO<sub>2</sub>  
459 (mostly < 1 wt. %) and Cr<sub>2</sub>O<sub>3</sub> (0–0.07 wt. %) contents.

460 Fig. 13b shows REE patterns normalized to chondrite values (McDonough and  
461 Sun, 1995) of clinopyroxenes from Tasraft and Tassent F-rich apatite veins, and

462 Tirrhist-Inouzane host syenites (Table 5). All clinopyroxene REE concentrations are 10–  
463 160 times enriched relative to chondrite abundances ( $\Sigma\text{REE} \sim 80\text{--}147$  ppm). They have  
464 parallel REE patterns [with slight high values in light REE in clinopyroxene from Tasraft  
465 apatite veins with  $(\text{La}_n/\text{Yb}_n)$  up to 1.3] and show strongly developed negative Eu  
466 anomaly ( $\text{Eu}/\text{Eu}^* \sim 0.19\text{--}0.37$ ). General high depletion for Cu, Rb, Ba, Nb, Ta, Pb, Th,  
467 and U ( $< 0.07$  ppm) is present for all analyzed clinopyroxenes. They have generally ~  
468 3–12 ppm Sc, 9–23 ppm Co, ~ 17–80 ppm Sr, ~ 26–98 ppm Zn, and ~ 23–166 ppm V.

### 469 6.2.2. Amphibole

470 Amphibole occurs as main gangue mineral in Cl-rich apatites veins (e.g., Ait-  
471 Daoud-Toumliline; Fig. 5) and as inclusions in almost all the Cl-rich apatites (e.g., Ait  
472 Daoud-Toumliline, Tasraft and Anemzi; Figs. 2, 5–6). It forms euhedral and acicular  
473 green to brown crystals up to 2 cm in size. Amphibole also occurs as euhedral to  
474 subhedral grains (~ 0.2–1 mm in size) in magmatic host rocks with colors similar to that  
475 of amphibole from apatite veins (e.g., Figs. 4–5). Table 4 shows representative major  
476 element compositions of the amphiboles. All amphiboles from apatite veins and their  
477 host rocks are calcic with high Ca content (~ 1.67–2.10 atoms per formula unit =  
478 a.p.f.u), Si (~ 6.87–7.78 a.p.f.u) and Mg ( $\text{Mg}\# \sim 0.60$  to 0.90) with moderate Fe (mostly  
479 ~ 0.54–1.20 a.p.f.u), low to moderate (Na+K) values (0.18–0.90 a.p.f.u) and low Al  
480 concentrations (~ 0.1–1 a.p.f.u), characteristics of magnesiohornblende-actinolite and  
481 edinite association (Fig. 13c; e.g., Leake et al., 1997). The amphibole trace element  
482 abundances have been analyzed for Anemzi apatite veins (dark gangue) and are given  
483 in Table 5. The REE abundance in amphiboles range from 46 to 101 ppm. In general,

484 their chondrite-normalized REE patterns are remarkably constant (Fig. 13d): they all  
485 display a gentle shape, flat pattern ( $La_n/Yb_n \sim 0.7-1$ ) with a pronounced negative Eu  
486 anomaly ( $Eu/Eu^* \sim 0.4-0.5$ ) similar to those of clinopyroxenes (Fig. 13c). They also  
487 have 32–650 ppm Cr and 364–405 ppm V, and Sc, Co, Zn, Sr, Y, Zr, and Nb  
488 abundances are less than 68 ppm, with low values ( $< 2$  ppm) for Cu, Rb, Ba, Ta, Pb,  
489 Th, and U (Table 5).

