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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.% Σ 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-</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 SO₃ 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 (TiO₂ and Cr₂O₃) 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/cm² 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 SiO₂ 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 SiO₂ 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 (K₂O + Na₂O) content from ~
307 4.6 to ~ 10 wt.%. All these facies have relatively large ranges in Fe₂O₃^{tot} (~ 2–10.3

308 wt.%), MgO (Mg# ~ 0.04–0.8), CaO (~ 1.7–10.6 wt.%), Al₂O₃ (14.2–16.5 wt.%), TiO₂ (~
309 0.4–2.3 wt.%) and P₂O₅ (~ 0.03–0.9 wt.%). Almost all these rocks are peralkaline [mol.
310 (Al₂O₃/CaO+Na₂O + K₂O) = ACNK = 0.51-0.94] with low K contents [K# = mol.
311 (K₂O/Na₂O+K₂O) = 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 (Na₂O 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 SO₃ 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.% SO₃) (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 P₂O₅— show rather little variations: CaO
334 and P₂O₅ 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). Na_2O contents can reach 0.55 wt.%, whereas FeO^{tot} and MgO never
355 exceed 0.11 and 0.08 wt.%, respectively. The S contents are particularly variable: 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 Na_2O
358 and MgO contents (up to 0.57 wt.% and 0.9 wt.%, respectively), FeO^{tot} contents
359 between 0.04 and 0.11 wt.% and low concentrations of SiO_2 (< 0.4 wt.%) and SO_2 (~ 0 –
360 0.3 wt.%). Similar FeO^{tot} and 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 Na_2O
362 (0.06–0.29 wt.%) and SO_2 (< 0.04 wt.%) contents and no MgO . The apatites from
363 Tirrhist-Inouzane and Tasraft (fine gems) have MgO below detection limit and they
364 generally contain similar low Na_2O (≤ 0.08 wt.%) and FeO^{tot} (≤ 0.04 wt.%)
365 concentrations, whereas SiO_2 (up to 0.76 wt.%) and SO_2 (0.30–0.57 wt.%) abundances
366 are relatively high. The Anemzi apatites from both white and dark gangues have
367 comparable Na_2O (0–0.15 wt.%), MgO (≤ 0.01 wt.%) and FeO^{tot} (0.02–0.09 wt.%)
368 concentrations. However, these two types of apatite are distinguished by their SiO_2 and
369 SO_2 contents: the apatites from the dark gangue have SiO_2 (0.12–0.32 wt.%) and 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., SiO_2 , FeO^{tot} , Na_2O , 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 SO₂ contents
377 decrease from core to rim while Na₂O 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.%), Na₂O (0.1–0.19 wt.%),
383 MgO (< 0.01) and FeO^{tot} (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, SiO₂ and SO₂ abundances in the syenite
386 apatites are similar to those found in the gemmy apatites in veins from Tassent and
387 Anemzi (dark gangue) (SO₂ < 0.04 wt.%, SiO₂ ~ 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 (ΣREE : 3846–17432 ppm), moderate contents in
397 Tassent apatites (ΣREE : 13140–13554 ppm) and low to moderate values (ΣREE :
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 Na₂O 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., SiO₂ and CaO— show little variations;
449 the representative values are ~ 49–54 wt.% SiO₂ 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_{44–49}, En_{12–26}, Fs_{27–43}, see Fig. 13a and Table 4; e.g., Morimoto et al.,
453 1988) with low Mg# [$\text{Mg}/(\text{Mg} + \text{Fe}^{2+}) \sim 0.32\text{--}0.66$], high total FeO content (~ 16–23 wt.
454 %), and moderate to high Na₂O 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_{46–}
456 ₅₁, En_{27–41}, Fs_{8–25}) with high Mg# ~ (up to 0.95), and low total FeO content (~ 5–14 wt.
457 %), MnO (~ 0.1–0.4 wt. %) and Na₂O values (mostly < 2 wt. %). All these clinopyroxene
458 types have similar variable compositional range, with, Al₂O₃ (~ 0.1–2.6 wt. %), TiO₂
459 (mostly < 1 wt. %) and Cr₂O₃ (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 SiO_2 , TiO_2 , and 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.% SiO_2 , 29–54 wt.% TiO_2 and 19–29 wt.% 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 SiO_2 wt.%,
523 18.50–22.70 Al_2O_3 wt.%, 21–25 CaO wt% wt.%, with slight variation FeO^{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) (Eu/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.** ΣREE (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** [= -0.06461logMn - 1.56logSr + 2.609logY + 0.3631logLa -
 969 1.766logCe + 0.6243logEu - 3.642logDy + 0.7086logYb - 1.178logPb + 0.4161logTh +
 970 0.963logU + 6.589]. **DP1-2** [= 0.2073logMn - 1.035logSr + 15.1logY + 4.995logLa -
 971 5.804logCe + 0.1741logEu - 8.771logDy - 4.326logYb + 2.022logPb - 0.6719logTh +
 972 0.02096logU - 10.45]. **DP4-1** [= 5.379·logV + 1.0285·logMn + 1.0004·logSr -
 973 0.0447·logCe + 3.169·logEu - 5.412·logYb - 0.3302·logPb - 0.1080·logTh + 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.