### 490 6.2.3. Feldspars

491 Feldspars are commonly present in all studied vein-type apatite ores. Table 6  
492 displays representative feldspar analyses. In all apatite ores and magmatic host rocks  
493 (Figs. 2–6), albite is the most abundant feldspar: the crystals are euhedral to subhedral  
494 and of centimeter size (up to 5 cm). Their compositions are similar, and they correspond  
495 almost pure to albite ( $Ab \sim 90$  to 100). K-feldspar is also present but less abundant than  
496 albite: the crystals are euhedral to subhedral in the veins, subhedral to anhedral in  
497 magmatic host rocks and with sizes varying between 0.1 and 1 cm. Their compositions  
498 are between  $Or_{94}$  and  $Or_{99}$ . Anorthoclase compositions ( $Ab \sim 70-75$ ,  $An \sim 1-3$ ,  $Or \sim$   
499  $23-29$ ) are exclusively found in the magmatic host rocks (see Table 6).

### 500 6.2.4. Other minerals

501 Magnetite is found in all apatite veins: the crystals are euhedral to subhedral with  
502 up to 1 cm in size. This oxide is also observed in magmatic host rock as subhedral to  
503 anhedral grains up to 2 mm in size (e.g., Fig. 3). Representative compositions are  
504 reported in Table 7. Analyses of magnetite indicate moderate to high  $FeO^{tot}$  contents of

505 about ~ 91–95 wt.% in the apatite veins. In the magmatic host rock, the magnetite  
506 composition varies from magnetite ( $\text{FeO}^{\text{tot}} \sim 90\text{--}95$  wt.%) to Ti-magnetite [ $\text{FeO}^{\text{tot}} \sim 87$   
507 wt.% and  $\text{TiO}_2 \sim 3$  wt.%], sometimes with ilmenite exsolutions ( $\text{TiO}_2 \sim 45\text{--}54$  wt.% and  
508  $\text{FeO}^{\text{tot}} \sim 46\text{--}50$  wt.%) (Table 7).

509 Titanite is also present in both apatite veins and host rocks and the analyses are  
510 reported in Table 7. It is associated with apatite veins and forms greenish yellow  
511 euhedral crystals up to 2 cm in size; in magmatic host rocks, it occurs as dark brown to  
512 colorless euhedral to subhedral crystals (0.5–2 mm). Titanite from veins has relatively  
513 homogeneous contents in  $\text{SiO}_2$ ,  $\text{TiO}_2$ , and  $\text{CaO}$  (~ 29.5, ~ 41.5 and 29 wt.%  
514 respectively), whereas in the host rocks the titanite compositions are more variable: 20–  
515 32 wt.%  $\text{SiO}_2$ , 29–54 wt.%  $\text{TiO}_2$  and 19–29 wt.%  $\text{CaO}$ . On the other hand, rutile is  
516 observed and analyzed exclusively in the magmatic host rock, forming small euhedral to  
517 subhedral dark-red grains (up to 0.2 mm in size) and its analyses are also given in Table  
518 7.

519 In the apatite mineralized veins, epidote (pistachite) occurs as green to greenish  
520 yellow euhedral crystals up to 3 cm in size, whereas in the alkaline host rocks it appears  
521 as yellow subhedral to anhedral grains of small size (0.02–0.5 cm). As shown in Table  
522 8, the epidote compositions are similar in veins and their host rocks: 36–38  $\text{SiO}_2$  wt.%,  
523 18.50–22.70  $\text{Al}_2\text{O}_3$  wt.%, 21–25  $\text{CaO}$  wt.% wt.%, with slight variation  $\text{FeO}^{\text{tot}}$  content (~  
524 10–18 wt.%).

525 Prehnite is observed in the Tassent and Ait Daoud-Toumliline apatite mineralized  
526 veins. It occurs as radiating fibrous aggregates and is commonly translucent with

527 colorless to white greenish color (up to 4 cm in size, e.g., Figs. 4–5). Prehnite can occur  
528 also in the magmatic host rocks where it appears both as veinlets and as microcavity  
529 filling products. Analyses of this calcic mineral (~ 28 wt.% CaO) are shown in Table 8  
530 and are homogenous and similar in both mineralized veins and host alkaline rocks.

## 531 7. Discussion and conclusions

532 According to our field observations, and petrological, mineralogical, and  
533 geochemical study, all Moroccan Central High-Atlas vein-type apatite ores (mostly gem-  
534 quality) are hosted by differentiated igneous intrusions ranging from syenite–quartz  
535 monzonite (dominant) to monzodiorite–gabbrodiorite (Fig. 7 and Table 2). The spatial  
536 association with Mesozoic (Jurassic-Cretaceous) alkaline/transitional rocks is typical of  
537 phosphates derived from silicate alkaline–carbonatite complexes (e.g., Simandl and  
538 Paradis, 2018 and references therein). However, carbonatites have not been found in  
539 the studied area, although the ore bodies are not so far to the well-known Tamazeght  
540 Eocene carbonatite complex (see Fig. 1b).

541 Our new data lead us to distinguish two types of apatite ores: (i) high gem-quality F-rich  
542 apatites associated with clinopyroxene (diopside to hedenbergite-augite) and (ii) gemmy  
543 Cl-rich apatites mainly formed in an amphibole matrix (hornblende–edenite to actinolite)  
544 (Figs. 2–6, 8, Table 3). Both ore bodies are also associated with feldspars (albitic  
545 plagioclase dominant and orthoclase), quartz, magnetite, titanite, epidote, calcite, and  
546 prehnite. All studied apatites of both deposit types have similar REE and multi-trace  
547 element patterns which suggest that they could have a relatively close origin (Figs. 9,  
548 10, 11, Table 3). A comparison between the Central High-Atlas apatites and those of

549 the well-known Sweden Kiruna-type iron ores which formed by late-stage of magmatic  
550 differentiation of felsic magmas (pink field in Fig. 10; data from Frietsch and Perdahl,  
551 1995) shows that the REE compositions of apatites from both deposits are similar with  
552 pronounced negative Eu anomaly. Conversely, they differ from the REE patterns of the  
553 worldwide carbonatite apatites from the Oka (Canada), Kaiserstuhl (Germany),  
554 Jacupiranga (Brazil), Fen (Norway), Sokli (Finland), Hörningsholmen-Alnö (Sweden),  
555 Siilinjärvi (Finland) and Loolekop-Phalaborwa (South Africa) complexes (green field in  
556 Fig. 10; data from Ingrid, 1998; Chen and Simonetti, 2013; Doroshkevich et al., 2018;  
557 Decrée et al., 2020) ( $\text{Eu}/\text{Eu}^*$  up to 0.63 in our apatites and around 1 in worldwide  
558 carbonatite apatites). The pronounced Eu negative anomaly exhibited by all the Central  
559 High-Atlas apatites studied strongly suggests that feldspar crystallized from phosphate-  
560 rich melt prior to apatite crystallization. Such negative Eu anomalies in apatite have also  
561 been reported in fenites ( $\text{Eu}/\text{Eu}^* \sim 0.73\text{--}0.84$ , e.g., Decrée et al., 2020) as a result of  
562 cooling and crystallization of carbonatitic and alkaline magmas in the crust and the  
563 presence of alkali-rich fluids that metasomatize the surrounding country rocks (e.g.,  
564 Elliott et al., 2018 and references therein). Although many of the mineralogical features  
565 observed in studied apatite veins (evidence of e.g., K-feldspar, albite, clinopyroxenes  
566 and calcic amphiboles and other accessory minerals, Figs. 2–6) and the extensive sodic  
567 metasomatism affecting the magmatic country rocks (Fig. 7) suggest that fenitization  
568 may have occurred during the crystallization of apatite-rich veins (e.g., Zharikov et al.,  
569 2007; Elliott et al., 2018). The pronounced Eu anomaly together with Sr, Y, and Mn  
570 contents (Fig. 14a–b; Belousova et al., 2002) strongly indicate that Central High-Atlas  
571 apatites are co-genetic and chemically similar to the magmatic apatites of granitoids of

572 Kiruna type iron ores (Fig. 14a–b). Indeed, the mineralogical and geochemical signature  
573 of the Central High-Atlas apatite vein-type ores may be related to host-alkaline magma  
574 crystallization and evolution rather than carbonatite magma. These observations are  
575 corroborated by apatite discrimination diagrams based on the concentrations of several  
576 trace elements in apatite (Figs. 14c–d; Mao et al. 2016), showing that the Central High-  
577 Atlas apatites have composition close to apatites from iron oxide copper gold (IOCG)  
578 ore deposits and unmineralized rocks (likely syenites) (Fig. 14c). Generally, the IOCG  
579 systems are clearly interpreted as origin from magmatic hydrothermal systems (mainly  
580 porphyries, e.g., Barton, 2009). In addition, the mineralogy, texture, alteration halos and  
581 geochemistry data also suggest that the Central High-Atlas vein-type apatite ores could  
582 be largely linked to the Central High-Atlas differentiated magmatic rocks (mostly  
583 syenites) (Figs. 9, 10, 11, 12, 14). Plagioclases in both veins and host-rocks indicate  
584 that both derived from sodic-rich melt or/and underwent general albitization (see above  
585 and Table 6, Fig. 7b–c). The clinopyroxene and amphibole observed in both units (veins  
586 and host rocks) refer both to Ca-rich minerals and support a common igneous origin  
587 (Table 4–5 and Fig. 13). Altogether, field observations and mineralogical and  
588 geochemical signatures of veins and host-rocks suggest that Moroccan Central High-  
589 Atlas apatite ores might be related to the magmatic processes that lead to the formation  
590 of differentiated rock (likely syenites), although hydrothermal/continental contamination  
591 may have also been involved.

592 The absolute age of these apatite deposits is elusive, but they are observed in  
593 veins crosscutting the Jurassic-Cretaceous (ca. 165–125 Ma) alkaline igneous rocks of  
594 the Central High-Atlas and are probably coeval or post-date the crystallization of the

595 magmatic intrusions. The geochemical observations suggest that the crystallization  
596 environment of the high quality F-rich apatite ores (e.g., Figs. 2–3) remained chemically  
597 stable and that the intensive parameters controlling their crystallization process did not  
598 fluctuate (Fig. 8, Table 3). In contrast, the Cl-rich apatite ores enclosing intense  
599 randomly distributed amphibole inclusions in apatite gems indicate slight fluctuation of  
600 magma and/or fluid compositions. The presence of low temperature minerals (albite,  
601 quartz, epidote, prehnite, K-feldspar, magnetite and calcite) in both apatite deposits and  
602 in their host alkaline rocks strongly suggest that crystallization occurred during late  
603 magmatic stages and that the main geochemical signatures of apatite are most  
604 probably related to crystallization of alkaline magma rich in P and incompatible  
605 elements such as F (e.g., Fig. 8). It is also important to note that interactions of evolved  
606 magma with country sedimentary formations rich in Cl, Na and Ca most probably  
607 participated to the Cl-Na enrichment of apatites cannot be ignored. In Fig. 15, we  
608 provide the suitable fluids involved during the formation of the Central High-Atlas apatite  
609 mineralization. In the igneous systems the apatite is initially enriched in F owing to the  
610 high apatite–melt partition coefficient for F relative to Cl (e.g., Harlov, 2015; Webster  
611 and Piccoli, 2015). Throughout progressive crystallization of the magmas, the F  
612 contents decrease due to the crystallization of F-rich apatite resulting in progressive Cl  
613 enrichment of apatite (e.g., Fig. 8a and Table 3), Anemzi amphibole-bearing apatites  
614 being an intermediate case between the purely F-rich (e.g., Anemzi white gangue,  
615 Tirrhist-Inouzane, and Tassent) and Cl-rich apatite deposits (e.g., Tasraft and Ait  
616 Daoud-Toumliline). As mentioned above, the genesis of the Cl-rich apatites could also  
617 result from contamination of magma by Cl-enriched hydrothermal fluids due to

618 percolation through Triassic evaporates (mainly halite) buried beneath the Moroccan  
619 Central High-Atlas belt (Fig. 1c). This might be related to halokinetic tectonic on the  
620 Central High-Atlas magmatic intrusions emplacement (e.g., [Michard et al., 2011](#); [Saura  
621 et al., 2014](#)) and a general Na-metasomatism/fenitization associated with the  
622 emplacement of the alkaline magmatic rocks directly hosting these vein-type apatite  
623 ores (see Fig. 7; e.g., [Zirner et al., 2015](#); [Essaifi and Zayane, 2018](#)). The presence of  
624 solid and fluid organic inclusions in Anemzi gemmy apatites ([Dumańska-Słowik et al.,  
625 2018](#)), also supports a contamination of magmas by country sedimentary rocks. In  
626 addition, the hydrothermal fluids could also percolate through Mg-rich rocks (likely  
627 gabbros) present in the same area as attested by the presence of MgO in Cl-apatites  
628 (e.g., Fig. 8, Table 3). Altogether, this study highlighted the role of magma/rock and  
629 fluid/rock interactions and the complex chemical reactions involved during magma  
630 emplacement and the genesis of phosphate ores deposits. Further detailed field,  
631 petrological, structural, geochemical, geochronological and isotope investigations are  
632 however necessary in order to better quantified the igneous and sub-solidus processes  
633 as well as the exchange reactions and environmental conditions that lead to igneous  
634 phosphate ores deposits and the crystallization of gem-quality apatite crystals.

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922 **Figure captions**

923 **Fig. 1. (a)** Sketch map showing the location of Central High-Atlas (CHA) segment in the  
924 north Morocco. **(b)** Geological map and **(c)** NW-SE cross-section of the Central High-  
925 Atlas [adapted from [Teixell et al. \(2003\)](#), [Michard et al. \(2011\)](#) and references therein],  
926 showing the spatial relationships among the Jurassic-Cretaceous alkaline magmatism  
927 and the Mesozoic sedimentary series, and location of the studied apatite deposits.

928 **Fig. 2. (a–b)** Hand specimen and **(c–e)** microscopic photographs of the Anemzi apatite  
929 ores showing gangue- and host rock-forming minerals: **(a, c, d)** white and **(b, e)** dark  
930 gangue types. Apatite (Ap), plagioclase (Pl), K-feldspar (Kfs), calcite (Cal), magnetite  
931 (Mag), quartz (Qz), amphibole (Amp).

932 **Fig. 3. (a, b)** Hand specimen and **(c, e, f, d)** microscopic photographs of the Tirrhist-  
933 Inouzane apatite mineralization showing gangue- and host rock-forming minerals.  
934 Clinopyroxene (Cpx), epidote (Ep), stilpnomelane (Stp).

935 **Fig. 4. (a, b)** Hand specimen and **(c, d, e)** microscopic photographs of the Tassent  
936 apatite mineralization showing gangue- and host rock-forming minerals. Prehnite (Prh),  
937 titanite (Ttn).

938 **Fig. 5. (a, b, c)** Hand specimen and **(d, e, f)** microscopic photographs of the Ait Daoud-  
939 Toumliline apatite mineralization showing gangue- and host rock-forming minerals.

940 **Fig. 6. (a)** Amphibole-bearing apatite gem in mount, **(b)** hand specimen and **(c, d, e)**  
941 microscopic photographs of the Tasraft apatite mineralization (fine gems), showing  
942 gangue- and host rock-forming minerals.

943 **Fig. 7.** Chemical classification for the lithologies hosting the Central High-Atlas vein-  
944 type apatite deposits. **(a)** TAS classification diagram (Middlemost, 1994). **(b)**  $K_2O$  vs.  
945  $SiO_2$  diagram of Rickwood (1989). **(c)**  $(K_2O+Na_2O)$  vs.  $[K_2O/(K_2O+Na_2O)]*100$  plot (after  
946 Hughes, 1973). Literature data compiled from Armando (1999), Lhachmi et al. (2001),  
947 Zayane et al. (2002) and Essaifi and Zayane (2018).

948 **Fig. 8. (a–d)** Major element concentrations in Central High-Atlas apatites. **(e–h)** Results  
949 of electron microprobe analytical profiles, and trends within images of apatite crystals  
950 corresponding to locations of these profiles.

951 **Fig. 9.** Compositional variation diagrams for the most abundant trace elements in  
952 Central High-Atlas apatites.

953 **Fig. 10.** REE patterns normalized to chondrites (McDonough and Sun, 1995) for the  
954 Central High-Atlas apatites. Worldwide carbonatite apatite data are from Ingrid (1998),  
955 Chen and Simonetti (2013), Doroshkevich et al. (2018) and Decrée et al. (2020). Iron  
956 ore apatite data are from Frietsch and Perdahl (1995).

957 **Fig. 11.** Spider diagrams normalized to chondrites (McDonough and Sun, 1995) for the  
958 Central High-Atlas apatites.

959 **Fig. 12.**  $\Sigma REE$  (ppm) versus  $Na_2O$  content (wt.%) in apatite from veins and host rocks.

960 **Fig. 13.** Classification and REE patterns normalized to chondrites (McDonough and  
 961 Sun, 1995) of studied clinopyroxenes and amphiboles: (a, b) clinopyroxenes and (c, d)  
 962 amphiboles.

963 **Fig. 14. (a-b)** Trace element discrimination diagrams for studied Central High-Atlas  
 964 apatites, compared to those of reference rock types (Belousova et al., 2002).  
 965 Carbonatite field from Ingrid (1998), Chen and Simonetti (2013), Doroshkevich et al.  
 966 (2018) and Decrée et al. (2020). Field of carbonatite related fenite apatites is plotted  
 967 using data from Decrée et al. (2020). (c-d) Apatite metallogenic discrimination diagrams  
 968 of Mao et al. (2016). **DP1-1** [=  $-0.06461\log\text{Mn} - 1.56\log\text{Sr} + 2.609\log\text{Y} + 0.3631\log\text{La} -$   
 969  $1.766\log\text{Ce} + 0.6243\log\text{Eu} - 3.642\log\text{Dy} + 0.7086\log\text{Yb} - 1.178\log\text{Pb} + 0.4161\log\text{Th} +$   
 970  $0.963\log\text{U} + 6.589$ ]. **DP1-2** [=  $0.2073\log\text{Mn} - 1.035\log\text{Sr} + 15.1\log\text{Y} + 4.995\log\text{La} -$   
 971  $5.804\log\text{Ce} + 0.1741\log\text{Eu} - 8.771\log\text{Dy} - 4.326\log\text{Yb} + 2.022\log\text{Pb} - 0.6719\log\text{Th} +$   
 972  $0.02096\log\text{U} - 10.45$ ]. **DP4-1** [=  $5.379\cdot\log\text{V} + 1.0285\cdot\log\text{Mn} + 1.0004\cdot\log\text{Sr} -$   
 973  $0.0447\cdot\log\text{Ce} + 3.169\cdot\log\text{Eu} - 5.412\cdot\log\text{Yb} - 0.3302\cdot\log\text{Pb} - 0.1080\cdot\log\text{Th} + 5.385$ ].

974 **Fig. 15.** A schematic model for the two main Central High-Atlas vein-type apatite ores.

976 **Tables**

977 **Table 1.** Selected samples of the studied apatite deposits for mineralogical and  
978 geochemical analysis.

979 **Table 2.** Whole-rock XRF analyses of alkaline rocks directly hosting the Central High-  
980 Atlas vein-type apatite ores.

981 **Table 3.** Representative major and trace element compositions of the studied Central  
982 High-Atlas apatites.

983 **Table 4.** Representative compositions (wt.%) of the clinopyroxenes and amphiboles in  
984 the Moroccan Central High-Atlas apatite mineralized veins and magmatic host rocks.

985 **Table 5.** Trace element compositions (ppm) of the studied clinopyroxenes and  
986 amphiboles.

987 **Table 6.** Representative compositions (wt.%) of the feldspars in the studied Central  
988 High-Atlas apatite ores and their host alkaline rocks.

989 **Table 7.** Representative compositions (wt.%) of the magnetite, ilmenite, titanite and  
990 rutile from studied apatite veins and their host alkaline rocks.

991 **Table 8.** Representative compositions (wt.%) of the epidote and prehnite associated the  
992 apatite ores and their host alkaline rocks.